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FOREWORD

The 1989 edition of *Geological Fieldwork: A Summary of Field Activities and Current Research* is the fifteenth in this publication series. It covers a year during which the Geological Survey Branch maintained its extensive program of mapping and research activity to improve the geoscience database in British Columbia.

The base budget of the Branch for the 1989-90 fiscal year is $6.22 million, with an additional $984 000 provided by the Canada/British Columbia Mineral Development Agreement (MDA). The government's policy of encouragement for mineral exploration is solidly backed by funding for geoscience research. In 1989, the combined base and MDA budgets funded 35 in-house field programs and provided support for an additional 28 projects by university researchers. This has been the last year of fieldwork under the MDA program.

Highlights of the program are:

- Eight 1:50 000 mapping projects continued in the Taseko–Bridge River, Whitesail Lake, Telkwa Range, Manson Creek, Stikine River, Iskut River, Tagish Lake and Atlin areas. Regional mapping projects in the Sicker Group on Vancouver Island and the Sylvester allochthon are now in the write-up phase and will generate major publications in the coming year.
- Metallogenic studies include mapping in the Stewart–Sulphurets–Iskut "Golden Triangle" and the Rossland Group volcanic rocks of the East Kootenay District, and on-going investigation of precious metal skarns. A separate section of this edition of "Fieldwork" is devoted to Alaskan-type ultramafic complexes, ophiolites and related precious metal deposits. Detailed mineral deposit studies by university researchers include investigations of the Shasta deposit in the Toodoggone epithermal precious metal district, the Golden Bear deposits near Tatsamenie Lake, the newly reopened Silbak Premier mine in the Stewart mining camp and the Silver Queen epithermal veins at Owen Lake in central British Columbia.
- Reconnaissance geochemical surveys were completed over four 1:250 000 map sheets covering southern Vancouver Island and the Lower Mainland.
- A surficial geology subsection was established in the late summer and preliminary results of work on placer gold deposits are published in this volume.
- A comprehensive program to evaluate the mineral potential of the Purcell Wilderness Conservancy was begun.

The success of the British Columbia Geoscience Research Grant Program is evidenced by the large number of external papers published in this and previous editions of "Fieldwork". In 1989 the MDA funded eight research grants totalling $60 000; a further 20 grants, totalling $130 000, were made from the base budget to researchers at 15 institutions spanning the breadth of Canada from Memorial University of Newfoundland to the University of Victoria. Research projeets cover such diverse topics as isotopic dating, fluid inclusion studies, depositional controls in placer gold deposits, platinum geochemistry, palynological dating and the potential for using zeolites to increase the solubility of phosphate rock and permit direct application of low-grade phosphorite in agriculture.

Although the number of papers in the edition of "Fieldwork" is down slightly from the previous two years, preparing it for publication against tight deadlines remains a significant achievement. The efforts of our editorial staff: Brian Grant who managed the process, Doreen Fehr and Janet Holland who formatted the text and did page layout, and John Newell who edited manuscripts, are gratefully acknowledged. Appreciation is also extended to the management and staff of the Queen's Printer who, as always, came through in the crunch; without their whole-hearted cooperation timely delivery would not be possible.

W.R. Smyth
Chief Geologist
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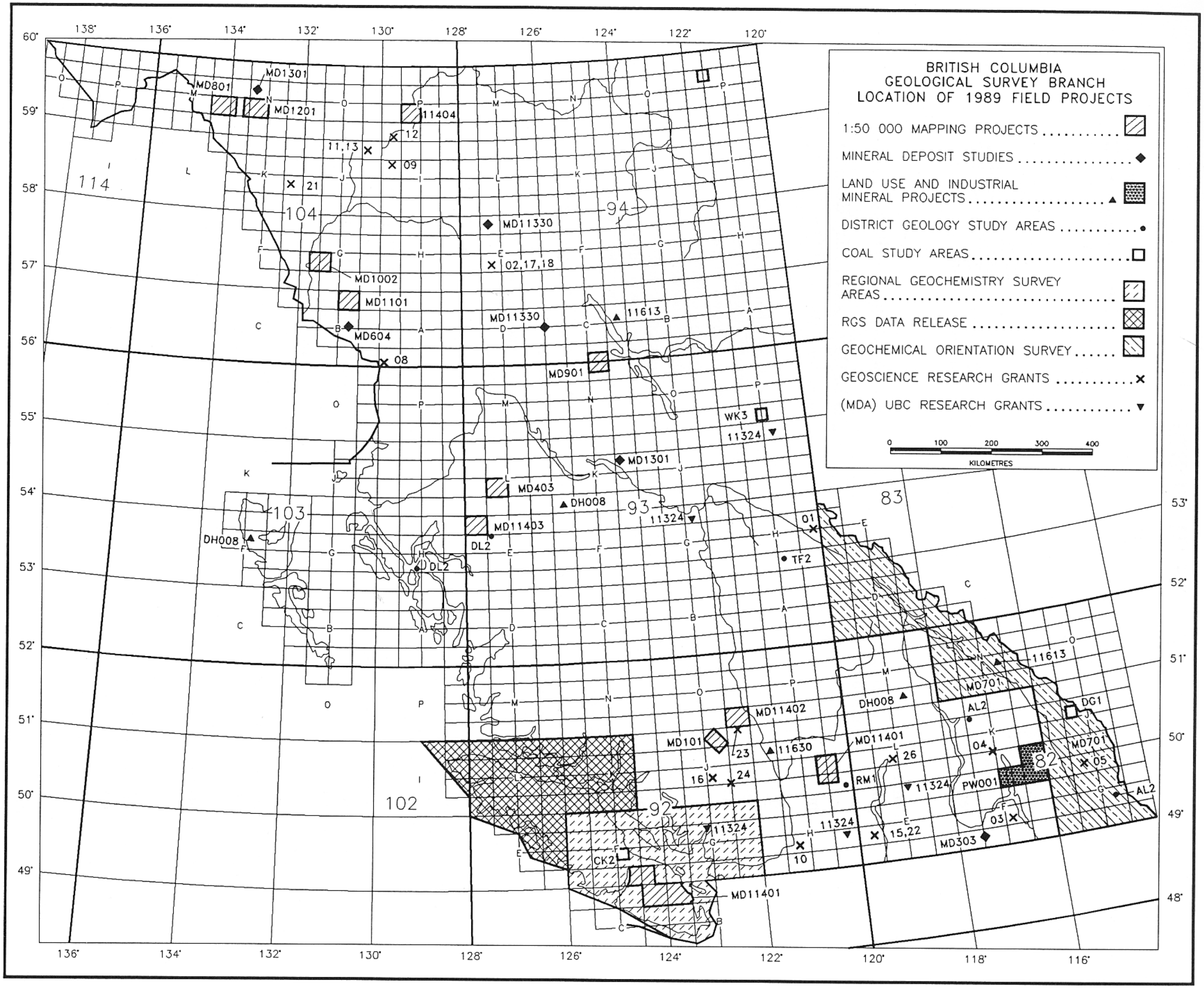
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GEOLOGICAL SURVEY BRANCH
LOCATION OF 1989 FIELD PROJECTS
1:50 000 MAPPING PROJECTS
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DISTRICT GEOLOGY STUDY AREAS
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Regional and District Mapping

STRUCTURE AND TECTONIC SETTING OF THE ROSSLAND GROUP, MOUNT KELLY – HELLROARING CREEK AREA, SOUTHEASTERN BRITISH COLUMBIA (82F/3W)

By Trygve Höy and Kathryn P.E. Andrew

KEYWORDS: Regional geology, Rossland Group, Archibald Formation, Elise Formation, Hall Formation, Mount Kelly syncline, Hellroaring Creek syncline, Waneta fault, shear-related gold deposits, vein deposits.

INTRODUCTION

The Mount Kelly–Hellroaring Creek area is situated in the Salmo map area south of the Nelson sheet (Little, 1960; Höy and Andrew, 1989a) and east of the Rossland–Trail sheet (Little, 1982; Andrew *et al.*, 1990a, b). The area (Figure 1-1-1) has been mapped at 1:50 000 scale by Little (1964) and formed the basis of a structural-stratigraphic thesis by Fitzpatrick (1985). This report focuses on a new structural interpretation of the area and on subdivision and interpretation of the Archibald and Elise formations of the Rossland Group. Regional correlations of the Rossland Group and tectonostratigraphic models are detailed in Andrew *et al. (op. cit.)*. A number of mineral occurrences in the area and favorable exploration results in Rossland Group rocks both to the north and in the Rossland camp have resulted in a considerable increase in the level of exploration, with the

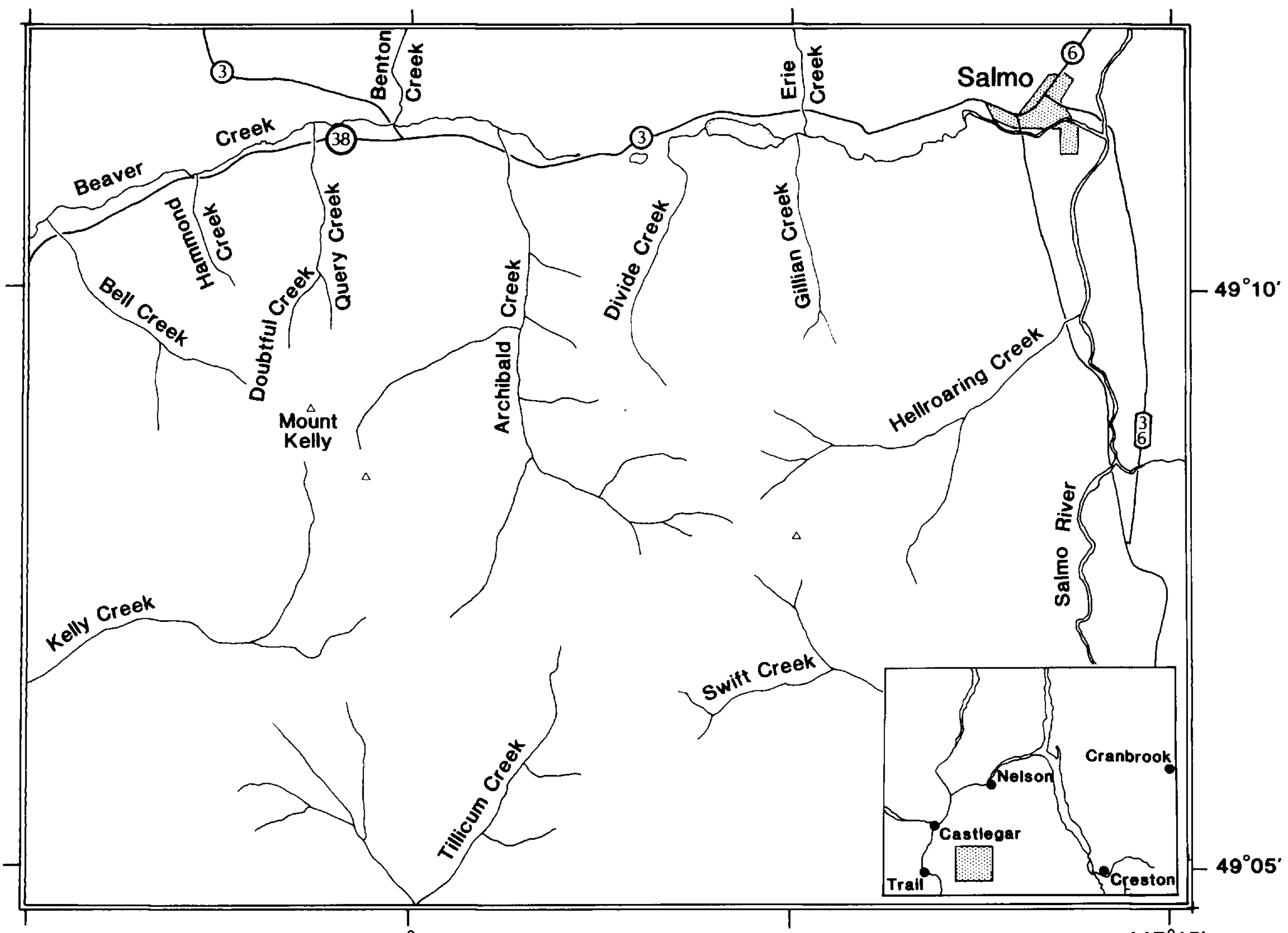


Figure 1-1-1. Location map showing physiographic features of the Mount Kelly–Hellroaring Creek area.

primary target being shear-related gold and copper mineralization.

ROSSLAND GROUP

The Lower Jurassic Rossland Group comprises dominantly clastic rocks of the Archibald Formation, overlain by a succession of volcanic rocks of the Elise Formation and clastic rocks of the Hall Formation. The base of the Rossland Group is not exposed in the Salmo map area. To the southeast, it is in contact along the Waneta fault with Paleozoic rocks of the Kootenay arc. To the north, in the Nelson map area, the Elise Formation is underlain by non-fossiliferous, fine-grained clastic rocks of the Ymir Group, which are correlative with the Archibald Formation. In the Rossland area, the Archibald Formation is missing (Fyles, 1984; Little, 1982) and Elise volcanic rocks unconformably overlie late Paleozoic rocks of the Mount Roberts Formation.

ARCHIBALD FORMATION

The Archibald Formation is exposed in a panel in the Divide Creek area east of Archibald Creek (Figure 1-1-2). Numerous bedding-cleavage intersections and sedimentary structures indicate that this panel is a homoclinal, west-facing succession. The base of the formation, exposed in the headwaters of the east fork of Archibald Creek, comprises dominantly argillite with thin interbeds of siltstone (Ja1). To the northeast, exposures are limited to scattered outcrops of altered argillite and siltstone near the faulted contact with the Elise Formation. The upper part of the Archibald comprises interbedded siltstone, argillite and minor conglomerate (Ja2). A prominent layer of plagioclase-rich lapilli and crystal tuff in the Archibald Formation (Jav) has been traced from the slopes south of Erie Lake in the northeast to the upper reaches of Archibald Creek in the central part of the map area (Figure 1-1-1).

A detailed section of the Archibald Formation, exposed in the hinge of the Mount Kelly syncline, is illustrated in Andrew *et al.* (1990a, this volume). It comprises essentially a basal succession of dominantly argillite, overlain by well-layered siltstone and wacke beds which are interpreted to have been deposited from turbidity currents. Farther west, near the western edge of the map area (Figure 1-1-2), conglomerates are prominent at a number of stratigraphic levels in the upper part of the Archibald Formation.

ELISE FORMATION

The Elise Formation comprises a thick succession of augite-phyric flows, tuffs, some epiclastic deposits and minor interbedded siltstone and argillite sequences. South of Nelson, on the east limb of the Hall Creek syncline, the formation is readily subdivided into a basal unit of mafic flows and flow breccias overlain by a thick accumulation of intermediate pyroclastic rocks (Andrew and Höy, 1988). On the west limb of the Hall Creek syncline, the distinction between the lower and and upper Elise is less evident (Höy and Andrew, 1989a). Here, the formation comprises primarily coarse, mafic pyroclastic breccia interlayered with minor flows and waterlain crystal tuff.

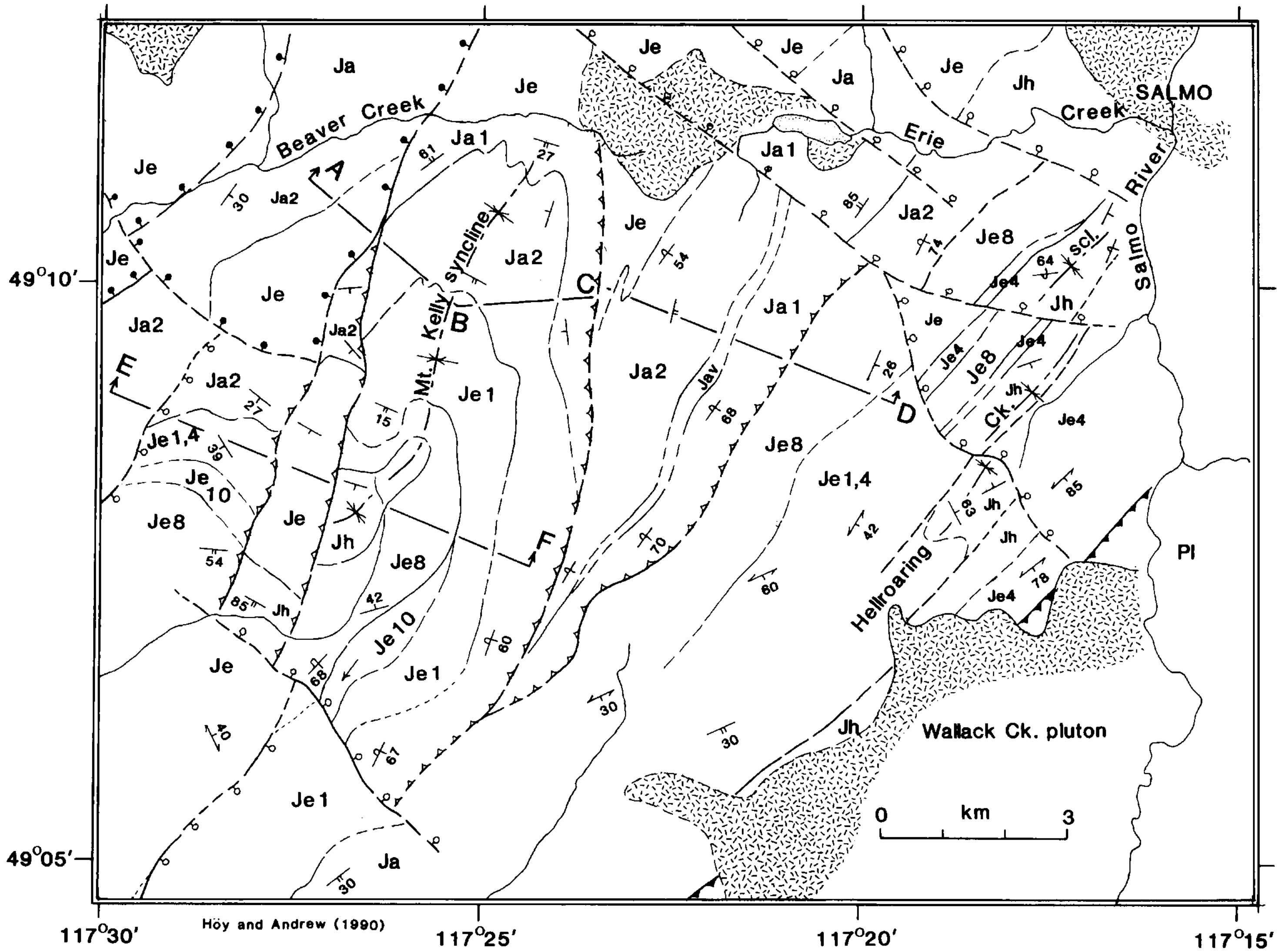
The distinction between the upper and lower Elise in the Salmo map area is apparent only in the east limb and hinge zone of the Mount Kelly syncline. Near the headwaters of Tillicum Creek, a thick succession of amygdaloidal, massive to pillowed augite porphyry flows and flow breccias (Je1) immediately overlies siltstones of the Archibald Formation. Locally, heterolithic lapilli tuff occurs near the base. The basal Elise is overlain by several hundred metres of interbedded argillite, siltstone, grit and polymictic pebble conglomerate (Je10), exposed on the switchbacks on the Tillicum Creek road.

The lower Elise at Mount Kelly (Je1) comprises mafic, locally pillowed flows (Plate 1-1-1), overlain by a succession of interlayered mafic tephra, tuffites (comprising mixed pyroclastic and epiclastic fragments) and epiclastic deposits. Chemical analyses of similar mafic volcanic rocks in the Nelson area indicate they are predominantly shoshonites (Höy and Andrew, 1989b; Beddoe-Stephens and Lambert, 1981). Pyroclastic deposits include well-bedded, commonly graded sequences of waterlain pyroclastic breccia, lapilli tuff and crystal tuff. Tuffaceous conglomerate and sandstone, locally with a calcareous cement, are prominent near the peak of Mount Kelly. Lahars, comprising poorly sorted, subrounded augite porphyry and plagioclase porphyry fragments, occur locally.

The upper Elise in the Mount Kelly syncline (Je8), comprises predominantly heterolithic lapillistone, lapilli tuff and pyroclastic breccia. The pyroclastic rocks contain subrounded to subangular clasts of plagioclase-rich volcanic and intrusive fragments and lesser augite porphyry in a fine tuffaceous to crystal-rich matrix.

In the Hellroaring Creek area, mafic flows and tuffaceous rocks are prominent in the upper part of the Elise, in contact with argillite and siltstone of the Hall Formation (Figure 1-1-2). However, considerable shearing occurs in these rocks and therefore their stratigraphic position is not exactly known. Structurally beneath this succession, northwest of Hellroaring Creek, are mixed mafic (augite-phyric) and intermediate (plagioclase-phyric) lapilli tuff, minor crystal tuff and some mafic flows. The structurally lowest succession, adjacent to the faulted Archibald contact, comprises predominantly intermediate tuffs (Je8) more typical of upper Elise rocks.

Plate 1-1-1. Amygdaloidal pillow basalt, Unit Je1, Elise Formation, southeast of Mount Kelly.



LEGEND

JURASSIC - CRETACEOUS

granite, granodiorite

LOWER JURASSIC

ROSSLAND GROUP

Jh — HALL FORMATION: argillite, siltstone

Je — ELISE FORMATION

Je10 — siltstone, argillite

Je8 — intermediate to mafic tuff

Je1,4 — mafic flows, tuff

Ja — ARCHIBALD FORMATION

Ja2 — Siltstone, conglomerate

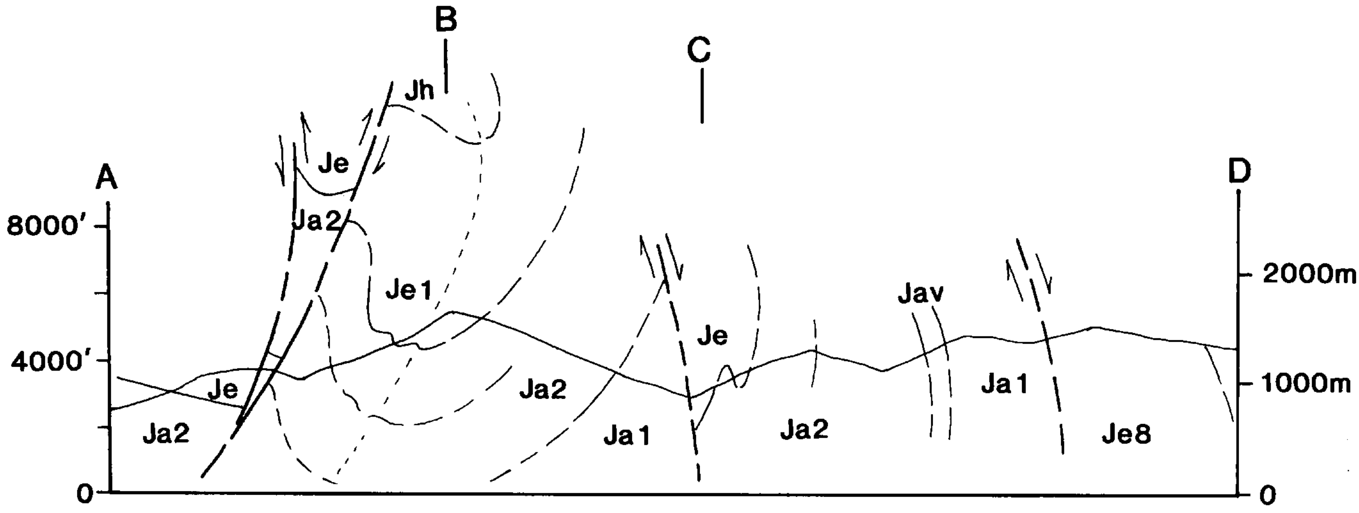
Ja1 — Argillite, siltstone

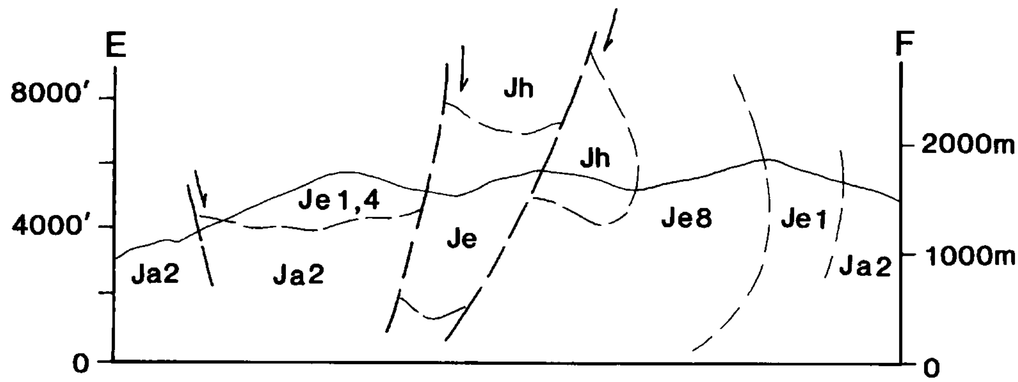
Jav — Lapilli tuff

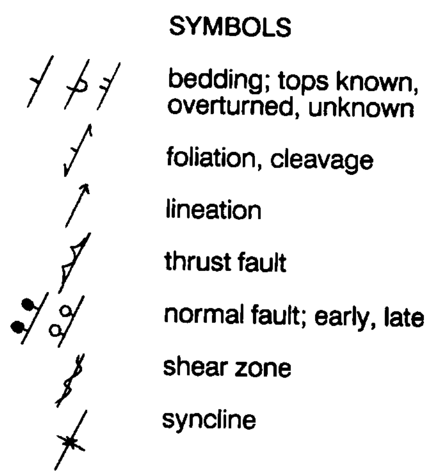
PALEOZOIC

Pl — LAIB FORMATION (?)

Figure 1-1-2. Geological map of the Mount Kelly–Hellroaring Creek area, Salmo map sheet, southeastern British Columbia (after Höy and Andrew, 1990; Fitzpatrick, 1985 and Little, 1964).







In summary, marked facies changes in the Elise Formation, similar to those recognized in the Nelson area to the north, characterize the Elise Formation in the Mount Kelly–Hellroaring Creek area. In the Mount Kelly area, as in the eastern part of the Nelson area, Elise rocks record inital effusive mafic volcanism followed by eruptions of mafic then more intermediate pyroclastic deposits. Northwest of Mount Kelly, and in the Hellroaring Creek area, intermixed mafic and intermediate pyroclastic deposits characterize much of the Elise succession. Epiclastic deposits near the middle of the formation comprise dominantly siltstone and argillite and record a hiatus in volcanic activity.

HALL FORMATION

The Hall Formation (Jh) is exposed in the cores of synclines in the Hellroaring Creek and Mount Kelly areas. Exposures in Hellroaring Creek comprise black, fissile, locally graphitic argillite with only minor thin siltstone laminations. Similar exposures occur in the Mount Kelly syncline. However, southwest of Mount Kelly, a fault-bounded structural panel of Hall rocks comprises a basal black argillite succession overlain by a succession of grit, arenite and wacke, siltstone and only minor argillite.

STRUCTURE

The earliest structures recognized in the Hellroaring Creek–Mount Kelly area are tight folds, locally associated with a penetrative mineral foliation and intense shearing and thrusting. In general, the intensity of this compressive strain increases to the southeast. The Waneta fault, near the southeast edge of the map area, is a steeply dipping, west-verging thrust fault that marks the boundary of Quesnellia with North American rocks. A number of essentially layer-parallel faults or shear zones in the vicinity of the Waneta fault are associated locally with an intense penetrative foliation. An overturned, east-dipping syncline with Hall Formation in its core and sheared Elise Formation in its limbs is exposed in the footwall of the Waneta fault. It is possible that this syncline, referred to as the Hellroaring Creek syncline, is the continuation of the Hall Creek syncline in the Nelson map area. The Hall succession on its eastern limb is locally overturned, dipping toward the southeast. Only a few stratigraphic tops were determined in the southeast-dipping Elise panel west of the syncline and these indicate that this panel is right-way-up. It is in contact with a west-facing Archibald succession in Gillian creek, necessitating a fault between the Elise and Archibald formations (Figure 1-1-2).

A large upright to overturned, south-plunging syncline, the Mount Kelly syncline (Fitzpatrick, 1985), is exposed in the Mount Kelly area (Section A-D, Figure 1-1-2). Archibald Formation in its hinge zone is gently to tightly folded and locally thickened (*see* Mt. Kelly ridge, northeast section; Andrew *et al.*, 1990a, this volume). The more competent Elise Formation rocks are concentrically folded and the Hall Formation in the core is strongly cleaved. The Mount Kelly syncline is cut by a steep west-dipping thrust fault that merges with a west-side-down fault to the north.

At least four generations of faulting are recognized. Intense shearing, particularly along the limbs of the Hellroaring Creek syncline and southeastward towards the Waneta fault, may be related to movement along the Waneta fault. This faulting predates intrusion of the Wallack Creek pluton; locally, however, mylonitic zones parallel to the Waneta fault, and associated shears developed in the pluton, indicate some post-intrusive movement.

North-trending, easterly verging thrust faults may be associated with development of more open folds such as the Mount Kelly syncline. These are later than shearing along the Waneta fault but are older than the intrusion of Late Jurassic plutons and normal faulting. They include the two thrust faults southwest of Mount Kelly and a fault in the Archibald Creek valley. The Archibald Creek thrust dips steeply to the west and places the west-facing Archibald succession on the east limb of the Mount Kelly syncline on an overturned succession to the east. A splay of the Archibald Creek thrust trends northeasterly towards Gillian Creek, juxtaposing exposures of the lower Archibald Formation (Ja1) against the Elise Formation (Je8). This eastern splay dips steeply to the east, perhaps due to overturning at higher structural levels or to regional tilting related to Eocene extensional tectonics (T.E. Irving, personal communication, 1989).

A number of north to northeast-trending, west-dipping normal faults occur northwest of Mount Kelly. They cut the early folds and faults but are cut and sealed by Late Jurassic intrusions. They may be southern extensions of the Red Mountain fault in the Nelson area (Höy and Andrew, 1989a). The most prominent, along Doubtful and Query creeks (Figure 1-1-1), follows the locus of an earlier thrust fault and juxtaposes Elise Formation on the west with Archibald Formation in Query Creek. It has a rectilinear shape, swinging sharply northwestward in Bell Creek, thus forming a down-dropped angular block. To the northwest in the Beaver Creek valley, a northeast-trending, west-side-down normal fault is inferred with folded Archibald Formation (Ja2) in the eastern footwall and Elise Formation (Je) in its hangingwall.

A complex array of late normal faults cuts all earlier structures. They offset intrusive rocks and some may be related to Eocene extensional tectonics documented to the northwest (Parrish *et al.*, 1988; Corbett and Simony, 1984). Steeply dipping, northwest-trending faults, down-dropped to the northeast, are prominent southwest of Salmo (Figure 1-1-2).

In summary, the earliest recognised structures in the Mount Kelly–Hellroaring Creek area involve west-verging thrusts, shears and tight folds along the Waneta fault, the tectonic boundary between Quesnellia and ancestral North America. Farther west, in the Mount Kelly and Archibald Creek areas, more open folds and east-verging thrusts are prominent. West-side-down normal faults that locally follow the trace of earlier thrusts, record extensional tectonics prior to the intrusion of Late Jurassic plutons. Late, post-intrusive, east and northeast-dipping normal faulting may be related to an Eocene extensional tectonic event.

MINERAL OCCURRENCES

Metallic mineral occurrences in the map area include shear-related conformable gold-copper zones and gold-silver-lead-zinc, gold-silver-copper and molybdenum veins (Table 1-1-1). Conformable gold occurrences, those where mineralization is conformable with either foliation or bedding, are closely associated with Rossland Group lithologies and early structures. Vein occurrences are post-tectonic and formed during intrusion of Middle Jurassic granites (Höy and Andrew, 1988).

The shear-related conformable gold-copper zones are characterized by intense carbonate-sericite-chlorite alteration in mafic flows and flow breccias. They occur southwest of Salmo in a zone of intense shearing more than a kilometre in width and extending from the headwaters of Tillicum Creek to the west fork of Hellroaring Creek (Figure 1-1-2). These alteration zones are 5 to 10 metres wide, over 20 metres long and are often cored by 1 to 2-metre sericite-silica altered felsic intrusions (or volcanics). Examples of these zones include the Gus and Jim showings in the headwaters of Hellroaring and Swift Creeks. Their structural and stratigraphic position, in the sheared limbs of a syncline cored by the Hall Formation (Figure 1-2-2), is a similar environment to conformable gold mineralization at the Great Western–Star property south of Nelson (Höy and Andrew, 1989c).

Vein occurrences are distributed throughout the Rossland Group and Middle Jurassic granitic intrusions. With the exception of the carbonate-hosted Silver Dollar (MINFILE 082FSW207), the veins are quartz-rich. Accessory silicate minerals include epidote, muscovite and tourmaline. Common sulphide minerals include pyrite, chalcopyrite, sphalerite and galena; molybdenum and minor tungsten occur at the Meadows showing (MINFILE 082FSW268) and arsenopyrite and stibnite at the Armstrong showing (MINFILE 082FSW267). The tenor of veins seems to have a lithologic control; precious metal rich veins that carry lead and zinc occur preferentially in sedimentary rocks of the Archibald or Hall Formations, copper-bearing veins are more

TABLE 1
MINERAL PROPERTIES WITH PAST PRODUCTION OR EXTENSIVE DEVELOPMENT, MOUNT KELLY-HELLROARING CREEK AREA

MINFILE No.	Name	Commodities	Type	Host	Status
82FSW207	Silver Dollar	Au,Ag,Pb,Zn	Vein	Jh	Past producer
82FSW267	Armstrong	Au,Ag,Pb,Zn	Vein	Jn	Past producer
82FSW268	Meadows	Mo	Vein	Jn	Showing
82FSW283	Allouez	Au,Ag,Cu	Vein	Je1	Showing
82FSW290	Jim	Au,Cu	Shear-related	Je1	Showing
82FSW291	Gus	Au,Cu	Shear-related	Je1	Showing

common in Elise Formation volcanic rocks and molybdenum-tungsten veins are restricted to late granitic intrusions.

EXPLORATION ACTIVITY

Ground underlain by the Rossland Group has had a steady increase in claims staked since the onset of the Rossland Project in 1987 (Figure 1-1-3). The area covered by claims on the Salmo west-half sheet alone has increased from 56 to 83 per cent. Exploration targets include conformable gold (shear-related and alkaline porphyry gold-copper), gold skarn and gold-silver-lead-zinc vein deposits. Although much of the staking may be due to increased activity generated by flow-through-share financing, re-evaluation and re-interpretation of classical vein and shear zone deposits has led to the development of new exploration models for past producers.

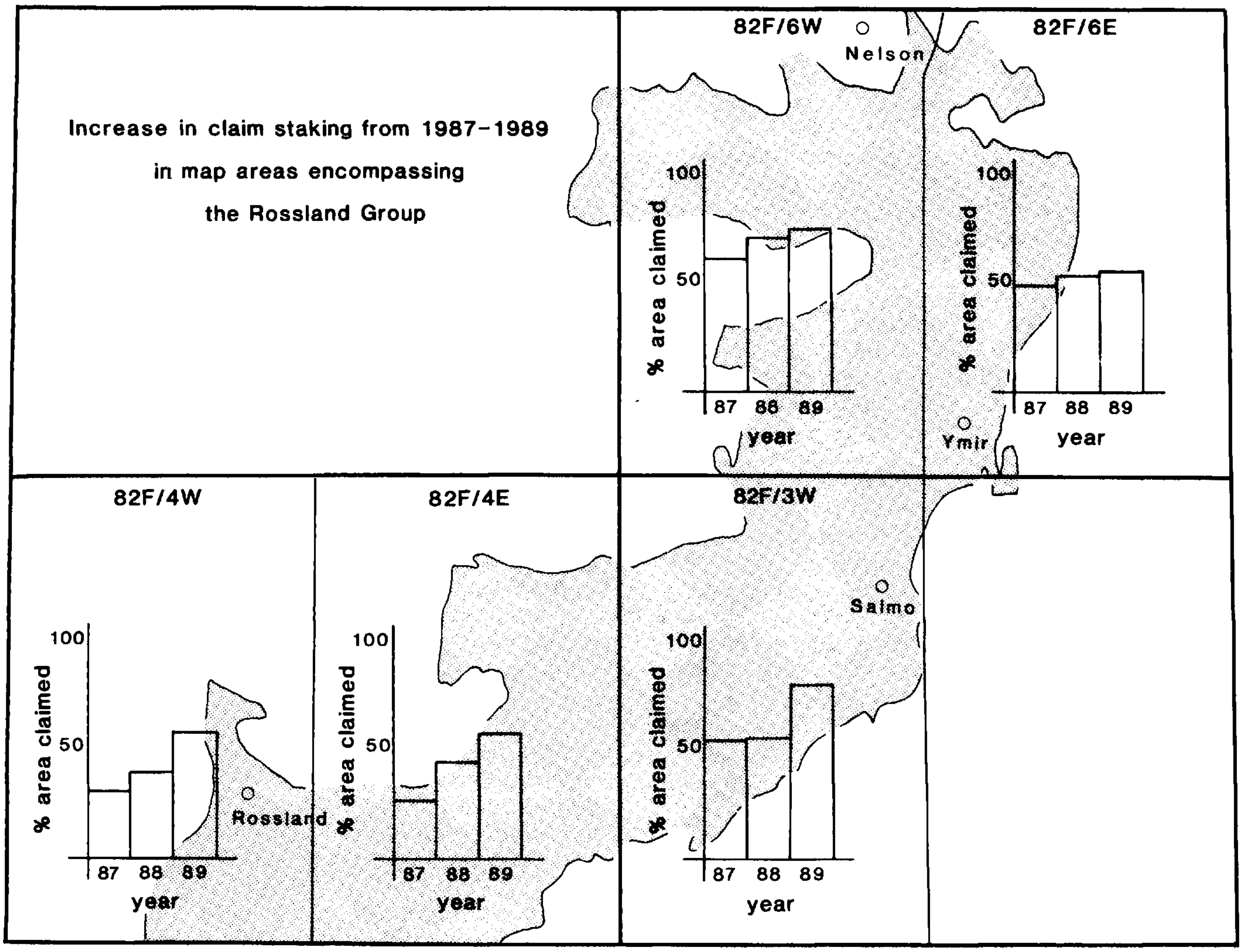


Figure 1-1-3. Map and graphs showing the increased level of staking in areas underlain by the Rossland Group, southeastern British Columbia, 1987 to 1989.

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NOTES

STRATIGRAPHY AND TECTONIC SETTING OF THE ARCHIBALD AND ELISE FORMATIONS, ROSSLAND GROUP, BEAVER CREEK AREA, SOUTHEASTERN BRITISH COLUMBIA (82F/4E)

By Kathryn P.E. Andrew, Trygve Höy and John Drobe

KEYWORDS: Regional geology, Rossland Group, Archibald Formation, Elise Formation, Hall Formation, conglomerate, turbidite, submarine fan, growth faults, volcanism, Champion Lake fault, Beaver Creek fault.

INTRODUCTION

The 1989 Rossland project focused on 1:20 000 regional mapping of the Rossland Group and study of selected mineral deposits in the Rossland–Trail east-half and Salmo west-half map areas (Figure 1-2-1). The mapping complements previous work in the Nelson map area to the northeast (Höy and Andrew; 1988, 1989a, b). The aim of the Rossland project is to develop a better understanding of the stratigraphic and tectonic setting of the Rossland Group and the controls of gold and silver mineralization. The project includes systematic whole-rock and trace element analyses of volcanic rocks, and fluid inclusion, stable and radiogenic isotope studies of mineral occurrences. This is the third year of the project and completion of mapping in 1990 is planned to cover the Rossland camp and south to the U.S. border.

This report deals primarily with overall distribution of the Archibald and Elise formations and facies changes within them. The geology of Rossland Group rocks in the Salmo area to the east is described in Höy and Andrew (1990a, b).

Regional mapping in the Rossland–Trail map area (Little 1960, 1962, 1982) included the Beaver Creek area (Figure 1-2-1). A geochemical and petrological study of volcanic rocks in the Rossland Group has been completed by Beddoe-Stephens and Lambert (1981) and Beddoe-Stephens (1982). The Beaver Creek area has been mapped most recently by Fitzpatrick (1985). The area to the south of Beaver Creek, in the vicinity of the Waneta fault zone, is currently being mapped as part of a Ph.D. thesis by J. Einerson, University of Calgary.

STRATIGRAPHY

The Rossland Group in the Beaver Creek area comprises a basal succession of fine and coarse-grained clastic rocks of the Archibald Formation, volcanic and epiclastic rocks of the Elise Formation, and overlying fine-grained clastic rocks of the Hall Formation (Figure 1-2-2). These rocks are Early Jurassic in age, bracketed by Sinemurian fossils in the Archibald (Frebold and Tipper, 1970; Tipper, 1984) and Pleinsbachian and Toarcian macrofossils in the Hall (Frebold and Little, 1962; Tipper, 1984). The Rossland Group is intruded by numerous small stocks that are probably correlative with either the Middle to Late Jurassic Bonnington or Trail plutons (Ghosh, 1986) and also by many diorite or rhyolite dikes and Coryell alkalic intrusions of Eocene age.

ARCHIBALD FORMATION

The Archibald Formation, named after exposures in Archibald Creek southwest of Salmo, is the lowermost unit of the Rossland Group. It generally comprises a succession of interbedded siltstones, sandstones and argillites with prominent sections of interbedded coarse conglomerate. The formation is exposed as a west-facing homoclinal panel on the slopes northwest of Montrose and Fruitvale and in the core of an anticline on the slopes opposite Marsh and Hudu creeks. In the Salmo area, it is exposed in the limbs of the Mount Kelly syncline and as east and west-facing panels east of Archibald Creek (Höy and Andrew, 1990a, this volume).

Previously referred to as "the Sinemurian beds", the Archibald Formation has yielded several macrofossil collections with *Arnioceras* indicating both early and late Sinemurian ages (Little, 1960; 1964; 1982; Tipper, 1984). Several new ammonite localities discovered during the 1989 season may further constrain the age of the formation. The base of the formation is not exposed; it is cut by either faults or by Middle Jurassic intrusions.

Detailed stratigraphic sections of the Archibald Formation in the Trail–Salmo–Nelson map areas are shown in Figure 1-2-3. The total exposed thickness of the formation varies from 825 to 2550 metres (Figure 1-2-4). It is thickest in the Mount Kelly–Archibald Creek–Gillian Creek sections and thins substantially to the west near Fruitvale and to the north in the limbs of the Erie Creek anticline (Höy and Andrew, 1989b). The oldest exposed rocks in the formation, the basal part of the Mount Kelly section, are dark grey to black, rusty weathering argillite and minor siltstone over 600 metres thick. They are massive to finely laminated. An interbedded sequence of thick-bedded, graded wacke, siltstone and silty argillite comprises the upper 1500 metres of the Mount Kelly section. Graded beds, sharp basal contacts and basal scours suggest that the sequence consists largely of turbidites (Plate 1-2-1). Crossbedding is seen only rarely (Plate 1-2-2). A number of augite porphyry sills or flows (?) occur at the top of the section, near the contact with the overlying Elise Formation.

In the Beaver Falls and Copper Creek sections, the lower argillite unit is either missing, not exposed or replaced by a coarser facies (Figure 1-2-3). The Beaver Falls section generally coarsens upward with approximately 540 metres of massive thick-bedded siltstone at the base, overlain by 300 metres of dominantly graded pebble sandstone and 430 metres of chaotic matrix-supported conglomerate and pebble sandstone. The conglomerate is poorly lithified and contains approximately 10 per cent limestone and siltstone clasts

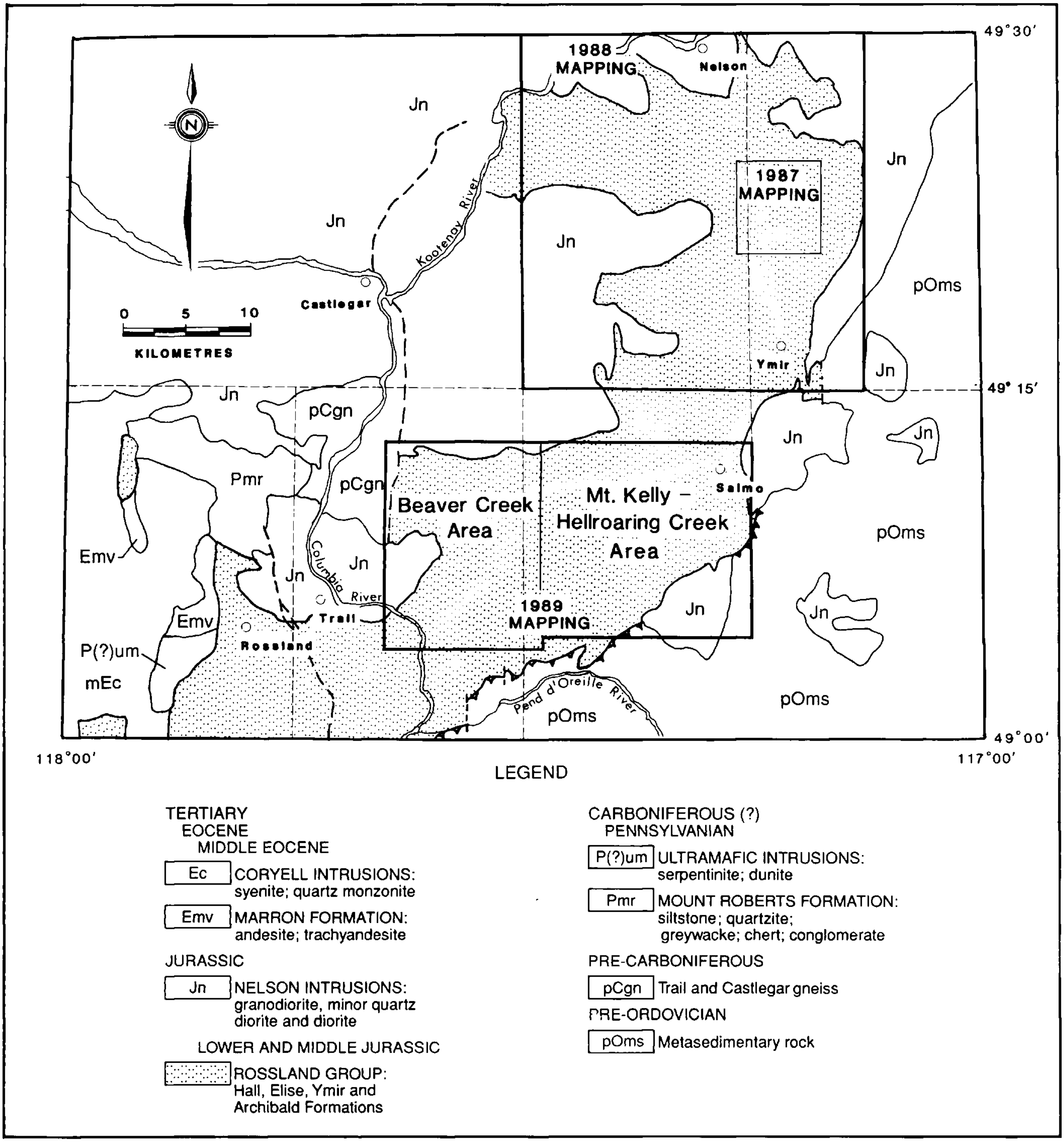


Figure 1-2-1. Location map and main physiographic features, Beaver Creek area, southeastern British Columbia.

(Plate 1-2-3); fossils in the limestone clasts have been identified as Permian, indicating that they are probably derived from the Mount Roberts Formation (Little, 1982). Similar limestone clasts occur in a pyroclastic breccia at the base of the Elise Formation. The Beaver Falls section is capped by 100 metres of maroon siltstone and pebble sandstone. Although mafic sills are absent from the upper part of the Beaver Falls section, they do occur in the Copper Creek section.

The Copper Creek section (Figure 1-2-3) is characterized by 870 metres of thin-bedded turbidite sandstone and intercalated thick-bedded siltstone and argillite. The sequence is lithologically similar to the upper part of the Mount Kelly section but is generally more thinly bedded.

Facies changes in the Archibald Formation are summarized in the stratigraphic columns of Figure 1-2-4. A wedge of very fine grained clastic rocks characterizes the base of the formation in the Mount Kelly block. This wedge pinches out to the west, supplanted and overlain by graded, coarse and fine clastic beds. Laterally discontinuous coarse clastics occur near the top of the formation in the Beaver Falls and Park Siding blocks.

The contact between the Archibald and Elise is gradational. It is mapped (Höy and Andrew, 1989a; Figure 1-2-2) where fine-grained interbedded siltstones and argillites with occasional thin flows give way to massive augite porphyry flows with fine argillaceous partings. In the Beaver Creek area, the upper(?) Archibald Formation contains a few mafic flows or sills but farther east and northeast in the Mount Kelly–Hellroaring Creek and Erie Creek areas, flows are more abundant.

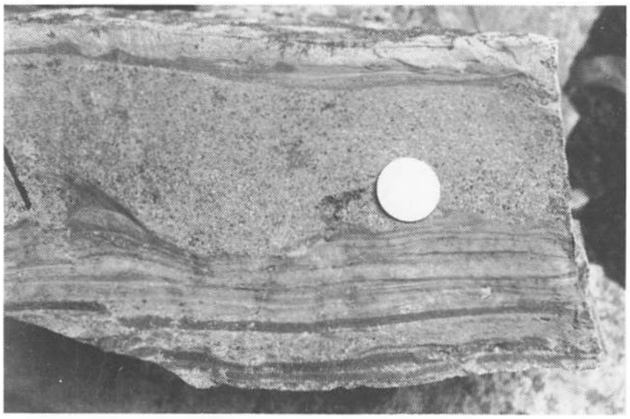


Plate 1-2-1. Thick-bedded, graded turbidite wacke with sharp basal contacts and basal scours, Archibald Formation, southwest of Salmo.

Plate 1-2-2. Crossbedding in thick-bedded, graded turbidite wacke, siltstone and silty argillite, Archibald Formation, east of Mount Kelly.

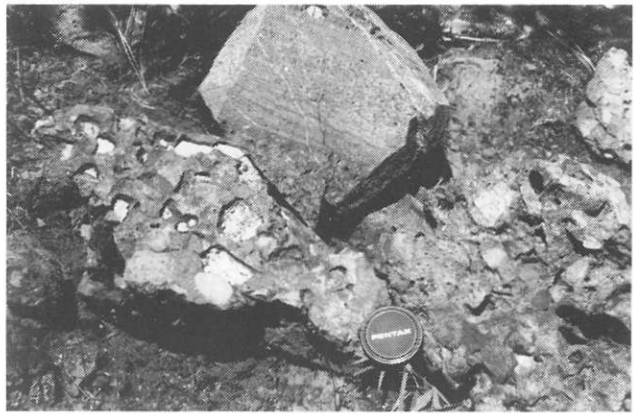


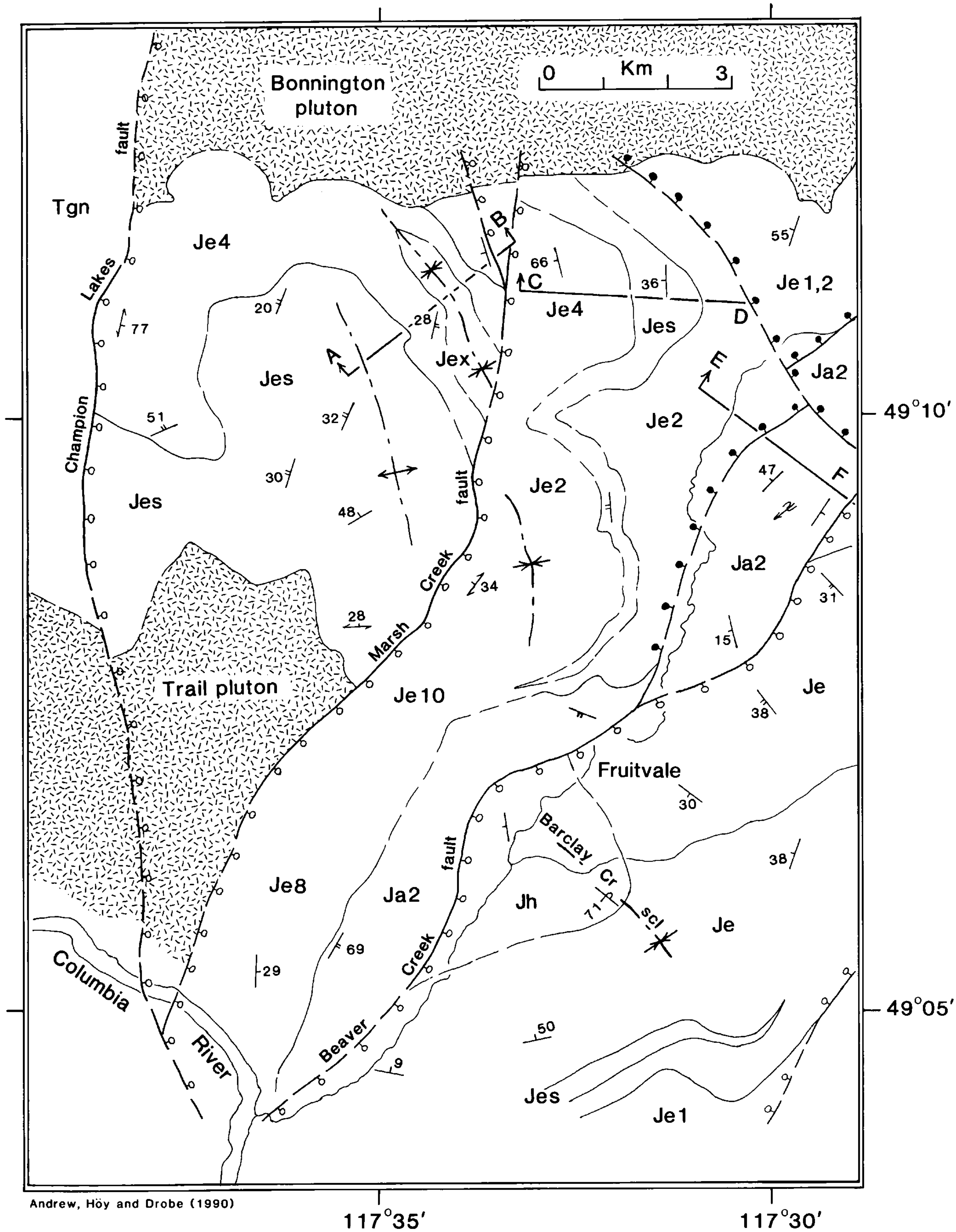
Plate 1-2-3. Matrix-supported conglomerate with limestone and siltstone clasts (0.5-10 centimetres in diameter), Archibald Formation, Beaver Creek area.

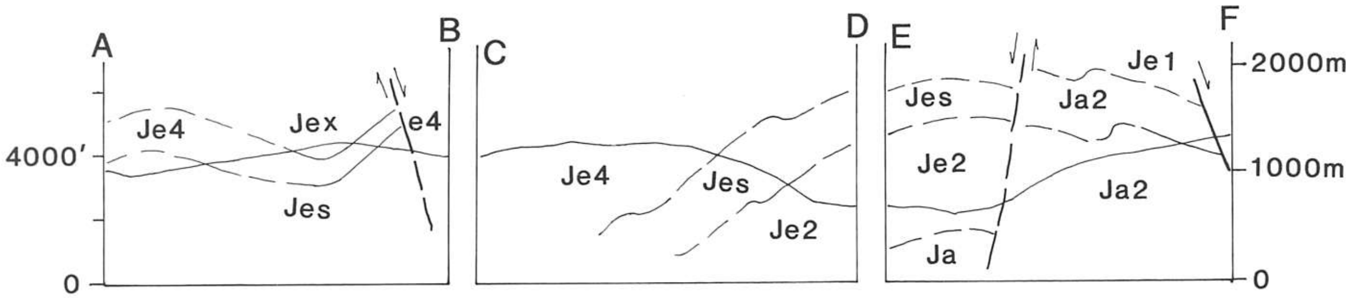
FACIES TRENDS AND INTERPRETATION

Systematic facies changes in the Archibald Formation, from dominantly conglomerates in the westernmost sections, through intermixed conglomerates and sandy turbidites in the Mount Kelly area, to fine-grained clastics in correlative Ymir Group rocks south of Nelson, indicate deposition in a large submarine fan with a western source area. The coarse clastic facies, restricted to the Beaver Falls–Fruitvale area, are laterally discontinuous subaqueous debris flows or slumps deposited on higher angle slopes than the classical turbidity current deposits. Thick sequences of A-E turbidites in the Mount Kelly–Hellroaring Creek and Erie Creek areas were deposited in a middle to lower fan environment.

The coarse conglomerates, rapid facies changes and locally subaerial siltstone and pebble sandstone in the Beaver Falls-Fruitvale area suggest deposition near a faulted basin margin. Absence of the Archibald Formation west of the Columbia River, with Elise Formation resting unconformably on Paleozoic Mount Roberts Formation (Little, 1982; Fyles, 1984) indicates a tectonic high. Erosion of an uplifted fault block west of Montrose, exposing older Mount Roberts Formation, provides a source for clasts of Mount Roberts limestone in fanglomerates of the Archibald Formation.

In summary, a model for the deposition of the western facies of Archibald Formation by growth faulting is proposed. The argillite succession in the lower Archibald was deposited as muds in a marine basin that stretched from Archibald Creek to Ymir and perhaps northward to include part of the Slocan Group. The initiation of growth faulting near the southwestern basin margin produced coarse clastic facies in the Montrose–Fruitvale area and proximal turbidites in the Mount Kelly–Hellroaring Creek and Erie Creek areas. These Early Jurassic growth faults are located between the basin margin near Montrose and the uplifted tectonic high near Rossland. Final subaerial deposition in the Montrose–Beaver Falls area produced a more oxidized facies immediately prior to deposition of Elise Formation pyroclastic breccia.


Bonnington pluton
0 Km 3
Tgn
Je4
fault
Lakes
Champion
Jes
Jex
Je2
Je1,2
Ja2
Jes
Trail pluton
Marsh Creek fault
Je10
Je
Fruitvale
Barclay Cr
Je8
Ja2
Jh
Je
Columbia River
Beaver Creek fault
Jes
Je1
A
B
C
D
E
F
49°10′
49°05′
117°35′
117°30′
Andrew, Höy and Drobe (1990)



Andrew, Hoy and Drobe (1990)

LEGEND

JURASSIC-CRETACEOUS

granite, granodiorite

LOWER JURASSIC

ROSSLAND GROUP

Jh — HALL FORMATION: argillite, siltstone

Je — ELISE FORMATION

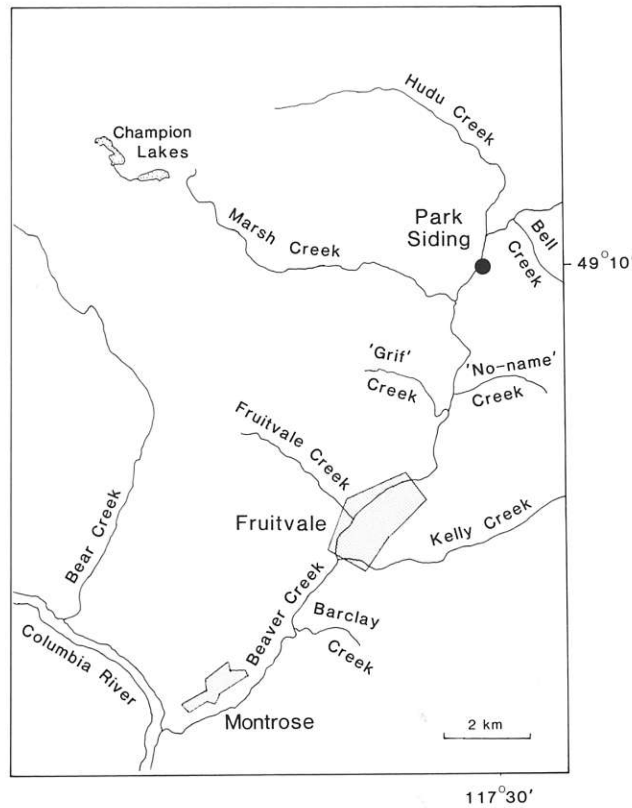
- Je10 — siltstone, argillite
- Je8 — intermediate to mafic tuff
- Jes — siltstone
- Je2 — mafic pyroclastic breccia
- Je1,4 — mafic flows, tuff

Ja — ARCHIBALD FORMATION

- Ja2 — siltstone, conglomerate

PALEOZOIC

Tgn — TRAIL GNEISS



SYMBOLS

bedding; tops known, overturned, unknown

foliation, cleavage

minor fold

normal fault; early, late

syncline

Figure 1-2-2. Geological map of the Beaver Creek area, Rossland-Trail east-half map sheet (after Andrew *et al.*, 1990; Fitzpatrick, 1985; Little, 1962, 1982).

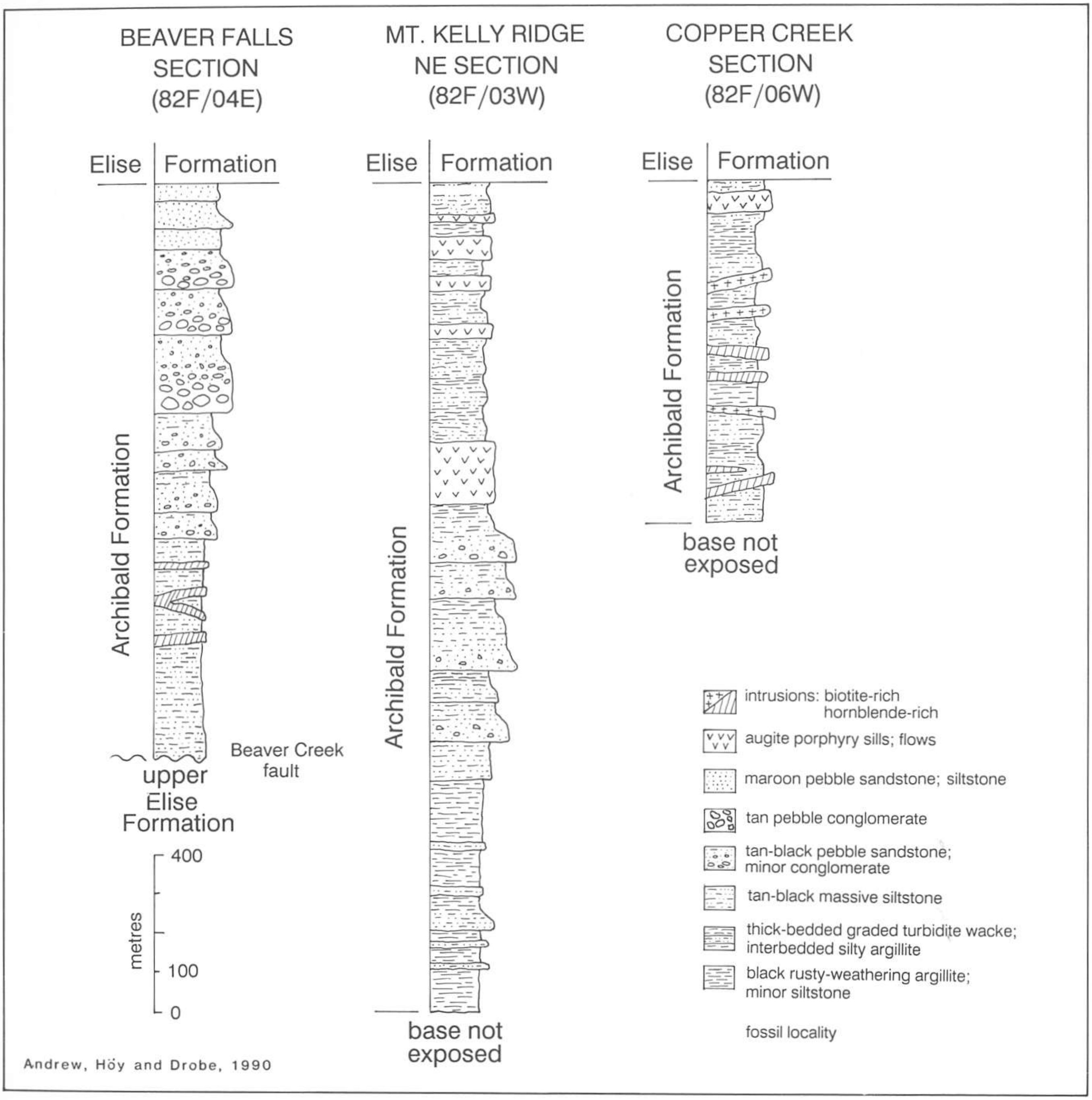


Figure 1-2-3. Stratigraphic sections of the Archibald Formation, Beaver Creek, Mount Kelly and Erie Creek areas.

ELISE FORMATION

The Elise Formation, named for exposures on the western slopes of Mount Elise south of Nelson, is characterised by massive and brecciated flows, subvolcanic porphyries, pyroclastic, tuffite and minor epiclastic deposits. The broad subdivision of the Elise Formation into a lower succession of mafic flows and an upper succession of intermediate to mafic pyroclastic rocks, recognized locally in the Nelson area (Andrew and Höy, 1988), is also applicable in the Beaver Creek, Salmo and Tillicum Mountain areas (Höy and Andrew, 1990a, b; W. Roberts, personal communication, 1988).

Although the Elise Formation conformably overlies the Archibald Formation in the Beaver Creek area, fossil evidence and, as previously mentioned, lithologic correlation indicate that the contact shows temporal transgression on a regional scale. Sinemurian fossils, diagnostic of the Archibald Formation, are also found in siltstone intervals in the basal part of the Elise Formation (Little, 1962, 1982).

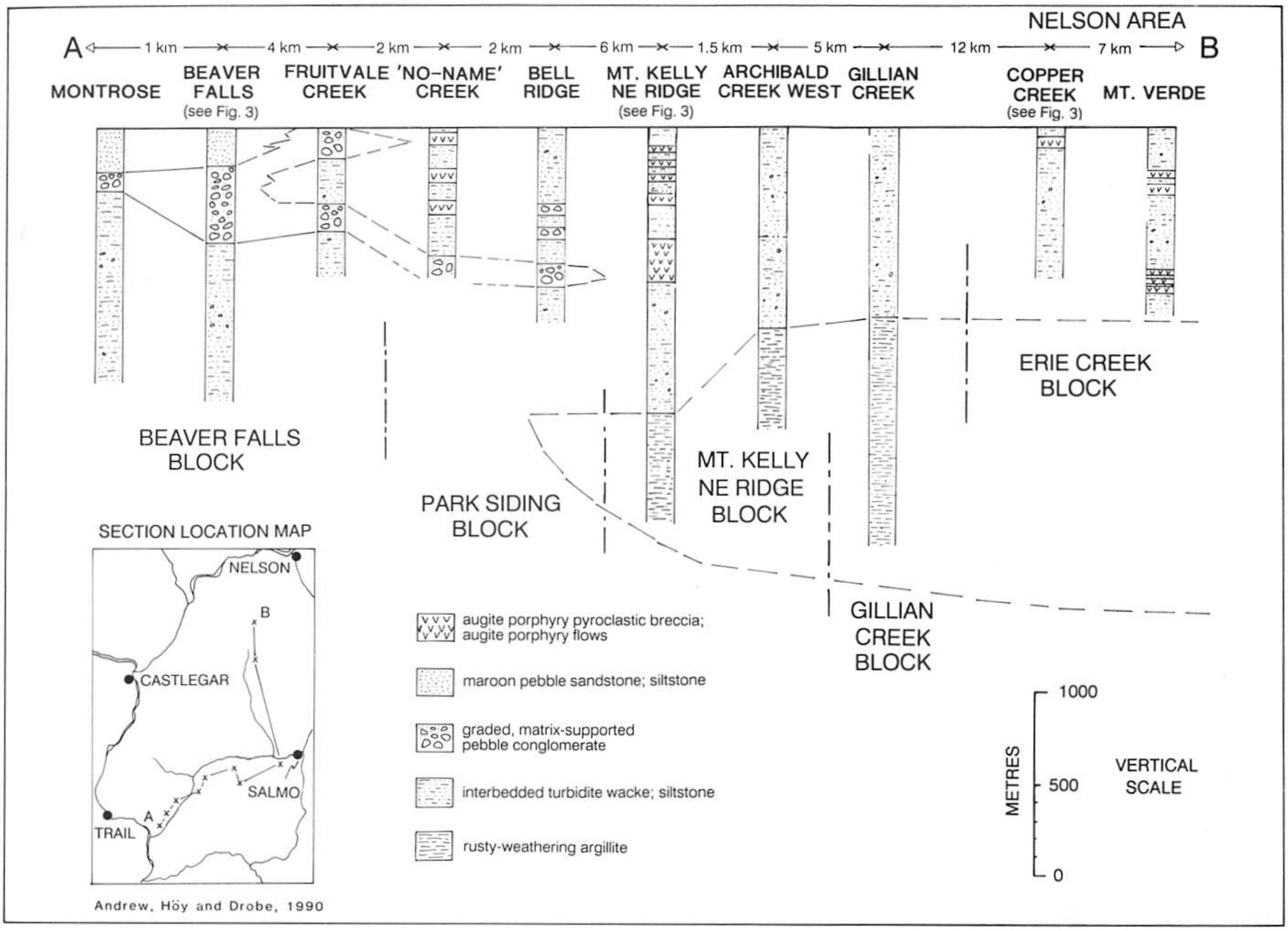
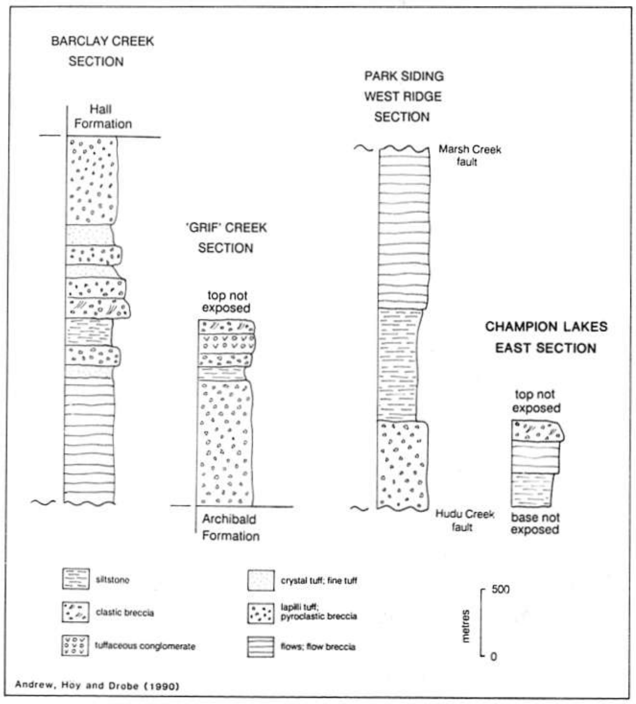


Figure 1-2-4. Correlation chart of the Archibald Formation showing main lithologic and thickness changes.

The basal part of the Elise Formation conformably overlies the Archibald Formation in the Beaver Creek area. Lateral facies changes are rapid; it comprises mafic pyroclastic breccia north of Fruitvale whereas intermediate pyroclastic breccia predominates west of Montrose (Figure 1-2-2). Elsewhere the base of the Elise is not exposed or is cut by faults. The top of the formation is exposed in a broad syncline southeast of Fruitvale. It comprises interlayered mafic and intermediate lapilli tuff, ash tuff and a prominent section of interbedded siltstone and argillite that is conformably overlain by the Hall Formation.

Facies changes in the Elise Formation in the Beaver Creek area complicate division into lower and upper members. Near Barclay Creek, southeast of Fruitvale, the formation may be separated into a lower section of augite porphyry flows and flow breccia approximately 950 metres thick, overlain by 1800 metres of lapilli, crystal and fine tuff (Figure 1-2-5). Autoclastic fragments in flow breccia outcrops at the headwaters of Nine Mile Creek and north of Doubtful Creek contain calcite-filled amygdules.

Near 'Grif' Creek north of Fruitvale, however, the distinction between upper and lower Elise is less evident. The total thickness of the formation here is approximately 1400 metres. It comprises dominantly mafic pyroclastic breccia overlain by tuffaceous conglomerate and clastic breccia



containing several per cent angular siltstone and volcanic clasts. In contrast, the Elise Formation in the Park Siding section to the northeast is characterised by at least 2750 metres of mafic pyroclastic and effusive rocks interbedded with a thick succession of siltstone (Figure 1-2-2). In the Champion Lakes area, exposures of the Elise Formation are flat-lying to gently folded and comprise flows, flow breccias and clastic breccias that overlie a succession of thinly bedded and finely laminated siltstone of unknown thickness (Figures 1-1-2 and 5). It is possible that this siltstone succession, included as part of the Elise Formation (Little, 1982; Figure 1-1-2), may be the upper part of the Archibald Formation.

Growth faulting, initiated during deposition of the Archibald Formation, probably continued in the early Elise. The occurrence of Mount Roberts limestone clasts and granitic clasts in pyroclastic breccia at the base of the Elise Formation indicates proximity to an uplifted fault block or tectonic high to the west. Volcanic facies in the lower Elise indicate explosive volcanism began in the 'Grif' Creek and Park Siding west ridge areas while effusive eruptions occurred in the Champion Lakes, Barclay Creek and 'No-name' Creek areas.

HALL FORMATION

Approximately 300 metres of the Hall Formation is exposed in the core of the Barclay Creek syncline south of Fruitvale. It rests conformably on upper Elise mafic pyroclastic breccia, lapilli tuff and fine tuff and is characterised by dominantly black carbonaceous argillite and tan siltstone, similar to the basal part of the formation in the Nelson and Salmo areas. Coarse clastic facies that typify much of the upper Hall Formation are not exposed here. The formation is dated by Toarcian pelecypods found on Bath Creek road southwest of Fruitvale (Little, 1982).

STRUCTURE

The Beaver Creek area lies north of the Waneta fault, a thrust that juxtaposes Rossland Group rocks in Quesnellia against North American miogeoclinal rocks, and east of the southern extension of the late Eocene Champion Lakes–Slocan normal fault. Two periods of pre-Middle to Late Jurassic compressional tectonics are recognised to the southeast (Höy and Andrew, 1990a). Near the Waneta fault, northeast-trending tight folds, associated shearing and a penetrative foliation are related to shearing along the fault. A series of northeast-trending folds in the Archibald Formation, in a fault-bounded structural block northeast of Fruitvale, may also be a result of this early deformation (Figure 1-2-2).

More open, south to southeast-trending folds with an axial planar cleavage and, locally, associated crenulation cleavage are superposed on the earlier folds (Plate 1-2-4). They are cut by two generations of normal faults and by the Late Jurassic plutons.

A northeast-trending, west-dipping normal fault follows the Beaver Creek valley north of Fruitvale. It juxtaposes Elise Formation in its hangingwall against folded Archibald Formation to the southeast. It is cut by a northwest-trending fault that parallels Hudu and Bell Creeks (Figure 1-2-2; Höy and

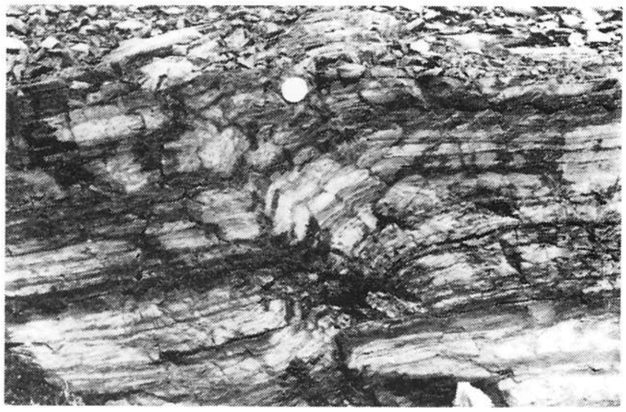


Plate 1-2-4. Crenulation cleavage superposed on early folds in interbedded siltstone and argillite, Elise Formation, Champion Lakes area.

Andrew, 1990a, this volume, Figure 1-2-2), which is also assumed to be an early (pre-intrusion) normal fault.

The latest structures are post-intrusive, northeast-trending normal faults and north-trending dikes. The Champion Lake fault (Little, 1962; Simony, 1979) is the southern extension of the Slocan fault (Parrish, 1984; Parrish *et al.*, 1988) along which east-side-down normal displacement decreases systematically southward, from approximately 10 kilometres at Slocan Lake (Carr *et al.*, 1987) to 1 to 2 kilometres near the Trail pluton (Corbett and Simony, 1984) to its termination in the Columbia River valley north of the Waneta fault. Some of this displacement may be transferred to a complex, but dominantly northeast-trending array of normal faults in the Beaver Creek area (Figure 1-2-2) and to the east in the Mount Kelly area (Höy and Andrew, 1990a, b). The Marsh Creek fault is a splay of the Champion Lake fault that forms the eastern boundary of the Trail pluton and cuts second generation, northwest-trending folds farther north (Figure 1-2-2). The Beaver Creek fault to the southeast has several kilometres of displacement along it, locally placing Hall Formation against Archibald Formation. North-trending diorite and rhyolite dikes of Eocene age(?) form a swarm extending from Montrose to just east of Champion Lakes (Andrew *et al.*, 1990).

In summary, structures in the Rossland–Trail east-half area record at least four tectonic events. Tight folds, shears and a penetrative foliation are associated with the Waneta thrust; open south-plunging folds superposed on these earlier folds record continued compressive strain. They are cut by west-side-down normal faults, possible extensions of the Red Mountain fault in the Nelson area (Höy and Andrew, 1989a). These structures predate the Late Jurassic Bonnington, Trail and Nelson plutonic rocks. East-dipping normal faults, including the Champion Lake fault, and north-trending dikes record an Eocene extensional event that is related to development of core complexes along the eastern margin of the Shuswap complex.

MINERAL OCCURRENCES

Despite potential for shear-related conformable gold, gold-copper skarn, and precious metal vein deposits, little systematic exploration has been documented in the Beaver Creek area prior to 1988. The Canada Day Red showing, a north-trending zone of rusty, pyritic, chlorite-epidote-muscovite-altered tuffaceous conglomerate and clastic breccia over a kilometre long, occurs at the headwaters of 'Grif' Creek. Geophysical, geochemical and geological surveys of this zone were undertaken by Noranda Exploration Company, Limited in 1988 and 1989. Elise Formation siltstone is locally hornfelsed and pyritized around Tertiary diorite dikes east of Montrose and southeast of Champion Lakes. Several small abandoned workings were found during the course of regional mapping (Andrew *et al.*, 1990); typically these are mineralized shear zones and quartz veins a few centimetres to a metre wide that contain variable sulphide content.

ACKNOWLEDGMENTS

We would like to thank Heather Blyth and Darryl Lindsay for their reliable and enthusiastic assistance during field mapping. Discussions with W. Howard, A. Skupinski and G. Gill provided useful information on the Rossland Bear claims. Field trips and discussions with J.T. Fyles, P.S. Simony, W.J. McMillan, J. Einerson and R. Robeck were both helpful and stimulating; the manuscript was reviewed and edited by J.M. Newell, B. Grant and W.J. McMillan.

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NOTES

GEOLOGY AND MINERAL POTENTIAL OF THE PURCELL WILDERNESS CONSERVANCY (82F/16; 82K/1, 8)

By G.P. McLaren, G.G. Stewart and R.A. Lane

KEYWORDS: Regional geology, Purcell Wilderness Conservancy, mineral potential, Purcell Supergroup, geochemistry, sedimentary exhalative deposits, veins, skarns.

INTRODUCTION

The Purcell Wilderness Conservancy was created in 1974 to preserve approximately 1320 square kilometres of the Purcell Mountains as a roadless tract of recreational wilderness area. It is located in the rugged mountain range between Kootenay Lake and the Columbia Valley, centred approximately 30 kilometres north of Kimberly (Figure 1-3-1) and is adjoined on the south by St. Mary's Alpine Provincial Park. Resource development, including mineral exploration and mining, was prohibited and any existing claims were frozen by a mineral reserve when the conservancy was established. Prospecting and mining are traditional land uses in this region of British Columbia as mineral discoveries were made at the turn of the century and continue to be made in the 1980s. The stratabound Sullivan massive sulphide orebody is located 30 kilometres to the southeast and the rocks hosting the orebody are known to extend into the conservancy. Despite this history of exploration and mining in the region, no systematic assessment of the mineral potential of the Purcell Wilderness Conservancy was undertaken prior to withdrawing it from the exploration land base.

In 1986 the Wilderness Advisory Committee studied the conservancy and recommended that resource assessments, specifically including a mineral potential study, should be completed prior to any final boundary decisions (Wilderness Advisory Committee, 1986). The Geological Survey Branch then summarized existing geological knowledge of the area preparatory to planning a mineral potential study (Grant, 1987). The Ministry of Parks subsequently identified a planning area for resource assessments surrounding the conservancy (Ministry of Parks, 1989). In the 1989 field season a mineral potential evaluation was initiated to provide the information required to settle the outstanding issues of mineral resource management in the conservancy.

The 1989 field project focused on the eastern side of the conservancy and the adjacent planning area (Figure 1-3-2). Work included geological mapping and a detailed stream-sediment survey together with prospecting and rock chip geochemical sampling. Approximately 1000 square kilometres were covered by the stream sediment sampling survey. Mapping at 1:50 000-scale was completed in much of this area, however, gaps remain where previous mapping and assessment report data have been assimilated into the summary map. Rock chip samples were collected wherever

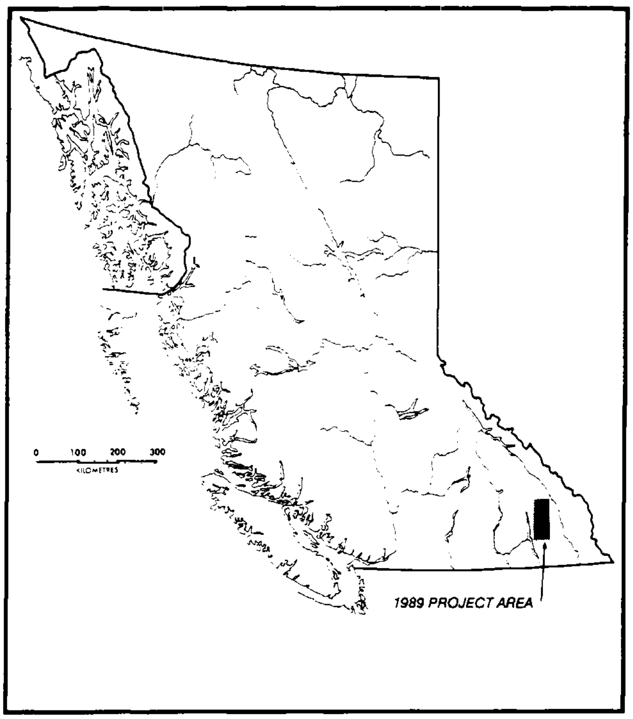

Figure 1-3-1. Location of 1989 study area.

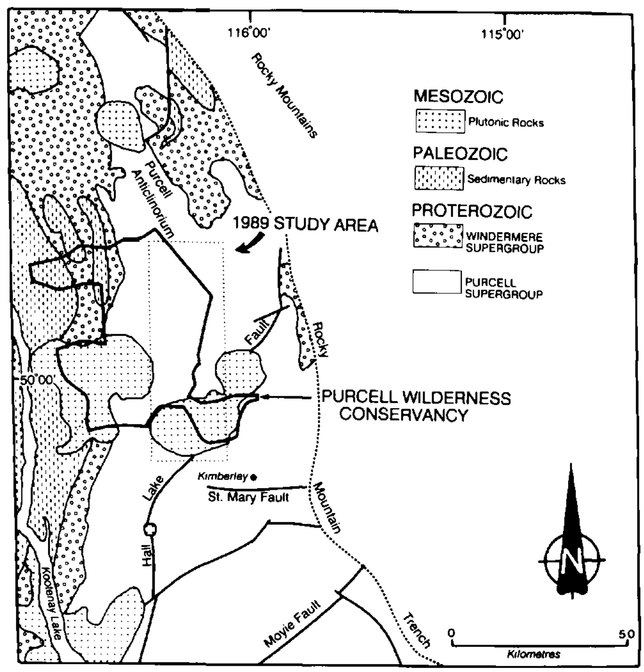

Figure 1-3-2. Regional geological setting of the Purcell Wilderness Conservancy and the 1989 study area.

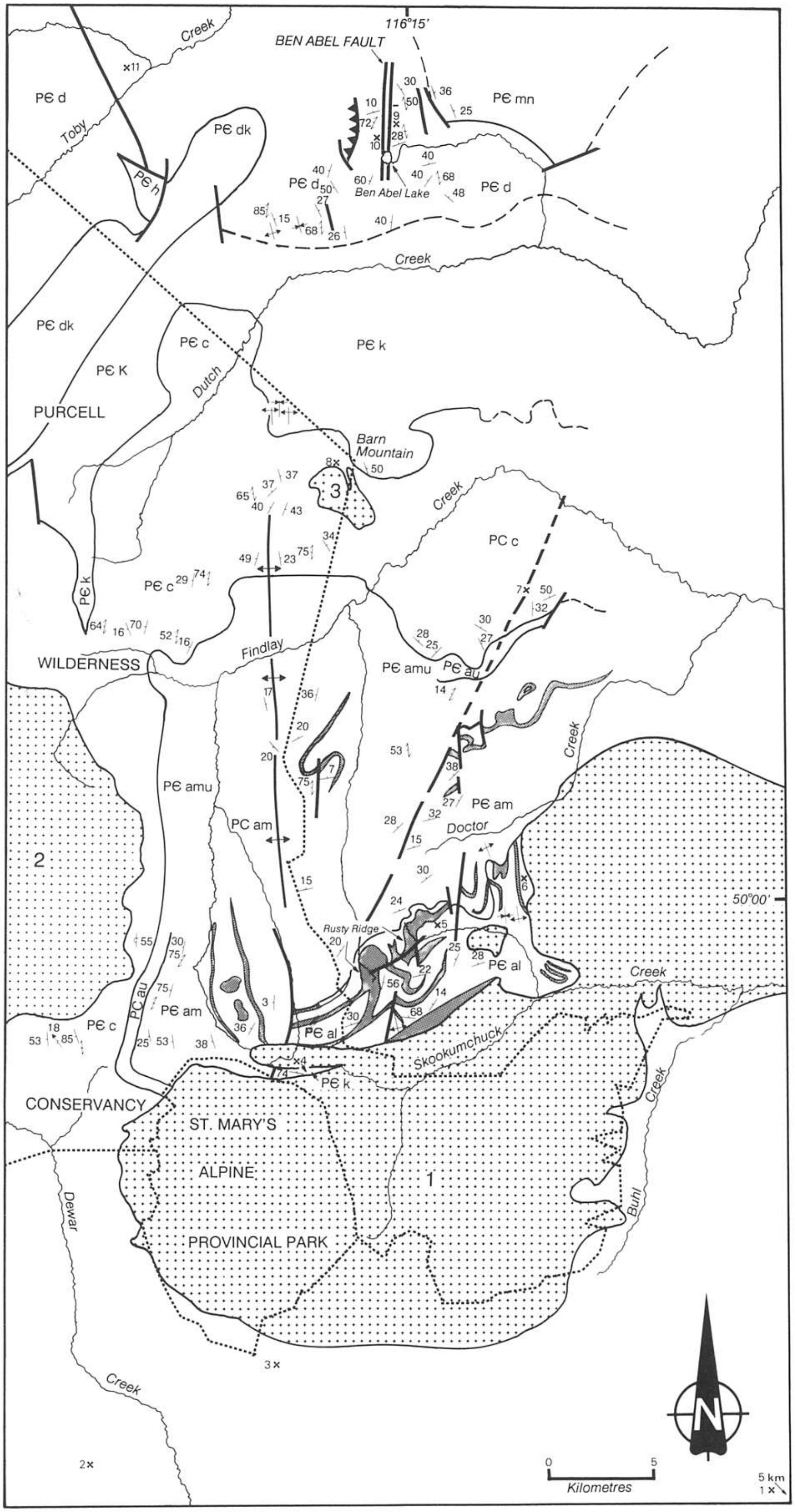

Figure 1-3-3. General geology of the project area.

LEGEND

INTRUSIVE ROCKS

MESOZOIC

CRETACEOUS

1 — WHITE CREEK BATHOLITH: GRANODIORITE, QUARTZ MONZONITE, PEGMATITE

2 — FRY CREEK BATHOLITH: QUARTZ MONZONITE

3 — FRYING PAN CREEK STOCK: QUARTZ MONZONITE, GRANODIORITE

STRATIFIED ROCKS

PROTEROZOIC

WINDERMERE SUPERGROUP(HADRYNIAN)

P€h — HORSETHIEF CREEK GROUP: PEBBLE CONGLOMERATE, GRITS, QUARTZITE AND SLATE

PURCELL SUPERGROUP(HELIKIAN)

P€mn — MOUNT NELSON FORMATION: WHITE QUARTZ ARENITE,GREEN SILTSTONE, DOLOMITIC QUARTZ WACKE AND SILTSTONE, MAROON ARGILLITE, BUFF DOLOMITE AND GREY LIMESTONE

P€dk — DUTCH CREEK AND KITHCHENER FORMATIONS: UNDIVIDED

P€d — DUTCH CREEK FORMATION: GREEN SILSTONE, BROWN DOLOMITIC SILTSTONE, GREY ARGILLITE, BUFF WEATHERING, ALGAL DOLOMITE, MINOR QUARTZ WACKE

P€k — KITCHENER FORMATION: BUFF-WEATHERING, DOLOMITIC SILTSTONE AND DOLOMITE, GREY AND GREEN ARGILLITE AND SILTSTONE, INOR LIMESTONE

P€c — CRESTON FORMATION: GREY AND GREEN QUARTZ SILTSTONE AND ARGILLITE, GREEN OR GREY-WHITE QUARTZITE, MINOR GREEN QUARTZ WACKE, MINOR DOLOMITIC SILSTONE

ALDRIDGE FORMATION

P€amu — UPPER AND MIDDLE DIVISIONS: UNDIVIDED

P€au — UPPER DIVISION: GREY ARGILLITE AND SILSTONE, MINOR QUARTZ WACKE

P€am — MIDDLE DIVISION: MASSIVE GREY QUARTZ ARENITE AND QUARTZ WACKE INTERBEDDED WITH THIN-BEDDED ARGILLITE

P€al — LOWER DIVISION: THIN-BEDDED, RUSTY WEATHERING, QUARTZ WACKE, QUARTZ ARENITE,SILTSTONE AND ARGILLITE

MOYIE SILLS: GABBRO AND DIORITE

NAME	MINFILE NUMBER	COMMODITIES
1. Sullivan	082F NE 011	Pb,Zn,Ag
2. Great Dane	082F NE 051	Ag,Pb,Cu
3. Vulcan	082F NE 093	Pb,Zn
4. Molly	082F NE 073	W,Mo
5. Pico	082F NE 089	W
6. Silver Key	082K SE 053	Ag,Pb,Zn
7. Doc(Alpine)	082K SE 060	Pb
8. Barn	082K SE 025	Cu,Mo,W
9. Yornoc	082K SE 009	Pb,Ag,Ba
10. Shelly Carolle	082K SE 059	Pb,Cu,Ba
11. Mineral King	082K SE 001	Zn,Pb,Ag,Cu,Cd,Ba

potential for mineralization was identified. All the geological and geochemical data are being compiled for release as Open File publications in 1990.

REGIONAL GEOLOGY AND PREVIOUS WORK

The Purcell Wilderness Conservancy is underlain in the east by Proterozoic rocks of the Purcell and Windermere supergroups exposed in the Purcell anticlinorium, and in the west by Paleozoic strata of the Kootenay arc (Figure 1-3-2). Mafic sills and dikes intrude the lower Purcell stratigraphy. The sedimentary units are cut by two major Cretaceous batholiths and a number of Jurassic and Cretaceous stocks. The Fry Creek batholith is a large, relatively homogeneous, quartz monzonite intrusion underlying much of the southwestern part of the conservancy. In contrast the White Creek batholith in the southeast is a well-differentiated and zoned intrusion.

Previous mapping in the conservancy itself is limited to that of Reesor (1958, 1973) in the Lardeau map area (82K) east-half, and in the Dewar Creek map area (82F/16). More recently, mapping to the east and southeast by Höy and Diakow (1982), Höy (1984) and Carter and Höy (1987a, b) has refined the Purcell stratigraphy. Stratigraphic descriptions by Höy (1985 and in preparation) have aided considerably in mapping during this project.

Government sponsored regional geochemical surveys have been conducted in map sheets 82F and 82K (National Geochemical Reconnaissance Program, 1977a, b), however, there has never been an opportunity to follow-up these results within the conservancy. Aeromagnetic data are available only for map sheet 82F/16 within the study area (Geological Survey of Canada, 1971).

LOCAL GEOLOGY

Figure 1-3-3 outlines the general geology of the eastern half of the Purcell Wilderness study area. As mapping at 1:50 000-scale in this season could not cover the entire area, the previous work by Reesor has been adopted in some parts, particularly in the northwest, to provide continuity across the map. Reesor (1958) has completed a much more detailed survey of the White Creek batholith than shown here. Most of the study area is underlain by Proterozoic Purcell Supergroup strata represented by the Aldridge, Creston, Kitchener, Dutch Creek and Mount Nelson formations. Time did not allow detailed mapping of the lithologies attributable to the Siyeh Formation overlying the Kitchener strata; as Reesor (1973) indicated that it is doubtful whether the Siyeh can be recognized as a separate formation, this entire sequence is mapped as Kitchener Formation here. Most of the contacts between stratigraphic units are gradational and therefore their positions are interpretative. Particular facies used to define contacts are identified in the following sections. The Kitchener – Dutch Creek contact is particularly difficult to identify and no clear contact was established in this project. Similar problems were encountered by Reesor and even though his contacts have been adopted in part in Figure 1-3-3, they are questionable in some areas. No outcrops of the volcanic lithologies of the Nicol Creek Formation, identified in the Skookumchuck area to the southeast (Carter and Höy, 1987a), were noted in this project.

STRATIGRAPHY

PURCELL SUPERGROUP

LOWER ALDRIDGE FORMATION: UNIT P€al

Strata belonging to the lower Aldridge Formation crop out immediately north of the White Creek batholith and are of limited extent in the map area. Dominant lithologies include quartz wacke, quartz arenite, siltstone and lesser argillite that are intruded by thick gabbroic sills. The sedimentary rocks are characteristically rusty weathering, fine to medium grained and thin to medium bedded. Finely disseminated pyrrhotite is common and imparts the rusty weathering to the rocks. Individual beds range from a few millimetres to 30 centimetres thick; thin black argillaceous laminae are common in some beds. Grading, crossbedding and basal scours, typical of the lower Aldridge elsewhere, are not readily evident in the present study area, primarily due to the degree of regional metamorphism. Discontinuous horizons of intraformational conglomerate were noted in a number of locations within lower Aldridge strata. These horizons are massive to poorly bedded. Clasts, generally less than 10 centimetres across, comprise angular to rounded pebbles of the surrounding lithologies set in an argillaceous matrix. Conglomerate float containing angular tourmalinite clasts in a siliceous matrix was found in one location. The conglomeratic horizons, and the tourmalinite fragments in particular, may be significant due to the occurrence of similar rocks in the footwall of the Sullivan and North Star orebodies. Medium to coarse-grained gabbroic to dioritic intrusions known as the Moyie sills are common within the lower Aldridge section (Plate 1-3-1) and will be discussed in more detail subsequently.

The sedimentary rocks in the lower Aldridge have undergone both thermal and regional metamorphism to at least greenschist facies. Biotite alteration in the argillaceous units and quartz-sericite alteration in the arenites and wackes have generated widespread phyllitic and schistose textures. Locally a spotted porphyroblastic texture has developed where knots of biotite and sericite alteration occur.

The base of the lower Aldridge is not exposed and the contact with the overlying middle Aldridge is gradational and difficult to identify. In accordance with other workers, this contact is placed beneath the first thick section of grey-weathering quartz wacke beds (Höy, in preparation).

MIDDLE ALDRIDGE FORMATION: UNIT P€ma

The middle Aldridge Formation comprises a thick succession of quartzite, quartz siltstone and argillite that is folded across the Purcell anticlinorium in the centre of the map area.

Plate 1-3-1. Interbedded lower Aldridge quartz arenite, quartz wacke and siltstone intruded by Moyie sills (m); Rusty Ridge area.

Pale grey to white, quartz-rich beds predominate. Coarse quartz arenite and quartz wacke beds may be over 1 metre thick; repetitive sequences of quartz siltstone beds 10 to 20 centimetres thick, separated by a few millimetres of argillite, are common. Many of the observed features are typical of extensive proximal turbidite deposition, however, bedding features are obscured by the penetrative foliation near the axis of the anticlinorium. Argillaceous and impure siltstone beds have undergone biotite and chlorite alteration and have developed schistose textures. Sericite and phyllitic textures are common in the cleaner, quartz-rich horizons.

The upper and lower contacts are both gradational. The lower part of the middle Aldridge may be rusty weathering and locally difficult to distinguish from lower Aldridge lithologies. The contact with the upper Aldridge is also gradational but is easier to identify at the last thick, pale grey weathering quartzite bed. This contact was not traced throughout the map area; where it was not mapped the undivided rocks are attributed to Unit PCamu.

UPPER ALDRIDGE FORMATION: UNIT P€au

A relatively thin (less than 200 metres) succession of limonitic grey argillite and siltstone with minor quartz wacke forms the upper Aldridge Formation. These rocks are thinly bedded and often thinly laminated. They are typically altered to dark grey phyllites. The contact with the overlying Creston Formation is gradational but is chosen where green phyllites and siltstones characteristic of the Creston begin to appear.

CRESTON FORMATION: UNIT P€c

Rocks of the Creston Formation occur in a broad band across the centre of the map area and along the western margin where they are cut by the Fry Creek batholith. This formation consists of an interbedded sequence of quartz siltstones and argillites with some quartz arenites and minor quartz wackes. The regular occurrence of green quartzites and quartz siltstones is characteristic of the Creston Formation, however grey-white quartzites and grey to black siltstone and argillite are common.

Traverses across most of the Creston Formation west of Barn Mountain revealed a lower section dominated by grey and green quartz siltstones and argillites with lesser quartz wackes and black limonitic argillites that passes upwards into a section dominated by thicker bedded (25 to 30 centimetres) pale grey to white quartzite and 6 to 8-centimetre beds of siltstone and argillite. This in turn grades back to grey-green quartz siltstone and argillite containing thin, brown, dolomitic siltstone beds. Disseminated pyrite and disseminated magnetite octahedra were noted in some beds, particularly in the middle quartzite member. Insufficient work was completed to confidently map these three members across the study area, however, this subdivision is similar to that reported by Carter and Höy (1987b) in the Skookumchuk area and is likely present throughout the study area.

Wherever mapped, the Creston Formation rocks are tightly to isoclinally folded (Plate 1-3-2) with local overturned beds. Along the axis of the Purcell anticlinorium axial plane cleavage is intense and often obscures all bedding features. Alteration to quartz-sericite phyllites and biotite schists is very common.

The contact with the overlying Kitchener Formation is marked either by the first thick, pure dolomite bed or at a point where brown-weathering dolomitic beds predominate over the green siltstones of the Creston.

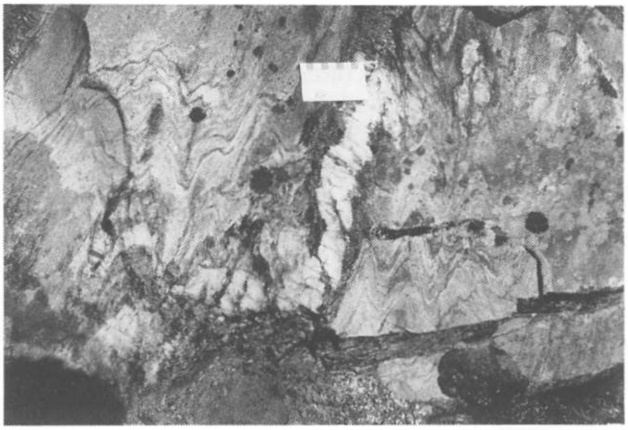

Plate 1-3-2. Tightly folded Creston Formation sediments cut by quartz veins following axial plane fractures. Thin dark grey beds are biotite-altered argillite interbedded with green phyllitic siltstones (pale grey).

KITCHENER FORMATION: UNIT PЄk

The Kitchener Formation is exposed in a broad belt across the north-central part of the map area, however, only a lower section of this unit was studied, mainly in the vicinity of Barn Mountain. At this point, Kitchener rocks consist of brown to buff-weathering dolomitic siltstone and impure dolomite. They are well bedded with individual beds up to 50 centimetres thick. Grey argillite and green chloritic phyllite occur as thin interbeds in this sequence. Dolomite and dolomitic siltstone are also seen at the southern tip of a long narrow strip of Kitchener Formation (Figure 1-3-3) previously mapped in the centre of the Purcell Wilderness Conservancy by Reesor (1973). A thin wedge of dolomitic marble and dolomitic siltstone and argillite was mapped between a lobe and the main body of the White Creek batholith along its northern border. Similar rocks attributed to the Kitchener Formation were previously mapped on the southern margin of the batholith (Reesor, 1958).

In the Skookumchuk area, Carter and Höy (1987b) described the entire Kitchener Formation as consisting of a lower dolomitic siltstone member and an upper carbonaceous dolomite and limestone member with molar tooth structures. Further mapping is required to determine if this division can be applied to the Purcell Wilderness Conservancy.

The contact with the overlying Dutch Creek Formation was not observed in this project. It is a gradational contact very difficult to recognize in this area. In the absence of a defined Siyeh Formation or the volcanic rocks of the Nicol Creek Formation there are no distinct markers between the top of the Kitchener and the base of the Dutch Creek. Reesor (1973) has identified a large area as undivided Kitchener and Dutch Creek rocks between Toby and Dutch Creeks in the northwest part of the map area (Unit PCdk in Figure 1-3-3). This contact, as shown in Figure 1-3-3, must be considered approximate.

DUTCH CREEK FORMATION: UNIT PЄd

The Dutch Creek Formation underlies much of the northern margin of the field area. It has never been subdivided in the area of the Purcell Wilderness Conservancy, however, it is thought to be correlative with the Sheppard, Gateway and Roosville formations (Höy, in preparation).

In the vicinity of Ben Abel Lake rocks mapped as Dutch Creek Formation can be subdivided into a lower sequence of interbedded pale green to brown-weathering dolomitic siltstone, green siltstone, buff-weathering silty dolomite and minor quartz wackes; a prominent buff-weathering algal-laminated dolomite approximately 300 metres thick (Plate 1-3-3); and a thick upper sequence of interbedded green siltstones, brown dolomitic siltstone, dark grey argillite and minor oolitic dolomite. The upper sequence contains at least two distinct sections of dark grey argillite and may be further subdivided in the future. All of these rocks tend to be very well bedded. The siltstones commonly display upward grading, ripple marks and horizons with rip-up clasts.

Northeast of Ben Abel Lake the Dutch Creek rocks pass quickly, but conformably, into white to buff-weathering quartz arenite of the Mount Nelson Formation. Elsewhere this conformable contact is obscured by faulting. These strata are repeatedly folded and locally cut by thrust faults across the northern part of the study area.

MOUNT NELSON FORMATION: UNIT PЄmn

One section of the Mount Nelson Formation was studied in the northeast corner of the map area. Strata observed conform to the divisions described by Höy (in preparation) in the Fernie west-half map area. The basal unit is a white quartz arenite with minor interbedded green siltstone and weakly dolomitic siltstone and quartzite. Disseminated pyrite occurs locally in some of these beds. These rocks pass upwards into a sequence of buff-weathering dolomitic quartz wacke and siltstone and distinctive maroon argillites and pink siltstones.

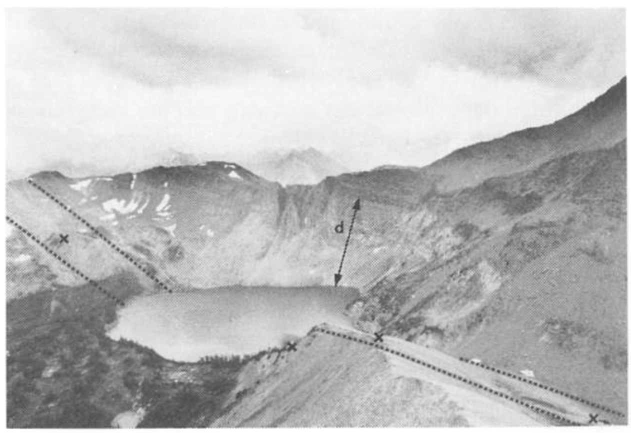

Plate 1-3-3. Dutch Creek Formation at Ben Abel Lake. Ben Abel fault zone contains copper-bearing quartz-barite veins (x). Agal-laminated dolomite (d) west of lake (to right) has been uplifted approximately 300 metres on the Ben Abel fault (photo looks southwest).

These in turn grade quickly into a unit of buff dolomite and grey limestone with minor grey argillite. Stromatolitic laminae and oolitic horizons are common in the carbonates. Another distinctive maroon to red-weathering argillite unit was seen to overly the previously mentioned rocks but mapping did not extend to this unit. Bennett (1986) describes this upper maroon argillite unit at the top of a redefined Mount Nelson Formation in the Mount Forster map area to the northeast.

WINDERMERE SUPERGROUP

Windermere Supergroup rocks were not studied in the mapping phase of the 1989 season, however, some geochemical samples were collected from basins draining these strata. They unconformably overly Purcell Supergroup rocks near the northern boundary of the study area. A basal polymictic conglomerate, the Toby Formation, is generally conformably overlain by sedimentary rocks of the Horsethief Creek Group (Reesor, 1973). A small fault-bounded wedge of Horsethief Creek sediments, shown in Figure 1-3-3, is included from Reesor's mapping.

INTRUSIVE ROCKS

MAFIC SILLS AND DIKES

The lower Purcell Supergroup rocks have been intruded by a large number of mafic sills and dikes. The Moyie sills cutting the lower Aldridge (Plate 1-3-1) and lower part of the middle Aldridge Formation have been shown to be distinct from mafic sills higher in the Purcell succession (Höy, 1989). Figure 1-3-3 outlines the general distribution of Moyie sills mapped in this project. These intrusions are sill-like in overall form but often crosscut bedding or appear as irregular lenses. Some are in excess of 100 metres thick and can be traced almost 10 kilometres in the map area. The thicker sills have coarse-grained gabbroic cores and finer dioritic margins. They are all primarily composed of hornblende and plagioclase phenocrysts set in a matrix of similar composition that has often undergone considerable chloritic alteration. Due to their irregular nature, the large number of sills in some areas, and the folding in the surrounding rocks, these sills are often difficult to correlate accurately across valleys or fault zones. Höy (1989) suggests they are an integral part of the Aldridge depositional environment and are therefore of the same Proterozoic age.

Dioritic sills and dikes of similar appearance in the field are common in the remainder of the Aldridge, Creston and parts of the Kitchener formations. These intrusions tend to be only a few metres thick and can only be traced for a few hundred metres at most. Their precise age is unknown.

WHITE CREEK BATHOLITH

The White Creek batholith is a well-differentiated granitic intrusion underlying much of the southern boundary of the study area. Only the marginal phases were mapped as a detailed discussion of this Cretaceous batholith is provided by Reesor (1958).

Along the northern border of the batholith a megacrystic granodiorite phase is common. Plagioclase phenocrysts are commonly 3 to 5 centimetres long, set in a matrix of fine to medium-grained plagioclase, potassium feldspar, quartz and biotite. Magnetite and pyrite occur locally. Aplitic and pegmatitic dikes are common. Reesor has mapped a number of pegmatitic bodies within lower Aldridge strata north of Skookumchuk Creek. Pegmatitic float with coarse beryl crystals was noted east of Rusty Ridge. Greisen-type quartz veins with muscovite selvages are common along the northern border of the batholith; further discussion of these veins is given in the mineral occurrences section. The aeromagnetic map for sheet 82F/16 clearly defines the outer boundary of the White Creek batholith (Geological Survey of Canada, 1971).

FRY CREEK BATHOLITH

The Fry Creek batholith is a large, essentially uniform Cretaceous quartz monzonite body (Reesor, 1973) in the centre of the Purcell Wilderness Conservancy. It is a distinctive pale grey, blocky weathering unit cutting the sedimentary rocks. This intrusion was not mapped in any detail in the present study.

FRYING PAN CREEK STOCK

The Frying Pan Creek stock cuts Creston Formation sediments near Barn Mountain on the boundary of the Purcell Wilderness Conservancy. This irregularly shaped intrusion is a grey-weathering granodiorite to quartz monzonite consisting of variable percentages of quartz, plagioclase, potassium feldspar, biotite and hornblende. Pegmatitic border phases are locally present. Numerous quartz veins, often with muscovite selvages, cut the stock and the surrounding sedimentary rocks. These veins carry minor amounts of chalcopyrite, chalcocite, molybdenite, galena and pyrite. Near the eastern contact, dolomitic units in the Kitchener Formation have been partially converted to epidote-magnetite skarns.

STRUCTURE

A broad north-plunging anticlinal fold, referred to as the Purcell anticlinorium, dominates the structural geology of the map area. Regionally the outcrop pattern of all stratigraphic units reflects this wide, open fold. On a more local scale, tight isoclinal folds are characteristic, particularly in the finer grained lithologies. A prominent axial plane cleavage that completely obscures bedding has developed within the core of the anticlinorium. Elsewhere the strata commonly take on a phyllitic or schistose texture in response to the tight minor folds.

Across the northern part of the study area a series of tight folds and related easterly directed reverse faults are present. The Ben Abel fault zone (Plate 1-3-3) is a high-angle reverse fault with approximately 300 metres of displacement across Ben Abel Lake. This zone is up to 100 metres wide, can be traced for 5 kilometres, and can be seen cutting stratigraphy a further 5 kilometres to the north. Even further to the north, in Toby Creek valley, the trace of the Mount Forster syncline and fault zone (Atkinson, 1976) lies along the projected strike extension of the Ben Abel fault. One kilometre west of the lake thrust faulting repeats the distinctive algal-laminated dolomite unit of the Dutch Creek Formation. Mapping has just begun to identify the complexities in this northern area but has shown that copper-bearing quartz-barite veins have developed in some of the faults. Another strongly deformed zone extends from the headwaters of Dutch Creek northwards toward the Mineral King mine in Toby Creek valley. This northern area will be studied in more detail during the next season.

Displacements on north to northeast-trending faults in the central part of the study area are difficult to establish due to their location or to the gradational nature of the stratigraphic contacts. The dramatic change in bedding attitudes and truncation of lower Aldridge strata across Alton Creek valley, a headwater tributary of Findlay Creek, strongly suggests a fault in this valley, however it is not exposed anywhere. It is likely that tight folding along the axis of the Purcell anticlinorium led to left lateral and/or west-side-down displacement of the lower and middle Aldridge rocks in this area.

A fault zone with little visible displacement, but marked by a line of limonitic and carbonate-altered outcrops, was traced from north of Rusty Ridge to Findlay Creek, a distance of 15 kilometres. This fault truncates some dioritic sills, however, no major displacements of the stratigraphic contacts were noted. Further discussion of this fault is given in the mineral occurrences section.

GEOCHEMISTRY

A total of 183 stream sediment samples were collected from an area of approximately 100 150 hectares in and adjacent to the eastern boundary of the Purcell Wilderness Conservancy. The density of sampling is approximately 1 site per 5.5 square kilometres. Samples are being analyzed for 30 elements using an inductively coupled plasma (ICP) technique, for gold by fire assay and neutron activation analysis. In excess of 100 samples were also collected in this region during the 1977 government regional geochemical survey (National Geochemical Reconnaisance Program, 1977a,b). These samples were not originally analyzed for gold or rare earth elements, however, the archived samples are now being re-analyzed for these and other elements and results will be made available in Open File publications.

Rock chip samples were collected from all locations containing mineralization or alteration potentially related to mineralization. A total of 157 rock samples will be analyzed for 14 elements, including base and precious metals and indicators.

MINERAL OCCURRENCES

Exploration and mining in this region of the Purcell Mountains began at the turn of the century with discoveries of the stratabound silver-lead-zinc Sullivan orebody; lead-zinc-silver vein occurrences in Toby Creek valley north of the conservancy and near Dewar Creek to the south; and of the replacement and vein lead-zinc-silver-barite orebody at Mineral King mine. Subsequent discoveries include skarn mineralization, porphyry molybdenum occurrences and greisen-vein tin and tungsten occurrences. Beryllium has been located in pegmatite along the north side of the White Creek batholith. The regional geology also suggests potential for stratiform copper-silver occurrences similar to that at Spar Lake, Montana, and possibly for stratiform barite-lead-zinc mineralization within thrust-emplaced Paleozoic carbonates. The potential for Kootenay arc type silver-lead-zinc deposits exists primarily to the west of the current study area and will be evaluated in the future.

In the Purcell Wilderness Conservancy planning area and surrounding environs, known mineral occurrences can be grouped into the following types: Sullivan-type sedimentary exhalative (sedex) deposits, structurally controlled silver-lead-zinc vein deposits, replacement and vein deposits similar to the Mineral King, skarn occurrences, and veins associated with felsic intrusions.

SULLIVAN-TYPE SEDEX DEPOSITS

The hostrocks for the Sullivan ore body (MINFILE 82FNE011) are exposed immediately south of the study area and in it at Rusty Ridge (Figure 1-3-3). Minor showings of laminated Sullivan-type lead-zinc mineralization occurring on the Vulcan property (MINFILE 082FNE093) have been explored by geophysical surveys and limited diamond drilling. This prospect, just south of the conservancy, appears to have excellent potential for lead-zinc sulphide mineralization. In the Rusty Ridge area the lower-middle Aldridge stratigraphy favorable for Sullivan-type mineralization extends through the study area and into the Purcell Wilderness Conservancy itself. Although no stratabound mineralization has been discovered, quartz-tourmaline veins are common, conglomeratic horizons similar to those in the footwall of the Sullivan mine are present, as are isolated veinlets with pockets of galena and minor sphalerite. Exploration of this area is continuing just outside the conservancy boundary.

STRUCTURALLY CONTROLLED SILVER-LEAD-ZINC VEIN DEPOSITS

Lodes of massive galena with associated sphalerite and chalcopyrite occur within vein-like structures on the Great Dane property (MINFILE 082FNE051) south of the conservancy (Figure 1-3-3). This silver-bearing mineralization, hosted by quartzite of the Creston Formation, is concentrated in reef structures in the axial zones of tight folds (Scott, 1986). Little work has been completed on the property in recent years but further evaluation appears warranted. Barren quartz veins within axial plane fractures or crests of folds were noted in many locations in the Creston Formation (Plate 1-3-2).

A similar structural control is shown at the Silver Key occurrence (MINFILE 082KSE053) where quartz-carbonate veins within tightly folded and sheared lower Aldridge sediments contain galena, sphalerite, pyrite and minor tetrahedrite and arsenopyrite. This occurrence is located very close to the edge of the White Creek batholith, however, the relationship to the intrusion is unknown. Two adits and a number of open cuts have been developed on the property (Minister of Mines Annual Report, 1938).

The Doc (Alpine) occurrence (MINFILE 082KSE060) consists of disseminations and streaks of galena, sphalerite, pyrite and ankerite within a strongly albitized and silicified fault zone. The main showing averages 3.5 metres thick over an 80 metre length (Mawer, 1986). Discontinuous patches of similar alteration can be traced along strike as far as the Rusty Ridge area to the southwest (Figure 1-3-3). Limonitic and ankeritic alteration in middle Aldridge sediments occurs in an aligned series of recessive saddles where this fault zone crosses ridges. Partially leached pyrite mineralization was noted but no other lead or zinc mineralization was found along this fault.

Copper mineralization was found in quartz or quartz-barite veins at a number of locations in the northeast-trending Ben Abel fault zone. Near Ben Abel Lake chalcopyrite, tetrahedrite, pyrite and traces of galena occur in veins or irregular lenses of intense silicification up to 3 metres wide; barite-rich pods are common (Plate 1-3-2). Malachite and azurite are abundant and fluorite crystals and manganese staining are locally present. Elsewhere along the fault, veins 15 to 30 centimetres wide carry pockets of similar mineralization. This trend is not documented in MINFILE records however the Yornoc (MINFILE 082KSE009) and Shelly Carolle (MINFILE 082KSE059) occurrences are described to the east and west of this fault respectively.

At the Yornoc prospect a number of pits have been cut on an east-trending zone of quartz-barite veining up to 2 metres wide and contain chalcopyrite, galena and pyrite with considerable malachite and azurite. The mineralized section of this vein system extends over 500 metres, however, outcrops of vein material can be traced across an overburden-covered valley for over a kilometre. Similar mineralization was seen in two small pits at the Shelly Carolle occurrence, however the veins here trend north to northeasterly and follow the regional axial plane fractures.

Mineral King Vein/Replacement Deposits

The Mineral King mine (MINFILE 082KSE001), located 6 kilometres northeast of the conservancy boundary, produced 2.1 million tonnes of silver, copper, lead, zinc and cadmium ore between 1953 and 1967. More recently the property has been operated as a barite producer. Mineralization occurs as veins and replacements in tightly folded and faulted dolomitic sediments of the Mount Nelson Formation (Fyles, 1959). No similar mineralization was found in the study area however the complex structural zone containing the deposit extends southwards, well inside the conservancy, to the headwaters of Dutch Creek. This area will be prospected in the future. Stratiform mineralization similar in appearance to that at the Mineral King mine is known within thrust-emplaced wedges of Devonian rocks to the northeast (Redmac occurrence, MINFILE 082KNE059). The thrusting identified in this northern part of the study area suggests this style of mineralization should be evaluated along the Mineral King trend.

Skarn Occurrences

The Molly occurrence (MINFILE 082FNE073) is located within the Purcell Wilderness Conservancy. A section of dolomitic marble and dolomitic siltstones was mapped on a ridge at the headwaters of Skookumchuk Creek between two lobes of the White Creek batholith. These sediments are approximately 350 metres thick on the ridge but are truncated by the intrusion in upper Skookumchuk valley. Calcsilicate alteration of the sediments to epidote, tremolite and calcite is common along the ridge. Tungsten and molybdenum mineralization are reported in skarn occurrences in the valley and have been trenched and diamond drilled to a limited extent (Minister of Mines Annual Report, 1969). No work has been permitted on this property since the conservancy was established.

Minor pockets of epidote-magnetite skarn are present in Kitchener Formation sediments along the eastern border of the Frying Pan Creek stock. Traces of scheelite were also noted in samples from this area. Virtually all of the prospective alteration zones around this stock are within the Purcell Wilderness Conservancy.

Veins Associated With Felsic Intrusions

Greisen-type veins with muscovite selvages are common in the sedimentary rocks along the northern margin of the White Creek batholith. Tin and tungsten mineralization has been reported from a number of locations (MINFILE 082FNE089, 90, 92). Veins sampled in this study contain variable amounts of scheelite and, at one location, considerable fluorite. Overall, the veins are thin (5 to 20 centimetres wide), appear scattered and mineralization is sparse.

Within and to the west of the Frying Pan Creek stock quartz veins up to 2 metres wide can be traced across a ridge for hundreds of metres. The veins generally have muscovite selvages or contain vugs filled with quartz and muscovite crystals. Some veins have brecciated contacts with the wallrocks. Sparsely disseminated mineralization in the veins includes molybdenite, chalcopyrite, pyrite, chalcocite and traces of galena.

Beryl crystals have been identified in pegmatite bodies along the northern contact of the White Creek batholith (MINFILE 082FNE107, 112). These occurrences were not examined in this project and their extent is unknown, however, coarse beryl crystals were noted in float east of the Rusty Ridge area.

SUMMARY: MINERAL POTENTIAL

In the vicinity of the Purcell Wilderness Conservancy the Proterozoic Purcell Supergroup comprises a thick succession of clastic and carbonate sedimentary lithologies. These rocks reflect both deep and shallow-water depositional facies within the larger Belt-Purcell basin. Syndepositional faulting has controlled facies changes in some areas of this basin (Höy, in preparation). Subsequent folding, faulting and intrusion by granitic batholiths and stocks have further modified the original sedimentary lithologies. Phyllitic and schistose textures and structurally controlled veins are common. The primary sedimentary depositional environments and the subsequent igneous and structural events combine to provide a variety of metallogenic settings and a varied mineral potential.

A good potential for Sullivan-type massive sulphide deposits has been identified in the Rusty Ridge area; this potential extends at least 1 to 2 kilometres into the Purcell Wilderness Conservancy. There is also potential for structurally controlled vein deposits, skarn deposits, tin or tungsten vein occurrences and possibly some pegmatite-related industrial mineral occurrences in this part of the 1989 study area. Mineral claim staking and exploration have been common activities in this locality for decades and can be expected to continue in the future.

Further analysis of the structural geology is required to determine the potential for vein and replacement deposits. Numerous vein occurrences are present within the planning area but outside the boundaries of the Purcell Wilderness Conservancy. However major structural zones not previously defined in detail, such as the zone including the Mineral King mine, extend well inside the conservancy and require evaluation.

The first year of this mineral potennial assessment has identified a number of favourable metallogenic environments and has provided reconnaissance data for further prospecting. This work will be supported by interpretation and follow-up of the 1989 geochemical data. Evaluation of the mineral potential of the Kootenay arc stratigraphy will be completed in the future.

ACKNOWLEDGMENTS

Capable assistance in the field was provided by Brian Nielsen. Trygve Höy added greatly to our knowledge of the area by introductory discussions and a field trip through the Purcell stratigraphy. The staff of Cominco Ltd. at Kootenay Exploration, Cranbrook and at the Sullivan Mine, Kimberly, are thanked for sharing their knowledge of the regional geology and for an informative tour of the Sullivan mine. Communications while in the field were greatly facilitated by the Ministry of Forests communication system and their staff in Invermere. Helicopter support was provided by Frontier Helicopters Limited, Fairmont Hotsprings.

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NOTES

PRELIMINARY REPORT ON THE LILLOOET LAKE MAPPING PROJECT SOUTHWESTERN BRITISH COLUMBIA (92J/1, 2, 7)

By J.M. Riddell
University of Montana

KEYWORDS: Regional mapping, Lillooet Lake, Fire Lake Group, Cadwallader Group, stratigraphic correlations.

INTRODUCTION

Fieldwork conducted during the summer of 1989 covered a map area of approximately 200 square kilometres west of Lillooet Lake near Pemberton, southwestern British Columbia (Figure 1-4-1). This report discusses the observations of the first of two field seasons of a mapping project that will form the basis of a Master's thesis at University of Montana.

The area has been mapped as Triassic Cadwallader Group by Cairnes (1925), Roddick and Hutchison (1973) and Woodsworth (1977). Mapping by Journeay and Csontos (1989) indicates that the map area is transected by a north-northwest-striking structure, probably a thrust fault, with rocks of the Cretaceous Fire Lake Group lying to the west, and Triassic rocks, possibly of the Cadwallader Group, lying to the east. The purpose of this project is to map this area in detail, to study the nature of the contact between these two units, and to establish local and regional correlations with equivalent stratigraphic sections in the Coast Belt. This year's mapping has confirmed the existence of a major north-northwest-striking westerly directed thrust fault, and supports correlation of rocks to the west with the Fire Lake Group. Analyses of microfossil and radiometric samples collected east of the fault will help to determine whether these rocks are correctly correlated with the Cadwallader Group.

Further work will include petrology and structural analysis of oriented samples and field data. Fieldwork in 1990 will expand the map area north of Mount Currie and concentrate on improving the understanding of the Triassic stratigraphy east of the fault.

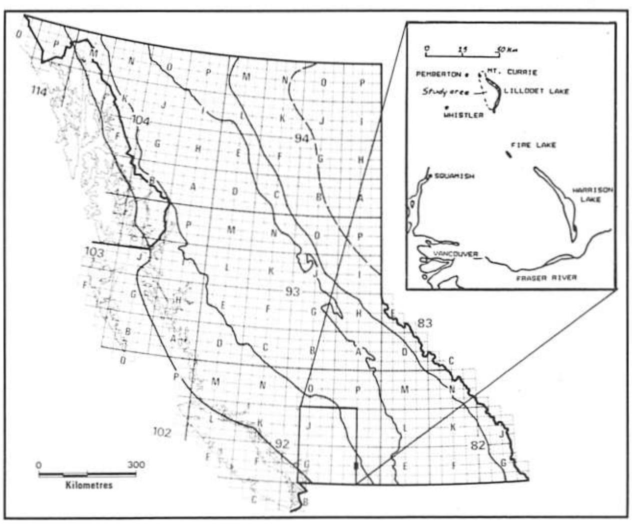

Figure 1-4-1: Location map.

TRIASSIC STRATIGRAPHY

Rocks to the east of the major thrust fault (main fault, Figure 1-4-2) are interpreted as Triassic, based on lithological similarities with rocks in the Tenquille Lake area which contain Norian microfossils (Woodsworth, 1977). No rocks from the study area have been dated.

Four major lithological units have been identified within this Triassic section. At this stage of the project contacts between these units are not well mapped and a good internal stratigraphy has not been developed. The following tentative stratigraphy is suggested and will be tested during the 1990 field season.

UNIT 1A (OLDEST): MASSIVE ANDESITE

Unit 1a is dominated by massive andesite and greenstone, and contains feldspar-porphyritic andesite, limestone pods and minor andesite breccias. Basalt is rare. The massive andesite is commonly speckled with rounded epidote clots and contains abundant epidote veins. Disseminated pyrite is ubiquitous. No pillow structures were seen. The unit is well exposed on the Lill property at the north end of Lillooet Lake (Figure 1-4-2), and along the Darcy road near Spetch, about 7 kilometres north of Mount Currie. This probably represents the base of the Triassic section. Mineral exploration in the map area has been concentrated within Unit 1a at the north end of Lillooet Lake. Brownish orange iron oxide staining can be seen in the cliffs above the lake, from the Duffy Lake road. Most of the mineralization that has been found is associated with skarn alteration of limestone pods by small hornblende diorite bodies, although mineralization associated with felsic fragmental rocks has been reported in the area (Day, 1987).

UNIT 1B: BANDED ANDESITE AND CHERT

A thickness of several hundred metres of banded andesite and chert is exposed at the north end of the easternmost ridge south of Ure Creek. The bedding strikes north-northwest and dips steeply to moderately to the west. The cherts are dominantly green, but locally white or black, and form beds 5 to 6 centimetres thick. Interbedded andesite is medium to dark green and massive, with bands centimetres to metres thick. Feldspar-porphyritic andesite was not seen in this section.

Unit 1c: Andesitic Fragmental Rocks

Unit 1c comprises a variety of andesitic fragmental rocks and greenstone. The fragmental rocks are consistently poorly sorted but show a wide variation in matrix type, clast type and clast size.

Feldspar crystal tuffs and lithic-lapilli tuffs with feldspar crystal tuff matrix are the dominant rock types at the south end of the easternmost ridge south of Ure Creek (Figure 1-4-2). Clasts are mainly white-weathering felsic volcanic fragments with lesser amounts of andesitic fragments. Green chert clasts are rare. Clasts make up anywhere from 0 to 80 per cent of the rock and average 3 to 4 centimetres in diameter.

The centre of the ridge is partly underlain by a hornblende and feldspar-phyric andesitic autobreccia; clasts average 5 or 6 centimetres in diameter and have deep purple reaction rinds. Distinct from this is another fragmental unit with an aphanitic, dark green andesitic matrix and a variety of clast types, including andesitic and felsic volcanic rocks, rare green or black chert, diorite, and basalt. Clast size in this breccia also averages 3 to 4 centimetres.

All of the andesitic fragmental rocks contain interbedded andesitic flows. Felsic flows are present but are not as common.

Unit 1d: Sedimentary Rocks

Sedimentary rocks are exposed on the west shore of Lillooet Lake, on the logging road that extends part way up the lake from the south. Black argillite and phyllite dominate the section, purple and green siltstones, volcanic sandstones and minor chert and limestone are also present.

The validity of the correlation of this Triassic section with the Cadwallader Group has been questioned because of the lithological differences between these rocks and the Cadwallader Group rocks described by Rusmore (1985) in the Eldorado Mountain area north of Gold Bridge, British Columbia. In contrast to the stratigraphy described above, the Eldorado Mountain section comprises a lower basaltic unit, a middle transitional unit and an upper sedimentary unit dominated by turbidites. It is interpreted to be an accumulation that formed adjacent to an active island arc (Rusmore, 1985). However, the lithological differences alone are not convincing evidence that the correlation is invalid. The rock types mapped east of the main fault in this study are also typical of those found in island arc assemblages. Because island arcs evolve and change rapidly, it is to be expected that abrupt lateral facies changes will occur within a single continuous arc (Hamilton, 1988).

CRETACEOUS STRATIGRAPHY

Unit 2a: Peninsula Formation

The lowermost unit seen west of the main fault is commonly well bedded and graded, and comprises interbedded quartz-bearing, white-weathering, feldspar-rich tuffaceous sandstone, siltstone, volcanic wacke and black shale. The section is topped by a pale-green-weathering tuffaceous sandstone that is conformable with the overlying breccia unit. Trough crossbeds are observed at this contact. Fossil and lithological similarities support correlation of these rocks with the Peninsula Formation as described by Arthur (1986). This unit has been mapped in the Fire Lake area by Lynch (in press).

This unit has several distinguishing characteristics. The white-weathering tuffaceous sandstones are quartz bearing and usually have a slightly calcareous matrix. Shaly beds are usually millimetres to centimetres thick and are a minor component of the section, but north of Ure Creek they can be tens of metres thick and make up a major proportion of the section. Plant fossils are often found in thin shaly beds in the tuffaceous sandstone. Shaly rip-up clasts are common (Plate 1-4-1). Rounded pebbles of feldspar-phyric volcanic rocks, coaly fragments and white to grey chert are found in pebble bands in the volcanic wacke. Brown-weathering, dark grey, massive limestone concretions 10 to 30 centimetres across are also found in the volcanic wacke (Plate 1-4-2). These concretions contain belemnites at one locality.

Unit 2b: Mauve and Green Andesitic Breccia, Brokenback Hill Formation

An andesitic unit, dominated by coarse autobreccia, conformably overlies the lowermost unit. This unit also contains pale green feldspar crystal tuffs and beige-weathering feld-

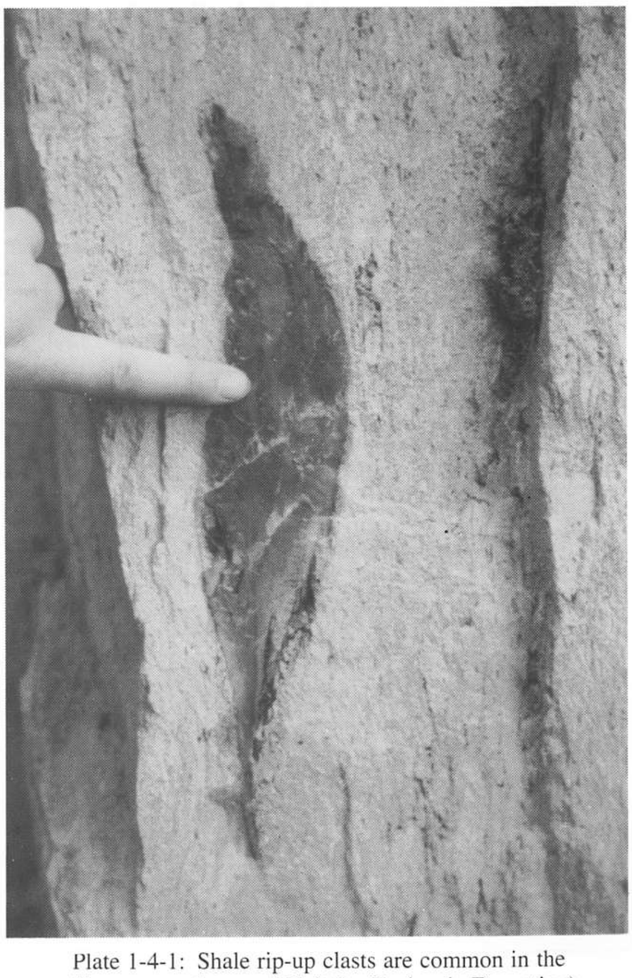

Plate 1-4-1: Shale rip-up clasts are common in the tuffaceous sandstone of Unit 2a (Peninsula Formation).

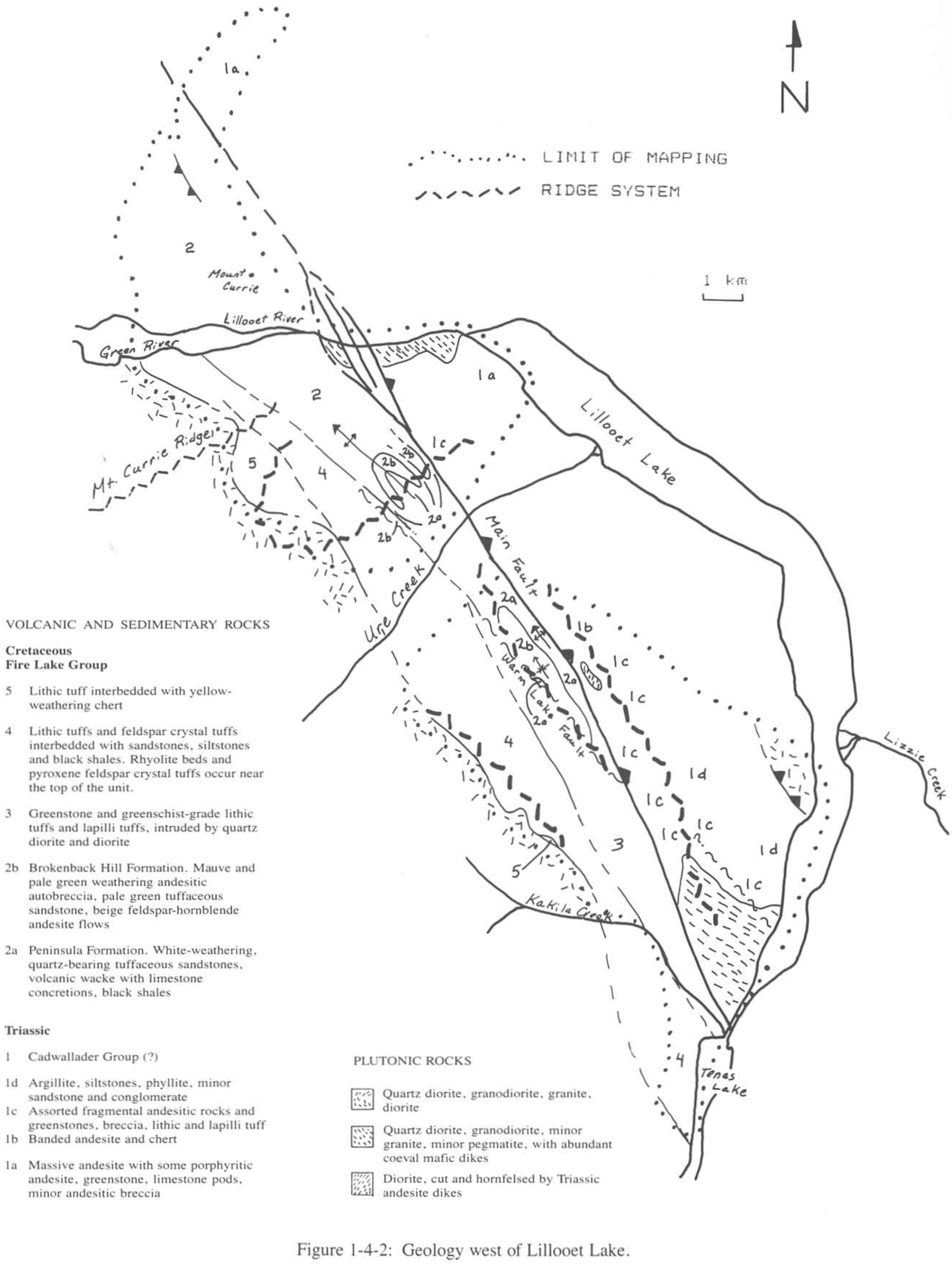

Figure 1-4-2: Geology west of Lillooet Lake.

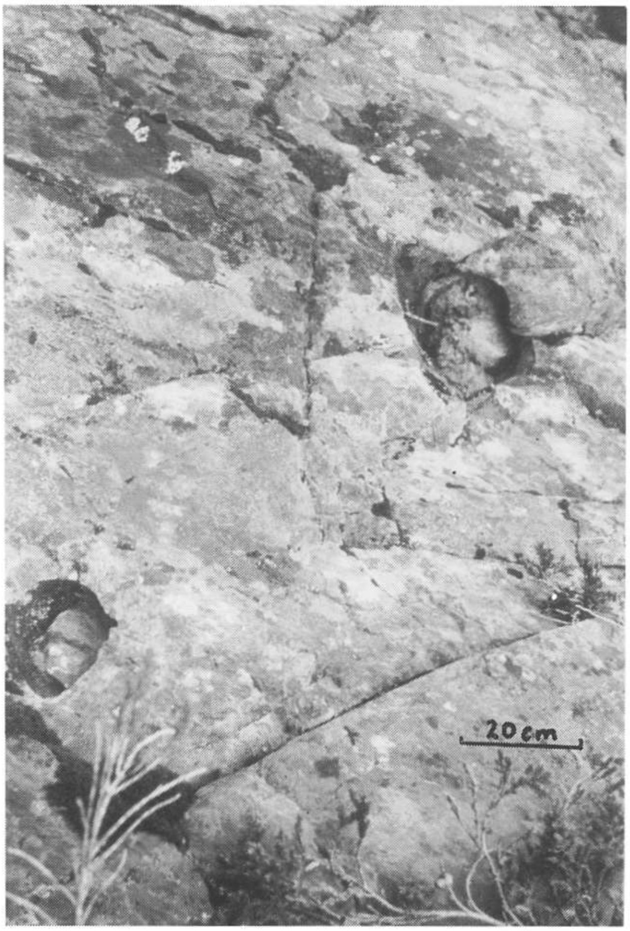

Plate 1-4-2: Belemnite-bearing limestone concretions in volcanic wacke of Unit 2a (Peninsula Formation), north of Ure Creek.

Plate 1-4-3: Clasts in the mauve and pale green andesitic breccia of Unit 2b are commonly 3 to 6 centimetres in diameter, but are locally much larger, as seen in this photograph from south of Ure Creek.

spar and hornblende-phyric andesitic flows. It is correlative, on the basis of lithological similarity, to the Brokenback Hill Formation as described by Arthur (1986) and was mapped by Lynch (in press) in the Fire Lake area.

The breccia is marked by its distinctive pale pastel-green and mauve weathering colours. Breccia clasts are commonly 3 to 6 centimetres across, but are locally much larger. Clasts over a metre across are common south of Ure Creek (Plate 1-4-3). Clasts may be rounded with reaction rims, or angular with distinct edges. Angular and rounded clasts are seen together in some outcrops. Clasts are dominantly feldspar phyric and commonly contain hornblende phenocrysts. Jasper clasts occur locally in finer grained breccia (*i.e.* where clasts average 0.5 centimetre) and jasper less frequently occurs in interstices between large closely packed blocks. Clast-to-matrix ratios range from 80:20 to 30:70. The matrix is compositionally equivalent to the clasts, although hornblende phenocrysts are typically not as abundant in the matrix. This unit is very resistant and forms ridges and benches.

Unit 3: Greenstone

The Warm Lake fault (informal) separates Units 2a and 2b from the overlying Units 3 through 5.

The greenstone unit is exposed west of the main fault in the south half of the map area, north of Kakila Creek (Figure 1-4-2). It is cut off by the Warm Lake fault within the Fire Lake Group rocks and is not present north of Ure Creek. Unit 3 is composed primarily of greenstone and contains lapilli and lithic tuff units which are metamorphosed to greenschist facies by abundant quartz diorite and aplite dikes.

Unit 4: Interbedded Lithic Tuffs and Sedimentary Rocks

Unit 4 is well exposed west of Tenas Lake, in the saddle between the two westernmost ridges south of Ure Creek, and west of a deep saddle north of Ure Creek (Figure 1-4-2). The unit contains several indistinguishable lithic tuff units that are usually 20 to 30 metres thick, that contain feldspar-phyric volcanic and green siliceous clasts in a feldspar-crystal-rich matrix. These tuffs are interbedded with black shales, siltstones, tuffaceous sandstones and minor conglomerates. Near the top of the section, thin, frothy white rhyolite bands were seen. These are usually less than a metre thick. Pyroxene feldspar crystal tuffs are also common near the top of the section. The bands may be several metres thick and commonly contain euhedral pyroxene crystals up to 1 centimetre across. Black chert is present at the top of the section on the westernmost ridge. Thick quartz veins striking parallel to foliation are common in the sedimentary rocks.

UNIT 5: INTERBEDDED LITHIC TUFF AND YELLOW-WEATHERING CHERT.

This unit is well exposed only near the south end of the westernmost ridge in the map area and was found in float and rubbly outcrop on the ridge directly south of Mount Currie Ridge. The lithic tuff is similar to that found in Unit 4 but also contains angular purple and brown chert fragments. The yellow-weathering chert is pale grey on the fresh surface, with laminae that are often convoluted.

PLUTONIC ROCKS

The oldest plutonic rocks in the area are diorites which outcrop at the top of the easternmost ridge south of Ure Creek and predate the andesites there. The diorites are cut and hornfelsed by andesite dikes, and clasts of diorite are found in an andesitic breccia.

Quartz diorite, granodiorite and granite, and coeval mafic dikes are exposed near the north end of Lillooet Lake along the south shore of the Lillooet River, and at the south end of the lake. The age of these rocks is not known.

Strongly foliated quartz diorite is exposed on the west shore of Lillooet Lake, opposite the Lizzie Creek alluvial fan. Near the contact with Triassic sedimentary rocks to the south, the granitic rocks are mylonitized and contain northeast-side-up kinematic indicators. Mafic phases are metamorphosed to amphibolite grade. The sedimentary rocks south of the contact are unmetamorphosed. A northeast-side-up thrust fault relationship is inferred.

STRUCTURE

The dominant structural feature in the map area is the steep north-northeast striking thrust fault that places Triassic rocks to the east in contact with the Cretaceous Fire Lake Group rocks to the west. The fault zone is well exposed along the logging road that runs along the south bank of the Lillooet River. Kinematic indicators in this exposure consistently give an east-side-up sense of movement. The fault zone is marked by intense brittle deformation across a width of almost 2 kilometres at this location, and contains zones of sericite and talc schist, but is not as wide in the rest of the map area. The vast majority of outcrops in the map area exhibit moderate to intense, steeply-dipping north-northwest foliation fabrics.

Structures on the east side of the main fault are not well understood. In the relatively rare bedded rocks in the Triassic section, bedding and foliation orientations are parallel, indicating that at least some of the section has undergone high shear strain or isoclinal folding.

West of the main fault the Fire Lake Group rocks display well-developed bedding/cleavage relationships (Plate 1-4-4). North of Ure Creek, rocks are deformed into broad, open upright folds with near-horizontal fold axes which parallel the strike of the fault. This pattern is consistent with the east-side-up thrust direction indicated in the north road fault-zone exposure.

Deformation prior to the folding associated with the main fault is apparent in the Fire Lake rocks. A north-northwest-striking fault, the Warm Lake fault (informal), cuts out an unknown amount of section. Pre-existing folding is apparent south of Ure Creek where the overprint of the main fault folding event has produced a more complex interference pattern.

Late stage north-striking extension cracks are common throughout the map area, forming 1 to 10 metre wide chasms and rubble pits.

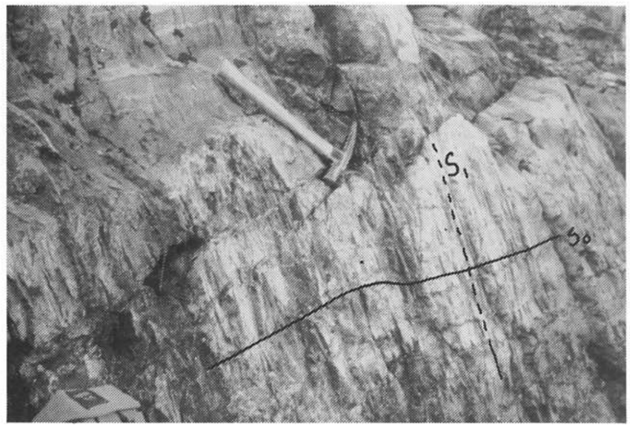

Plate 1-4-4: Well developed bedding/cleavage relationships displayed in a hinge zone in the volcanic wacke of Unit 2a (Peninsula Formation), north of Ure Creek.

ACKNOWLEDGMENTS

This project is supported by the Geological Survey of Canada (Pemberton Project), and by funding from Geoscience Research Grant Number RG89-24 from the British Columbia Ministry of Energy, Mines and Petroleum Resources. I am most grateful to Murray Journeay for suggesting the project and for providing encouragement and excellent leadership during the field season. Field assistants Robin Shropshire, Laurie Welsh, Shelley Higman and David Bilenduke worked hard and provided moral support. Greg Lynch provided an introduction to the Fire Lake Group stratigraphy. I am grateful to John and Patricia Goats of Pemberton Helicopters for safe and dependable transportation and expediting.

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$^{40}Ar/^{39}Ar$ DATING AND THE TIMING OF DEFORMATION AND METAMORPHISM IN THE BRIDGE RIVER TERRANE, SOUTHWESTERN BRITISH COLUMBIA* (92O/2; 92J/15)

By D.A. Archibald, Queen's University
P. Schiarizza, Geological Survey Branch
and J.I. Garver, University of Washington

KEYWORDS: Geochronology, Ar-Ar dating, Bridge River, Tyaughton Creek, Bralorne, Noaxe Creek, Yalakom fault, blueschist, deformation, metamorphism.

INTRODUCTION

The Tyaughton Creek area lies approximately 200 kilometres north of Vancouver on the eastern margin of the Coast plutonic complex and west of the Yalakom fault. This area is the focus of a regional program of $^{40}Ar/^{39}Ar$ dating which was initiated in 1987 in the Warner Pass (92O/3) and Noaxe Creek (92O/2) map areas (Archibald *et al.*, 1989). In 1988 and 1989, the program continued with sampling in the Bralorne (92J/15) and Bridge River (92J/16) map areas (Figure 1-5-1). These latter areas are underlain by rocks of the Bridge River Terrane, and include a fault-bounded panel of blueschist facies metamorphic rocks and the Shulaps ultramafic complex. In this note we report $^{40}Ar/^{39}Ar$ step-heating data for white mica from the blueschist rocks and a hornblende from a porphyry which intruded the Shulaps ultramafic complex near the Yalakom fault.

GEOLOGIC SETTING OF THE SAMPLES

The regional and detailed geology has been outlined in a series of B.C. Ministry of Energy, Mines and Petroleum Resources publications. Of particular relevance to this study are the reports of the blueschist locality in the Eldorado Mountain area (Garver *et al.*, 1989a) and the summary of the geology of the Bridge River Terrane near the Shulaps ultramafic complex (Schiarizza *et al.*, 1989a, 1990, this volume; Calon *et al.*, 1990, this volume). Both samples selected for dating are from the Bridge River Terrane (Potter, 1986) which comprises imbricated Permian(?) to Jurassic oceanic sedimentary and volcanic rocks of the Bridge River complex together with the Shulaps ophiolite complex (Figure 1-5-2).

BLUESCHIST FACIES ROCKS

In the Tyaughton Creek area, blueschist and greenschist facies metamorphic rocks are structurally interleaved with rocks of lower metamorphic grade within a narrow northwest-trending belt that has been traced for 30 kilometres (Schiarizza *et al.*, 1989b). The best exposures occur in the drainage of North Cinnabar Creek where this package is unconformably overlain by middle Albian rocks of the Taylor Creek Group which contain boulders of blueschist, chert and greenstone.

The blueschist is strongly flattened, locally records isoclinal folding, and commonly has a pronounced crenulation cleavage (Garver *et al.*, 1989a). Two principal mineral assemblages have been recognized in this area: crossite/glaucophane + lawsonite and crossite/glaucophane + garnet + epidote ± white mica. These two assemblages, which represent slightly different pressure-temperature conditions during metamorphism, occur in the same area but are probably separated by low-angle faults. Prehnite is present as crosscutting veins in both rock types (Garver *et al.*, 1989b). The sample dated in this study contains the second mineral assemblage and the medium-grained white mica lies in the plane of the schistosity. Previous K-Ar and Rb-Sr dating of rocks and white mica separates from the same structural panel yielded dates between 195 and 250 Ma (Garver *et al.*, 1989b). Step-heating experiments were undertaken to refine the primary cooling age and to determine the magnitude and timing of later thermal events that are thought to have affected the area.

HORNBLENDE PLAGIOCLASE PORPHYRY

The other sample selected for step-heating is an amphibole from a hornblende plagioclase porphyry dike which outcrops along the northeast margin of the Shulaps ultramafic complex. The hornblende from this sample yielded a $^{40}Ar/^{39}Ar$ total-fusion date of 75.6 ± .8 Ma (Archibald *et al.*, 1989). This area was remapped in 1989 and the interpretation of the geology and the previous date for this sample are revised.

This area was mapped first in 1987 as part of the Noaxe Creek map area (Glover *et al.*, 1988). At that time, the porphyry was considered to have intruded the Yalakom fault zone which was viewed as a zone, some 500 metres wide, characterized by penetratively deformed, serpentinized ultramafic rock, locally altered to a carbonate-quartz-fuchsite assemblage. The porphyry was mapped as a narrow dike, 1500 metres long, between and approximately parallel to two splays of the Yalakom fault. Although the margins of the dike are weakly foliated, it was inferred to have been emplaced after major movement on the fault.

In 1988, the imbricate zone along the south side of the Shulaps ultramafic complex was mapped in detail by T.J. Calon (Memorial University of Newfoundland) and the

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

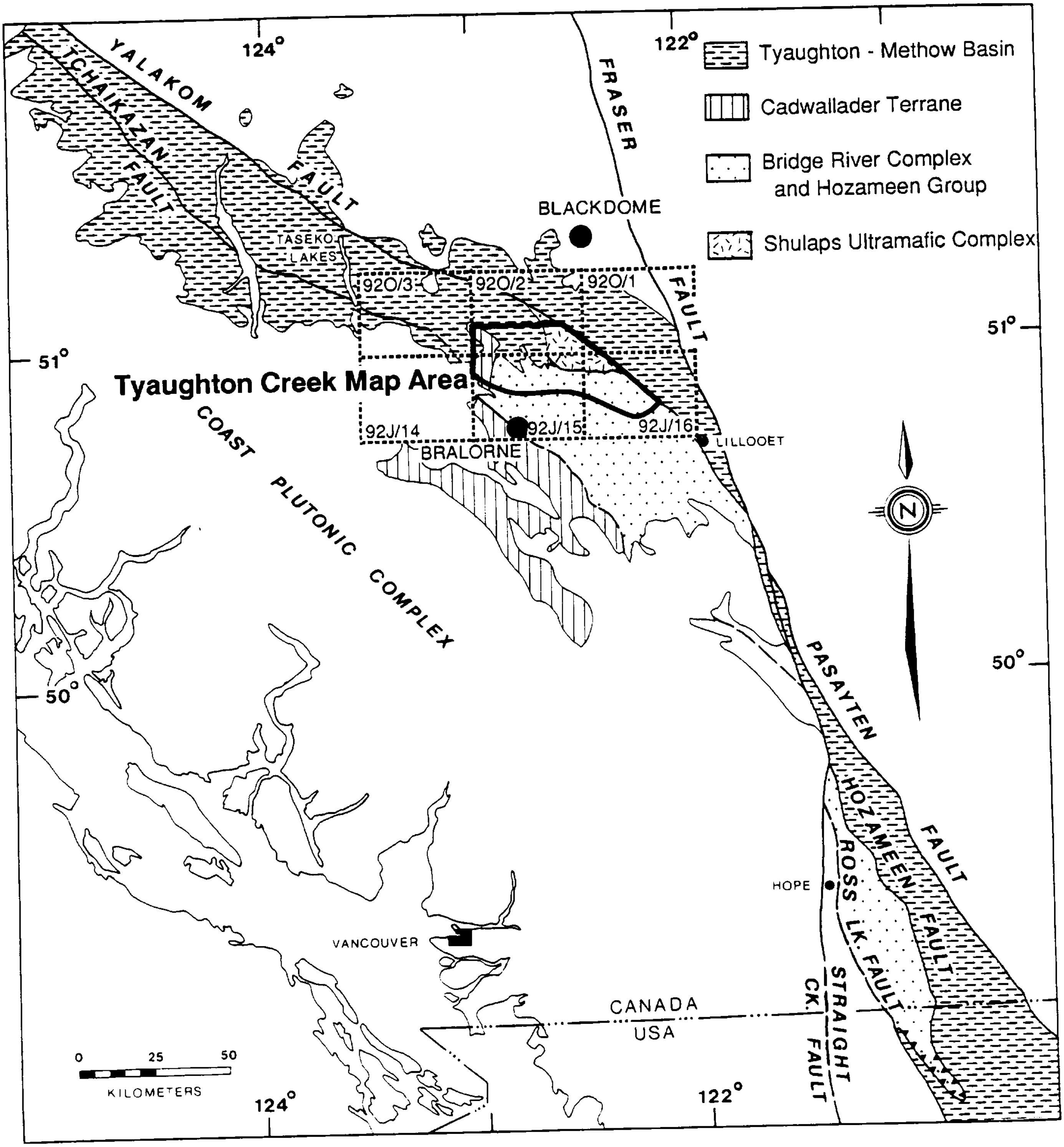

Figure 1-5-1. Location and geological setting, Tyaughton Creek map area.

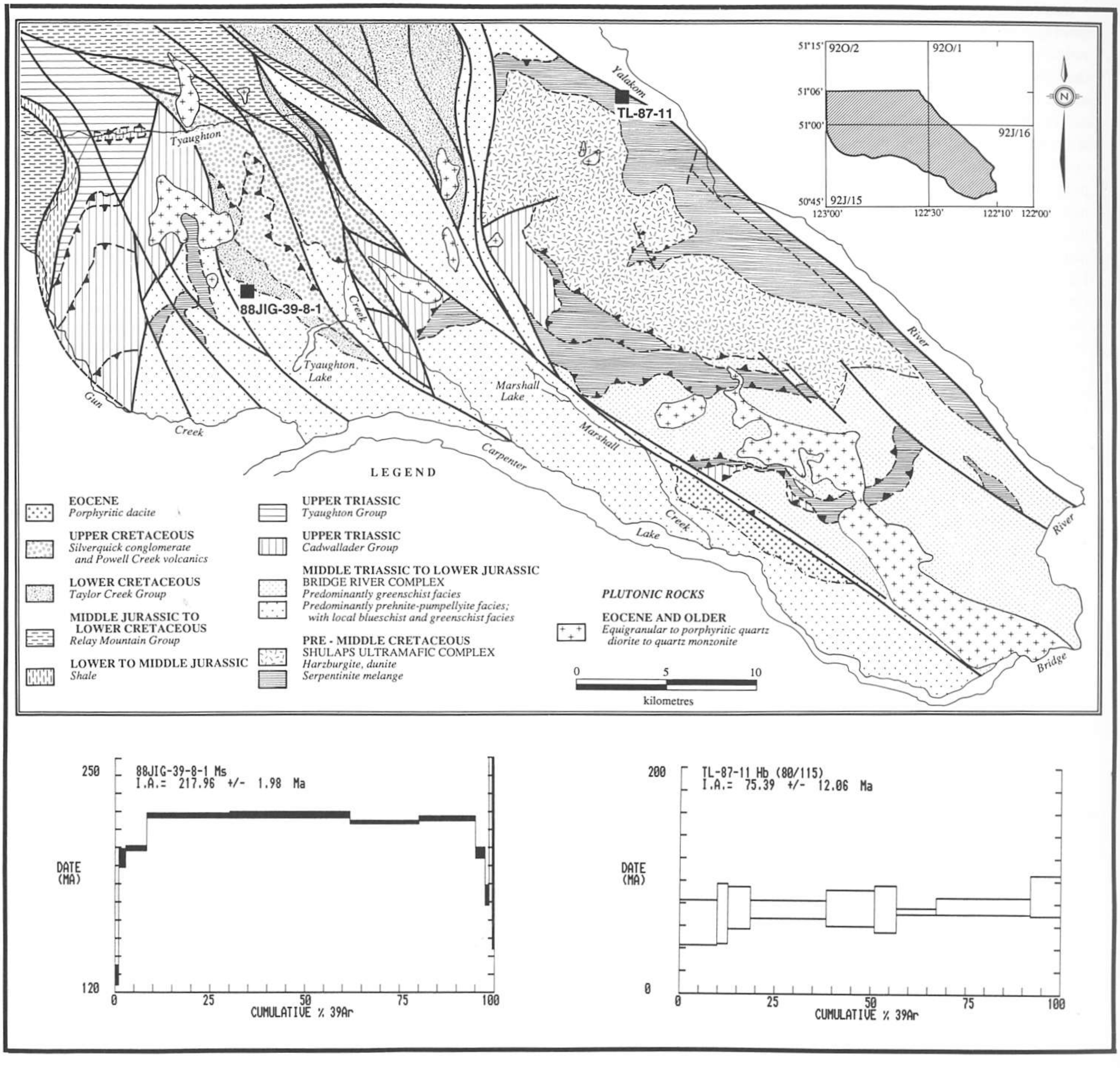

Figure 1-5-2. Sample locations and $^{40}Ar/^{39}Ar$ age spectra for rocks from the Bridge River Terrane, southwestern British Columbia.

authors. This work combined with that by Potter (1986) revealed a zone with up to 1000 metres of structural thickness of penetratively deformed serpentinite mélange which contains knockers of a variety of rock types. This zone is exposed along the southern margin of the Shulaps ultramafic complex where it lies structurally beneath harzburgite with a mantle tectonite fabric (Calon *et al.*, 1990). Although most of the knockers are ultramafic and mafic from the cumulate section of an ophiolite, some low-grade metasedimentary rocks and dioritic plutonic rocks are also present. The larger dioritic bodies occur as flattened pods with variably rodingitized margins. Locally, olivine has been regenerated from serpentine in the mélange. The smallest granitoid pods are completely replaced by rodingite but outcrop in linear (planar) zones suggesting the form of a discontinuous dike. The best-preserved dioritic rocks are characterized by blocky to acicular hornblende phenocrysts and are internally undeformed. In one knocker, an undeformed dike cuts penetratively deformed metasedimentary rocks but does not extend into the enclosing serpentinite.

In 1989 mapping along the north and northeast sides of the Shulaps ultramafic complex (Schiarizza *et al.*, 1990, this volume) revealed that the complex is broadly synformal such that the imbricate zone also outcrops in these areas. Contrary to the interpretation of Glover *et al.* (1988) and Archibald *et al.* (1989) it now appears that this penetratively deformed serpentinite containing knockers of ultramafic rocks, gabbro/amphibolite and aligned pods of hornblende plagioclase porphyry is part of the Shulaps imbricate zone and is not necessarily related to the Yalakom fault. A narrow zone of quartz-carbonate-fuchsite alteration a short distance to the northeast of the sample site may mark the position of the fault, which juxtaposes the Shulaps imbricate zone and Jura-Cretaceous sedimentary rocks to the northeast. Since the imbricate zone is cut by the Yalakom fault, the dike chosen for dating predates at least the latest movement on this fault. As low-potassium amphiboles commonly contain excess argon, step-heating was done to provide a more reliable date for this dike.

$^{40}Ar/^{39}Ar$ ANALYTICAL METHODS

Mineral separates were prepared using a Frantz magnetic separator, heavy organic liquids and, where appropriate, by hand-picking.

Samples and six flux-monitors (standards) were irradiated with fast neutrons in position 5C of the McMaster nuclear reactor (Hamilton, Ontario) for 30 hours. The monitors were distributed throughout the irradiation container and J-values for individual samples were determined by interpolation.

Both step-heating experiments and analysis of the monitors were done in a quartz tube heated using a Lindberg furnace. The bakeable, ultra-high vacuum, stainless steel argon-extraction system is operated on-line to a substantially modified A.E.I. MS-10 mass-spectrometer run in the static mode. Measured mass-spectrometric ratios were extrapolated to zero-time, corrected to an $^{40}Ar/^{36}Ar$ atmospheric ratio of 295.5, and corrected for neutron induced ^{40}Ar from potassium, and ^{39}Ar and ^{36}Ar from calcium. Dates and errors were calculated using formulae given by Dalrymple *et al.* (1981), and the constants recommended by Steiger and Jäger (1977). The errors shown in Table 1-5-1 were used to plot the age spectra in Figure 1-5-2; these represent the analytical precision at 2σ assuming that the error in the J-value is zero.

TABLE 1-5-1
$^{40}Ar/^{39}Ar$ STEP HEATING DATA

88JIG-39-8-1Ms
J = 0.00717

Temp °C	40/39 (1)	36/39 (1)	37/39 (1, 2)	Vol. ^{39}Ar $\times 10^{-8}$ cm^{3}NTP (3)	f39	%^{40}Ar Rad.	Date $\pm 2\sigma$ (4, 5) Ma
500	26.518	0.0548	1.664	0.291	0.0102	39.34	130.4±5.6
575	18.778	0.0100	3.411	0.508	0.0178	85.56	197.3±5.0
675	17.555	0.0033	0.842	1.611	0.0563	94.64	203.3±1.4
750	18.456	0.0007	0.019	6.208	0.2170	98.78	221.8±1.1
800	18.391	0.0004	0.008	9.144	0.3196	99.31	222.2±1.5
850	18.174	0.0008	0.024	5.208	0.1820	98.66	218.4±0.9
900	18.304	0.0007	0.011	4.267	0.1492	98.83	220.2±1.2
960	17.463	0.0037	0.053	0.659	0.0230	93.60	200.1±3.0
1020	18.773	0.0151	0.086	0.283	0.0099	76.11	175.8±5.7
1070	25.391	0.0093	0.202	0.248	0.0087	89.19	271.7±8.0
1200	30.426	0.0428	0.334	0.181	0.0063	58.48	216.8±72.6

Weight = 90 mg
Total $^{39}Ar = 28.608 \times 10^{-8}$ cm^{3}NTP
I.A. = 218.1 ± 1.9 Ma
P.A. = 221.0 ± 1.2 Ma; 750 to 900°C

TL-87-11 Hb (80/115)
J = 0.00718

Temp °C	40/39 (1)	36/39 (1)	37/39 (1, 2)	Vol. ^{39}Ar $\times 10^{-9}$ cm^{3}NTP (3)	f39	%^{40}Ar Rad.	Date $\pm 2\sigma$ (4, 5) Ma
725	52.752	0.1630	4.401	2.853	0.0978	9.27	62.4±20.1
825	19.175	0.0494	12.065	0.843	0.0289	28.58	70.2±26.8
925	18.164	0.0433	8.749	1.751	0.0600	33.04	76.5±18.8
975	7.548	0.0089	12.086	5.759	0.1975	77.12	74.4±8.2
1000	8.540	0.0116	11.837	3.656	0.1254	69.98	76.4±15.8
1025	11.485	0.0216	11.346	1.691	0.0580	51.72	75.9±20.6
1050	10.654	0.0197	12.528	3.067	0.1052	53.83	73.4±2.8
1100	8.585	0.0116	12.921	7.233	0.2480	71.10	78.0±6.8
1200	11.424	0.0182	11.156	2.313	0.0793	60.11	87.5±17.8

Weight = 200 mg
Total $^{39}Ar = 29.166 \times 10^{-9}$ cm^{3}NTP
I.A. = 75.4 ± 12.1 Ma
P.A. = 77.0 ± 10.7 Ma; 925 to 1200°C

(1) True ratios corrected for fractionation and discrimination.
(2) $^{37}Ar/^{39}Ar$ is corrected for the decay of ^{37}Ar during and after irradiation using a ^{37}Ar half-life of 35.1 days.
(3) Volume of ^{39}Ar is determined using equilibration peak height and mass spectrometer sensitivity.
(4) Isotope production ratios for the McMaster reactor are from Masliwec (1981).
(5) Ages calculated using the constants recommended by Steiger and Jäger (1977). Errors represent the analytical precision only (i.e., error in J value = 0). Flux monitor used: LP-6 Biotite at 128.5 Ma.
(6) I.A. – integrated age for all steps. P.A. = plateau age.

RESULTS AND DISCUSSION

BLUESCHIST FACIES ROCKS

RESULTS

The age spectrum for the white mica from the blueschist locality (88JIG-39-8-1) is shown in Figure 1-5-2; analytical data are listed in Table 1-5-1. The integrated age of this sample is 218.1 ± 1.9 (2σ) Ma which is in the middle of the range of previous dates for this site (Permo-Triassic; Garver *et al.*, 1989b). The 500°C step yields a date of 130 Ma, and dates for subsequent steps increase to 222 Ma for the 750°C step. The steps from 750° and 900°C yield a well-defined plateau date of 221.0 ± 1.2 (2σ) Ma for 87 per cent of the ^{39}Ar released from the sample. In this temperature range, the $^{37}Ar/^{39}Ar$ ratio ranges from 0.008 to 0.024 which corresponds to the low Ca/K ratio of phengite (*e.g.*, Sisson and Onstott, 1986). The age spectrum (at least to the 960°C step) approximates the form of a volume diffusion loss-profile. Comparison of the spectrum to computed loss-profiles suggests that the sample has only lost a small amount (<10 per cent) of argon through post-Triassic reheating. Thus, it appears that blueschist facies metamorphism in the Bridge River Terrane was over by at least 221 Ma (Late Triassic time).

DISCUSSION

The white mica from the blueschist contains a record of two tectonothermal events. The earlier event is Late Triassic or older and is revealed in the broad plateau segment of the age spectrum. Blueschist facies metamorphism is a low-temperature event (300–400°C; Sisson and Onstott, 1986) and the mica may have grown in this temperature range. As the mica grew syntectonically, the 221 Ma plateau date implies that both deformation and metamorphism are at least this old.

The low dates for the initial steps indicate that the area was affected by a thermal event in post-Late Triassic time. This fact may provide an explanation of the wide range of dates reported by Garver *et al.* (1989b). The older K-Ar dates [244 ± 7 (1σ) Ma] may be from samples which experienced overprinting to a lesser degree. Ideally, the age of the overprint should be given by the first fraction of argon released in a step-heating experiment; for small losses, the higher temperature plateau segment should approach the "true" cooling age of the sample. In practice, however, the age of the initial step is a relatively inaccurate estimate and only serves as a guide (usually a maximum) to the timing of the event. The region contains several small, high-level, mid-Cretaceous and younger plutons which may have overprinted the white mica; however, there are none within a kilometre of the sample site. It is more probable that the overprint is the result of a more pervasive thermal event that may have been coincident with structural thickening associated with contractional deformation between 110 and 85 Ma (Garver, 1989).

As noted by Garver *et al.* (1989a), in this area, overturned strata of the mid-Cretaceous Taylor Creek Group and overlying Silverquick conglomerate are in apparent thrust contact with upright strata of the Silverquick conglomerate. Unconformably underlying the Taylor Creek Group are the blueschist and greenschist rocks of the Bridge River Terrane. The overturning and thrusting occurred during a widely-recognized phase of mid-Cretaceous compressional tectonics that ended prior to emplacement of several Late Cretaceous plutons of the coast plutonic complex (85 Ma; Garver *et al.*, 1989a). The 130 Ma date from the first step of the age spectrum suggests that this event was the source of the thermal disturbance recorded in the age spectrum of the white mica. Based on modelling of the age spectrum, the thermal event must have been of short duration and/or low temperature (the closure temperature of argon diffusion in white mica is 350°C); an original cooling age only slightly greater than the 221 Ma plateau date (225–230 Ma) is favoured for this sample. The two K-Ar dates older than this (reported by Garver *et al.*, 1989b) remain problematic and additional analyses of white mica are planned.

HORNBLENDE PLAGIOCLASE PORPHYRY

RESULTS

The age spectrum for the amphibole from the hornblende plagioclase porphyry (TL-87-11) is shown in Figure 1-5-2; analytical data are listed in Table 1-5-1. The integrated date for this age spectrum is 75.4 ± 12.1 (2σ) Ma which is identical to the previously reported $^{40}Ar/^{39}Ar$ total-fusion date (Archibald *et al.*, 1989). The three lowest temperature steps are characterized by low radiogenic content, erratic $^{37}Ar/^{39}Ar$ ratios and large errors. In contrast, the remaining, higher temperature steps, representing 81 per cent of the total ^{39}Ar, have a consistent $^{37}Ar/^{39}Ar$ ratio in the range 11.2 to 12.9, are much more radiogenic and represent the main pulse of argon release from the amphibole. These steps define a plateau date of 77 ± 11 (2σ) Ma. Although the analytical errors are large (due to the small volume of argon released in each step), there is no evidence in the age spectrum of a later thermal overprint or of the presence of initial argon, and thus, we consider the plateau date to be a reliable cooling age. This date indicates that the dike is not younger than Late Cretaceous.

DISCUSSION

The amphibole from the diorite dike in the imbricate zone within the lower structural levels of the Shulaps ultramafic complex yielded a cooling age of 77 ± 11 Ma. This may or may not be the time of emplacement, however, the date does indicate that the dike cooled through 500°C (the approximate closure temperature for argon diffusion in hornblende) in Late Cretaceous time. These dikes are common in the imbricate zone, were emplaced synkinematically and appear to have been the source of major reheating in this structural panel. Evidence of reheating in the imbricate zone is preserved in the age spectrum for amphibole in a knocker of brecciated amphibolite (Archibald *et al.*, 1989). This rock is from the southern part of the imbricate zone and presumably is the same age as the oceanic crust that formed part of the protolith of the Shulaps complex. The plateau segment in this age spectrum is, within analytical error, identical to the age from the diorite dike. There is no evidence of the undoubtedly older, seafloor metamorphism that produced the rock, or of cooling related to mid-Jurassic obduction as postulated by Potter (1986). Thus, the brecciated amphibolite was reheated to at least greenschist-facies conditions and cooled through 500°C in Late Cretaceous time.

The duration of this reheating event is, at present, unknown. Within the study area, most magmatic rocks that postdate the 85 Ma and older Coast plutonic complex are Paleocene or younger (Archibald *et al.*, 1989) rather than Late Cretaceous. However, on a regional scale, there is evidence of plutonic/thermal activity in the interval from 78 to 73 Ma. In the Warner Pass map area, sericitic alteration and alunite associated with mineralization yield total-fusion $^{40}Ar/^{39}Ar$ dates in this range (Archibald *et al.*, 1989). Further, rocks with a similar thermotectonic history are present in the North Cascades crystalline core of Washington and southern British Columbia, which by mid-Cretaceous time shares the same tectonic history and was probably continuous along strike to the south of the study area. The crystalline core contains granitoid rocks of this age, many of which have both primary and cooling ages in this interval (Tabor *et al.*, 1989; Journeay and Csontos, 1989). Some of these are epidote tonalites which were emplaced at deep crustal levels. Although circumstantial, the scattered occurrences of dates in this range in the Tyaughton Creek area may reflect this short-lived episode of magmatic activity. It is possible that in British Columbia plutonic rocks of this age were emplaced where the crust was cut by major faults such as the basal décollement of the Shulaps ultramafic complex or the Yalakom and Marshall Creek strike-slip faults which bound the western and northeastern margins of the Shulaps complex.

SUMMARY AND CONCLUSIONS

Several conclusions may be drawn concerning the timing of deformation and metamorphism in the Bridge River Terrane:

- Blueschist facies metamorphism and attendent deformation within the fault-bounded panels of blueschist are Late Triassic events (>221 Ma). As the age of the Bridge River complex is in part Permo-Triassic, subduction may have involved relatively young oceanic crust.
- Imbrication of the blueschist-bearing rocks and adjacent rocks, both of which contain metamorphic prehnite, may have been an Early Cretaceous event (≈130 Ma) or a mid-Cretaceous event (≈110–95 Ma) related to major structural thickening which produced increased thermal activity. This thermal event was (locally) short lived and temperatures were probably much less than 350°C.
- Hornblende diorite dikes intruded the imbricate zone at the base of the Shulaps ultramafic complex in Late Cretaceous time (>77 Ma) during the final stages of emplacement and uplift of the complex. This event predates, or was synchronous with at least some of the movement on the Yalakom fault. The dikes appear to be restricted to the Shulaps complex which suggests that bounding faults acted as a conduit that tapped (deep?) sources of magma.
- Although Rusmore *et al.* (1988) argue for Middle Jurassic terrane accretion during closure of the Bridge River–Cache Creek ocean basin, to date, there is no isotopic evidence of thermal activity related to this event in the Tyaughton Creek map area.

ACKNOWLEDGMENTS

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NOTES

GEOLOGY AND MINERAL OCCURRENCES OF THE YALAKOM RIVER AREA* (92O/1, 2, 92J/15, 16)

By P. Schiarizza and R.G. Gaba,
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KEYWORDS: Regional mapping, Shulaps ophiolite, Bridge River complex, Cadwallader Group, Yalakom fault, Mission Ridge fault, Marshall Creek fault.

INTRODUCTION

The Yalakom River area covers about 700 square kilometres of mountainous terrain along the northeastern margin of the Coast Mountains. It is centred 200 kilometres north of Vancouver and 35 kilometres northwest of Lillooet. Our 1989 mapping provides more detailed coverage of the northern and western Shulaps Range, partly mapped in 1987 (Glover *et al.*, 1988a, 1988b) and 1988 (Schiarizza *et al.*, 1989a, 1989b), and extends the mapping eastward to include the eastern part of the Shulaps Range, the Yalakom and Bridge River valleys and the adjacent Camelsfoot Range. In addition, several weeks were spent re-examining critical areas in the Tyaughton Creek area and traversing the area south of Gun Creek in an effort to mesh our work with the maps produced by B.N. Church during his mineral deposit studies of the Bridge River mining camp (Church, 1987; Church and MacLean, 1987a; Church *et al.*, 1988a, 1988b; Church and Pettipas, 1989).

Mapping in the Yalakom River area was carried out in cooperation with Meg Coleman of Carleton University who extended her 1988 mapping of the Mission Ridge area (Coleman, 1989) northwestward to Shulaps Creek. It also incorporates detailed mapping of the Shulaps ultramafic complex in the Jim Creek–East Liza Creek area, begun in 1988 by Tom Calon and continued this field season by Calon, John Malpas and Rob Macdonald, all from the Memorial University of Newfoundland (Calon *et al.*, 1990, this volume). Geological mapping and sampling by D.A. Archibald of Queen's University extends a geochronology study begun in 1987 and continued in 1988 (Archibald *et al.*, 1989, 1990, this volume).

This is the final year of a 4-year regional mapping project, initiated east of Taseko Lakes in 1986 and funded by the Canada/British Columbia Mineral Development Agreement. Open File geology and mineral potential maps covering this season's study area will be released in February, 1990. A final report covering the entire 4-year program, including updated 1:50 000 maps, will be prepared during the 1990/91 fiscal year.

REGIONAL GEOLOGY

The regional geologic setting of the Taseko–Bridge River project area is described by Glover *et al.* (1988a) and Schiarizza *et al.* (1989a). The distribution and relationships of the major tectonostratigraphic assemblages are summarized in Figures 1-6-1 and 1-6-2.

The Yalakom River area, comprising the southwestern segment of the project area, encompasses the whole of the Shulaps ultramafic complex which is interpreted by Nagel (1979), Potter (1983, 1986) and Calon *et al.*(1990) as a dismembered ophiolite. The area south and west of the Shulaps complex is underlain mainly by oceanic rocks of the Permian(?) to Jurassic Bridge River complex, and arc-derived volcanic and sedimentary rocks of the Upper Triassic Cadwallader Group. These two assemblages are structurally interleaved over a broad area extending from west of Gold Bridge eastward to the slopes northeast of the Yalakom and Bridge rivers. In the Bralorne–Gold Bridge area they are imbricated with the Permian Bralorne diorite complex and associated ultramafic rocks.

Sedimentary rocks exposed north and west of the Bridge River–Cadwallader belt range from Late Triassic to mid-Cretaceous in age. The base of the section comprises Upper Triassic clastic rocks and limestone of the Tyaughton Group and overlying Lower to Middle Jurassic sandstone and shale of the Last Creek formation (Tipper, 1978; Umhoefer, 1989). These rocks are not seen in depositional contact with the slightly older Cadwallader Group, but are inferred to represent a continuation of the same arc-derived sedimentation, and are included within the Cadwallader Terrane of Rusmore *et al.* (1988).

Younger sedimentary rocks within the region are assigned to the Tyaughton basin (Jeletzky and Tipper, 1968; Kleinspehn, 1985). Southwest of the Yalakom fault these include shallow-marine clastic rocks of the Middle Jurassic to Lower Cretaceous Relay Mountain Group together with conglomerate and associated finer grained clastics and volcanic rocks of the Albian Taylor Creek Group. The Relay Mountain Group outcrops most extensively in the Warner Pass and Noaxe Creek map areas where it is locally in depositional contact with the underlying Last Creek formation (Umhoefer, 1989). To the south and southeast the Relay Mountain Group occurs as local fault-bounded slivers in contact with either the Cadwallader Group or the Bridge

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

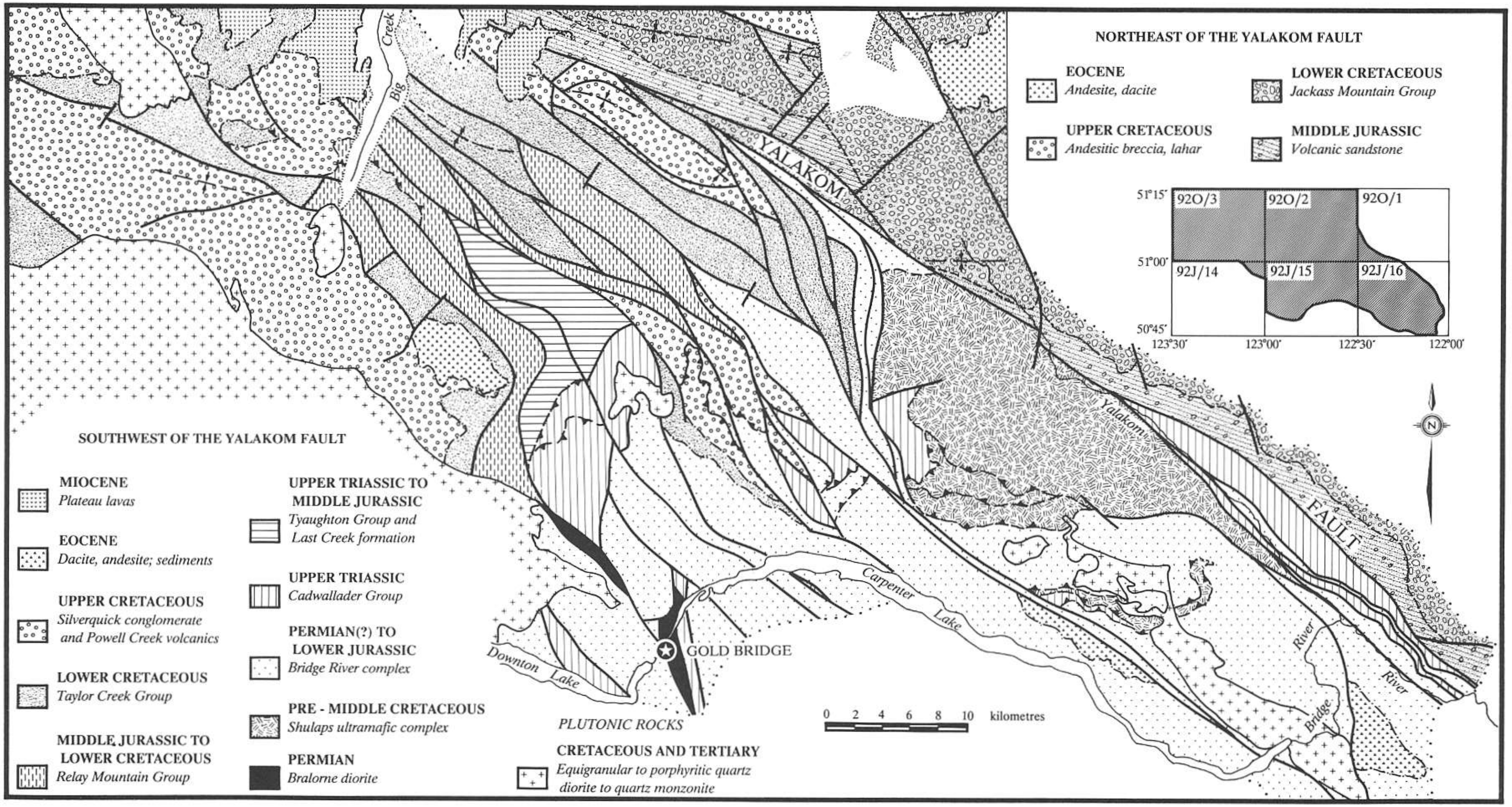

Figure 1-6-1. Generalized geology of the Taseko–Bridge River project area.

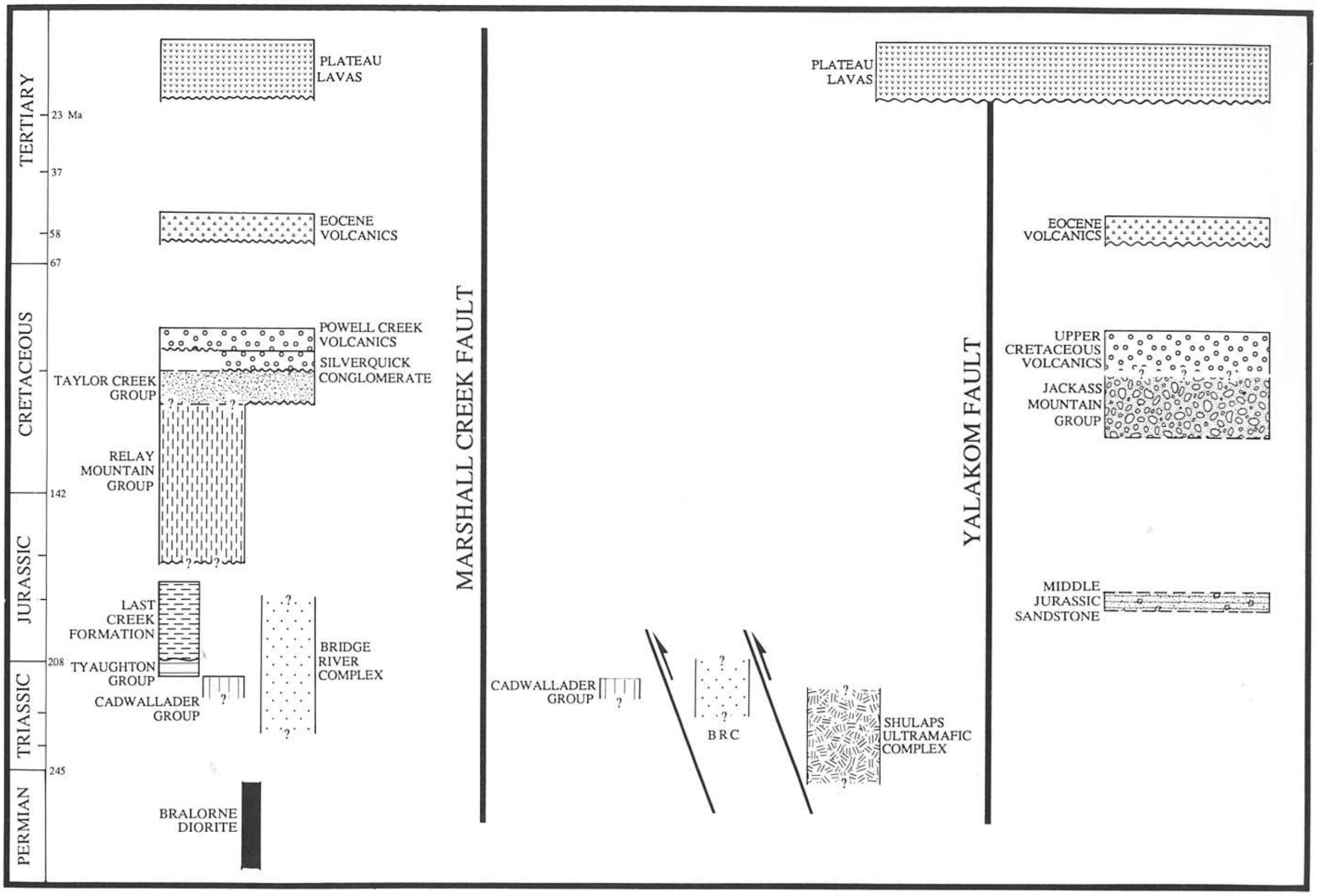

Figure 1-6-2. Tectonostratigraphic assemblages of the Taseko–Bridge River area.

River complex. Synorogenic deposits of the Taylor Creek Group sit stratigraphically above the Relay Mountain Group in the Relay Mountain area, and above deformed Bridge River rocks in the Taylor Creek area (Garver *et al.*, 1989a). Clasts within the Taylor Creek Group provide the first evidence of regional uplift and erosion of the Bridge River complex.

Upper Cretaceous andesitic breccias and flows of the Powell Creek volcanics are widespread in the northwestern part of the area, where they sit above the Taylor Creek Group and older rocks with pronounced angular unconformity. To the southeast Upper Cretaceous deposits of the nonmarine Silverquick conglomerate rest unconformably above the Taylor Creek Group and pass gradationally upward into andesitic volcanic breccia correlated to the Powell Creek volcanics (Garver *et al.*, 1989a).

Northeast of the Yalakom fault Mesozoic sedimentary rocks are distinctly different from those to the southwest. The base of the succession comprises Middle Jurassic volcanic-rich sandstones and associated shale and conglomerate. These are overlain by a thick succession of arkosic sandstone, conglomerate and shale of the Lower Cretaceous Jackass Mountain Group. Andesitic volcanic and volcaniclastic rocks, similar to the Powell Creek volcanics southwest of the Yalakom fault, occur locally as fault-bound slivers within this belt, but were not observed in stratigraphic contact with the Jackass Mountain Group.

Mesozoic strata throughout the region are intruded by felsic to intermediate stocks and dikes ranging from Late Cretaceous to Oligocene in age (Archibald *et al.*, 1989). Locally the strata are unconformably overlain by Eocene volcanic and sedimentary rocks and by Miocene to Pliocene plateau lavas of the Chilcotin Group (Mathews, 1989). Late Cretaceous granite to quartz diorite of the Coast plutonic complex intrudes the Mesozoic strata in the southwestern part of the Warner Pass map area and along the western edge of the Bralorne map area (Figure 1-6-1).

LITHOLOGY

SHULAPS ULTRAMAFIC COMPLEX

The Shulaps ultramafic complex covers most of the northern half of the study area (Figure 1-6-3). It is bounded by the Yalakom fault to the northeast, and is juxtaposed against Bridge River and Cadwallader Group rocks across thrust and high-angle faults on the north, west, south and southeast. The complex was first studied in detail by Leech (1953), who concluded that it was an intrusive body, emplaced in the Late Triassic or Early Jurassic, and later redistributed along fault

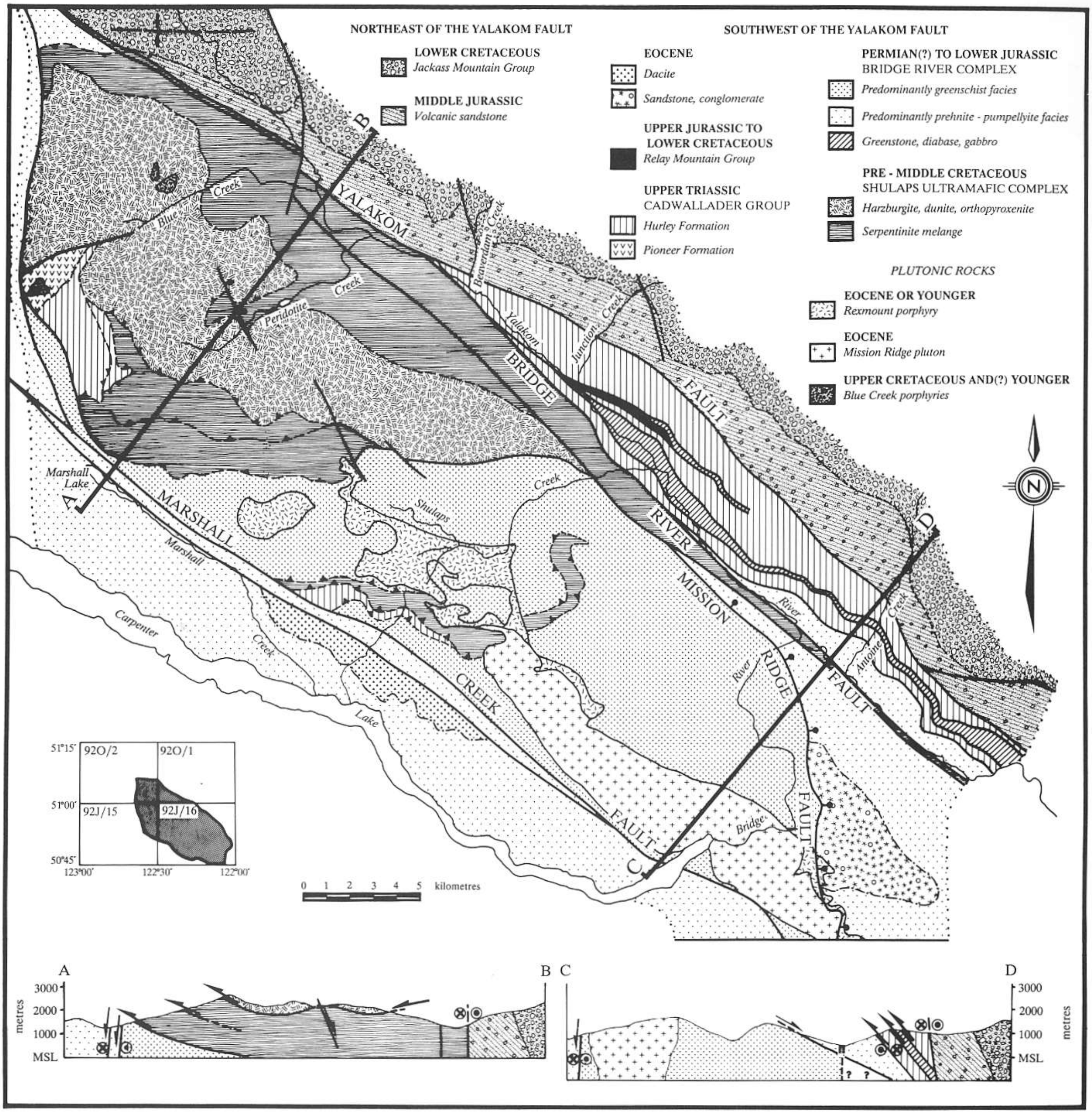

Figure 1-6-3. Generalized geology map and cross-sections, Yalakom River map area.

zones to the west and northwest. Later workers (Monger, 1977; Nagel, 1979; Wright *et al.*, 1982; Potter, 1983, 1986) suggested that the Shulaps and Bridge River complexes together constitute a dismembered ophiolite. The present study, and in particular the detailed mapping by Tom Calon and coworkers along the southwestern margin of the complex (Calon *et al.*, 1990) has confirmed this interpretation.

The structurally and topographically highest portion of the Shulaps complex comprises variably serpentinized harzburgite with lesser dunite and orthopyroxenite. The harzburgite is locally layered, with layering defined by centimetre-wide bands of orthopyroxenite and rarely by wider bands of dunite, orthopyroxenite and harzburgite. A penetrative mineral foliation and lineation are locally evident; the foliation is typically parallel, or at a low angle to, compositional layering. This foliation is interpreted by Calon *et al.* to be a mantle tectonite fabric. A spectacular mylonitic foliation displayed by harzburgite 2 kilometres north-northeast of Serpentine Lake is also thought to be a mantle fabric. Dunite within the upper harzburgite unit locally defines layering, but is more common as unoriented pods and lenses, some of which crosscut layering and foliation within the harzburgite. This may reflect an upper mantle origin for the harzburgite unit, in the lower part of the transition zone to overlying ultramafic-mafic cumulates (T. Calon, personal communication, 1988). Where the harzburgite unit sits structurally above cumulate-derived serpentinite mélange, it does so across a sole of harzburgite-derived foliated serpentinite that can generally be distinguished from the underlying cumulate-derived serpentinite (Calon *et al.*, 1990). The thick harzburgite unit itself is apparently imbricated across thrust contacts defined by similar foliated serpentinite, but these have not been mapped in detail.

A well-exposed belt of cumulate-derived serpentinite mélange structurally underlies the mantle harzburgite unit along the southwestern margin of the Shulaps complex, between East Liza and Hog creeks. It is in turn structurally underlain by Bridge River schists and locally by Cadwallader Group metasediments. This belt was studied in detail by Calon *et al.* and the following summary is based largely on their work. The mélange comprises foliated serpentinite containing blocks of ultramafic, gabbroic, volcanic and sedimentary rock. The largest knockers, up to hundreds of metres in size, derive from an igneous complex which includes layered ultramafic cumulates, layered gabbro and varitextured gabbro, all cut by swarms of mafic to intermediate dikes. Gabbro at the base of the mélange in the western part of the belt grades into a dike complex which in turn grades into pillowed volcanic rocks. The mélange thus contains remnants of a plutonic-volcanic suite characteristic of the upper part of an ophiolite complex. Volcanic and sedimentary knockers occur throughout the mélange and presumably represent a sampling of the footwall succession across which the Shulaps complex was emplaced. Sedimentary knockers include bedded chert, limestone, sandstone and pebble conglomerate. These in part resemble rocks found in the Bridge River complex, but in part may have been derived from the Cadwallader Group or an unknown clastic sequence.

Serpentinite also dominates the poorly exposed northeastern part of the Shulaps complex. In exposures extending from the northern tip of the complex southeastward to Peridotite Creek this serpentinite contains small knockers and boudinaged dikes of diabase, amphibolite and gabbro. The serpentinite mélange exposed along upper Peridotite Creek clearly sits structurally beneath mantle harzburgite. The two serpentinite mélange belts are therefore inferred to be continuous beneath the intervening mantle harzburgite unit that comprises the backbone of the Shulaps Range, indicating that the Shulaps complex is broadly synformal in nature.

A separate belt of serpentinite mélange outcrops between 4 and 7 kilometres south of the main part of the Shulaps complex and has been traced for about 12 kilometres eastward from the Marshall Creek fault (Figure 1-6-3). Serpentinite within this belt encloses knockers of ultramafic, gabbroic and dioritic rocks similar to those within the cumulate-derived mélange exposed between East Liza and Hog creeks. The lower mélange belt is structurally overlain by penetratively deformed Bridge River rocks, and is underlain by Cadwallader Group conglomerates and sandstones in the west and Bridge River rocks in the east. The occurrence of ophiolitic serpentinite mélange at this lower structural level suggests that emplacement and imbrication of the Shulaps complex was a complex process involving some out-of-sequence thrusting or folding.

BRIDGE RIVER COMPLEX

The Permian(?) to Jurassic Bridge River complex includes variably metamorphosed and structurally imbricated chert, mafic extrusive and intrusive rocks, limestone, clastic rocks and serpentinite (Potter, 1983, 1986; Schiarizza *et al.*, 1989a, Garver *et al.*, 1989a). It underlies much of the area southwest of the Yalakom fault, where it is structurally interleaved with rocks of the Cadwallader Group and Shulaps ultramafic complex. A central block of penetratively deformed schists and phyllites exposed in the Shulaps Range is separated from lower grade rocks to the northeast and southwest by the Mission Ridge and Marshall Creek faults respectively (Figure 1-6-3).

Bridge River rocks southwest of the Marshall Creek fault are described by Potter (1983, 1986) and Schiarizza *et al.* (1989a). They consist mainly of prehnite-pumpellyite-grade chert and greenstone, together with lesser amounts of argillite, limestone, tuff, chert and volcanic-rich sandstone, pebble conglomerate, diabase and gabbro. Similar rocks characterize the belt east of the Mission Ridge fault although clastic rocks and limestone are uncommon in this area. Structural slivers of serpentinite and diabase-gabbro-bearing serpentinite mélange are common in the area southwest of the Bridge and Yalakom rivers. Bridge River rocks northeast of the Bridge River comprise chert and greenstone that are structurally overlain by the Hurley Formation across a moderately northeast-dipping fault. Similar northeast-dipping faults bound two persistent slivers of pillowed and brecciated greenstone with lesser diabase, gabbro and serpentinite that occur within the Hurley belt farther to the northeast. These are tentatively assigned to the Bridge River complex, but in part may be equivalent to the Pioneer Formation and/or Shulaps serpentinite mélange.

The Bridge River complex between the Marshall Creek and Mission Ridge faults is represented mainly by phyllites and schists that were penetratively deformed under predominantly greenschist-facies metamorphic conditions (Potter, 1983, 1986). The most common rock types are medium to dark grey phyllite, quartz phyllite and biotite-bearing schist (locally garnet-bearing) derived from argillite and chert, and chloritic schist (locally biotite-bearing) derived from mafic volcanic rock. These are locally intercalated with crudely foliated phyllosilicate-bearing metasandstone, marble, and chlorite-actinolite-carbonate schists probably derived from impure calcareous sediments. Serpentinite is commonly interleaved with the schists for several kilometres south of the Shulaps complex. Non-penetratively deformed rocks similar to those which characterize the Bridge River complex elsewhere within the map area occur locally in the block, particularly along the upper reaches of Hell and LaRochelle creeks.

The Bridge River schists are bounded by the Shulaps ultramafic complex on the north; they sit structurally beneath serpentinite mélange on the west, but are juxtaposed directly against mantle harzburgite to the east (Figure 1-6-3). Farther south they enclose an imbricate belt of serpentinite mélange and Hurley Formation which was traced for more than 12 kilometres eastward from the Marshall Creek fault. The schists are intruded by foliated granodiorite of the Eocene Mission Ridge pluton, which crosscuts the imbricate belt, as well as by undeformed to foliated and folded dikes and sills of similar composition. They, together with the Mission Ridge pluton, are also intruded by the undeformed Rexmount porphyry and associated dikes.

CADWALLADER GROUP

The Upper Triassic Cadwallader Group, as redefined by Rusmore (1985, 1987), comprises mafic volcanic rocks of the Pioneer Formation and conformably overlying clastic sediments of the Hurley Formation. These rocks, inferred to be volcanic-arc related, are the same age as parts of the Bridge River complex with which they are structurally interleaved over a broad area extending from the Coast plutonic complex west of Gold Bridge to the Yalakom River valley (Figure 1-6-1). The Pioneer Formation consists of green to purplish weathering, commonly amygdaloidal, pillowed and massive greenstone and greenstone breccia. The overlying Hurley Formation consists mainly of thin-bedded sandstone and siltstone turbidites, but commonly includes distinctive pebble to cobble conglomerates containing limestone, mafic to felsic volcanic and granitoid clasts.

Within the study area, the Cadwallader Group is exposed in two areas on the northeast side of the Marshall Creek fault, as well as within an extensive, but previously unrecognized belt along the northeastern slopes of the Yalakom and Bridge rivers (Figure 1-6-3). Cadwallader rocks northeast of the Marshall Creek fault are structurally imbricated with the Shulaps complex and Bridge River phyllites. They were penetratively deformed under lower greenschist(?) facies metamorphic conditions such that fine-grained sediments are typically cleaved and clasts in conglomerate are locally highly flattened. The most extensive exposures are in the East Liza Creek area, where the Hurley Formation is structurally overlain, across a gently dipping and locally folded thrust contact, by a pillowed volcanic-dike-gabbro complex at the base of the Shulaps serpentinite mélange. Pillowed greenstone that contacts the Hurley to the north is tentatively asigned to the Pioneer Formation, although the nature of the contact has not been established. The greenstones are juxtaposed against the Shulaps harzburgite by a steeply dipping east-northeast-trending fault that defines a prominent left-stepping jog in the western boundary of the Shulaps complex. The Hurley Formation also outcrops 12 kilometres southeast of the East Liza Creek exposures, where it occurs as a narrow lens structurally overlain by the lower serpentinite mélange and underlain by Bridge River schists (Figure 1-6-3). This thrust-imbricated package is truncated on the west by the Marshall Creek fault; the Hurley lens pinches out 7 kilometres southeast of the fault, whereas the overlying serpentinite mélange belt was traced an additional 7 kilometres eastward before apparently pinching out within Bridge River schists.

The extensive belt of Hurley Formation sedimentary rocks mapped along and northeast of the Yalakom and Bridge rivers (Figure 1-6-3) was largely unrecognized by previous workers, although Leech (1953) describes the rocks, including the distinctive conglomerate lenses, where they outcrop along the Yalakom River north of Shulaps Creek, and Roddick and Hutchison (1973) mapped a small patch of Hurley Formation at the mouth of Antoine Creek. Coleman (1989) noted the similarity between sedimentary rocks northwest of Applespring Creek and the Cadwallader Group described by Rusmore (1987), but tentatively included them in the Lillooet Group. We presently map the Hurley Formation within a belt up to 4 kilometres wide and more than 30 kilometres long that extends from Beaverdam Creek southeastward to at least Applespring Creek. This belt includes rocks assigned to an unnamed Lower Cretaceous unit by Leech (1953), to the the Relay Mountain and Jackass Mountain groups by Roddick and Hutchison (1973) and to the Lillooet Group by Coleman (1989). The Hurley Formation within this belt is strongly deformed by southwesterly overturned folds and northeast-dipping transpressional faults. It is imbricated with two mappable lenses of pillowed greenstone, volcanic breccia, diabase and gabbro tentatively assigned to the Bridge River complex and with one or more slivers of *Buchia*-bearing Relay Mountain Group (Figure 1-6-3).

The Hurley Formation within this belt consists largely of grey siltstone to fine-grained sandstone that occurs as thin, commonly graded and crosslaminated beds intercalated with dark grey mudstone. These are interbedded with thin to thick, locally graded beds of sandstone, calcareous sandstone and gritty sandstone, and rare thin to medium beds of laminated and crosslaminated limestone. Conglomerate occurs locally as lenticular beds several metres to more than 10 metres thick that commonly cut into underlying beds. It consists of angular to rounded pebbles, cobbles and blocks of light grey weathering limestone that occur with variable proportions of rounded felsic to intermediate volcanic and plutonic clasts within a limy matrix. These conglomerates, together with all of the other associated lithologies, are typical of the Hurley Formation elsewhere in the region, and form the basis for our correlation. Collections of limestone are presently being processed for conodonts in an attempt to confirm the inferred Late Triassic age of the rocks.

RELAY MOUNTAIN GROUP

Middle Jurassic to Lower Cretaceous shallow-marine sedimentary rocks of the Relay Mountain Group outcrop extensively in the Warner Pass and Noaxe Creek map areas (Figure 1-6-1). There, they are locally in depositional contact with the underlying Last Creek formation and are in turn overlain by the Taylor Creek Group (Figure 1-6-2). Within the Yalakom River area, Upper Jurassic and Lower Cretaceous *Buchia*-bearing rocks outcrop locally near Ore Creek (Leech, 1953; Jeletzky, 1967). These fossiliferous rocks apparently led Roddick and Hutchison (1973) to map most of the lower slopes northeast of the Yalakom and Bridge rivers between Applespring and Junction creeks as Relay Mountain Group. Most of these rocks are now interpreted as Hurley Formation on the basis of the lithologic correlation discussed in the previous section. The Relay Mountain Group is thought to be restricted to one or more narrow fault-bound slivers of mainly shale and siltstone structurally interleaved with the Hurley Formation northeast of the Yalakom River between Ore and Junction creeks (Figure 1-6-3).

Middle Jurassic Volcanic Sandstone Unit

Middle Jurassic rocks outcrop along the northeastern side of the Yalakom River area where they form a continuous belt bounded on the southwest by the Yalakom fault (Figure 1-6-3). They comprise a steep to moderately dipping (locally overturned along the Yalakom fault) east to northeast-facing succession of volcanic sandstones intercalated with lesser amounts of granule to pebble conglomerate, siltstone and shale. To the northeast, these rocks sit stratigraphically beneath the Jackass Mountain Group with no apparent angular discordance.

The Middle Jurassic section consists mainly of coarse to medium-grained, locally gritty, green to grey volcanic-lithic sandstone. The volcanic grains are commonly accompanied by lesser amounts of feldspar and fine-grained sedimentary rock fragments, and locally by several per cent glassy quartz grains. The sandstone commonly occurs as medium to thick, locally graded beds with relatively thin caps or interbeds of grey shale. Locally it is massive and apparently unbedded over intervals of several tens of metres. Thick beds of granule to small-pebble conglomerate are not uncommon and contain mainly volcanic and fine-grained sedimentary clasts, including lenticular clasts of grey argillite and siltstone that were probably local rip-ups. Thin-bedded, commonly laminated or crosslaminated siltstone and mudstone locally define intervals up to several tens of metres thick within the coarser rocks. These are commonly carbonaceous, as are some portions of the coarser grained intervals. Brown-weathering beds of calcareous sandstone, conglomerate or siltstone occur locally, and thin to medium beds of silty limestone were noted rarely.

The rocks described in the previous paragraph are in part Middle Jurassic in age on the basis of Aalenian and Bajocian ammonites collected from the lower part of the exposed interval near the mouth of Blue Creek (Leech, 1953; Frebold *et al.*, 1969; Tipper, 1978). The upper part of the succession has not been dated, but is lithologically similar to the underlying Middle Jurassic rocks and therefore may be separated from the overlying Lower Cretaceous Jackass Mountain Group by a significant disconformity. Correlative rocks in the Noaxe Creek map area were mapped as Unit 3v by Glover *et al.* (1988a, 1988b), and tentatively assigned to the lower part of the Jackass Mountain Group; ammonites collected from the unit along Dash Creek, however, were subsequently identified as Bajocian (T. Poulton, written communication, 1988) supporting the present correlation. The Middle Jurassic sandstones are lithologically similar to the Lillooet Group described by Duffell and McTaggart (1952) and Trettin (1961) along the Fraser River near Lillooet, and occupy the same stratigraphic position beneath the Jackass Mountain Group. The rocks near Lillooet were assigned a Lower Cretaceous age by Duffell and McTaggart (1952) but an ammonite recently discovered within them may be Middle Jurassic (J.W.H. Monger, personal communication, 1989), supporting their correlation, at least in part, with the Middle Jurassic sandstone unit described here.

Jackass Mountain Group

The Jackass Mountain Group (Selwyn, 1872; Duffell and McTaggart, 1952) comprises Lower Cretaceous clastic sedimentary rocks that sit stratigraphically above Middle Jurassic strata along the northeastern margin of the study area. This area is part of the southwestern margin of a continuous belt of Jackass Mountain Group exposures that extends from south of Lillooet northwestward almost 150 kilometres to Big Creek. Rocks within this belt have yielded sparse collections of Early Cretaceous fossils ranging from Barremian to Albian in age (Duffell and McTaggart, 1952; Trettin, 1961; Jeletzky and Tipper, 1968; Roddick and Hutchison, 1973).

Only the lower part of the Jackass Mountain Group was examined in the Yalakom River area (Figure 1-6-3). It consists mainly of olive-green to blue-green medium to coarse-grained feldspathic-lithic wackes, commonly with sparsely scattered granules to small pebbles of volcanic, sedimentary and plutonic rock fragments. The sandstones are typically massive, with bedding only locally defined by pebble concentrations or trains of siltstone intraclasts. Intervals of siltstone and shale, ranging from less than a metre to as much as 300 metres thick, occur locally and are characterized by distinct thin beds that may be graded or crosslaminated. Pebble to cobble conglomerates were observed only at or near the contact with the underlying Middle Jurassic section. These comprise predominantly rounded clasts of mainly volcanic and granitic to dioritic plutonic rocks, with relatively minor proportions of sedimentary and metamorphic clasts.

The basal contact of the Jackass Mountain Group was observed at one locality, 5 kilometres east of the mouth of Blue Creek, and was closely approached in several other places to the southeast. No angular discordance is apparent between Jackass Mountain Group and underlying rocks in any of these locations. The contact is typically marked, however, by a thin conglomerate unit at the base of the Jackass Mountain Group; the conglomerate, together with overlying sandstones, contains abundant plutonic detritus that is nowhere seen in the underlying section. This observation, in combination with the Middle Jurassic age of at least the lower part of the underlying unit as compared to the Barremian to Albian age of the Jackass Mountain Group, suggests that the contact is a significant disconformity.

Eocene Volcanic and Sedimentary Rocks

Sedimentary and volcanic rocks of probable Eocene age outcrop in two areas within the map area. On Mission Ridge they comprise sedimentary rocks exposed in a northwest-plunging syncline that is truncated on the west by the Mission Ridge fault (Figure 1-6-3). They apparently lie unconformably upon low-grade Bridge River rocks in the hangingwall of the fault, although the contact is poorly exposed. Eocene(?) rocks exposed 15 kilometres to the west-northwest, on the slopes northeast of Carpenter Lake, apparently also lie unconformably above low-grade Bridge River rocks. These comprise gently northeast-dipping intermediate volcanic and volcaniclastic rocks with lesser sediments that are truncated to the northeast by the Marshall Creek fault.

The sedimentary rocks on Mission Ridge comprise several hundred metres of thick-bedded, moderately well sorted, volcanic and chert-rich conglomerates interbedded with siltstone and sandstone in a fining-upwards sequence. Low-angle cross-stratification and basal scour in some of the conglomerates, and the presence of wood fragments and *Metasequoia* leaf fossils in the fine-grained intervals, indicate that deposition probably occurred in a fluvial environment. The composition of the sandstone and conglomerate suggest that two different source terrains supplied detritus to this sequence; one rich in chert and another rich in felsic volcanics. Mixing of the two types of detritus is common, but the felsic volcaniclastic detritus also occurs undiluted. Thin sections reveal that these white-weathering sandstones are replete with felsic volcanic clasts, beta quartz, plagioclase and biotite. Minor amounts of strained quartz, feldspar, muscovite and sedimentary rock fragments are present and may represent partial input from a plutonic and sedimentary source. Chert-rich sandstones and conglomerates contain minor amounts of felsic volcanicastic material but are dominated by clasts of veined and unveined chert and metachert with lesser quantities of volcanic rock and sandstone.

The Eocene(?) rocks northeast of Carpenter Lake comprise nearly 1000 metres of light grey to buff-weathering volcanic flows and breccias locally underlain by several tens of metres of sedimentary rocks. The volcanic rocks are mainly hornblende, biotite, quartz and feldspar-phyric dacites. The sediments comprise conglomerate, sandstone and shale, locally with narrow seams of lignite. Clasts in the conglomerate include chert with lesser amounts of granitic and felsic volcanic rock.

The chert-rich detritus within both the Mission Ridge and Carpenter Lake sedimentary sections was probably derived from the Bridge River complex. The intimate mixing of chert and felsic volcanic detritus, together with the presence of exclusively volcanic-derived interbeds on Mission Ridge and the thick volcanic succession along Carpenter Lake, suggests that volcanism was contemporaneous with deposition and comprised periodic eruptions that punctuated the erosion of the Bridge River complex. Both the Mission Ridge and Carpenter Lake sections are cut by normal faults that record relative uplift of an intervening belt of relatively high-grade metamorphic rocks during a complex period of Tertiary strike-slip and extensional faulting. Bridge River detritus within the Eocene sediments is, however, derived almost exclusively from low-grade parts of the complex, although rare clasts of quartz-biotite schist in one thin section from Mission Ridge may have been derived from Bridge River schists. This suggests that the presently exposed Eocene sediments were deposited relatively early in the uplift history, prior to unroofing of the higher grade rocks.

Intrusive Rocks

BLUE CREEK PORPHYRIES

Hornblende feldspar porphyry, diorite and quartz diorite that occur in and adjacent to the Blue Creek drainage area were referred to as Blue Creek porphyries by Leech (1953). They occur as abundant dikes and small plugs that intrude both mantle harzburgite and serpentinite mélange in the northern part of the Shulaps complex. Two of the largest plugs cut the harzburgite unit 6 kilometres west of the mouth of Blue Creek and host the Elizabeth and Yalakom gold-quartz veins. One of these has recently been dated at 58.4 ± 2.0 Ma by the whole-rock K-Ar method (Church and Pettipas, 1989). Fresh hornblende separates from hornblende feldspar porphyry that intrudes serpentinite mélange along the Yalakom fault, 4 kilometres to the northwest, however, yielded ^{40}Ar-^{39}Ar total fusion and step-heating dates of 75.6 ± 2.8 and 77.0 ± 10.7 Ma respectively (Archibald *et al.*, 1989, 1990). These mineral-separate dates may be a more reliable estimate of the cooling age or, alternatively, rocks included within the Blue Creek porphyries may represent more than one intrusive suite.

Although they are most abundant within the northern part of the Shulaps complex, dikes of hornblende feldspar porphyry and porphyritic diorite to quartz diorite also occur within the southern Shulaps complex and structurally underlying Bridge River schists. These include a suite of hornblende $\pm$ feldspar porphyry dikes that intrudes the mélange belt and caused local synkinematic metamorphism (Archibald *et al.*, 1989, 1990; Calon *et al.*, 1990). These may be related to the Blue Creek porphyries since ^{40}Ar-^{39}Ar step-heating of an amphibolite knocker within the same mélange belt suggests cooling of the mélange at about 73 Ma (Archibald *et al.*, 1989).

MISSION RIDGE PLUTON

The Mission Ridge pluton (Potter, 1983) is a markedly elongate body of coarse-grained biotite granodiorite that extends from 5 kilometres south of the study area (Coleman, 1989) for about 30 kilometres northwestward to the head of Holbrook Creek (Figure 1-6-3). It intrudes Bridge River schists within the belt of metamorphic rocks that is bounded by the Marshall Creek and Mission Ridge faults. Uranium-lead dating of zircon and monazite fractions from the pluton indicates an age of 47.5 ± 0.2 Ma (Coleman, M.Sc. thesis in progress) which is slightly older than a previously reported K-Ar biotite date of 44 Ma (Woodsworth, 1977). The Mission Ridge pluton displays strongly foliated margins and is accompanied, within the Bridge River schist belt, by both deformed and undeformed dikes of similar composition. It is inferred to be synkinematic with respect to the latest stages of deformation within the belt.

REXMOUNT PORPHYRY

Rexmount porphyry (Drysdale, 1916; Leech, 1953) refers to a light grey weathering rock comprising phenocrysts of hornblende, biotite, quartz and feldspar in an aphanitic to fine-grained groundmass. It outcrops mainly in the Shulaps Range between LaRochelle and Hog creeks (Figure 1-6-3) and may be an intrusive equivalent of the dacitic volcanics that occur on the other side of the Marshall Creek fault, 3 kilometres to the southwest (Drysdale, 1916; Roddick and Hutchison, 1973). Although many earlier maps (Roddick and Hutchison, 1973; Potter, 1983; Schiarizza *et al.*, 1989b) do not differentiate between the Rexmount porphyry and coarse-grained granodiorite of the Mission Ridge pluton, our 1989 mapping indicates that the porphyry is a discrete, later intrusive phase, as indicated by Woodsworth (1977). In the southwestern part of the belt it occurs mainly as dikes and sills cutting Mission Ridge granodorite and adjacent Bridge River schists along the northeastern margin of the pluton. The younger porphyry becomes the dominant intrusive phase to the northwest, and extends from the head of Holbrook Creek 10 kilometres northwestward as a moderately northeast-dipping sheet cutting, from south to north, the southern serpentinite mélange belt, Bridge River schists and part of the main Shulaps serpentinite mélange belt. Separate bodies of porphyry make up the Hog Creek stock to the west (Figure 1-6-3) as well as a series of small plugs extending several kilometres to the east (Leech, 1953). Hornblende-phyric felsite that intrudes both hangingwall and footwall rocks along the Mission Ridge fault south of the Bridge River (Coleman, 1989) may also be correlative.

STRUCTURE

OVERVIEW

The regional structure is dominated by a system of northwest to north-trending faults that reflect a complex history of mid-Cretaceous to Tertiary compressional, strike-slip and extensional deformation (Figure 1-6-1). Our earlier interpretations attributed most of the through-going faults to dextral strike-slip in Late Cretaceous time (Glover *et al.*, 1988a; Schiarizza *et al.*, 1989a). Systems of folds and thrust faults which are preserved locally were attributed mainly to an earlier compressional event of mid-Cretaceous age, reflected in a pronounced angular unconformity in the northwestern part of the project area (Glover and Schiarizza, 1987). Our 1989 mapping suggests, however, that many of the important faults in the region, including the Tyaughton Creek and Castle Pass systems, have a history of sinistral transpressional deformation. This deformation probably occurred in early Late Cretaceous time and produced both sinistral strike-slip faults and compressional structures.

Dextral strike-slip is recorded along the Marshall Creek–Relay Creek and Yalakom fault systems and is, at least in part, Tertiary in age. Southwesterly directed thrust emplacement of the Shulaps ophiolite complex occurred prior to and/or during the Late Cretaceous, and apparently predated most or all of the dextral strike-slip faulting. Northerly trending oblique normal faults within the western Shulaps complex are related to a transfer zone linking the Marshall Creek and Yalakom fault systems along the southeastern margin of an extensional strike-slip duplex. Farther to the southeast, extensional faulting is reflected by the gently northeast-dipping Mission Ridge normal fault (Coleman, 1989) and by later vertical displacement on the Marshall Creek fault. These normal faults truncate Tertiary rocks and structures, including fabrics related to dextral strike-slip faulting.

The mid-Cretaceous to Tertiary structures which dominate the map pattern of the region are superimposed on older structures which are not well understood. Triassic subduction-related deformation and metamorphism is reflected in penetratively deformed blueschist-facies Bridge River rocks which occur locally (Garver *et al.*, 1989a, 1989b; Archibald *et al.*, 1990). Imbrication of the blueschist-facies rocks with greenschist and prehnite-pumpellyite grade rocks occurred sometime after Late Triassic metamorphism and prior to (or during) mid-Cretaceous uplift and erosion when the metamorphic rocks were incorporated as clasts in Albian Taylor Creek conglomerate. The brittle faulting and lenticularity of lithologic units that characterizes the Bridge River complex elsewhere may also be attributed to an earlier deformational history, perhaps in a subduction zone or accretionary prism setting. A Middle Jurassic deformational event has also been postulated (Potter, 1986; Rusmore *et al.*, 1988) and is inferred to mark the juxtaposition of Cadwallader Terrane with the Bridge River and Shulaps complexes. Structures that can be unequivocally assigned a Middle Jurassic age have not been found, however, and it is possible that amalgamation of these tectonostratigraphic elements was a mid-Cretaceous event (Schiarizza *et al.*, 1989; Garver, 1989).

TYAUGHTON CREEK – CASTLE PASS FAULT SYSTEMS

Mapping in the Gold Bridge area was aimed at establishing the southern extensions of the Tyaughton Creek and Castle Pass fault systems. These are important through-going structures that had previously been traced from the Warner Pass map area through the southwestern corner of the Noaxe Creek area and into the Bralorne map area as far south as Gun Creek (Glover *et al.*, 1988a, 1988b; Garver *et al.*, 1989a; Schiarizza *et al.*, 1989a, 1989b). Within this area the two faults enclose a lens of Cadwallader and Tyaughton Group rocks and juxtapose them against younger Jura-Cretaceous strata (Figure 1-6-4). Our 1989 mapping suggests that these fault systems were the locus of early Late Cretaceous sinistral transpressional deformation, and that the uplifted block of older rocks they enclose is separated from adjacent younger rocks by predominantly inward-dipping reverse-sinistral faults.

The structure near Gold Bridge (Figure 1-6-4) is dominated by a complex system of mainly northwest to north-trending faults (Cairnes, 1937, 1943; Church *et al.*, 1988b) informally referred to as the Bralorne fault zone (Rusmore, 1985; Leitch, 1989). Previous detailed work has concentrated along a northwest to north-trending belt of structurally interleaved Bralorne diorite, Cadwallader Group and Bridge River complex that hosts the Bralorne, Pioneer, and numerous smaller gold-quartz vein systems. These studies have established a complex history of faulting that included west-directed thrusting of the Bridge River complex over the Cadwallader Group along the Ferguson thrust fault; oblique thrusting along the north to north-northeast-dipping ribboned gold-quartz veins; and later dextral and vertical offset along steeply dipping north-trending faults (Cairnes, 1937; Joubin, 1948; Leitch, 1989). Observations made along the northern part of this system, near Gold Bridge (Figure 1-6-4), support the proposed west-directed thrusting of Bridge River complex over Cadwallader Group; these include westerly overturned folds in footwall Cadwallader Group, asymmetric west-verging mesoscopic folds in hangingwall Bridge River complex, and shear bands in foliated rocks along the fault itself. Kinematic indicators were also observed along two northwesterly trending faults, one that apparently marks the northern termination of the Cadwallader Group and Bralorne diorite panels, and one that separates the Bralorne diorite from adjacent Bridge River rocks near Gold Bridge. These are northeast-dipping structures along which the hangingwall has moved to the west; this sinistral transpression is the same as the documented movements along the Bralorne and Pioneer vein systems farther south.

The northeast-dipping transpressional fault that crosses the Bridge River at Gold Bridge apparently extends northwest-

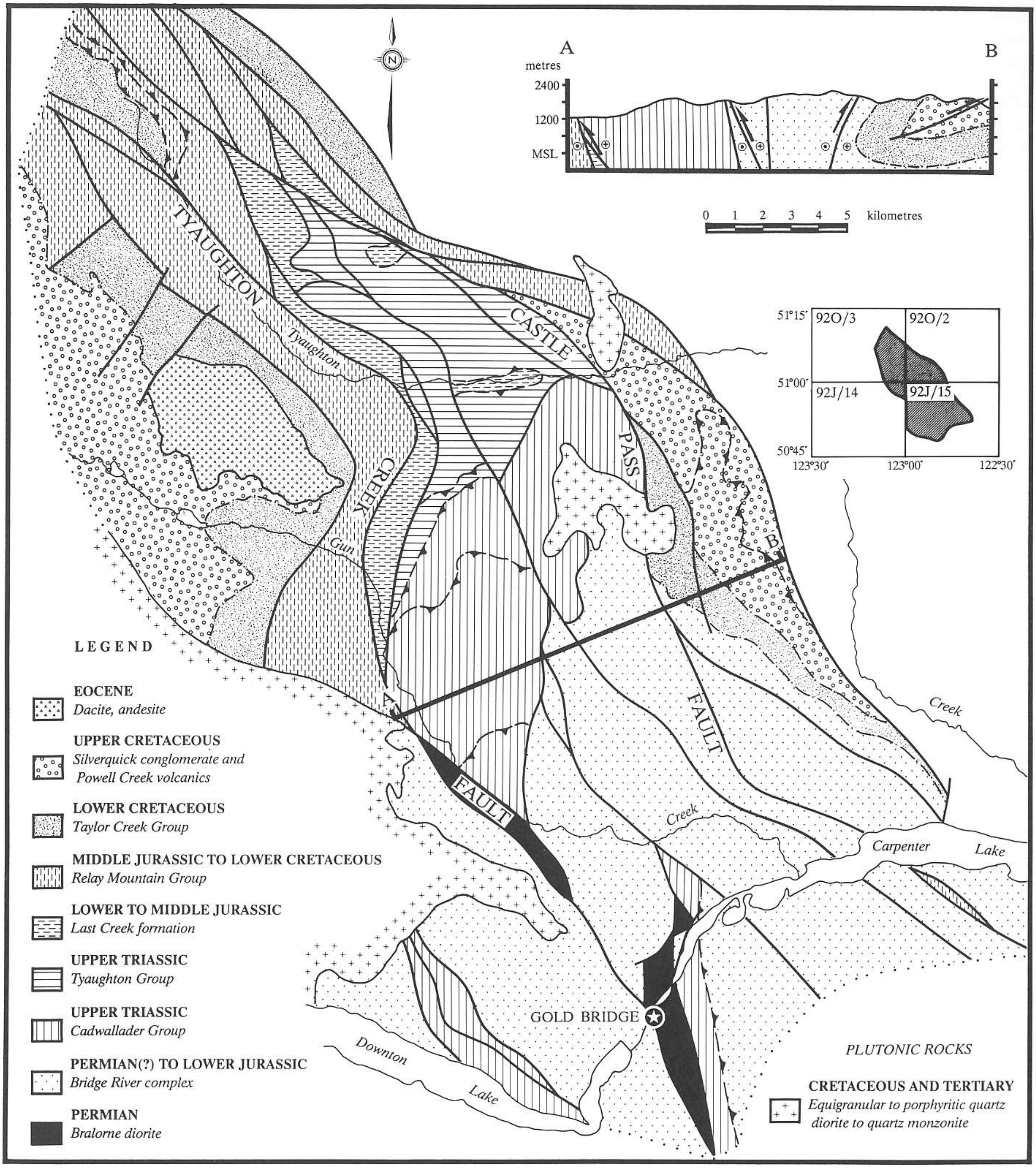

Figure 1-6-4. Generalized geology of the Tyaughton Creek and Castle Pass fault systems.

ward to Gun Creek (Figure 1-6-4). From there it is continuous with the north-trending segment of the Tyaughton Creek fault, although an important splay continues northwestward to separate the Bridge River complex from younger strata to the north. This splay is truncated by the Coast plutonic complex but may have been continuous with the Tchaikazan fault which abuts the granitic rocks along strike to the northwest (Figure 1-6-1).

The southern extension of the Castle Pass fault is inferred to be a southwest-dipping fault identified on the south side of Carpenter Lake 10 kilometres northeast of Gold Bridge (Figure 1-6-4). It separates contrasting Bridge River packages, comprising mainly greenstone and chert on the west, and variably foliated and sheared sandstone, chert, argillite and serpentinite on the east. It encloses a lens of conglomerate and sandstone identified as Relay Mountain Group by Church and MacLean (1987b) on the basis of fossil *Buchia* collected from a thin interbed of laminated siltstone and argillite. Foliation and mesoscopic shear zones within the Bridge River rocks on the northeast side of the fault dip moderately southwest to west. Shear bands and outcrop-scale transpressional duplex structures indicate a top-to-the-east sense of shear. This segment of Castle Pass fault is therefore inferred to have been the locus of oblique sinistral shear; the east-west movement direction is the same as that demonstrated for the Cadwallader–Tyaughton Creek system to the west, but it has the opposite sense of vergence because the fault and associated shears dip in the opposite direction.

The observations summarized here are consistent with the map-scale geometry of the Tyaughton Creek and Castle Pass fault systems from Carpenter Lake to the head of Tyaughton Creek, as they enclose a composite lens of relatively old, presumably uplifted rocks and separate it from younger rocks to both the northeast and southwest (Figure 1-6-4). Furthermore, east-verging folds and thrust faults are documented directly east of the Castle Pass fault while northwest to west-verging thrust faults occur within the Cadwallader and Tyaughton groups along the Tyaughton Creek fault. These are consistent with the outward-directed vergence of structures related to this uplifted block implied by fault dips and kinematics observed in the Gold Bridge area. The uplifted lens enclosed by the Tyaughton Creek and Castle Pass faults resembles the "positive flower" or "palm tree" structures documented along numerous strike-slip fault systems (Sylvester, 1988). The uplift may have been localized by the change from northwest to northerly trends of the faults between Tyaughton and Gun creeks since this corresponds to a restraining bend in a sinistral fault system (Woodcock and Fisher, 1986; Sylvester, 1988).

The Castle Pass fault cuts the Albian Taylor Creek Group and the overlying (Albian or Cenomanian ?) Silverquick conglomerate, and is itself cut by the 64 Ma Eldorado pluton (Garver *et al.*, 1989a). The Tyaughton Creek fault also cuts the Taylor Creek Group, and is apparently continuous with the Bralorne fault zone, within which reverse-sinistral mineralized shear veins probably formed between 86 and 91 Ma (Leitch, 1989). Farther south within the Bralorne fault zone penetrative deformation and metamorphism of the Chism Creek schists occurred between 85 and 100 Ma (Rusmore, 1985). Important strands of the Bralorne–Tyaughton Creek fault system are truncated by the Coast plutonic complex, but may in part continue as the Tchaikazan fault farther to the northwest (Figure 1-6-1); a lower limit of about 87 Ma for major movement along the Tchaikazan fault is provided by radiometric dating of the adjacent Coast plutonic rocks (McMillan, 1976; Archiabald *et al.*, 1989). The timing constraints outlined above indicate that sinistral transpressional deformation occurred in early Late Cretaceous time and produced steep regionally persistent faults as well as folds and thrust faults. This requires a re-evaluation of the complex network of anastomosing northwest-trending faults that pervades the area within and adjacent to the Tyaughton Creek–Castle Pass systems as most of these were previously attributed to a wide band of dextral faulting related to the Yalakom system. Although dextral offsets are locally apparent in this area, they may be of relatively minor importance compared to early Late Cretaceous transpressional structures. The dextral faults are presumably related to younger dextral faulting that was concentrated along the Relay Creek–Marshall Creek and Yalakom systems farther to the northeast.

THRUST FAULTS WITHIN THE SHULAPS ULTRAMAFIC COMPLEX

Detailed mapping along the southwestern margin of the Shulaps ultramafic complex by Tom Calon and coworkers indicates that the Shulaps harzburgite and underlying serpentinite mélange together define a major southwest-verging linked thrust system comprising a number of hinterland-dipping duplexes (Calon *et al.*, 1990). The large-scale two-fold division of the Shulaps complex, comprising mantle harzburgite sitting structurally above serpentinite mélange derived from ultramafic-mafic cumulates, reflects structural stacking of lower over higher stratigraphic elements of an original ophiolite suite. A similar stacking order occurs within the serpentinite mélange itself, as upper structural levels contain knockers of ultramafic-mafic cumulates whereas a large block of gabbro and pillowed volcanics linked by an intervening dike swarm occurs along the structural base of the mélange. Serpentinite forming the matrix of the mélange commonly displays a penetrative, steeply northeast-dipping S_1 foliation cut by discrete, more gently northeast-dipping S_2 slip surfaces spaced several millimetres to several centimetres apart; sigmoidal deflection of S_1 at S_2 boundaries typically suggest a top-to-the-southwest sense of shear. Mylonite, possibly synchronous with the S_1 serpentinite foliation, commonly occurs within gabbro and serpentinite along the margins of large knockers. These mylonites display a variety of kinematic indicators, including S-C foliations, shear bands and rotated mineral grains, that indicate a top-to-the-southwest sense of shear. Serpentinite mélange along the northeastern margin of the Shulaps complex is not as well exposed as the belt to the southwest, but is inferred to comprise part of the same imbricate zone and to be continuous with the southwestern mélange beneath the intervening mantle harzburgite. S-C foliations within moderately northwest-dipping serpentinite mylonite along the contact between serpentinite mélange and overlying mantle harzburgite near upper Peridotite Creek support this interpretation as they also indicate a top-to-the-west sense of shear.

Deformation within the Shulaps serpentinite mélange was in part synchronous with intrusion of a suite of dioritic hornblende porphyry dikes. These dikes caused prograde metamorphism of serpentinite to talc-serpentine-magnesite schist, locally with regenerated olivine porphyroblasts. The dikes are locally boudinaged within the serpentinite mélange matrix and locally occur in knockers where they are truncated at the contact with the enclosing serpentinite. They therefore predate some movement within the mélange but because they caused prograde metamorphism of previously serpentinized ultramafic rock, and at one locality cut the foliation in a penetratively deformed metasedimentary knocker, are interpreted to have been intruded during the late stages of deformation (Archibald *et al.*, 1989; Calon *et al.*, 1990). These dikes in part resemble the Blue Creek porphyries; one which occurs as aligned pods within serpentinite mélange along the northeast margin of the Shulaps complex has yielded a ^{40}Ar-^{39}Ar plateau age of 77 Ma (Archibald *et al.*, 1990). Heating of the mélange at this time is also suggested by a 73 Ma ^{40}Ar-^{39}Ar date obtained from an amphibolite knocker within the southwestern mélange belt which may date cooling following intrusion-related heating (Archibald *et al.*, 1989).

The Shulaps serpentinite mélange, comprising the structural base of the imbricated Shulaps ophiolite complex, sits structurally above Bridge River schists in the area of Jim and Hog creeks. The contact was not observed, but is apparently concordant to the moderately to steeply dipping foliation within both the mélange and underlying schists, and is inferred to be a thrust contact, as suggested by Potter (1983). To the east, along East Liza Creek, the serpentinite mélange was apparently thrust over the Cadwallader Group. In this area, a greenstone-gabbro complex at the base of the mélange sits structurally above the Hurley Formation across a narrow mylonitic zone which is deformed by east-verging folds and associated slaty cleavage (Calon *et al.*, 1990). A thrust contact also defines the northern margin of the Shulaps complex (Figure 1-6-3) where southerly dipping serpentinite mélange at the base of the complex overlies subgreenschist grade Bridge River rocks. The contact was not observed, but southerly dipping striated shear surfaces locally bounding a sigmoidal flattening(?) foliation within Bridge River rocks near the contact suggest southerly directed thrusting of the Shulaps complex over the Bridge River complex.

INTERNAL STRUCTURE OF THE BRIDGE RIVER SCHISTS

The Bridge River schists comprise several kilometres of structural thickness of foliated and lineated schist and phyllite intruded by abundant syntectonic to post-tectonic granitic to felsic porphyry intrusions. The schists are truncated by the Mission Ridge fault on the northeast and by the Marshall Creek fault on the southwest. They are structurally overlain by the Shulaps complex to the north, and several kilometres south of the Shulaps enclose an imbricate lens of Shulaps serpentinite mélange and penetratively deformed metasedimentary rocks of the Hurley Formation.

Foliation in the northern part of the Bridge River schist belt dips to the north, beneath the structurally overlying Shulaps complex. It is axial planar to gently east or west-plunging mesoscopic folds and is locally crenulated and folded about later open to tight folds that are approximately coaxial with the earlier ones. The foliation within the schists is concordant to foliation and thrust contacts within the overlying Shulaps serpentinite mélange as well as to those within the southern imbricate zone of serpentinite mélange and Hurley metasediments. Where observed southwest of Rex Peak, the contact between Bridge River schists and underlying mélange is marked by a mylonitic fabric which appears to grade upward into the schistosity in the overlying schists, and is folded about later upright, gently east-plunging, south-verging asymmetric folds (Schiarizza *et al.*, 1989). These relationships suggest that foliation and later deformation fabrics displayed by the Bridge River schists in the northern part of the belt relate primarily to a complex history of thrust imbrication and folding during south to southwest-directed emplacement of the Shulaps complex. The timing of this event is not well constrained but, as discussed in the previous section, the latest stages of deformation may have occurred in the Late Cretaceous. It predated intrusion of the Eocene Mission Ridge pluton, which crosscuts the southern imbricate zone (Figure 1-6-3).

Farther south, foliation within the Bridge River schists has a consistent northwest strike with shallow to moderate northeast dips. Stretching and intersection linearions are subhorizontal and consistently trend northwest, as do the fold hinges of both early and late folds. Kinematic indicators, including shearbands, S-C planes and rotated porphyroblasts, give a consistent upper-member-to-the-southeast sense of shear (Coleman, 1989). Two late-syntectonic granitic dikes which crosscut foliation in the enclosing schists but have the same S-C mylonitic fabrics have yielded 47 ± 1 Ma U-Pb zircon ages (Coleman, M.Sc. thesis in progress). This suggests that the top-to-the-southeast kinematic indicators formed in the Eocene, probably during dextral strike-slip faulting, and are relatively late products of a protracted history of ductile deformation within the Bridge River schists.

YALAKOM FAULT

The Yalakom fault is the most prominent structural feature of the region as it separates areas of sharply contrasting stratigraphy and structural style (Figure 1-6-1). A major steeply dipping fault zone bounding the Shulaps complex along the Yalakom River was first described and named the Yalakom fault by Leech (1953). It was traced northwestward through the Taseko Lakes and Mount Waddington map areas by Tipper (1969, 1978) who postulated that it was the locus of major right-lateral displacement. It extends southeastward to Lillooet, where it is truncated by the more northerly trending Fraser fault system (Monger and McMillan, 1984), along which it is separated by about 90 kilometres from its probable offset equivalent, the Hozameen fault, to the south (Monger, 1985). Dextral offset of more than 100 kilometres has been postulated along the Yalakom fault based on a number of different piercing point correlations. These include: 130 to 190 kilometres offset of Middle Jurassic volcanic rocks that outcrop within the Mount Waddington, Anahim Lake and Bella Coola map areas (Tipper, 1969); 125 to 175 kilometres offset of similar submarine-fan facies within the Albian Jackass Mountain Group between the Camelsfoot Range and

Chilco Lake (Kleinspehn, 1985); and about 100 kilometres separation between the Shulaps ultramafic complex and the Petch Creek serpentinite along the Hozameen fault, after first accounting for 85 kilometres of dextral offset along the Fraser fault (Umhoefer, 1989). While none of these reconstructions is unequivocal, they are consistent with the dextral sense inferred within the Taseko–Bridge River project area, and with the large offset implied by the juxtaposition of contrasting Albian lithologies of the Jackass Mountain and Taylor Creek groups.

During the present study the Yalakom fault has been traced southeastward across the Noaxe Creek map area (Glover *et al.*, 1988a, 1988b), through the southwest corner of the Big Bar Creek sheet to the southeast corner of the Bridge River map sheet (this report). Northwest of Blue Creek the fault is a steeply dipping structure commonly marked by a zone of serpentinized to listwanite-altered ultramafic rocks up to several hundred metres wide. It juxtaposes broadly folded and faulted Middle Jurassic and Jackass Mountain Group rocks on the northeast against more complexly deformed Powell Creek, Taylor Creek, Bridge River and Shulaps rocks to the southwest (Figure 1-6-1). Dextral strike-slip movement is suggested by fibrous minerals and slickensides along fault surfaces within and near the fault zone and by east-trending folds within the Middle Jurassic sandstone–Jackass Mountain Group succession that are truncated by the fault (Glover *et al.*, 1988a). Shear bands cutting foliated serpentinite along the fault zone near the mouth of Blue Creek also indicate dextral movement, as does the orientation of extensional veins and slickensided surfaces in listwanite-altered ultramafite along the fault in the same area.

Southeast of Blue Creek, the Yalakom fault apparently splays into two sub-parallel strands that enclose a wedge of imbricated Hurley Formation and Bridge River complex. The northeastern strand separates the Middle Jurassic sandstone–Jackass Mountain Group belt on the northeast from the imbricated Hurley–Bridge River wedge and is here considered the extension of the Yalakom fault. The southeastern strand separates the latter package from the Shulaps and Bridge River complexes to the southeast and is referred to as the Bridge River fault (Figure 1-6-3).

The northeastern strand is most readily related to the Yalakom fault since it bounds the same package of structurally simple Middle Jurassic and Jackass Mountain Group rocks that occur along the fault to the northwest, and juxtaposes it against Bridge River and Hurley rocks which elsewhere in the region occur only on the southwest side of the Yalakom fault. This fault is not exposed and was not recognized by previous workers; its trace crosses the wooded ridges northeast of the Yalakom and Bridge River valleys. Its presence is corroborated by contrasting structural styles since the Middle Jurassic sandstone and Jackass Mountain Group comprise a structurally simple, steeply dipping east-northeast facing belt, while the Hurley–Bridge River belt is characterized by west to southwest-verging overturned folds and imbrication across northeast-dipping faults. Nevertheless, the fault is in places poorly constrained because of poor exposure and the difficulty in differentiating between the Middle Jurassic rocks and Hurley Formation where only the finer grained facies are represented. The fault was apparently the locus of igneous intrusion as quartz feldspar porphyry, hornblende feldspar porphyry, granodiorite and diorite were noted at several localities along or near its inferred trace.

BRIDGE RIVER FAULT

The Bridge River fault apparently diverges from the Yalakom fault between the mouths of Blue and Beaverdam creeks, from where it extends southeastward along the slopes southwest of the Yalakom River to follow the lower coarse of the river and the adjoining Bridge River to the vicinity of Camoo Creek. It bounds the imbricated Hurley–Bridge River wedge to the northeast and separates it from rocks of the Shulaps and Bridge River complexes to the southwest. The fault zone is locally marked by conspicuous exposures of serpentinite mélange along the Bridge River; this mélange includes knockers of peridotite, gabbro and diabase similar to those seen within the mélange zone at the structural base of the Shulaps complex. This suggests that the Shulaps may have been translated in a dextral sense along the fault zone, although the dip, timing and nature of movement have not been established. The relationship of this fault (typically identified as the Yalakom fault by earlier workers) to the major fault bounding the Middle Jurassic sandstone–Jackass Mountain Group succession farther northeast (and here considered the main strand of the Yalakom fault) is also uncertain, but will be discussed further in the section dealing with the Mission Ridge fault.

NORTHEAST-DIPPING FAULTS ALONG THE YALAKOM AND BRIDGE RIVERS

The 4-kilometre-wide lens enclosed by the Yalakom and Bridge River faults southeast of Beaverdam Creek includes Hurley Formation, Bridge River complex and Relay Mountain Group rocks deformed by southwesterly overturned folds and northeast-dipping faults. The contacts that bound the major lithologic packages were identified as moderately northeast-dipping faults by Coleman (1989) who suggested that they might be oblique thrusts. The orientation of the faults was confirmed by our 1989 fieldwork, and sparse but consistent kinematic evidence suggests that they record sinistral transpressional deformation. The kinematic indicators include: sinistral shear bands cutting foliated serpentinite along a northeast-dipping Hurley–Bridge River fault contact west of Applespring Creek; outcrop-scale fault systems with oblique east to east-northeast plunging striations preserved on northeast-dipping faults and top-to-the-west sense of movement indicated by offset marker beds; and west to southwest-verging folds with axes locally trending more northerly than the strike of the northeast-dipping faults.

The relationship of the northeast-dipping transpressional faults within this wedge to adjacent structures is not certain. Coleman (1989) correlated the easternmost of these faults, which separates pillowed greenstone, gabbro and diabase of the Bridge River complex from overlying Hurley Formation (her Lillooet Group) with the Yalakom fault. Our 1989 fieldwork has established, however, that the Yalakom fault is farther to the northeast, where it separates the Hurley Formation from a structurally simpler Middle Jurassic sandstone–Jackass Mountain Group panel. Although neither the dip nor kinematics along this section of the Yalakom fault have been established it is readily correlated with the Yalakom fault farther northwest, which is a steeply dipping dextral strike-slip fault. Since the panel of imbricated Hurley and Bridge River rocks containing these northeast-dipping transpressional faults apparently pinches out between the Yalakom and Bridge River fanlts near Beaverdam Creek, they may be pre-Yalakom structures isolated within this fault-bound wedge. It is noteworthy that the sinistral transpressional movement along these faults is similar to that documented along early Late Cretaceous structures in the Gold Bridge area, and also that the imbricated Hurley–Bridge River–greenstone-diabase-gabbro panels within the wedge bear a strong resemblance to imbricate slices associated with the lower serpentinite mélange south of the Shulaps complex and its possible offset equivalent west of the Marshall Creek fault.

MISSION RIDGE FAULT

The Mission Ridge fault extends from 4 kilometres southeast of Lillooet, where it is truncated by the Fraser fault, northwest at least 40 kilometres to Shulaps Creek. It was first recognized and named by Coleman (1989) who established the geometry, kinematics and relative timing of movement on the segment of the fault between Lillooet and the Bridge River canyon.

The Mission Ridge fault strikes northwest and has a dip of 25° to 30° northeast. South of the Bridge River it juxtaposes low-grade, non-penetratively deformed Bridge River complex and Eocene nonmarine sedimentary rocks in the hangingwall against lower to upper greenschist-grade Bridge River schist and phyllite. Assuming a normal geothermal gradient, an estimated 12 kilometres of down-dip displacement is required to account for the contrasting metamorphic grade of hangingwall and footwall rocks (Coleman, 1989). The fault trace crosses the Bridge River about 1 kilometre west of the Yalakom River confluence and continues northwestward, parallel to the Yalakom, to at least Shulaps Creek. Between the Bridge River and LaRochelle Creek the fault trace follows the base of a distinctive planar slope. The slope has the same orientation as the Mission Ridge fault, consists of footwall schists and phyllites, and is interpreted to be the exhumed fault surface. The contrast in metamorphic grade across the fault diminishes to the northwest and is evidence for a decrease in the amount of normal displacement at this end of the fault. Its continuation beyond Shulaps Creek is uncertain, but it may swing westward and mark the boundary between Bridge River schists and Shulaps harzburgite, perhaps accounting for the pinching out of the intervening serpentinite mélange (Figure 1-6-3). Alternatively, or in addition, it may splay into a series of imbricate faults extending north and northwestward from Shulaps Creek as outcrop-scale, predominantly east-side-down, low and high-angle normal faults were observed at several localities in this area.

The Mission Ridge fault truncates Eocene(?) nonmarine sedimentary rocks in its hangingwall and the 47.5 Ma Mission Ridge pluton in its footwall (Coleman, 1989). Where the fault zone is exposed on the southeast side of the Bridge River canyon, an anastomosing fracture cleavage parallel to the fault is superimposed on foliation of the Bridge River schists in a zone 5 metres wide. This confirms that brittle normal movement on the Mission Ridge fault postdates penetrative strain in the Bridge River schists, part of which may be related to dextral strike-slip faulting. The Mission Ridge fault is in turn cut by the Fraser fault system, and was therefore probably active in the Middle to Late Eocene.

The major displacement suggested by the contrast in metamorphic grade across the low-angle Mission Ridge fault indicates that lithologic units and structures in its hangingwall have been displaced a considerable distance northeastward relative to adjacent footwall rocks. If pre-normal-faulting dextral shear recorded in footwall fabrics is related to movement on the Yalakom fault, then the Yalakom is an earlier structure that may have been cut by the Mission Ridge fault. This suggests that the southeastern segment of the Yalakom fault identified to the northeast of the Yalakom and Bridge rivers represents only a relatively high-level expression of an original Yalakom fault that was detached from its roots and translated eastward along the Mission Ridge fault. Perhaps continued or later strike-slip(?) movement along the root fault broke through the overlying cover (hangingwall of the Mission Ridge fault) at the position of the Bridge River fault and generated the present fault configuration.

MARSHALL CREEK FAULT

The Marshall Creek fault is a prominent northwest-trending structure that separates Bridge River schists on the northeast from lower grade Bridge River rocks on the southwest (Potter, 1983, 1986). It is a regionally persistent fault that extends from the Fraser fault system, 35 kilometres south of Lillooet (Monger and McMillan, 1984), for 90 kilometres northwestward to Marshall Lake. From there it extends an additional 45 kilometres northwest as the Relay Creek fault system (Glover *et al.*, 1988a, 1988b) before apparently merging with the Yalakom fault near Big Creek (Tipper, 1978).

Within the study area the Marshall Creek fault zone comprises two steeply dipping strands. The northeastern strand separates penetratively deformed greenschist-facies Bridge Riger complex and locally imbricated Cadwallader Group and serpentinite mélange on the northeast from prehnite-pumpellyite-grade Bridge River rocks on the southwest. A parallel strand to the southwest juxtaposes the low-grade Bridge River rocks against a similar Bridge River package and unconformably overlying Eocene(?) volcanics, indicating a component of Eocene or later southwest-side-down displacement. The two strands apparently merge to the southeast where, south of Carpenter Lake, the fault also cuts the Eocene Mission Ridge pluton which intrudes Bridge River schists on its northeast side. Farther to the southeast, beyond the limits of the present map area, the Marshall Creek fault truncates a low-angle fault on its southwest side which juxtaposes footwall Bridge River schist against hangingwall low-grade Bridge River rocks (Coleman, 1989). The low-angle fault is interpreted by Coleman to be part of the Mission Ridge fault; its restoration gives approximately 3.5 kilometres of southwest-side-down vertical displacement on the Marshall Creek fault (*see* Coleman, 1989, Figure 1-12-3).

The Marshall Creek fault is also inferred to have been the locus of significant dextral strike-slip movement. This inference is based partly on the prominent system of northerly trending faults that forms a transfer zone linking the Marshall Creek with the Yalakom strike-slip fault west of the

Shulaps complex (Figure 1-6-1). The regional fault pattern and distribution of lithologic units suggest that these northerly trending faults form the southeastern margin of an extensional duplex within a dextral strike-slip system (Schiarizza *et al.*, 1989a). Detailed mapping along the southwestern margin of the Shulaps complex supports this inference since northerly trending faults in this area are typically transtensional (Calon *et al.*, 1990). Dextral offset is further supported by tentative correlation of the thrust-imbricated package of Hurley Formation, serpentinite mélange and Bridge River schists that is truncated by the Marshall Creek fault 10 kilometres southeast of Marshall Lake with a similar (but lower metamorphic grade) package truncated on the other side of the fault 15 kilometres to the northwest at Liza Lake (Schiarizza *et al.*, 1989b). If strike-slip was synchronous with dextral shear along horizontal stretching lineations within the Bridge River schists, it predates the southwest-side-down normal faulting documented along the southeastern segment of the Marshall Creek fault

MINERAL OCCURRENCES

Metallic mineral prospects within the Yalakom River area occur mainly between the Marshall Creek and Yalakom–Bridge River faults. These include mesothermal gold-quartz veins within stocks of Blue Creek porphyry, as well as veins, disseminations and stockwork containing molybdenum, copper and gold along the northeastern margin of the Mission Ridge pluton. In addition, ultramafic rocks of the Shulaps complex contain small chromite concentrations and have been prospected for nephrite jade, magnesite and chrysotile (Figure 1-6-5). Cinnabar occurs locally as disseminations and veinlets near the Bridge River fault.

Gold-bearing quartz veins at the Yalakom and Elizabeth prospects occur within stocks of porphyritic quartz diorite (Blue Creek porphyry) that cut Shulaps harzburgite north of Blue Creek (Figure 1-6-5). The veins are typically ribboned and occupy steeply dipping, northerly trending shears (Gaba *et al.*, 1988). They contain visible gold and rarely more than a few per cent sulphide minerals, mainly arsenopyrite, pyrite and chalcopyrite. These veins are similar to the mesothermal veins at the Bralorne and Pioneer mines, which have yielded most of the gold produced from the Bridge River mining camp.

Most other metallic mineral occurrences are associated with granodiorite of the Mission Ridge pluton. These include high-sulphide auriferous veins at the Spokane and Broken Hill prospects as well as stockwork molybdenite at the Cub showing, which was discovered during the course of our mapping. These showings are described in a separate report by Gaba (1990, this volume).

Cinnabar veinlets and disseminations at the Eagle and Red Eagle prospects are within greenstone and greenstone breccia of the Bridge River complex. The cinnabar is associated with widely spaced carbonate veins that occupy shears parallel to the adjacent Bridge River fault. Similar mercury mineralization occurs along the Yalakom fault 30 kilometres to the northwest (Glover *et al.*, 1988a, 1988b), and along the Relay Creek fault system north of Tyaughton Lake (Schiarizza *et al.*, 1988a. 1988b).

SUMMARY

The Taseko–Bridge River project area was the locus of a complex history of mid-Cretaceous to mid-Tertiary deformation and intrusion that was superimposed on an earlier deformational history that in part included subduction-related deformation and metamorphism of the Bridge River complex. The main conclusions drawn from our 1989 fieldwork are summarized as follows:

- The northwest to north-trending Tyaughton Creek and Castle Pass fault systems were the locus of sinistral transpressional deformation. They were previously attributed to Late Cretaceous dextral strike-slip related to the Yalakom system, and inferred to be distinctly later than mid-Cretaceous thrust faults they locally truncate. Our revised interpretation suggests that steep faults and compressional structures formed together during early Late Cretaceous sinistral transpression. This provides a better explanation for several map-scale features within the Taseko–Bridge River project area including: the localization of compressional structures along the north-trending fault segments associated with the Tyaughton Creek–Castle Pass systems in the Spruce Lake–Eldorado Mountain area, since these reflect restraining bends in a sinistral fault system; and the abrupt change in structural style west of Big Creek where relatively undeformed Upper Cretaceous Powell Creek volcanics apparently rest unconformably above both low and high-angle faults (Glover *et al.*, 1987) that may be related to transpressional deformation. The Tyaughton Creek fault is apparently continuous with the Bralorne fault zone farther southeast, which was the locus of mesothermal gold-quartz veining during this deformation.

Imbrication and thrust emplacement of the Shulaps ophiolite complex over the Bridge River complex and Cadwallader Group occurred along southwesterly directed thrust faults. ^{40}Ar-^{39}Ar cooling dates from dikes and knockers within the Shulaps serpentinite mélange suggest a Late Cretaceous age for the latest pulse of deformation (Archibald *et al.*, 1990). Thrust faults that may be related to the Shulaps imbricate zone are also identified west of the Marshall Creek fault where they separate slices of serpentinite mélange, Cadwallader Group and Bridge River complex (Schiarizza *et al.*, 1989a, 1989b). West to southwest-verging overturned folds and transpressional faults also occur east of the Shulaps complex where they imbricate lenses of Cadwallader Group, Bridge River complex, serpentinite-diabase-greenstone and Relay Mountain Group between the Bridge River and Yalakom faults.

The structures described in the previous two paragraphs may have all formed during a protracted period of late Early Cretaceous to early Late Cretaceous compressive to transpressive deformation. The earliest manifestation of this event is the implied deformation and uplift related to deposition of the synorogenic Taylor Creek Group (Garver, 1989). Although middle to early Late Cretaceous contractional deformation is recognized throughout the region (*e.g.* Rusmore and Woodsworth, 1989; Journeay and Csontos, 1989), the

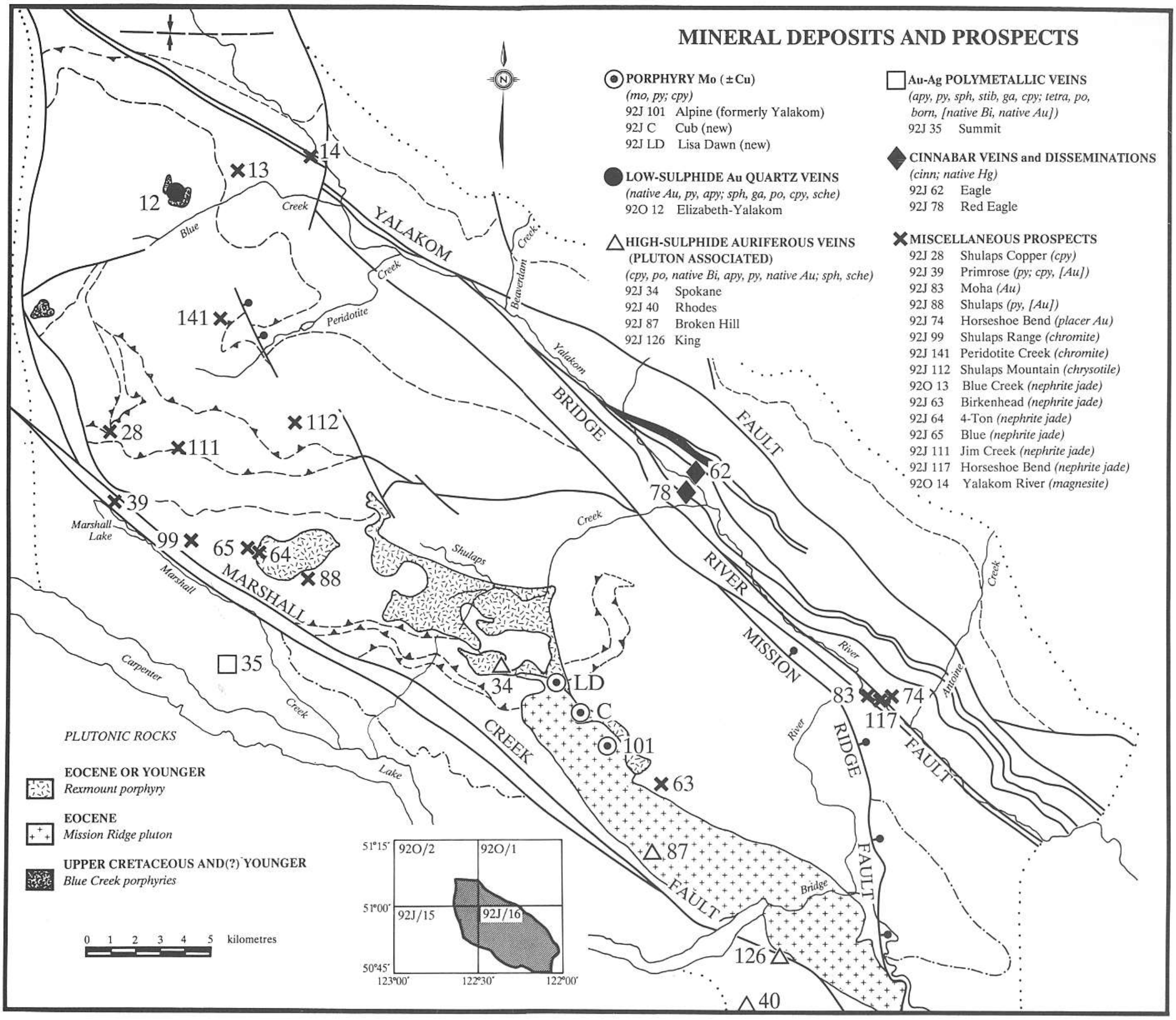

Figure 1-6-5. Mineral occurrences, Yalakom River map area. MINFILE number precedes deposit or prospect name. Mineral abbreviations: apy = arsenopyrite, born = bornite, cinn = cinnabar, cpy = chalcopyrite, ga = galena, mo = molybdenite, py = pyrite, po = pyrrhotite, sche = scheelite, sph = sphalerite, stib = stibnite, tetra = tetrahedrite.

regional extent or significance of the sinistral component is at present uncertain. It is of interest, however, that Greig (1989) has recently documented mid-Cretaceous sinistral transpression along the Pasayten fault to the southeast.

- Dextral strike-slip faulting along the Yalakom and Relay Creek–Marshall Creek fault systems postdates the compressional to transpressional deformation described above. Dextral movement along this system is suggested by east-trending folds within the Middle Jurassic sandstone and Jackass Mountain Group northeast of the Yalakom fault (Glover *et al.*, 1988a), and by a transfer zone of northerly-trending faults that links the Relay Creek–Marshall Creek and Yalakom fault systems northwest of the Shulaps ultramafic complex to define an extensional dextral-strike-slip duplex (Schiarizza *et al.*, 1989a). Shear bands in foliated serpentinite along the Yalakom fault near Blue Creek corroborate the dextral shear implied by these map-scale features. Top-to-the-southeast kinematic indicators associated with subhorizontal stretching lineations within northeast-dipping Bridge River schists also indicate dextral shear that may be related to movement along the bounding Yalakom and/or Marshall Creek faults. The same kinematic indicators are found in late synkinematic Eocene dikes, suggesting that dextral faulting was, at least in part, Eocene in age. The upper limit of dextral movement is not well constrained, however, and it is possible that the early stages of strike-slip faulting along the Yalakom and/or Marshall Creek systems coincided with the final pulse of (Late Cretaceous) deformation documented within the Shulaps serpentinite mélange.

- Miller (1987) conducted a detailed structural analysis of Lillooet and Jackass Mountain Group rocks exposed between the Yalakom and Fraser faults directly north of Lillooet. He concluded that structures within these rocks suggested a history of left-lateral followed by high-angle reverse movement along the Yalakom fault. The panel of oblique-sinistral faults and related folds documented here between the Yalakom and Bridge River faults is more or less along strike from Miller's study area, but contains distinctly different stratigraphic elements. Furthermore, these structures are inferred to be mid-Cretaceous in age and unrelated to the Yalakom fault which is here interpreted as a younger dextral strike-slip fault. Their relationship to the structures studied by Miller is therefore not readily apparent. However, Greig (1989) has also documented mid-Cretaceous sinistral transpression along the Pasayten fault, which marks the northeastern boundary of a panel of Jackass Mountain Group and related rocks that is probably the offset equivalent of the one studied by Miller. It is therefore suggested that the structures described by Miller may reflect mid-Cretaceous sinistral transpression, and be unrelated to (younger) movement along the Yalakom fault.
- The Bridge River schists and phyllites between the Marshall Creek and Mission Ridge faults were penetratively deformed under lower to upper greenschist-facies metamorphic conditions, in contrast to the predominantly prehnite-pumpellyite-facies that characterizes rocks to the northeast and southwest. Elevated metamorphic conditions along the north end of the belt apparently prevailed during imbrication and southwesterly directed thrusting of the Shulaps ophiolite complex over the Bridge river complex and Cadwallader Group. Metamorphism was in part synchronous with intrusion of a suite of late-kinematic hornblende feldspar porphyry dikes of Late Cretaceous age, but earlier deformation was also in part ductile and generated mylonites and greenschist-facies metamorphic mineral assemblages prior to intrusion of the dikes (Archibald *et al.*, 1989, 1990; Calon *et al.*, 1990). Potter (1983, 1986) suggested that the heat source for this metamorphism was hot upper mantle of the obducted Shulaps complex. The present study has not documented an inverted metamorphic gradient beneath the Shulaps complex where locally, as along the northern margin of the complex, Shulaps rocks are in thrust contact with subgreenschist facies rocks. However, a thick slice of serpentinite mélange occurs along the contact in areas where a thrust relationship between the Shulaps and underlying Bridge River complex is inferred or documented. Structures within and adjacent to the mélange record a complex deformational history that includes mixing of ophiolitie and lower to upper greenschist-facies supracrustal elements within the mélange as well as later imbrication and/or infolding of the mélange and underlying Bridge River and Cadwallader rocks. This deformation and attendant late metamorphism have clearly shuffled and overprinted the early metamorphic pattern such that its spatial relationship to the Shulaps ophiolite is unclear.

  Bridge River schists throughout most of the high grade belt, which extends from the Shulaps complex more than 40 kilometres southwestward to the Fraser River (Monger and McMillan, 1984; Coleman, 1989) are intruded by abundant synkinematic to postkinematic Middle Eocene granitic rocks, and record Middle Eocene ductile deformation (Price *et al.*, 1985; Potter, 1986; Coleman, 1989). Most of the belt was therefore at elevated temperatures during the Middle Eocene, perhaps directly or indirectly related to the granitie intrusions localized along the belt at that time. Later uplift of the belt was in part accommodated by extensional faulting along the gently northeast-dipping Mission Ridge fault (Coleman, 1989). Displacement along this fault seems to diminish to the north, suggesting that uplift may have been greatest in the south and imparted a northward tilt to the block. This is consistent with preservation of the structurally higher Shulaps ophiolite complex at the north end, and of structures related to its emplacement within Bridge River and Cadwallader Group rocks directly beneath it.
- Metallic mineral occurrences within the Yalakom River area are concentrated within the belt of relatively higher grade metamorphic rocks between the Marshall Creek and Yalakom faults. These include the Elizabeth-Yalakom mesothermal gold-quartz veins within stocks of Blue Creek porphyry, as well as base and precious metal bearing veins and disseminations within and adjacent to the Mission Ridge pluton. The metal prospects are therefore broadly related to the intrusive activity that characterized the belt in Late Cretaceous to Eocene time. Mineralization associated with the Mission Ridge pluton includes stockwork molybdenite that was discovered during the course of this summer's work; this type of mineralization was previously unrecognized and represents a new exploration target in the Bridge River mining camp (Gaba, 1990).
- The stratigraphic succession northeast of the Yalakom fault is represented by Middle Jurassic volcanic sandstone, granule to pebble conglomerate and less common siltstone and shale, together with disconformably overlying upper Lower Cretaceous arkosic sandstone, volcanic and plutonic-clast conglomerate, siltstone and shale of the Jackass Mountain Group. These units are lithologically distinct from age-equivalent strata on the southwest side of the Yalakom fault, where the Middle Jurassic is represented by mainly shales and siltstones of the Last Creek formation (Frebold *et al.*, 1969; Umhoefer, 1989) and the upper Lower Cretaceous is represented by mainly volcanic and chert-rich clastics of the Taylor Creek Group (Garver, 1989). The stratigraphic succession on the northeast side of the Yalakom fault compares more closely to the stratigraphy 100 to 200 kilometres to the south, where the Jackass Mountain Group is in part directly underlain by the Lower to Middle Jurassic Ladner Group, including lower Middle Jurassic volcanic sandstones and conglomerates, volcanic tuff, breccia and local flows of the Dewdney

Creek Formation (O'Brien, 1986). These rocks occur on the northeast side of the Hozameen fault, the probable southern extension of the Yalakom fault.

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GEOLOGY OF THE SWAKUM MOUNTAIN AREA, SOUTHERN INTERMONTANE BELT (92I/7)

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KEYWORDS: Lithoprobe transect, Intermontane Belt, Nicola Group, Nicola horst, Ashcroft Formation, volcanic stratigraphy, structure, alteration, mineral deposits.

INTRODUCTION

Fieldwork begun in 1988 along the LITHOPROBE Southern Cordillera Transect in the Intermontane Belt (Moore, 1989) was continued in 1989, with the dual objectives of resolving the stratigraphy and structure of the Nicola Group and gaining a better understanding of the architecture of the Nicola horst. West of the horst (Figure 1-7-1) the Nicola Group has been mapped at 1:25 000 scale as far north as 50°15′ (Preto, 1979; McMillan, 1981) but farther north there was only reconnaissance coverage (Cockfield, 1948; Monger and McMillan, 1984). Because of the large number of former small mines and mineral occurrences on Swakum Mountain (Figure 1-7-1), this area was selected for more detailed study. Excellent access is afforded from Merritt by the Coquihalla Highway and Mamit Lake road and a number of seasonal forest-access roads. Three weeks were devoted to mapping, using 1:15 000 aerial photographs and compiling at the same scale on a base enlarged from the 1:50 000 topograph map. The geology of the area has proven much more complex and diverse than expected, and the new data obtained this season have implications for regional structure and metallogeny. In this paper we present a preliminary account of the stratigraphy and structure; a more comprehensive treatment of the mineral occurrences and potential will be presented by the authors with R.E. Meyers in *Exploration 1989*.

REGIONAL SETTING

The study area lies within the Western Belt of the Nicola Group, comprising primarily calcalkaline arc volcanic rocks (Preto, 1979). It is bordered on the west by the Late Triassic to Early Jurassic Guichon Creek batholith (McMillan, 1978) and on the east by the Nicola horst. The Nicola horst (described in Moore, 1989 where it is referred to as the "Central Nicola horst") comprises Nicola Group rocks, sedimentary rocks of unknown age, tonalite and tonalite porphyry, all strongly deformed, metamorphosed to low amphibolite facies and intruded by granitoid rocks ranging in age from at least Early Jurassic to Paleocene. It is separated by normal faults from surrounding Nicola Group rocks that are of subgreenschist and greenschist grade and lack penetrative deformation. The Swakum Mountain rocks exhibit continuity with Nicola Group units mapped to the south (McMillan, 1981) but may be separated by a northwest-trending fault from those to the north on Mount Guichon (Figure 1-7-1).

LITHOLOGY

A generalized geological map of the area is presented in Figure 1-7-2; the legend is given in Figure 1-7-2a. To date there are no paleontological age determinations in the study area, so all age assignments are tentative, but the relative ages of the major units are evident from field relationships.

Rocks of the Nicola horst are not subdivided on the map, but will be treated in a later publication. As noted above, they are in part age-equivalent to units on Swakum Mountain.

The Nicola Group is divided into five units based on predominant lithology, without implication of relative age. There is very little continuity of any unit in the area and, given the limitations of exposure and traverse density, the nature of most of their contacts remains uncertain. Lava flows (NF) are most abundant in the western half of the area; they are predominantly plagioclase-phyric andesites. Phenocrysts reach 2 centimetres or more in length and constitute up to 30 per cent of flow rocks. Fresh augite phenocrysts are present in places, particularly around Revelle Lake and Saxon Lake, but are generally much subordinate or absent; a few samples contain hornblende phenocrysts. Most flows contain less than 5 per cent amygdules; where present these are filled with quartz, chlorite and/or calcite. Flow contacts are generally not visible; a few flows interbedded with breccia are 2 to 10 metres thick. Flows are in part intercalated with monolithologic flow or pyroclastic breccias, from which they are difficult to distinguish in the field.

Fragmental volcanic rocks are predominant in the Nicola Group. Breccias and tuffs (NB) are of similar composition to flows, and are distinguished from definite epivolcaniclastic rocks (NC) by their monolithologic character, coupled with the absence of layering or rounding of fragments. Some of the breccias contain abundant aphanitic chips, now converted to dark green chlorite, that resemble hyaloclastite, and many breccias may be epiclastic debris flows with a relatively homogeneous source. Agglomerate (NBA) *sensu stricto* is of mappable thickness only south of Dartt Lake, where it contains maroon scoriaceous, rounded and spindle bombs in a calcite-rich lapilli-tuff matrix. Most of the volcaniclastic rocks are probably laharic deposits. They are heterolithic, containing a variety of andesitic and in places more felsic clasts, massive and unsorted, angular to subrounded, with modal fragment size varying from less than 1 centimetre to 5 centimetres. In a few places the finer facies are well layered and show features of turbidite wackes. Distinctly felsic rocks are generally subordinate to the intermediate volcaniclastics. Laharic breccias in the southwest part of the area consist predominantly of quartz-feldspar-phyric fragments, and a lenticular rhyolite or dacite welded tuff (NTW) north of Dartt Lake is at least 500 metres thick.

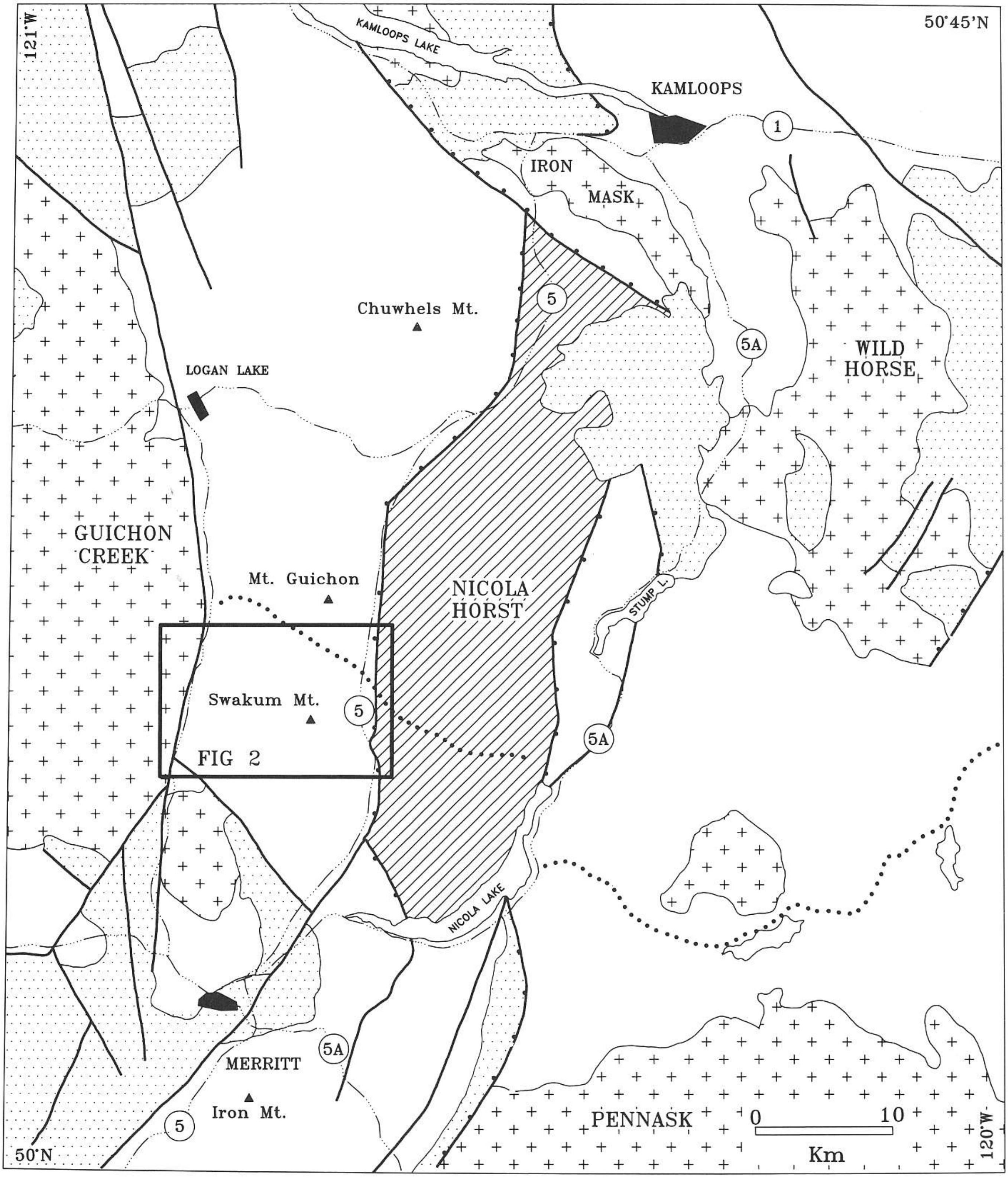

Figure 1-7-1. Location and access map of the Swakum Mountain area. Nicola Group and minor pre-Nicola stratified rocks are unpatterned; undifferentiated igneous and metamorphic rocks of the Nicola horst are hatched. Crosses: Late Triassic-Jurassic plutons, with names of batholiths. Stipple: post-Nicola volcanic and sedimentary rocks. Heavy lines are faults, with dots on downthrown side. Main roads and LITHOPROBE transect are also shown.

LEGEND

LITHOLOGY

QUATERNARY

Q	Glacial, fluvioglacial and fluvial gravel, sand and clay

TERTIARY

Miocene (?)

TB	Olivine basalt flows with ultramafic inclusions

Eocene (Kamloops Group) (?)

KR	Flow-laminated rhyolite flow rocks and breccia or dome

EARLY TO MIDDLE JURASSIC

Ashcroft Formation (?)

AC	Polymict boulder conglomerate (V: volcanic clasts; P: plutonic and volcanic clasts); subordinate sandstone
AS	Sandstone; pebble conglomerate
AL	Limestone; subordinate siltstone interbeds

LATE TRIASSIC AND YOUNGER (?)

Intrusive Rocks

G	Biotite granite with K-feldspar megacrysts (Rey Lake)
D	Diorite: subvolcanic (?) bodies in Nicola Group

LATE TRIASSIC

Nicola Group (Western Belt)

NL	Limestone; polymict volcanic conglomerate with abundant limestone clasts
NT	Dacite or rhyolite tuff, tuff-breccia (W: welded)
NC	Heterolithic andesite-dacite laharic breccia; wacke
NB	Monolithic andesite breccia, tuff (A: agglomerate)
NF	Andesite and basalt flows, flow breccia

TRIASSIC, JURASSIC (AND OLDER ?)

NH	Undifferentiated metamorphic and plutonic rocks of the Nicola Horst

SYMBOLS

Lithologic contact (defined, inferred)

Fault (defined, inferred; dots on downthrown side)

Topographic lineament

70 Strike and dip of bedding (inclined, vertical)

Extent of skarn alteration

▲ Mountain peak

Main road, highway

Lakes:		Former Mines:	
DL	Dartt	A	Old Alameda
EL	Eve	B	Bernice
HL	Helmer	C	Corona
REL	Revelle	L	Lucky Mike
RYL	Rey	S	Sunshine
SOL	Sophia	T	Thelma
SXL	Saxon		

Figure 1-7-2a. Legend for Figure 1-7-1 and 2.

Thin grey limestone lenses (NL) are a minor but distinctive part of the Nicola succession. The greatest thickness mapped, in the southeast corner of the area, is 100 metres. More typically, limy units consist of intercalated limestone up to a few metres thick and heterolithic volcanic breccia/ conglomerate, with limestone clasts up to metre-scale. The limestone is invariably bioclastic, containing in places well-preserved molluscs and coral fragments. One such layer, east of the Lucky Mike deposit, is hematitic with large coral heads, and resembles the red reefoid limestone mapped on Iron Mountain south of Merritt (McMillan, 1981).

All the Nicola volcanic rocks have fine-grained or aphanitic matrices with abundant chlorite and epidote; biotite and amphibole are not evident in hand specimen. An exception is the skarn alteration zone, approximately delineated in Figure 1-7-2, where limestone is converted to coarse pyroxene-garnet rock and volcanic facies to fine-grained magnetite-bearing epidote amphibolite, locally with garnet and pyroxene(?). At numerous localities within and beyond the skarn zone, the volcanic rocks are altered to rusty weathering carbonate-rich rocks containing fine ankerite and pyrite, with or without calcite. Generally these are elongate and associated with northerly trending topographic lineaments; the zone at Corona (Figure 1-7-2) is 600 metres long and up to 50 metres wide. Where carbonate alteration occurs within the skarn zone, magnetite is generally absent and appears to have been converted to pyrite.

The Nicola rocks are intruded by small bodies of augite or hornblende diorite (D) and, near Rey Lake, by coarse biotite granite(G). the diorite is massive, medium to coarse and magnetite bearing. Near Revelle Lake numerous small carbonate alteration zones occur near diorite bodies that generally appear less altered than the enclosing rocks. The granite is coarse and massive with subhedral quartz and potassic feldspar megacrysts. It resembles the Paleocene granite of the Nicola horst to the east (Moore, 1989) and has yielded a Paleocene K-Ar date of 68.9 ± 2.5 Ma (Preto *et al.*, 1979). In the light of early Tertiary K-Ar updating of Mesozoic plutonic rocks elsewhere in the region in and around the Nicola horst (*see*, for example, the isotopic age data in Monger and McMillan, 1984) this figure may not represent the age of intrusion. Granite dikes, that are similar to the Rey Lake

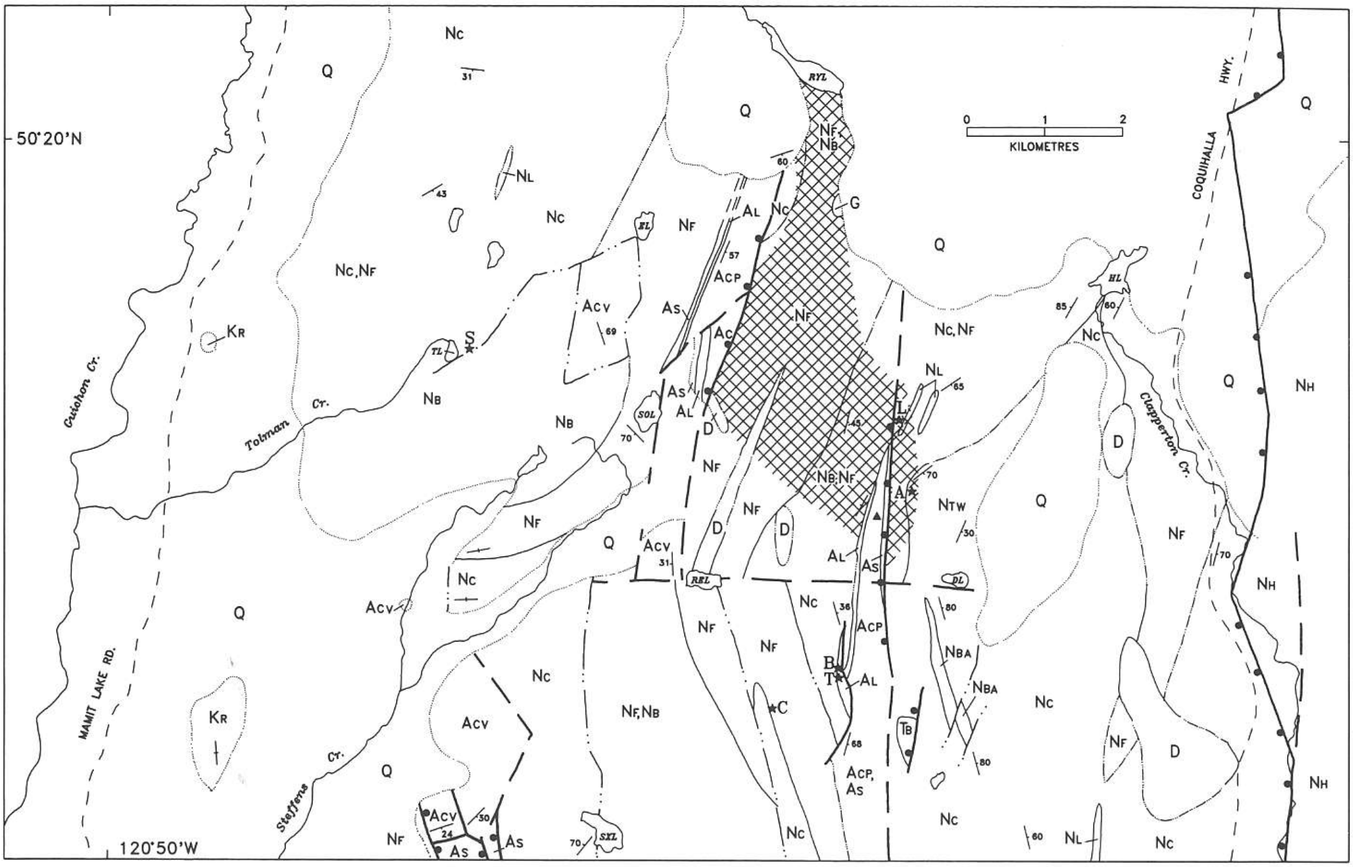

Figure 1-7-2. Geology of the Swakum Mountain area. For legend *see* Figure 1-7-2a.

granite but less coarse and more distinctly porphyritic, are seen cutting skarn in drill core at Rey Lake, and in outcrop at the northwest corner of the map area, where the host breccias are not conspicuously altered.

Clastic and carbonate rocks previously included in the Nicola Group, that occur in at least three and probably five separate locales in the map area, are tentatively correlated with the Early to Middle Jurassic Ashcroft Formation. The most striking examples are two steeply dipping, fault-bounded slices that extend through the crest of Swakum Mountain and northward from Sophia Lake. These contain similar successions that pass eastward and upward from limestone (AL) with thin pebbly, sandy and silty layers to thick, massive to weakly stratified coarse boulder conglomerate (AC) containing poorly sorted, but rounded to well-rounded clasts in a dark green matrix with abundant volcanic plagioclase. Clasts comprise mainly porphyritic intermediate and felsic volcanic rocks, with medium-grained diorite and biotite granite that locally predominate, and minor sedimentary rocks. The succession on Swakum Mountain is topped by up to 80 metres of uniform, siliceous, pyritic sandstone (AS). The stratigraphy as a whole is distinct from that of the adjacent Nicola rocks by virtue of its relative continuity. The limestones contain fine to coarse fossil debris like the Nicola limestone, but in contrast they weather buff and are consistently fetid, whereas the Nicola carbonates are rarely so. Conglomerate clasts are notably more rounded, coarser and the matrix less lithified than in typical coarse Nicola clastics; although some epidote is present the clastic plagioclase is milky white rather than grey or green as in the Nicola rocks. Many of the more felsic volcanic clasts appear less altered than typical Nicola rocks. The presence of plutonic rocks is also distinctive, as are abundant chert pebbles and sand in some layers. Altogether the conglomerate and finer clastics resemble those of the "Clapperton conglomerates" that occur to the south near Merritt (Cockfield, 1948; McMillan, 1981), that have been assigned to the Jurassic Ashcroft Formation by Monger and McMillan (1984). It is evident from Figure 1-7-2 that the skarn alteration of Nicola carbonate volcanic rocks around Swakum Mountain does not affect the immediately adjacent carbonate-clastic successions, indicating that they are younger than the alteration event.

In the extreme southwest corner of the area, west of Saxon Lake, is a fault-bounded succession of mainly coarse clastic rocks that resemble those on Swakum Mountain except that they are easterly striking, contain no plutonic clasts and have abundant coarse fossil debris in the matrix. At one locality they rest on hornblende-phyric (dacite?) flows and are interlayered with and succeeded by volcanic sandstone, also bioclastic, fining upward to black siltstone with minor limestone. One fault block at this locality contains distinctive grey felsic welded tuff and breccia and hornblende dacite(?), with a conformable layer of the volcanic sandstone. Near Revelle Lake and Eve Lake are coarse volcanic conglomerate and sandstone that resemble the Saxon Lake rocks, but lack the finer facies or carbonate rocks.

The Ashcroft succession has been intruded by a few augite-phyric mafic dikes west of Saxon Lake. At Sophia Lake and Swakum Mountain, distinctive dikes of tan-weathering, coarse quartz feldspar porphyry cut sandstone and conglomerate.

Tertiary volcanic rocks, also unrecognized before the present work, are of minor extent. They include a small outlier of olivine basalt south of Dartt Lake and two isolated exposures of rhyolite near Guichon Creek, at the western margin of the area. The basalt is downfaulted against Nicola volcaniclastics to the east; it is at least 30 metres thick and minor variations suggest the presence of several flows. Although no flow contacts are recognized, flow features indicate a moderate easterly dip. Some flows contain peridotite nodules a few centimetres across, similar to those seen in basalt mapped as Miocene north of Lac Le Jeune (Monger and McMillan, 1984). The rhyolite is best exposed on ridges near the southwest corner of the area where it is grey, strongly flow-laminated and contains open lithophysae up to 3 centimetres in diameter. The lamination is steeply-inclined and the rock is locally brecciated. As contacts are not exposed it is not possible to state whether the rocks are flows or a dome.

Areas largely underlain by unconsolidated Quaternary cover occupy all major depressions as well as the flanks and down-ice ends of ridges and mountains.

STRATIGRAPHY AND STRUCTURE

Nicola Group rocks in the area mostly strike northerly and dip steeply (Figure 1-7-2); scarce bedding indicators show that the beds dip predominantly toward the east and are upright. As a whole they are bounded on the east and west by major fault systems that occupy the valleys of Clapperton Creek and Guichon Creek, respectively (Monger and McMillan, 1984; Moore, 1989). The Clapperton fault system appears to be normal, with a net dip-slip of at least several kilometres, in order to have exhumed the relatively deep-seated rocks seen in the Nicola horst. The west-northwest trending linear valley containing Rey Lake, at the north side of the map area, may also contain a major break, as the Nicola Group on Mount Guichon to the north includes well-bedded wackes and coarse laharic deposits without close counterparts along strike to the south; poor exposure on the south flank of the mountain and in the valley precludes a definite conclusion.

The lack of continuity or consistent succession within the Nicola Group suggest strongly that the stratigraphy has been broken into a large number of easterly titled fault blocks, of unknown sense and displacement, hence an estimate of total thickness is not possible. In the study area there is a predominance of flows west of Swakum Mountain and volcaniclastic rocks to the east. However carbonates and thick felsic units occur to both sides. Most of the units south of a line through Revelle and Dartt lakes have a north-northwest trend, whereas those to the north strike north-northeast. There is also a lack of continuity in at least some of the units, notably near Dartt Lake, that reinforces the proposal of an easterly striking fault along this line. There is, however, sufficient similarity between the units to the north and south to preclude the drawing of a lithologic boundary in this vicinity, such as shown by Monger and McMillan (1984).

The Ashcroft rocks at Swakum peak and Sophia Lake strike northerly and dip moderately to steeply east. On their west sides the successions contain thin, immature clastic layers near the base that include fragments of Nicola volcanic rocks; these, and scarce top indicators, indicate eastward facing of the succession and suggest that the western contact may be an unconformity. As the conglomerate and sandstone are succeeded structurally to the east by Nicola rocks, they must be downfaulted against them on the east, thus the successions lie in east-facing half-grabens. Relationships at the Thelma and Bernice properties (Figure 1-7-2) indicate the limestone is repeated by normal faulting. Faulting across the main graben structure is also required to juxtapose the thick limestone segments with the sandstone/thin limestone/conglomerate sequence north of Sophia Lake. The clastic rocks west of Saxon Lake occupy a small graben enclosed by Nicola rocks; it is plausible that the other occurrences described are in a similar setting, and the one at Revelle Lake may occupy a southerly extension of the same structure that contains the Sophia Lake succession. The similarity of the Swakum and Sophia Lake stratigraphy demands correlation and suggests that they are parts of the same succession, dismembered by extensional faulting. It should be emphasized that this interpretation is distinct from that put forth by Cockfield (1948, pages 59-60) and commonly quoted in subseqnent exploration reports. He inferred that the limestones of Swakum Mountain and Sophia Lake, all of which he assigned to the Nicola Group, occupy the limbs of an asymmetric, southerly plunging anticline. The lack of continuity between these localities, coupled with the similar facing of the succession at each, does not support Cockfield's hypothesis. The differences between these and the other three occurrences, given their close proximity, argue that they are not simply lateral correlatives, but are of different age and may represent a different formation. It is possible, for example, that one correlates with the Cretaceous Spences Bridge Group. Paleontology may answer this question.

The Ashcroft succession on Swakum Mountain lies on a variety of Nicola rocks and is not displaced across the proposed fault between Revelle and Dartt lakes, indicating that it was deposited on a relatively flat erosion surface that postdates some of the deformation of the Nicola Group. Its less altered character, particularly the absence of skarn development adjacent to strongly altered Nicola rocks, also demonstrates a significant time gap between the two successions.

The occurrence of felsic tuff and flow rocks in conformable contact with the volcaniclastic rocks west of Saxon Lake, the general presence of euhedral clastic feldspar in all the Ashcroft rocks, and the relatively fresh appearance of some of the volcanic clasts, all indicate the existence of volcanic activity contemporaneous with clastic sedimentation in post-Nicola, possibly Early to Middle Jurassic time.

The Tertiary volcanic rocks also appear to occupy tilted fault blocks; flow lamination in the rhyolite may be in a steep primary orientation, but more probably has been rotated on a Tertiary fault separating the Guichon Creek batholith from the Nicola Group.

DEPOSITIONAL ENVIRONMENTS

Some Nicola volcanic features, such as red agglomerates and rounded clasts in debris flows, are clearly indicative of subaerial processes. The scareity of well-defined bedding in the volcaniclastic rocks and the prevalence of massive, ill-sorted deposits implies they are subaerial lahars. Other criteria, such as the presence of reefoid limestone and hyaloclastite, demonstrate subaqueous deposition, as does the low incidence of oxidized, ropy or brecciated flow tops. All of these features are consistent with a transitional subaerial to shallow submarine environment, characterized by tectonic instability and ephemeral shorelines. At least some lahars flowed into the sea, burying patch reefs and carrying shore-worked debris with them. Synvolcanic faulting must have been an important control on deposition and may explain the abrupt termination of some units, such as the welded tuffs north of Dartt Lake. It would also permit the accumulation of relatively thick successions of subaerial and shallow subaqueous rocks. A similar scenario is indicated by the Western Belt succession on Iron Mountain near Merritt, mapped by McMillan (1981), and was also envisioned by Preto (1979) for Nicola rocks to the south and east of the present area.

The strata assigned to the Ashcroft Formation also present evidence of a transition, upward in the succession, from a submarine to a subaerial environment, accompanied by a substantial increase of relief in the source area. The continuity of succession over at least two separate blocks, as well as the occurrence of sandstone adjacent to the fault on Swakum Mountain, suggest that sedimentation was not related to the present boundary faults. The tabular character and continuity of the finer units suggest deposition on a well-established, stable erosion surface, and the composition of the conglomerate clasts indicates unroofing of at least some (synvolcanic?) plutons. The structures in the conglomerate are consistent with high-energy fluvial deposition; although this environment cannot be conclusively established in the map area, high-angle planar crossbeds seen in similar Ashcroft sandstones to the south, near Merritt, are supportive.

METALLOGENIC IMPLICATIONS

Since the discovery of the Lucky Mike deposit in 1918, the Swakum area has been recognized as a mining camp that has yielded small but significant quantities of base and precious metals (Cockfield, 1948). Although none of the early discoveries remain in production, exploration is active to the present day. There are two principal deposit types, both polymetallic: copper-bearing skarns within the alteration zone shown on Figure 1-7-2 and lead-zinc-copper-silver-gold quartz-stockwork veins associated with iron-rich carbonate alteration zones, both within and outside the skarn zone. The former type is exemplified by the Lucky Mike, where copper is accompanied by subordinate tungsten, silver, gold, lead and zinc. Old Alameda and the other deposits shown on Figure 1-7-2 are of the latter type.

Pending a fuller account of these deposits, to be given in a later publication, a few important conclusions may be drawn. Field relationships described above show that the skarn alteration predates the Ashcroft sedimentary rocks. Similar reasoning indicates that the granite near Rey Lake, despite its spatial association with skarn, is also later than the alteration. In the absence of direct evidence, it may be suggested that an unexposed intrusive body is responsible for the alteration zone.

In contrast, the carbonate alteration and associated mineralization are younger than the Ashcroft limestone at the Thelma and Bernice properties and also north of Swakum peak, where limestone is mineralized and, together with Nicola rocks and post-Ashcroft porphyry, silicified and altered to iron carbonate. Dating the sedimentary rocks is required to place an upper limit on the age of this mineralizing event, but it is clearly younger and distinct from skarn formation.

ACKNOWLEDGMENTS

Richard Meyers, with Todd Huebner, took part in some of the fieldwork, especially at mineral properties, and provided helpful data, support and debate. The senior author also benefited from discussions with Ken Daughtry, Bill McMillan and Vic Preto, who were all generous with their data and time. Pip Moore gave invaluable care and support with field camp logistics and the database. Guy Rose and Elmer O'Hanly of Quilchena ranch kindly provided good accommodation and hospitality and, with other residents of the region, assisted the project in many ways.

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AGE OF THE BOOTJACK STOCK, QUESNEL TERRANE, SOUTH-CENTRAL BRITISH COLUMBIA* (93A)

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KEYWORDS: Geochronology, Bootjack stock, Polley stock, Quesnel Terrane, subvolcanic intrusive complex, $^{40}Ar/^{39}Ar$ dating, Early Jurassic.

INTRODUCTION

The Bootjack stock in south-central British Columbia is a zoned nepheline syenite intrusion related to a magmatic event which gave rise to the adjacent Polley stock, host to the Cariboo Bell porphyry copper-gold deposits. Hodgson *et al.* (1976) considered the Bootjack Stock to be a deeper level phase of the subvolcanic Polley stock. A K-Ar date of 184 ± 7 Ma published by Hodgson *et al.* (1976) from hydrothermal biotite related to sulphide mineralization was considered by these authors to be a "median" age of the Polley stock in that intrusive events bracketed the time of formation of the metasomatic biotite.

Bailey (1978, 1988), on the basis of a whole-rock K-Ar date obtained from the fine-grained margin of the Bootjack stock, and on geological grounds, considered the Bootjack stock to be much younger than the adjacent Polley stock.

This note reports new data obtained from a coarse-grained amphibole-rich phase of the Bootjack stock which support the conclusion of Hodgson *et al.*, that the Bootjack and Polley stocks are probably magmatically related.

GEOLOGICAL SETTING

The Bootjack stock and the adjacent Polley stock occur within the south-central part of the Mesozoic Quesnel Terrane, a tectonic belt bounded to the west by the mainly Paleozoic Cache Creek Group and to the east by Paleozoic rocks of the Barkerville Terrane (Bailey, 1988) over which it has been thrust (Struik, 1983) (Figure 1-8-1). Rocks of the Quesnel Terrane comprise a lower assemblage of mainly fine-grained epiclastics, overlain by a bimodal shoshonitic volcanic suite ranging from alkaline basalt (alkali olivine basalt to analcite-bearing basalt) at the base to felsic volcanic breccias (mainly laharic debris flows with abundant trachyte and latite clasts) toward the top. Overlying this assemblage is a younger analcite-rich basalt unit which, in turn, is overlain by remnants of a probably Middle Jurassic successor basin assemblage (Figure 1-8-2). From fossil evidence the age of the lower basalt is Carnian to Norian while the overlying felsic volcanic rocks are Sinemurian and younger (Bailey, 1988).

Observations discussed by Bailey and Hodgson (1979) indicate the Polley stock is a subvolcanic intrusive complex emplaced within a volcanic edifice coevally with volcanism and that the age of crystallization of the stock is the same as the Sinemurian and younger felsic volcanic rocks.

AGE OF THE BOOTJACK STOCK

Hornblende from a coarse-grained porphyritic phase of the Bootjack stock was collected for radiometric dating by the $^{40}Ar/^{39}Ar$ method to define the age of the stock and to determine whether the Polley and the Bootjack stocks have similar ages, as suggested by Hodgson *et al.* (1976).

ANALYTICAL METHOD

Samples and three flux monitors (LP-6 biotite; *see* Table 1-8-1) were irradiated with fast neutrons in position 5C of the

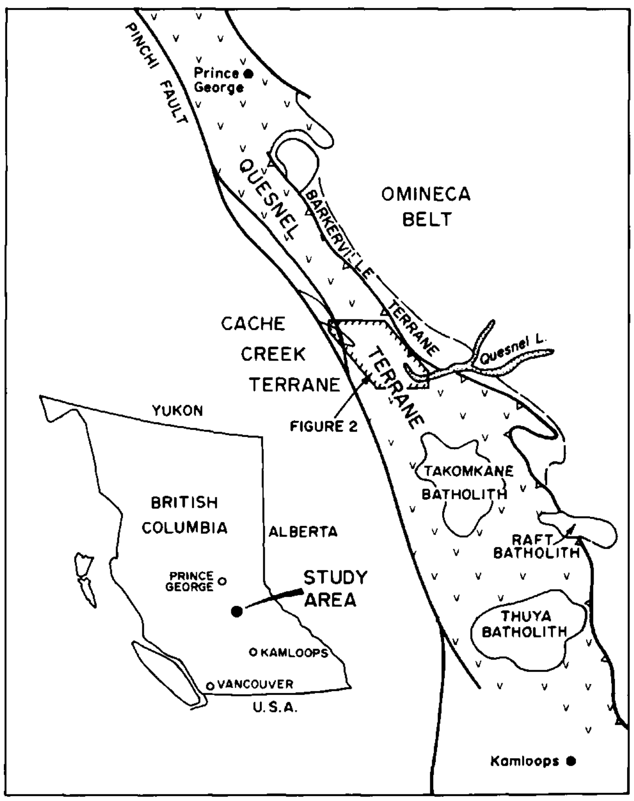

Figure 1-8-1. Location of Quesnel Terrane, south-central British Columbia.

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

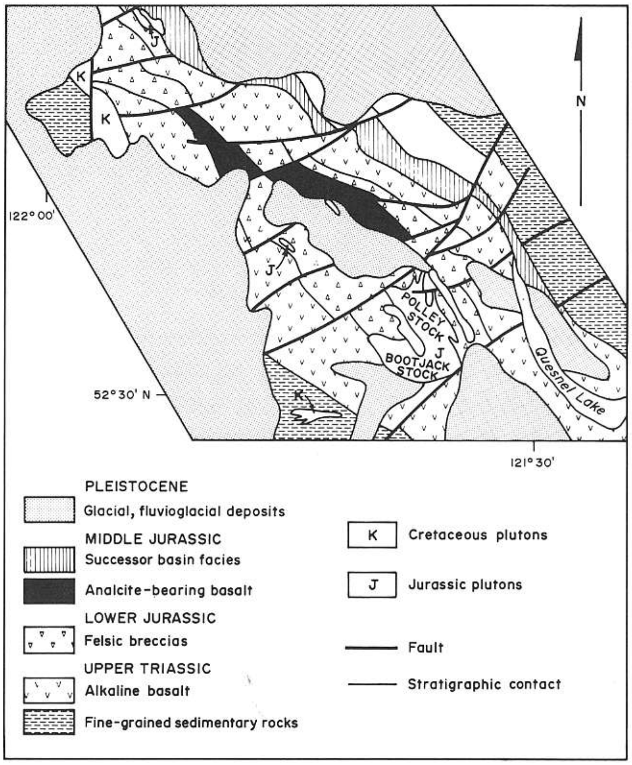

Figure 1-8-2. Generalized Triassic-Jurassic geology, central Quesnel belt. Younger Mesozoic and Tertiary strata not shown (after Bailey, 1988).

McMaster nuclear reactor (Hamilton, Ontario) for 1.04 days. The monitors were distributed throughout the irradiation container and the J-values for individual samples were determined by interpolation.

For step-heating experiments, irradiated samples were loaded into a quartz furnace-tube and heated using a Lindberg tube furnace. The bakeable, ultrahigh vacuum, stainless steel, argon extraction system is operated on-line with a substantially modified A.E.I. MS-10 mass spectrometer run in the static mode. Total fusion analyses were done using a custom five-position turret system and resistively heated tantalum-tube crucibles. Measured mass spectrometric ratios were extrapolated to zero time, corrected for an $^{40}Ar/^{36}Ar$ atmospheric ratio of 295.5 and corrected for neutron-induced ^{40}Ar from potassium and ^{39}Ar and ^{36}Ar from calcium (*see* Table 1-8-1). Ages and errors were calculated using the formula given by Dalrymple *et al.* (1981) and the constants recommended by Steiger and Jäger (1977). The errors shown in Table 1-8-1 represent the analytical precision at 2σ assuming that the error in J-value is zero.

RESULTS

The age spectrum for the amphibole from the Bootjack stock is shown in Figure 1-8-3; analytical data are listed in Table 1-8-1. The small, low-temperature (500° to 875°C) gas fractions, representing approximately 15 per cent of the total ^{39}Ar released, are characterized by low $^{37}Ar/^{39}Ar$ ratios (which correspond to low Ca/K ratios) and record the presence of a small amount of fine-grained biotite enclosed in the somewhat poikilitic hornblende. The age and thermal significance of these steps are not known. The five steps from 920° to 1020° C correspond to the principle release of ^{39}Ar from the amphibole. The dates for these steps yield a well defined plateau age of 203.1 ± 2.0 (2σ) Ma for 80 per cent of the ^{39}Ar released. The $^{37}Ar/^{39}Ar$ ratios for this segment of the age spectrum are in the range 2.09 to 2.66 (Ca/K ratios of about 3.8 to 4.9), which is typical of alkali amphiboles. The three highest temperature steps account for about 5 per cent of the ^{39}Ar released and have poorly defined dates and significantly higher $^{37}Ar/^{39}Ar$ ratios than the amphibole. The form of the age spectrum is more likely the result of mixing of three phases rather than thermal overprinting of the amphibole. Thus, we interpret the plateau age as being the best estimate of the age of emplacement of the Bootjack stock.

The new $^{40}Ar/^{39}Ar$ date invalidates the K-Ar whole-rock date (117.3 ± 2.7 (2σ) Ma (recalculated from 111 Ma using the constants of Steiger and Jäger, 1977), reported by Bailey (1978). The whole-rock sample is from the fine-grained margin of the stock (sample number BML-4/5, chemical analysis number 51 in Bailey, 1978) and is characterized by abundant biotite and feldspar. Although the K-Ar date has no age significance, the discordance between it and the amphibole date probably indicates that the area was affected by a thermal event (of unknown magnitude) in post-mid-Jurassic

TABLE 1-8-1
ANALYTICAL DATA, BOOTJACK STOCK AMPHIBOLE

DB-87-2 Hb (80/115) J = 0.006374

Temp. °C	40/39 (1)	36/39 (1)	37/39 (1, 2)	Vol.^{39}Ar x 10-9 cc NTP (3)	f39	%^{40}Ar Rad.	Date (5)	± 2σ Ma
600	22.574	0.0279	0.493	3.544	0.0691	63.45	157.6	± 7.2
700	20.613	0.0199	0.419	1.205	0.0235	71.49	162.0	± 13.4
775	24.852	0.0339	0.441	0.873	0.0170	59.67	163.0	± 24.9
825	18.786	0.0116	0.721	1.187	0.0231	81.89	168.8	± 7.6
875	18.777	0.0077	1.465	1.391	0.0271	88.33	181.5	± 5.2
920	19.217	0.0032	2.089	4.810	0.0938	95.74	200.3	± 2.5
950	19.148	0.0020	2.289	7.579	0.1478	97.71	203.5	± 2.0
980	18.974	0.0012	2.333	9.455	0.1843	98.88	204.0	± 1.4
1000	19.061	0.0017	2.534	8.819	0.1719	98.15	203.5	± 2.7
1020	19.157	0.0024	2.663	10.536	0.2054	97.27	202.8	± 1.7
1040	21.151	0.0183	7.903	0.648	0.0126	77.06	179.2	± 16.9
1070	22.927	0.0168	16.089	0.365	0.0071	83.45	209.7	± 21.7
1200	25.272	0.0497	168.400	0.882	0.0172	91.82	277.3	± 23.6

wt. = 150 mg
Total ^{39}Ar = 51.294 × 10-9 cm^3 NTP
I.A. = 197.9 ± 3.9 Ma
P.A. = 203.1 ± 2.0 Ma

(1) True ratios corrected for fractionation and discrimination using $^{40}Ar/^{36}Ar$ atmos. = 295.5
(2) $^{37}Ar/^{39}Ar$ is corrected for the decay of ^{37}Ar during and after irradiation using a ^{37}Ar half-life of 35.1 days.
(3) Volume of ^{39}Ar is determined using equilibration peak height and mass spectrometer sensitivity.
(4) Isotope production ratios for the McMaster reactor are from Masliwec (1981). (40/39)K = 0.0156
(36/39)Ca = 0.390169
(37/39)Ca = 1536.1
(5) Ages calculated using the constants recommended by Steiger and Jäger (1977). Errors represent the analytical precision only (i.e. error in J value = 0). Flux monitor used: LP-6 Biotite at 128.5 Ma.

time. The region borders the Omineca crystalline belt and contains several small plutons of possibly Cretaceous age (*e.g.* the Gavin Lake stock to the southwest) which may have overprinted minerals less argon retentive than amphibole.

DISCUSSION

The Polley and Bootjack stocks together comprise a number of intrusive phases which, from the southwest to the northeast, have been emplaced at successively shallower levels. The map distribution of rock types represents an oblique section through the intrusive complex, indicating it has been tilted down to the northeast (Figure 1-8-4).

Metasomatic biotite from the Polley stock yields a conventional K-Ar date of 184 ± 7 Ma (Hodgson *et al.*, 1976). As discussed above, the whole-rock K-Ar date may indicate post-Jurassic thermal overprinting of the region; biotite from the Polley stock may have been similarly affected. As hydrothermal activity, volcanism and plutonism were essentially coeval events (Bailey and Hodgson, 1979), this date must be viewed as a minimum age for magmatic crystallization at subvolcanic levels. On the other hand, the plateau age of 203.1 ± 2.0 (2σ) Ma for amphibole of the Bootjack stock indicates the age of magmatic cooling at a deeper crustal level of an early intrusive phase, possibly representing the time of the onset of felsic volcanism.

It may be speculated that, as the last period of basaltic volcanism, stratigraphically beneath the felsic volcanic rocks, is characterised by sodium enrichment (Bailey, 1978), the nepheline syenite of the Bootjack stock may represent the result of differentiation of a sodium-rich mafic magma. Successively younger phases of the intrusive complex, from diorite through monzonite to syenite, and their extrusive equivalents, latite and trachyte, may indicate increasing potassium enrichment of such an initially sodium-rich parent.

The age data reported here, together with the results of earlier dating (Hodgson *et al.*, 1976) suggest that plutonism and accompanying volcanism related to the Bootjack and Polley stocks occurred over a relatively long period of time, from lower Sinemurian to possibly the Aalenian stages, or through most of the Early Jurassic period.

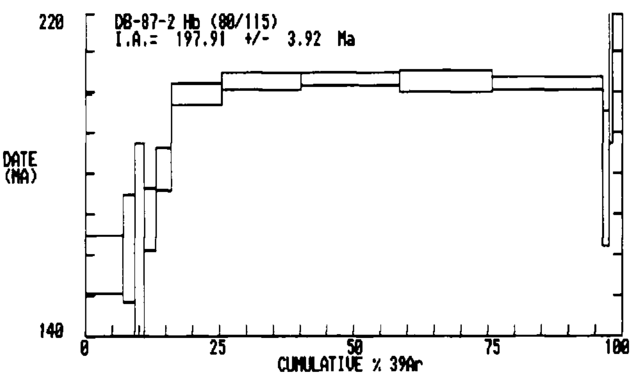

Figure 1-8-3. $^{40}Ar/^{39}Ar$ age spectrum of amphibole from the Bootjack stock. The vertical thickness of the bars represents the 2σ errors for each step.

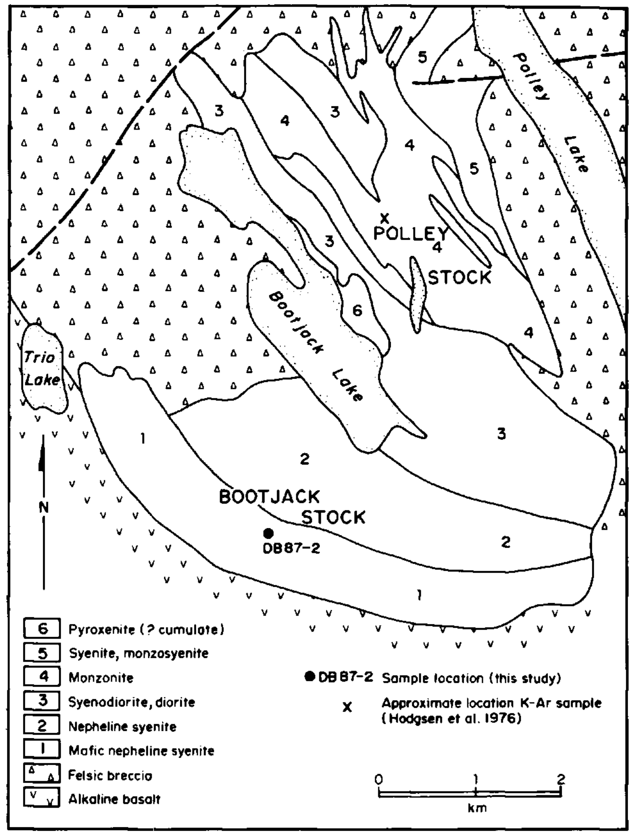

Figure 1-8-4. Generalized geology of Bootjack and Polley stocks (after Hodgson *et al.*, 1976; Bailey 1988 and unpublished) showing sample locations.

ACKNOWLEDGMENTS

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GEOLOGY OF NANIKA LAKE MAP AREA (93E/13)

By L.J. Diakow

KEYWORDS: Regional mapping, Nanika Lake, stratigraphy, Hazelton Group, quartz veins, stratabound sulphides, skarn, porphyry copper.

INTRODUCTION

In 1989, the Whitesail regional mapping project concluded with a survey of 800 square kilometres in the Nanika Lake map area (Figure 1-9-1). The project was initiated in 1986 with the objectives of refining stratigraphic divisions of Mesozoic and Cenozoic volcanic arc successions, and evaluating this stratigraphy for economic potential for precious metals. The project area encompasses roughly four 1:50 000 map sheets. The results of this mapping are summarized in a series of published reports (Diakow and Mihalynuk, 1987; Diakow and Koyanagi, 1988; Diakow and Drobe, 1989).

This report presents results of mapping conducted in Nanika Lake map area which is underlain by roughly equal volumes of plutonic and volcanic rocks. The characteristics and inter-relationship of plutonic rocks and Jurassic volcanic stratigraphy are discussed.

GENERAL GEOLOGY

The Nanika Lake area is bisected by a segment of the northwest-trending boundary between the Coast Belt and Intermontane Belt. In the southwest, plutons of probable late Cretaceous and Tertiary age cut plutonic and metamorphic

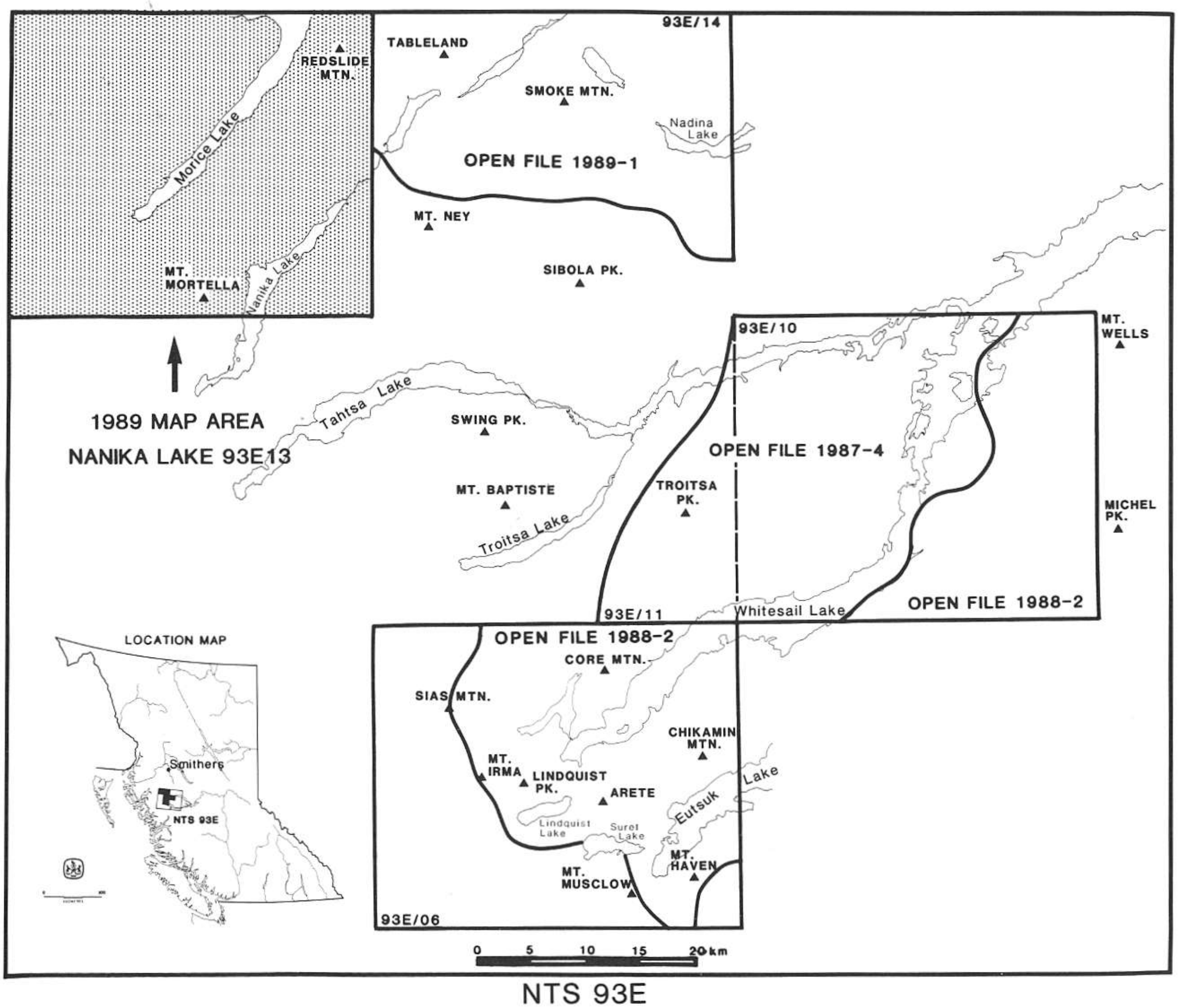

Figure 1-9-1. Index map showing the location of Nanika Lake map area and published Open File maps of the Whitesail Project.

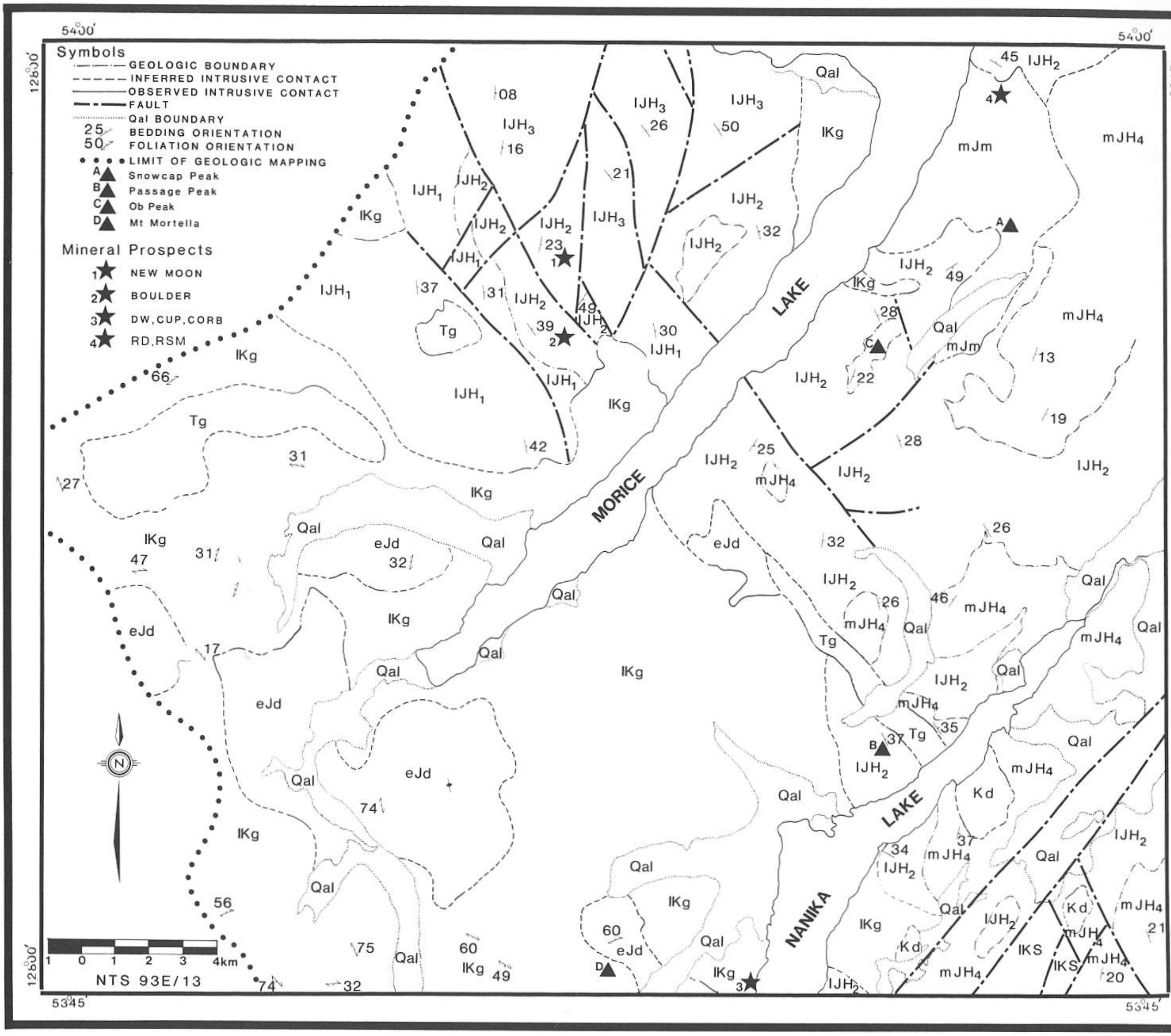

Figure 1-9-2. Generalized geology and mineral prospects in the Nanika Lake map area.

rocks of the Early to Middle Jurassic Gamsby complex (Figure 1-9-2). The Gamsby complex, renamed after the informal Gamsby group of Woodsworth (1978), is presently defined as an early-Late Jurassic metamorphic belt superimposed on Early and Middle Jurassic magmatic rocks of the Intermontane Belt (van der Heyden, 1989).

The Morice Lake pluton bounds volcanic and minor sedimentary rocks of the Lower and Middle Jurassic Hazelton Group to the northeast. Sedimentary rocks of the Lower Cretaceous Skeena Group crop out in several localities near the southeast boundary of the map area. The layered successions are cut by plutons and swarms of dikes of mid-Jurassic to Tertiary age. These intrusive episodes coincide with nearby coeval volcanic deposits which mark the end of Hazelton island arc development and construction of a younger continent-margin arc.

HAZELTON GROUP

Strata of the Hazelton Group comprise a northeast to east-facing homoclinal succession which has an estimated minimum thickness of about 4.5 kilometres in the Nanika Lake area. A lower stratigraphic contact with pre-Jurassic rocks was not observed. Instead the oldest Jurassic strata share an intrusive contact with the Morice Lake pluton west of Morice Lake. The youngest strata of the Hazelton Group cap mountain peaks in the eastern part of the area and continue into the adjoining Newcombe Lake map area where, near Mount Ney, they are apparently disconformable with overlying Cretaceous stratigraphy of the Skeena and Kasalka groups.

The Hazelton Group consists almost exclusively of subaerial volcanic rocks; rare intravolcanic sediments of lacustrine origin are confined to a few stratigraphic intervals.

	LEGEND
Qal	Alluvium and glacial till
	LAYERED ROCKS
	LOWER CRETACEOUS
lKs	Skeena Group: micaceous sandstone and siltstone, shale
	MIDDLE AND LOWER JURASSIC
	Hazelton Group
mJH4	Dacite and rhyolite lapilli tuff, lithic-crystal tuff, lava flows and minor ash-flow tuff
lJH3	Well-layered, maroon lapilli and ash tuffs of andesitic composition, interspersed flows of basaltic and dacitic composition.
lJH2	Basalt and rhyolite flows and related lapilli tuff, ash tuff, minor ash-flow tuff and local agglomerate; intravolcanic sedimentary rocks on Ob Peak ridge.
lJH1	Well-layered green lapilli and ash tuffs of andesitic composition, interspersed flows of basaltic and rhyolitic composition, rare limestone intercalated with chert and ash tuff occur at several intervals.
	INTRUSIVE AND METAMORPHIC ROCKS
Tg	Tertiary: miarolitic granite and megacrystic granodiorite.
lKg	Late Cretaceous(?): hornblende-biotite granodiorite of the Morice Lake pluton.
Kd	Cretaceous(?): fine-grained diorite
mJm	Middle Jurassic: porphyritic monzonite
eJd	Early to Middle Jurassic: Gamsby complex hornblende diorite; includes metavolcanic rocks and orthogneiss.

The volcanic succession is subdivided into four lithostratigraphic units, each established on the basis of composition, texture and volume of flows versus pyroclastic rocks. Collectively, these units reflect successive episodes of volcanic activity in an island arc environment.

UNIT lJH1

The lowest stratigraphic map unit of the Hazelton Group, lJH1, is at least 1 kilometre thick west of Morice Lake. It is a well-layered succession composed mainly of pyroclastic rocks with minor lava flows. In general, lapilli tuffs form massive beds up to 5 metres thick that commonly grade upward into sections composed of 1 to 30-centimetre layers of ash tuff. Accretionary lapilli are sometimes found within finer ash tuff layers. Rapid lateral variation in texture and thickness of layered pyroclastic sections is common, diminishing the usefulness of these rocks as reliable markers for correlation.

Basalt, andesite and dacite flows make up perhaps as much as 20 per cent of Unit lJH1. The mafic to intermediate flows are dark green, aphyric and often partly altered to chlorite and epidote. Dacitic flows, between 15 and 30 metres thick, appear mainly in the upper part of the unit in association with lapilli and block tuffs that contain abundant felsic pyroclasts. Rare sag structures are developed where airborne blocks have fallen and disrupted alternating parallel layers of lapilli and graded ash tuff.

Grey to black limestone is locally intercalated with chert and ash tuff at several intervals in the middle and upper part of the unit. The limestone comprises single and multiple beds 25 centimetres to 5 metres thick within sections up to 25 metres thick.

The upper contact of Unit lJH1 is gradational with overlying strata of Unit lJH2. West of Morice Lake, the contact is placed at the base of the first major section of basaltic flows.

UNIT lJH2

A significantly greater volume of lava flows in Unit lJH2 is the main feature which distinguishes it from Unit lJH1. Flows of basalt, andesite and rhyolite, and subordinate interbeds of tuff, characterize the unit. It has a maximum thickness of 1.4 kilometres west of Morice Lake and thickens to at least 2 kilometres east of the lake.

In the west, the section is comprised of alternating flows of basalt, andesite and rhyolite. Dark green mafic flows with porphyritic, aphyric and amygdaloidal textures dominate lower parts of the sequence. They generally pass npward into light green and cream-colored felsic flows with sparsely porphyritic, flow-laminated and spherulitic textures. Contacts between flows are sometimes indicated by flow breccia and vesiculated tops; otherwise they are difficult to recognize within homogeneous weathered sections that may be as much as 200 metres thick.

A variety of bedded pyroclastic deposits occupy stratigraphic intervals between the lava flows. These include green, mauve and maroon lapilli tuff with graded coarse to fine ash interbeds, and in places substantial accumulations of monolithic block tuff. Rhyolitic fragments and up to 5 per cent quartz crystals are common constituents in tuffs interlayered with the felsic flows. Compaction layering, indicated by flattened and aligned pyroclasts, is evident over several metres within local sections of well-layered and graded airfall tuffs. Lapilli-block tuff forms several spectacular lenticular deposits in contact with both mafic and felsic flows. For example, 1 kilometre southwest of the New Moon prospect, a section 100 metres thick is composed of 50 metres of tightly packed monolithic blocks of rhyolite, some as large as 75 centimetres in diameter. This deposit rests sharply upon about 150 metres of basaltic andesite flows at the base of Unit lJH2. Up-section the blocks are concentrated in thinner beds within related deposits of lapilli tuff and ash-flow tuff. This breccia is truly an agglomerate that was deposited close to a vent. At several other localities, breccias dominated by blocks and smaller fragments are interpreted as talus deposits at the over-steepened margin of felsic lava flows.

East of Morice Lake, Unit lJH2 is dominated by very thick basaltic and andesitic flows with intercalated tuffs and, rarely, intravolcanic sedimentary rocks. Rhyolitic flows and pyroclastic rocks that are common in Unit lJH2 to the west are notably absent from this succession. Black aphyric basalt flows, with a fresh appearance and subtle protruding parallel laminae produced by trains of plagioclase microlites, underlie the lower west-facing slope on the east side of Morice Lake. These flows pass upwards into porphyritic basaltic andesite which is the dominant lithology of Unit lJH2 as far southeast as Nanika Lake. Pristine augite is diagnostic of these flows; although, their small size (2 or 3 millimetres) and low abundance (1 to 3 per cent) allow them to be easily overlooked. Amygdaloidal flows with irregular calcite-quartz amygdules comprise discrete members or occasionally the upper part of texturally graded aphyric and porphyritic lava flows.

Monolithic breccia interleaves co-genetic basaltic flows over 250 metres of vertical section within an area of 1 square kilometre approximately 6 kilometres due south of Ob Peak. Less than 2 kilometres away, the breccia thins rapidly to just a few scarce beds, or is absent altogether from the enclosing succession of flows. The breccia is an aggregate of poorly sorted, subangular and angular fragments varying from lapilli to blocks 75 centimetres in diameter. The texture and composition of fragments is identical to basaltic flows nearby. The matrix consists of fine granular mafic debris or white quartz which imparts a diagnostic irregular net-like texture. The overall angularity and unimodal composition of fragments suggest they are proximal deposits of a nearby volcanic centre. The quartz-rich matrix is attributed to low temperature, postdepositional diagenetic fluids, as hydrothermal fluids would undoubtly result in a broad area of altered country rock.

Dark maroon lapilli tuff with graded crystal-ash and accretionary lapilli interbeds comprise well-layered sections between resistant lava flows. Individual sections exceed a thickness of 100 metres.

Intravolcanic sedimentary rocks at least 75 metres thick are conformable with underlying flows and tuffs of Unit lJH2 on the ridge that includes Ob Peak. The longest section extends for 2 kilometres; a small outlier caps the southern part of the ridge. A representative section near Ob Peak comprises a base of ash tuff and tuffaceous siltstone with scarce marlstone and limestone layers up to 10 centimetres thick. The bulk of this succession consists of interbedded green and grey sandstone and siltstone that commonly grade into thinly laminated mudstone. Structures in these rocks include parallel beds from several millimetres to 0.5 metre thick, scarce trough crosslaminations and rare thin beds with rip-up clasts and convolute bedding.

UNIT lJH3

Unit lJH3 is a heterogeneous succession of maroon and green pyroclastic rocks and subordinate lava flows between 1.2 and 2.1 kilometres thick. Superb sections crop out in the type area between Atna Bay in the north and the New Moon prospect in the south. Immediately north of the New Moon, maroon tuffs conformably overlie and in part interleave basaltic and rhyolitic flows of Unit lJH2.

Maroon and brick-red lapilli tuff are diagnostic of this unit. From a distance, a typical exposure is very well layered; close up, medium and thick beds of lapilli tuff are commonly internally graded or separated by thinly bedded coarse to fine ash and crystal-ash tuff. Accretionary lapilli are characteristic of many of the finer grained layers. Occasionally tuff beds composed primarily of accretionary lapilli occur over an interval of at least 50 metres. The most common pyroclasts are composed of dark maroon and green aphyric fragments. Light-colored fragments of rhyolitic composition are also sporadically present. Trace amounts of broken quartz crystals are ubiquitous in most tuff beds. Scarce welded tuffs generally have aligned and compacted pyroclasts. In thin section ash-flow tuff consists of varying proportions of lithic fragments, cuspate shards and broken crystals. Ash-flow tuff containing compacted dacitic fragments up to 14 centimetres in diameter is associated with nearby dacitic flows and maroon tuffs on several small islands in Atna Bay.

Lava flows in Unit lJH3 include aphyric and amygdaloidal basalt and sparsely porphyritic and flow-laminated dacite. They are distinguished from flows of similar composition in Unit lJH2 by their widely spaced and relatively thin character, absence of visible pyroxene phenocrysts and their occurrence within thick successions of pyroclastic rocks. Lithologies of Unit lJH1 resemble Unit lJH3, however, demonstrable gradational contacts with intervening flows of Unit lJH2 indicate that similar successions of tuffs recur within the stratigraphy.

UNIT mJH4

Unit mJH4 is a very distinctive succession, more than 600 metres thick, that is characterized by tuffs with rhyolitic fragments, massive lava flows and minor ash-flow tuff of rhyolitic composition. This succession is widely distributed along the east margin of the map area. The lower contact is abrupt and unconformable with underlying strata of Unit lJH2 approximately 3 kilometres east-southeast of Ob Peak. A similar stratigraphic relationship is inferred on a ridge leading to an unnamed 2130-metre peak in the southeast corner of the map area.

Light green lapilli tuff with interbeds of lithic-crystal tuff are the principal lithologies of the unit. A typical example of lapilli tuff contains fragments of off-white, light green and pink, aphyric and flow-laminated rhyolite. The matrix may be slightly recessed and charged with ubiquitous quartz (2 to 5 per cent) and plagioclase crystal fragments. Other fragments that have local prominence include fine-grained andesite, chloritized shards and rare quartz-feldspar intrusive. The size of fragments is generally between 0.5 and 2.0 centimetres in diameter, although blocks between 20 and 40 centimetres in diameter are not uncommon.

Rhyolite flows are interlayered with the felsic tuffs mainly in an area adjacent to the north end of Nanika Lake. Also, a section of massive flows, more than 200 metres thick, overlies a thick succession of rhyolitic tuffs in the southeast corner of the map area. In general, the flows underlie knolls and pass laterally into interfingering tuffs and local ash-flow tuffs. Massive and flow-laminated textures are dominant but are sometimes obscured by coalescing spherulites. In some rhyolite flows east of Nanika Lake miarolitic cavities up to 15 centimetres in diameter are lined by drusy quartz and feldspar crystals. Nearby, the ash-flow tuffs contain collapsed pumice fragments that define thin zones of moderate welding. Ash-flow tuffs are generally thin, areally restricted deposits that apparently grade laterally into thickened piles of interfingered rhyolitic flows and related tuffs. These deposits record small-volume explosive eruptions, perhaps associated with domes; they lack features indicative of caldera-forming eruptions.

Depositional Environment

Jurassic volcanic rocks underlying the map area are analogous with the complexly interlayered pyroclastic and flow deposits of composite volcanoes (Cas and Wright, 1987). In the Nanika Lake area volcanic rocks accumulated in a subaerial environment. A thick succession composed of parallel-layered and graded deposits attests to the explosive nature of volcanoes that erupted large volumes of airborne ejecta. The relative absence of intervening deposits of pyroclastic breccia in the thick successions characterized by tuffs charged with lapilli-size and finer pyroclasts may indicate that these deposits aggraded on a gentle topography some distance from the volcanic centres. Thick flows with few interspersed pyroclastic deposits record waning explosive phases in eruptive episodes. Volcanism was punctuated by intervals of fine clastic and limestone sedimentation in shallow lakes established on the volcanic terrain.

Except for felsic volcanic rocks at the top of the stratigraphy, the physical nature and composition of rock units changed little in space and time. This terminal episode of felsic volcanic activity relates to small rhyolitic volcanoes that apparently mark the end of Hazelton volcanism and island arc construction in the Whitesail Lake area.

Age and Correlation

The age of volcanic rocks underlying the Nanika Lake area is poorly constrained by a U-Pb age of 1844 Ma (van der Heyden, 1989). This date is from tuffs mapped as part of the uppermost felsic volcanic unit, mJH4. Felsic rocks with similar stratigraphic position in the Whitesail Lake area have been informally named the Whitesail formation (Woodsworth, 1980). They are contemporaneous, in whole or in part, with Aalenian to Bajocian shallow-marine sedimentary rocks of the Smithers Formation.

Based on lithology, strata of map units lJH1 and lJH2 are tentatively assigned to the Telkwa Formation. In particular, the general gradation up-section from andesitic tuffs with subordinate basalt and dacite flows in Unit lJH1, into thick basalt, andesite and rhyolite flows and interlayered tuffs in Unit lJH2 resembles sections of the Telkwa Formation reported to the north in the Thautil River area (Desjardins *et al.*, 1990, this volume). Well-layered tuffs with subordinate flows that characterize Unit lJH3 may be stratigraphically equivalent to the Red Tuff member of the Nilkitkwa Formation. Felsic volcanic rocks representative of map units lJH2 and mJH4 have been collected from three sites for U-Pb geochronology.

INTRUSIVE AND METAMORPHIC ROCKS

Gamsby Complex Diorite and Metamorphic Rocks – Unit eJd

Hornblende diorite and subordinate quartz diorite crop out intermittantly within an area that extends from the south end of Morice Lake to Nanika Lake. Variation in proportions of felsic and mafic minerals, and grain size over short distances impart an inhomogeneous weathered appearance to the diorite. The main variety of diorite contains as much as 60 per cent medium and coarse-grained hornblende. Typical exposures are dark green and mesocratic with the concentration of hornblende generally exceeding that of plagioclase. Quartz is rarely visible in hornblende diorite but it increases to approximately 15 per cent in quartz diorite. Diorite is typically unfoliated except in the northernmost exposure east of Morice Lake, where a weak to strong foliation is locally developed. In the same area segregations of light and dark minerals alternate in layers within sections up to 3 metres thick. Several generations of randomly oriented basaltic dikes cut the diorite.

Metamorphic rocks, including metavolcanics, schist and orthogneiss in the greenschist and amphibolite grade of regional metamorphism, are spatially associated wih dioritic rocks. They occur discontinuously within areas tens of metres to 2 kilometres across. Diorite clearly intrudes the metamorphic rocks at some localities. The contacts are varied; some contain xenoliths of schist incorporated into diorite and others are transitional contacts with greenschist alternating with narrow tabular injections of diorite.

The age of hornblende diorite in the map area is assumed to be early Jurassic. A typical hornblende diorite near the south boundary of the map area and on the west side of Nanika Lake apparently continues into the adjoining map area to the south. It has yielded a K-Ar age of 176 ± 40 Ma (Sample GSC 80-38; Stevens *et al.*, 1982).

Morice Lake Pluton – Unit lKg

The Morice Lake pluton underlies the entire southwest half of the map area. Satellite bodies crop out in the Atna River valley and south of the entrance to Atna Bay on Morice Lake. Midway down the west side of Morice Lake a sharp intrusive contact with early Jurassic volcanic rocks rises steeply from the shoreline; it flattens below the ridge crest at about 1950 metres elevation. In many places this pluton and related dike swarms cut across diorite and the foliation of metamorphic rocks of the Gamsby complex. Agmatite is locally extensive in irregular bodies adjacent to these contacts. It is comprised of angular blocks of diorite supported by a matrix of leucocratic granodiorite.

The Morice Lake pluton is composed of granodiorite with a contrasting fabric and subtly variable proportions of felsic and mafic minerals. Xenoliths of grey, fine-grained quartz diorite are both diagnostic and widely distributed. The main mass of the intrusion is light pink, inequigranular and contains between 20 and 40 per cent anhedral quartz up to 1.5 centimetres in diameter. Quartz has a distinctive elliptical shape in the granodiorite at Atna Bay. Biotite is typically more abundant than hornblende and combined they average 5 to 10 per cent of the rock and rarely exceed 20 per cent. Chlorite pseudomorphs and commonly comprises a felty aggregate that mantles the mafic minerals. Southwest of the head of Morice Lake the appearance and composition of granodiorite changes; here it is off-white in color, the quartz content decreases and a planar penetrative fabric is pervasive. In these rocks the foliation is defined by flattened aggregates of chlorite psedomorphous after biotite and hornblende. The foliation is weak to moderate and its orientation random throughout the pluton. At Mount Mortella and 3 kilometres to the west, a foliation in large rafts of schist is apparently concordant with that of the surrounding granodiorite.

The Morice Lake pluton is inferred by Woodsworth (1980) and van der Heyden (1989) to be the same age as a suite of Middle Jurassic plutons which are unroofed along the Coast-Intermontane Belt structural boundary that transects the Whitesail Lake map sheet. Intrusive relationships in the map area, however, suggest it may be as young as the tentative late Cretaceous age assigned in this report. Two samples, one foliated and the other unfoliated granodiorite, have been collected in an attempt to resolve the timing of deformation and emplacement of the Morice Lake pluton.

The stock of porphyritic monzonite (mJm) at Snowcap Peak, east of Morice Lake, is dated by the potassium-argon method and the original reported age (Carter, 1981) recalculated to 181 ± 8 Ma. This pluton is composed of subhedral plagioclase phenocrysts, averaging 5 millimetres in diameter, set within a hypidiomorphic granular aggregate of potassium feldspar. Biotite, partly altered to chlorite, varies in abundance between 3 and 7 per cent. In terms of composition and texture this pluton is distinctly different from the nearby Morice Lake pluton and they are believed to be unrelated.

Plutons of perceived Tertiary age (Tg) are distinguished by fresh mafic minerals and their unique textures. Several examples include elongate plutons northwest of Nanika Lake and at the headwaters of the Atna River. Northwest of Nanika Lake, pink, medium-grained equigranular miarolitic granite weathers recessively so that its contact with the Morice Lake pluton is a pronounced escarpment. Irregular cavities with a drusy lining of minute feldspar and quartz crystals are diagnostic of this pluton. The intrusion at Atna River is composed of granodiorite with 3 to 5 per cent pristine biotite and megacrysts of potassium feldspar between 1 and 2 centimetres in diameter. Although the rock is unfoliated, a weak foliation is developed along a subtle, steeply dipping contaet with rocks of the Morice Lake pluton. By contrast, the Morice Lake pluton in this area is a homogeneous white granodiorite with 15 per cent biotite and 5 per cent hornblende that are aligned on a pervasive moderate foliation. Small dioritic xenoliths present in the Morice Lake pluton are absent in the granodiorite containing megacrysts.

STRUCTURE

Stratified rocks comprise a homocline in which beds consistantly dip between 20° and 40° to the northeast and east. The homocline is dissected by numerous steeply dipping normal faults, particularily west of Morice Lake. Two dominant fault orientations recognized are a set trending northwest and an array trending north to northwest. The northwest faults correspond with swarms of basaltic dikes which strike 125° to 145° and dip steeply throughout the region. A complementary swarm of quartz feldspar porphyry and monzonitic dikes east of Morice Lake appears to be unrelated to mapped faults. In the southeast corner of the map area, several northwest-trending structures are segments of a larger fault system that apparently continues into the adjoining Newcombe Lake map area.

MINERAL PROSPECTS

Mineral prospects in the area can be catagorized as follows: (1) Quartz veins containing gold and silver in association with high concentrations of base metal sulphides; (2) banded copper-zinc-lead sulphides in thin lenses associated with interlayered limestone and chert; (3) iron-copper skarn associated with Tertiary granodiorite; (4) disseminated chalcopyrite, with or without molybdenite, assoeiated with Middle Jurassic and late Cretaceous(?) plutons. Locations of these mineral prospects are shown in Figure 1-9-2.

QUARTZ VEINS

Quartz veins at New Moon (MINFILE 093E 011) are the best known precious metal prospect in the map area. They are hosted by a thick succession of basaltic andesite and rhyolitic flows, and related tuffs of Unit lJH2. The veins are confined to local fractures and normal faults that dip steeply and trend north to northeast.

Quartz occurs in solitary veinlets that develop en echelon and in branehed sets of veinlets up to 3 metres wide and typically less than 50 metres long. Banded and comb textures are common, suggesting that silica-rich fluids were channelled along dilational fractures. Galena, sphalerite, minor chalcopyrite and pyrite are the typical opaque mineral assemblage. Reported assays of gold and silver indicate their concentrations vary significantly with the concentration of base metal sulphides. Calcite is present within vein quartz; epidote and chlorite are widespread in altered country rocks near the veins.

BANDED BASE METAL SULPHIDES

Boulders containing banded and massive sphalerite, galena, chalcopyrite, magnetite, pyrite and hematite are found in glacial debris at the base of an icefield about 2 kilometres south to southeast of the New Moon prospect. The sulphides are associated with thinly bedded limestone and chert, and sometimes jasperoidal chert. These rocks resemble the thin, laterally discontinuous sedimentary members that occur near the middle and upper parts of the dominantly flow-pyroclastic succession of Unit lJH1 to the south. The source of the mineralized boulders is unknown; a program of diamond drilling at the base of the icefield failed to locate mineralized rocks in place. Massive sulphide mineralization associated with chemical sedimentary rocks has not been recognized elsewhere in the Jurassic volcanic stratigraphy in the Whitesail Project area.

SKARN

Iron-rich skarn is localized in a thin layer of calcareous tuffs of Unit lJH1, 2.5 kilometres northwest of the aforementioned sulphide boulder occurrence. Magnetite, specular hematite, pyrite and minor chalcopyrite occupy a lens 1 metre thick and 35 metres long. Garnet, epidote and diopside are the main calc silicate minerals present. Dikes of granodiorite cut layered tuffs near the occurrence. These intrusions are apparently offshoots of a large Tertiary pluton exposed 1 kilometre to the south.

Porphyry Copper-Molybdenum

Pluton-related copper and pyrite, and sometimes molybdenite occur as disseminations and in fractures within gossanous volcanic rocks (lJH2) near the contact with porphyritic monzonite (mJm) east of Morice Lake. This mineralization was covered by the RD and RSM claims (MINFILE 093E 083), on the north side of Redslide Mountain. Similar mineralization occurs on the DW, CUP and CORB claims (MINFILE 093E 055), west of Nanika Lake near the southern border of the map area. The mineralized hostrocks are reported to be volcanics, however they were not examined during this study. Unfoliated granodiorite of the Morice Lake pluton (lKg) crops out immediately north and west of drill sites on the property.

SUMMARY

The Nanika Lake map area is underlain in part by a thick interlayered succession composed of basaltic to rhyolitic pyroclastic rocks and lava flows. This succession is subdivided into four lithostratigraphic units that are part of the Lower and Middle Jurassic Hazelton Group. These strata record prolonged subaerial eruptions of stratovolcanoes in an island arc environment. Jurassic volcanism culminated in a local episode of felsic eruptions that correlates with regionally distributed shallow marine deposits of the Smithers Formation.

Jurassic diorite and metamorphic rocks of the Gamsby complex are intruded by granodiorite of the Morice Lake pluton. This intrusion is tentatively assigned a Late Cretaceous age.

Mineralized propects occur in a variety of settings that include fracture-controlled precious metal veins, stratabound sulphide lenses hosted by early Jurassic strata, disseminated sulphides and rare skarn associated with Middle Jurassic and Late Cretaceous(?) intrusions.

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NOTES

GEOLOGY OF THE THAUTIL RIVER MAP AREA (93L/6)

By P. Desjardins, D.G. MacIntyre, J. Hunt, L. Lyons and S. Pattenden

KEYWORDS: Regional geology, Jurassic stratigraphy, Hazelton Group, Telkwa Formation, Telkwa Range, structure, Skeena Group, Bowser Lake Group, Ashman Formation, Smithers Formation, Thautil River sediments, Bulkley intrusions, Topley intrusions.

INTRODUCTION

This report discusses the geology and mineral occurrences of the Thautil River map area (93L/6). These observations are based on 1:50 000 mapping conducted as part of the Telkwa project (Figure 1-10-1) in 1989. The project area includes the Babine and Telkwa ranges; approximately five 1:50 000 map sheets have now been completed.

The Ashman, Skeena and Thautil River sedimentary units were studied by Julie Hunt, as part of an M.Sc thesis study at The University of British Columbia; her observations are included in this report.

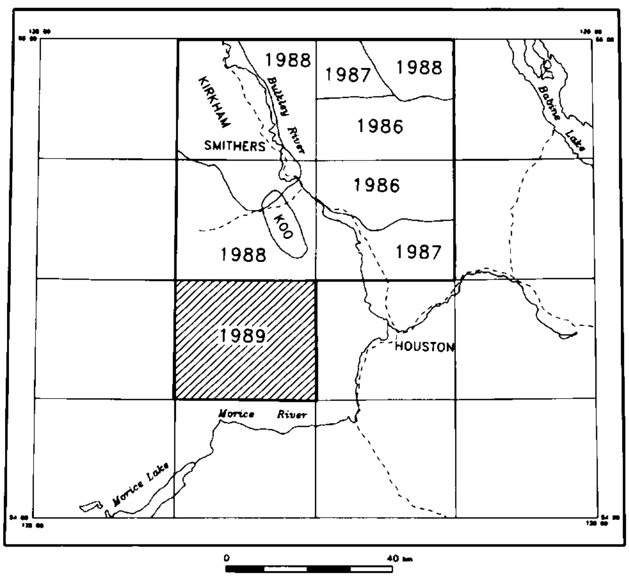

Figure 1-10-1. Location of the Thautil River map sheet, NTS 93L/6 relative to the area covered by the Babine and Telkwa projects to date.

REGIONAL GEOLOGIC SETTING

West-central British Columbia is part of the Stikine Terrane. This terrane, which is believed to have migrated northward from low paleolatitudes in late Cretaceous or early Tertiary time, includes: submarine calcalkaline to alkaline volcanic island arc rocks of the Late Triassic Takla Group; subaerial to submarine calcalkaline volcanic, volcaniclastic and sedimentary rocks of the Early to Middle Jurassic Hazelton Group; Late Jurassic and Early Cretaceous successor basin sedimentary rocks of the Bowser Lake, Skeena and Sustut groups; and Late Cretaceous to Tertiary calc-alkaline continental volcanic arc rocks of the Kasalka, Ootsa Lake and Goosly Lake groups. The younger volcanic rocks occur sporadically throughout the area, mainly in subsided fault blocks and grabens that may be the remains of cauldron subsidence complexes.

Potassium-argon isotopic dating has defined three major magmatic events. These are the Late Triassic to Early Jurassic Topley intrusions, the Middle to Late Cretaceous Bulkley intrusions and the Eocene Babine intrusions (Carter, 1981). Mineral deposits in the area are associated with emplacement of these intrusions. The most economically important exploration targets are porphyry copper and molybdenum deposits and related mesothermal and epithermal precious metal veins. A few small massive sulphide occurrences have also been discovered (MacIntyre *et al.*, 1989).

TECTONIC HISTORY

The tectonic history of the area is divisible into three distinct periods. From Early to Middle Jurassic time an extensive calcalkaline island arc evolved, with a possible back-arc basin located to the east. This was followed from late Middle Jurassic to Early Cretaceous time by development of the Bowser and Nechako successor basins. Thick deposits of molasse derived from an uplifted Skeena arch and Omineca crystalline belt accumulated within these basins. A major plate collision in Middle Cretaceous time resulted in uplift of the Coast Range and extensive folding of rocks to the east. Debris was shed eastward across the area from the rising metamorphic-plutonic complex and this was followed by the growth of a north-trending Andean-type volcanic arc in Middle to Late Cretaceous time. A transtensional tectonic regime in Late Cretaceous to Early Tertiary time produced the basin-and-range geomorphology that controls the current map pattern of the area. The latest tectonic event appears to be northeast shearing and tilting of fault blocks to the southeast. This shearing has offset northwest-trending grabens that developed in Late Cretaceous to Early Tertiary time (MacIntyre *et al.*, 1989).

LITHOLOGY OF THE STUDY AREA

The generalized geology of the Thautil River area is shown in Figure 1-10-2. The map area is underlain by a series of uplifted and tilted fault blocks containing rocks ranging from Early Jurassic to Eocene in age. In general the fault blocks are tilted to the southwest (Figure 1-10-3). The youngest rocks are probably Tertiary and are preserved within a north-trending graben which follows the Thautil River. Cretaceous rocks occur as downthrown blocks in the centre and the southeast part of the map area. A generalized stratigraphic column is shown in Figure 1-10-4.

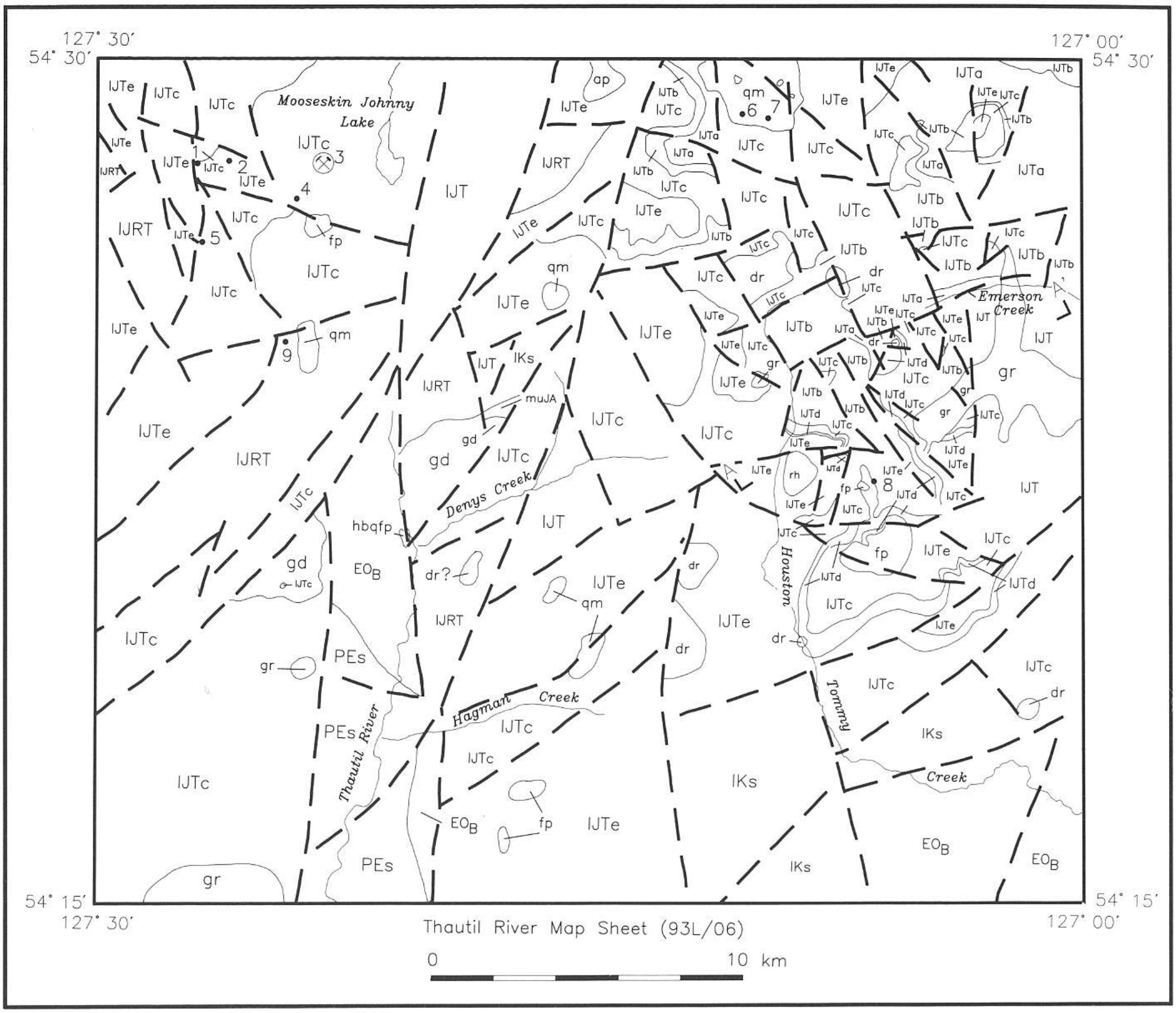

Figure 1-10-2. General geology of the Thautil River area, NTS 93L/6.

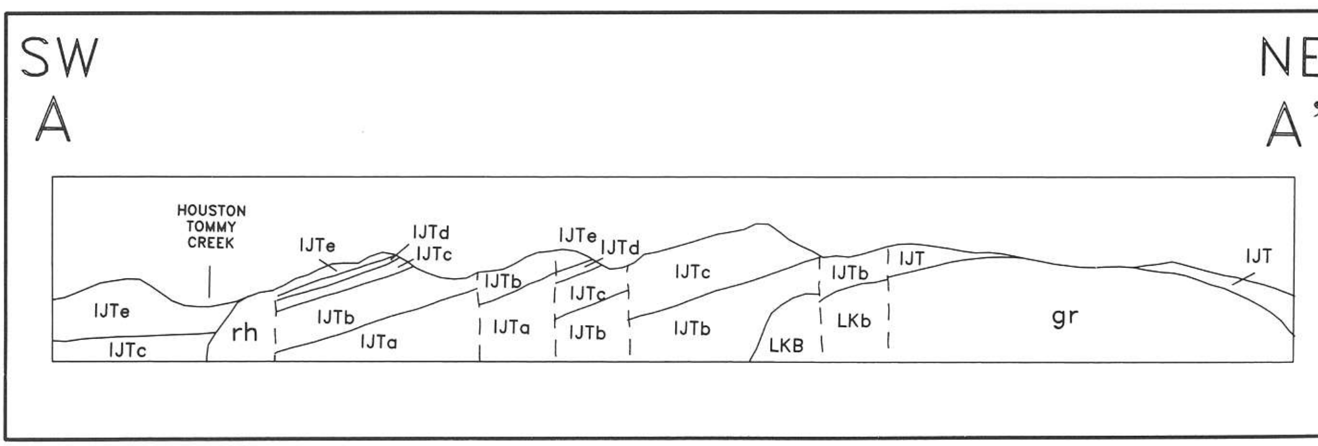

Figure 1-10-3. Cross-section through the Emerson Creek–Houston Tommy Creek area showing typical structural style of the map area. See Figure 1-10-2 for location of section.

LAYERED ROCKS

PALEOCENE TO MIOCENE

EOB — *basalt flows, breccia*

PEs — *Thautil River Sediments: heterolithic, poorly sorted conglomerate, sandstone, siltstone, minor coal*

LOWER CRETACEOUS (ALBIAN)

SKEENA GROUP

IKS — *sandstone, siltstone, shale, micaceous greywacke, coal bearing*

JURASSIC

BOWSER LAKE GROUP

MIDDLE TO UPPER JURASSIC (CALLOVIAN TO OXFORDIAN)

muJA — *Ashman Formation: marine black shale, siltstone, greywacke; fossiliferous*

HAZELTON GROUP

MIDDLE JURASSIC (BAJOCIAN TO CALLOVIAN)

mJS — *Smithers Formation: feldspathic sandstone, greywacke, siltstone, shale, minor pebble conglomerate, very fossiliferous*

LOWER JURASSIC (PLEINSBACHIAN TO TOARCIAN)

IJRT — *Nilkitkwa Formation, Red Tuff Member: red, well-bedded air fall tuff, minor ash flow tuff*

LOWER JURASSIC (SINEMURIAN TO LOWER PLEINSBACHIAN)

IJT — *Telkwa Formation: undivided andesite, dacite, rhyolite, basalt, flows and pyroclastics*

- **IJTe** — *Basalt-Red Tuff Unit: well-bedded, recessive, brick red air fall tuffs and related epiclastics with sporadic amygdaloidal basalt flows; minor grey welded tuff*
- **IJTd** — *Shallow Marine Sedimentary Unit: well-bedded limestone, calcareous sandstone, siltstone; interbedded with epiclastics and air fall tuff; fossiliferous*
- **IJTc** — *Siliceous Pyroclastic Unit: well bedded quartz-feldspar phyric ash flows, ignimbrite, breccia, siliceous air fall tuff, red tuff, basalt, rhyolite flows.*
- **IJTb** — *Basaltic Flow Unit: massive maroon to green augite-feldspar phyric to aphyric basalt flows; minor maroon tuff between flows; flow top breccia common; locally amygdaloidal*
- **IJTa** — *Andesitic Pyroclastic Unit: andesitic air fall tuff, breccia, feldspathic epiclastics, minor welded lapilli tuff*

INTRUSIVE ROCKS

LATE CRETACEOUS TO EOCENE

KEg — *gr - undivided granitic intrusions; gd - granodiorite; qd - quartz dioritie; dr - diorite; rh - rhyolite; fp - feldspar porphyry; bfp - biotite-feldspar porphyry; hfp - hornblende-biotite-feldspar porphyry; ap - augite feldspar porphyry; qm - quartz monzonite*

EARLY JURASSIC

EJT — *Topley Intrusions: undivided granitic intrusions*

Occurrence	Name	Minfile No.
1	Duchess	066
2	Evening	064
3	Santa Maria	063
4	War Eagle	062
5	Princess	061
6	Sunsets Creek Showing	045
7	Sunsets Creek Showing	046
8	Erin Claims	---

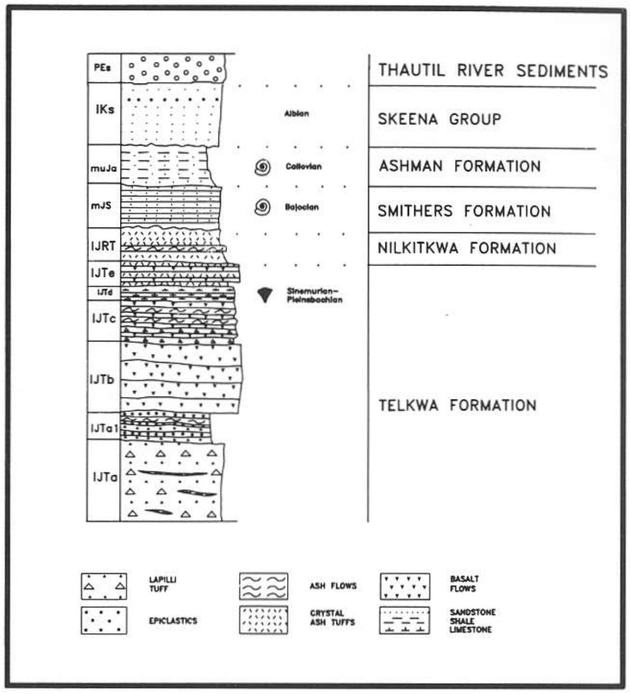

Figure 1-10-4. Generalized stratigraphic column for the Thautil River map area. See legend Figure 1-10-2 for explanation of symbols.

Hazelton Group

The Hazelton Group (Leach, 1910) is a calcalkaline island arc assemblage that evolved in Early to Middle Jurassic time. Tipper and Richards (1976) divide the group into three major formations in the Smithers map area (93L). These are the Sinemurian to early Pliensbachian Telkwa Formation, the early Pliensbachian to middle Toarcian Nilkitkwa Formation and the middle Toarcian to early Callovian Smithers Formation. For simplicity's sake we have adopted this terminology. However, our mapping in the type area of the Telkwa Formation suggests there are a number of lithostratigraphic map units that could stand alone as formations. Consideration should be given to raising the Telkwa Formation to group status and the Hazelton Group to supergroup.

TELKWA FORMATION

In the Telkwa Range, a thick section of Early Jurassic volcanic rocks constitutes the type area for the Telkwa Formation of the Hazelton Group. Tipper and Richards (1976) describe the typical lithology as "reddish, maroon, purple, grey and green pyroclastic and flow rocks". The formation varies from marine to nonmarine and ranges from Sinemurian to early Pliensbachian in age.

The Lower Jurassic volcanics in the study area are predominantly air-fall tuffs and basalt flows and constitute the Howson subaerial facies of the Telkwa Formation. The transition between this facies and the Babine shelf facies to the east trends through the map area and is defined by the western limit of late Sinemurian marine sedimentary rocks.

Our mapping in the Telkwa Range suggests the Telkwa Formation is divisible into five major lithostratigraphic units, each representing distinct cycles of arc volcanism. These units are characterized by their predominant lithologies although internal facies variations are common. In ascending stratigraphic order they are: (1) an andesitic pyroclastic unit comprised of thick-bedded, massive, maroon andesitic lapilli, crystal and ash tuffs with minor interbeds of siliceous banded ash flows and grey welded lapilli tuffs; (2) a flow unit which is predominantly massive, cliff forming, augite-feldspar-phyric to aphyric, dark green to maroon basalt; (3) a siliceous pyroclastic unit that is well bedded and includes felsic epiclastics, welded lapilli tuffs, flow-banded ash flows, feldspathic breccias, pebble conglomerates, lahars, sandstones, air-fall lapilli and crystal-lithic tuffs; (4) a shallow marine sedimentary unit characterized by well-bedded, near-shore, fossiliferous limy sandstone, siltstone, conglomerate, grey bioclastic and massive limestone; and (5) a recessive unit of well-bedded and maroon to red crystal, lapilli and ash tuff and associated epiclastics with lesser flows of amygdaloidal augite-feldspar-phyric basalt. A similar stratigraphic sequence underlies the Middle Jurassic Smithers and Ashman formations on Ashman Ridge, 60 kilometres northwest of the map area.

Andesitic Pyroclastic Unit (lJTa)

Andesitic pyroclastics and flows are the predominant lithologies of the lower part of the Howson facies of the Telkwa Formation. These rocks are well exposed in valleys along the east edge of the map area where the exposed thickness is in excess of 1000 metres. This unit is predominantly thick-bedded, massive maroon andesitic lapilli, crystal and ash tuff, volcanic breccia and associated epiclastics. Minor grey ash flows occur throughout the section. Maroon and green volcanic clasts predominate in a feldspathic matrix. Toward the top of the section the unit is characterized by thin beds of maroon, feldspar-phyric andesite flows, grey welded lapilli tuffs, rubbly weathering volcanic conglomerates and feldspathic sandstone, siliceous, banded ash-flow tuffs and grey welded lapilli tuffs.

Tipper and Richards (1976) describe a measured section from the Loljuh Creek area that we believe transects the upper contact of this unit. We would place the boundary between this unit and the overlying basaltic flow unit at the base of the first amygdaloidal basalt flow.

The upper contact of Unit lJTa is also exposed in the northeast corner of the map area. Here, as elsewhere, it is overlain by massive basalt flows. Unit lJTa also underlies the Howson Creek headwaters but here the top of the unit is not exposed.

Nowhere in the map area is the base of the Telkwa Formation exposed. However, a basal conglomerate has been recognized in the Babine Range (MacIntyre *et al.*, 1986). The andesitic pyroclastic unit is probably in excess of 1000 metres thick.

Basaltic Flow Unit (lJTb)

A distinctive unit of massive basalt flows with minor interflow maroon tuff overlies the andesitic pyroclastic package. This unit, which is up to 500 metres thick, forms the prominent cliffs and higher peaks of the Telkwa Range, particularly at the headwaters of Emerson and Dockrill creeks. It becomes progressively thinner to the south and southwest of Emerson Creek.

The basaltic flows that characterize the unit are fine to medium grained, dark green to maroon and aphyric to augite-feldspar phyric. Flow tops are typically amygdaloidal and brecciated with zeolite-calcite-epidote cement. Maroon crystal and lapilli tuff beds occur between flows.

Orientation of the basalt flows varies somewhat, possibly due to the influence of paleotopography. In general the flows dip moderately to the southwest. This unit is faulted off to the west and is missing in the area near the headwaters of Howson Creek and north of Starr Creek; it is restricted to the east and northeast part of the map area.

Siliceous Pyroclastic Unit (lJTc)

A well-bedded unit of siliceous ash flows with lesser amygdaloidal basalt flows, epiclastics and air-fall tuffs overlies the massive basalts. This unit, which varies from less than 10 to greater than 200 metres thick, is typically well bedded and consists of a mixture of maroon, pink and grey welded lapilli tuffs, flow-banded ash flows, minor green and maroon feldspathic volcanic-clast breccias, pebble conglomerate, lahars and volcanic sandstone. A channel filled with coarse volcanic debris cuts into well-bedded tuffs just south of Sunsets Creek. It is approximately 150 metres wide and trends perpendicular to the cliff south of the Creek.

The siliceous pyroclastic unit is widespread within the Thautil River map area where it varies from very thin or absent in the northeast corner to several hundred metres thiek in the core of the Telkwa and Howson ranges. In adjacent map areas this unit is only locally present. These thickness variations may reflect proximity to eruptive centres. Figure 1-10-5 illustrates the thickness variations of the major lithostratigraphic units perpendicular to the trend of the Telkwa Range.

Shallow Marine Sedimentary Unit (lJTd)

Well-bedded marine sediments and red tuffs overlie and are in part interbedded with volcanics of Unit lJTc. This unit, which varies from 5 to 100 metres thick, includes grey bioclastic to massive limestone and well-bedded near-shore

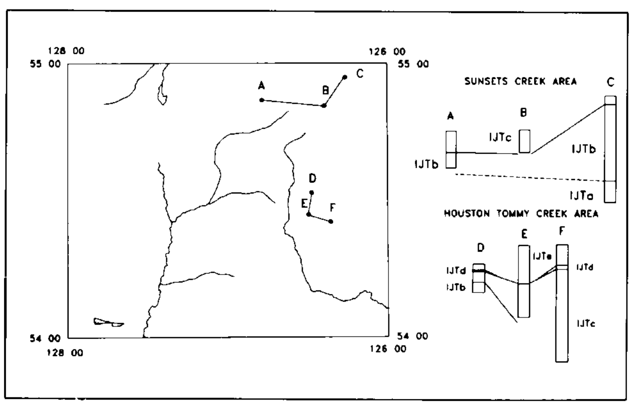

Figure 1-10-5. Stratigraphic sections illustrating unit thickness variations across the Thautil River map area. See legend Figure 1-10-2 for explanation of symbols.

limy sandstones, siltstones, conglomerates and maroon epiclastics. It is typically fossiliferous and is the only good marker bed in the Telkwa Range. Fossils include the late Sinemurian pelecypod *Weyla*. A small coral bioherm discovered by T.A. Richards and H.W. Tipper while mapping the Smithers area in 1971 to 1973 occurs just east of Houston Tommy Creek (Plate 1-10-1). The reef sits at the same stratigraphic position as the well-bedded late Sinemurian marine sediments. Dr. Terry Poulton of the Geological Survey of Canada, recently publish a description of the reef (Poulton, 1989) which is believed to be the only true Lower Jurassic bioherm reef in North America and perhaps the world. Dr. George Stanley, professor at the University of Montana and an expert on ancient reefs, examined the reef in detail during the 1989 field season.

The Sinemurian marine sediments are confined to the southeast part of the map area, particularly east of Houston Tommy Creek, where the most northerly occurrence is just south of the headwaters of Emerson Creek and extends over 9 kilometres to the south. These sediments are probably correlative with those exposed in the area east of Webster Creek as discussed in a previous report (MacIntyre *et al.*, 1989).

Plate 1-10-1. View looking west toward Houston Tommy Creek. White-weathering unit is a Lower Jurassic biohermal reef within Unit lJTd of the Telkwa Formation.

BASALT RED TUFF UNIT (lJTe)

Overlying and in part interbedded with the marine sedimentary unit is a poorly exposed unit of tuffs, epiclastics and minor flows of variable thickness. This unit is characterized by medium to thin-bedded, brick-red to maroon crystal, lapilli and air-fall ash tuffs with local green amygdaloidal basalt flows, grey welded lapilli tuffs and associated epiclastics. Graded bedding and crossbedding are common in tuffs and epiclastics. Tuff beds are recessive and often badly decomposed.

The red tuff–basalt flow unit occurs mainly within the western half of the map area where it stratigraphically overlies and is in part interfingered with the marine sedimentary or siliceous pyroclastic units. A lateral facies change occurs south of Emerson Creek where a section of epiclastics, lithic red tuffs and minor basalt flows grades into mainly basalts and minor limestone over an 8-kilometre strike length.

Basalt flows are most common near the base of the unit; up-section it is predominantly brick-red air-fall tuff and epiclastics. In some areas this unit has been mapped by Tipper and Richards (1976) as the Red Tuff member of the Nilkitkwa Formation; elsewhere it has been included with the Telkwa Formation. In this study the Red Tuff member is restricted to well-bedded red tuffs and epiclastics that lack amygdaloidal basalt flows.

AGE OF THE TELKWA FORMATION

The Telkwa Formation is Sinemurian or older based on sporadic fossil fauna within the stratigraphic succession (Tipper and Richards, 1976). In the map area, which covers part of the Skeena arch, the granitic Topley intrusions cut the formation. These intrusions give K-Ar ages between 195 and 205 Ma (Carter, 1981). This is further evidence that the Telkwa Formation is predominantly Sinemurian or older (MacIntyre *et al.*, 1989). Fossils collected from marine sedimentary rocks near the top of the formation are late Sinemurian to early Pliensbachian in age.

NILKITKWA FORMATION – RED TUFF MEMBER

The Red Tuff member of the Nilkitkwa Formation is comprised of medium to thin-bedded red to maroon ash, crystal and lapilli air-fall tuff and related epiclastic rocks, with subordinate beds of grey ash-flow tuff. It is exposed on the ridges south of Sunsets Creek and just west of Denys Creek. It appears to be conformable with underlying strata of the Telkwa Formation.

MIDDLE TO LATE JURASSIC SEDIMENTARY ROCKS

Heterolithic conglomerate and coarse-grained sandstone containing Jurassic ammonites and bivalves crops out on the western edge of the Thautil River map area near Denys Creek. The conglomerate is matrix supported with approximately 30 per cent subangular clasts which vary from 5 to 10 centimetres in diameter. Clasts are primarily fine-grained purple, grey or flow-banded volcanics. The matrix is composed of coarse-grained, poorly sorted, angular sand and pebbles up to 10 millimetres long and averaging 5 millimetres long. Fossiliferous feldspathic sandstone underlies the conglomerate. Sandstone contains bivalves, ammonites and trace fossils together with carbonized wood fragments and minor amounts of mica. These rocks underlie Cretaceous Skeena Group sediments and therefore are probably correlative to the Ashman Formation as defined by Tipper and Richards (1976). However, lithologically these rocks are similar to the shallow-marine sediments of the Smithers Formation.

SKEENA GROUP

Cretaceous Skeena Group sediments crop out in Houston Tommy Creek and near Denys Creek. Sedimentary strata consist of marine, transitional marine and nonmarine interbedded sandstone, siltstone, shale and coal. Cretaceous sediments, approximately 60 metres thick in total, occur in the Denys Creek area. They grade upward from medium-grained sandstones to interbedded siltstone, sandstone and coal. Sandstone is well sorted, pale grey-green and micaceous, weathering to a dark green or rusty brown colour. Sandstone is thinly bedded with individual beds varying from 0.5 to 2 centimetres thick. Beds are planar-laminated and often contain carbonized plant fragments and discontinuous siltstone lenses, especially along bedding planes. Overlying the medium-grained sandstone is an interbedded sequence of 60 per cent siltstone, 40 per cent sandstone. Here the sandstone is fine grained and beds vary from 2 to 4 centimetres thick. The siltstone is fine grained and dark grey, weathering to a rusty brown. It is also micaceous, carbonaceous and thinly bedded, with beds varying from 1 to 2 centimetres thick. The interbedded sequence contain two(?) coal seams. The lower seam is approximately 0.5 metre thiek and highly deformed by numerous fractures and tiny folds. The upper seam outcrops at a number of places and appears to be similar to the lower seam.

In Houston Tommy Creek, Skeena Group sediments are approximately 50 metres thick and in fault contact with Hazelton Group volcanics. The lower 20 metres of section is made up of interbedded sandstone and shale. Sandstone varies from fine grained, pale grey and well sorted to pale grey, poorly sorted pebbly sandstone with angular clasts up to 3 centimetres in diameter, composed of chert and pale pink or maroon feldspathic volcanics. Shale is fine grained and dark grey. The upper 30 metres of the section is composed of shale with minor sandstone beds. Shale is fine grained, dark grey-brown and weathers to a pale grey or rust colour. Shale beds average 3 metres thick. Sandstone occurs in thin beds approximately 0.3 metre thick and is seen primarily as concretions(?) up to 40 centimetres in diameter. Banded dull coal occurs in a seam approximately 0.5 metre thick, with approximately 85 per cent dull and 15 per cent bright bands. Bright bands average 1 centimetre thick and dull bands vary from 1.5 to 3 centimetres in thickness.

THAUTIL RIVER SEDIMENTS

A distinctive unit of clastic sedimentary rocks crops out in the Thautil River valley. Here the strata consist of friable conglomerate, sandstone, siltstone and minor coal. These rocks are distinct from the Early Cretaceous Skeena Group rocks and are believed to be Late Cretaceous or Tertiary in age. The predominant lithology is heterolithic, clast-supported, poorly sorted conglomerate. Clasts are subangular to subrounded and vary from 5 to 20 centimetres in diameter. Approximately 60 per cent of clasts are fine-grained or amygdaloidal basalt, 20 per cent are maroon volcanics, 10 per cent are augite porphyry and 10 per cent are chert and sandstone. The conglomerate matrix is coarse grained, poorly sorted muddy sandstone.

The finer sediments, sandstone, siltstone and coal, occur as beds or lenses within the conglomerate. Sandstone varies greatly along the length of Thautil River. In the southern part of the river, friable, fine grained, grey, poorly sorted feldspathic sandstone containing carbonized material is interbedded with very fine grained, dark grey micaceous siltstone and coal. Coal is black, glassy and very friable, occurring as laminations, blebs and thin beds approximately 3 millimetres thick within the siltstone.

In the central reaches of Thautil River interbedded sandstone and siltstone occur as small lenses within the conglomerate. Sandstone is medium grained, pale grey, well sorted and lithified. Siltstone is fine grained, grey and friable. Both sandstone and siltstone contain carbonaceous material. Farther up stream sandstone is coarse grained, dark green-brown, micaceous, well sorted and poorly lithified. Here sandstone is thinly bedded with individual beds approximately 1 centimetre thick. Still farther upstream conglomerate is overlain by sandstone. This sandstone is medium grained, light grey-brown, well sorted, fairly well lithified and contains approximately 2 per cent magnetite. The sandstone has been brecciated to form a very angular homolithic breccia in which average fragment diameter is 5 centimetres. In the northern part of Thautil River the sediments are cut by amygdaloidal basalt dikes and eventually come into contact with fine-grained, black, amygdaloidal basalt farther upstream. This basalt is also mapped as Tertiary.

The age of the sediments exposed in the Thautil River graben is not well established, but they are most likely younger than the Skeena Group. Tipper and Richards (1976) assigned a Paleocene to Eocene age, based on palynology dates from samples taken by Dr. W.S. Hopkins for the Geological Survey of Canada, Calgary (H.W. Tipper, personal communication, 1989).

INTRUSIVE ROCKS

Quartz monzonite (qm), granodiorite (gd), diorite (dr), granite (gr), feldspar porphyry (fp), hornblende-biotite-quartz-feldspar porphyry (hbqfp), augite-feldspar andesite porphyry (ap), and rhyolite (rh) intrusions are recognized within the Thautil River map area.

Two quartz monzonite plutons intrude Lower Jurassic stratigraphy in the Telkwa Range. The largest of these is exposed just north of Sunsets Creek and a smaller stock underlies the area north of Hagman Creek. These intrusions are medium to coarse grained, equigranular to porphyritic with a feldspar-rich groundmass.

There are many small feldspar and augite-feldspar porphyry plutons in the northwest, west and southeast parts of the map sheet. The largest occurs east of Mooseskin Johnny Lake; smaller intrusions occur southwest of the lake, just north of Starr Creek and east of Thautil River. These are typically medium to coarse grained, grey, commonly with potassium feldspar phenocrysts. The matrix is fine grained with locally minor amounts of accessory magnetite. Many are pyritic and gossanous and are cut by basaltic dikes and shear zones. Clay and sericite alteration occurs locally.

Diorite and granodiorite-diorite intrusions are common in the area east of Howson and Starr creeks. These rocks are medium to coarse grained and equigranular. In some areas the mafic minerals are altered to chlorite. Minor amounts of magnetite and pyrite are locally present. Most intrusions are subcircular stocks; a thick dioritic sill occurs west of the headwaters of Houston Tommy Creek. A pluton south of the Holland Lakes has both a granodiorite and monzonite phase.

A plug of siliceous feldspar porphyry with a fine grained, flow-banded groundmass occurs east of Houston Tommy Creek. It is the only rhyolitic intrusion exposed in the map area.

A dark maroon hornblende-biotite-quartz feldspar porphyry intrusion with a fine-grained groundmass is exposed along the Thautil River near the mouth of Denys Creek. It is most likely part of the Eocene Babine plutonic suite.

A large granitic intrusion underlies the area south of Emerson Creek. Its upper contact appears to be relatively flat. Overlying rocks are silicified and pyritized and contain minor chalcopyrite.

STRUCTURAL STYLE

The structural style of the Thautil River map area is typical of the Telkwa Range as described in a previous report (MacIntyre *et al.*, 1989). Extensive block faulting in Tertiary time has produced a typical basin-and-range geomorphology. Within the study area, fault blocks are generally tilted to the southwest with progressively higher stratigraphic levels exposed in this direction. Locally beds have been domed around major intrusions such as the Sunsets Creek pluton. Elsewhere, such as near the headwaters of Howson Creek, fault blocks have been rotated, resulting in divergent bedding attitudes between adjacent fault blocks. Such rotations may also be due to emplacement of intrusive bodies and later modification by Tertiary block faulting.

MINERAL DEPOSITS

As discussed in a previous report (MacIntyre *et al.*, 1987), mineral deposits in the Smithers area can be subdivided into four groups: (1) mesothermal and epithermal gold-silver-bearing quartz veins; (2) copper-silver veins and pods in mafic volcanic rocks; (3) copper-zinc-silver massive sulphide deposits associated with bimodal submarine volcanics; and (4) porphyry copper-molybdenum deposits associated with quartz monzonite to granodiorite intrusions.

The showings on the Thautil River map sheet are mostly Type 2 copper-silver veins and pods in mafic volcanic rocks which may in part be associated with porphyry copper mineralization at depth and Type 4 porphyry copper-molybdenum deposits.

The preferred host rocks for copper-silver occurrences, as elsewhere in the region, are the amygdaloidal basalt flows of the upper Telkwa Formation. Intense epidote-calcite-chlorite alteration is often associated with this type of occurrence.

Several showings occur in the area near the headwaters of Howson Creek. The country rocks are flows and pyroclastics of Units lJTa, lJRT and lJTc of the Telkwa Formation; numerous basic and acid dikes cut the volcanics.

DUCHESS (MINFILE 93L 066)

The Duchess showing consists of a northerly trending shear zone mineralized with chalcopyrite, pyrite, hematite, and quartz containing tetrahedrite. The mineralized zone is up to 4 metres wide. The shear zone lies near the contact between fine-grained green epidotized andesite to the west and fine-grained purplish brown to olive-brown tuff to the east. Highly broken and sheared, buff-coloured feldspar porphyry dikes cut the volcanic rocks and carry only very minor amounts of sulphides. North-trending faults cut this area and mineralization is spatially associated with them. These fractures may have served as conduits for hydrothermal fluids as indicated by the formation of skarn along and near the main fault and by the concordance in attitude between the fault and the mineralized shear zones. Post-mineral movement is indicated by a northwest-trending and southwest-dipping fault that truncates the main lode. A northeast-trending fault which dips 45° to the southeast is exposed a few metres west of the portal and probably represents a set of fractures which displace the mineralized veins. A lower adit was driven on the supposed location of the vein as projected from above but it and several crosscuts failed to intersect the lode.

EVENING (MINFILE 93L 064)

The Evening showing is characterized by irregular shears and quartz-sulphide veins in strongly epidotized and chloritized buff, reddish, and green fine-grained andesitic tuffs and flows of Unit lJTa of the Telkwa Formation. The mineralized shears trend northeast to east and dip at moderate angles to the north.

SANTA MARIA (MINFILE 93L 063)

The Santa Maria showing is a northwest-trending, southwest-dipping quartz vein carrying chalcopyrite, pyrite, chalcocite and bornite. The quartz vein follows a zone of intensely altered and sheared pyroclastic rocks of Unit lJTc of the Telkwa Formation. Buff, cream and brick-red rhyolite porphyry dikes subparallel the vein structure. Saussurite alteration, sparsely distributed silicification, and sulphide mineralization occur on both sides of the shearing.

WAR EAGLE (MINFILE 93L 062)

Easterly trending feldspar porphyry dikes, some intensely epidotized and sparsely mineralized, cut volcanic rocks at the War Eagle showing. Narrow northeasterly and northwesterly trending shear zones are mineralized with pyrite, chalcopyrite, hematite, bornite and locally low-iron sphalerite.

PRINCESS (MINFILE 93L 061)

The Princess showing is characterized by narrow mineralized shear zones and veinlets of hematite, iron-rich sphalerite and chalcopyrite in a gangue of white calcite and quartz. The hostrock is strongly sheared and epidotized fine-grained greenish flows of Unit lJTc of the Telkwa Formation (Preto, 1967).

SUNSETS CREEK SHOWINGS (MINFILE 93L 045, 46)

A large pluton occurs in the centre of the Telkwa Range north of Sunsets Creek. It has domed the surrounding pyroclastic rocks of the Telkwa Formation which dip away from the pluton in all directions. The rocks are predominantly lapilli to fine-grained tuffs. Dark greenish grey tuffs grade laterally into maroon tuffs outside the limits of hornfelsing. Close to the pluton the rocks have been altered to mosaics of secondary plagioclase, actinolite, quartz, opaque minerals and garnet. Hornfels grades outward in a concentric pattern through a biotite zone to a chlorite zone.

The pluton is a homogeneous body composed entirely of porphyritic quartz monzonite of nearly constant composition and texture. A pyritic gossan surrounds the stock and roughly corresponds to the area of intense hornfelsing. Pyrite is sparse yet the rocks weather an intense rusty colour. Widely spaced, banded, drusy quartz veins up to 2 centimetres wide occur within the pluton. The veins contain pyrite, chalcopyrite and minor molybdenite. In the hornfels, near the southwestern contact, there are isolated concentrations of chalcopyrite, pyrite and specular hematite in some garnet-epidote skarn bands that parallel bedding.

In the interior of the stock there are two altered zones associated with mineralization. The larger one to the west consists of a broad crescentic area, measuring 700 by 1000 metres, in which all rocks are abnormally pyritic. The pyrite is disseminated and also occurs as coatings on joints and irregular fractures. Chloritization of biotite and hornblende and a sericitization of feldspars are associated with pyrite. The core of the crescent is a zone of much more intense alteration with some rocks locally converted to aggregates of quartz, muscovite and pyrite. Molybdenite mineralization occurs in part of the southern arm of the crescent. Quartz veinlets in the joints are common and might be considered to be a wide-spaced stockwork. Veinlets and small faults and fractures contain abundant pyrite and molybdenite. The better mineralized joints are fairly flat gently dipping. The hostrocks are granitic and show some evidence of recrystallization and alteration of feldspars (Sutherland Brown, 1967)

MSJ Prospect (Minfile 93L 241)

A quartz monzonite porphyry intrusion north of Starr Creek is gossanous and mineralized with pyrite and variable amounts of tenorite, malachite, chalcopyrite and chalcocite. Mineralization is both disseminated and coatings on fracture planes. Showings to the north may represent halo mineralization surrounding the MSJ porphyry prospect. A small aeromagnetic low coincides with the area of most intense hydrothermal alteration. Hostrocks are mainly andesitic crystal-lapilli tuff and minor welded tuff of Unit lJRT. The country rock is cut by basalt, feldspar porphyry and feldspar quartz porphyry dikes and sills.

Erin Claims (Minfile 93L 227)

The Erin claims are located east of Houston Tommy Creek. Trenches on the central plateau, east of the creek, expose mineralized veins as well as pervasive manganese staining and minor rhodochrosite in basalt over a wide area. Copper mineralization is hosted within narrow veins cutting amygdaloidal basalts of Unit lJTc and carbonate-rich volcaniclastics of Unit lJTd. The carbonate-rich volcaniclastics locally contain various bivalves and ammonites. The veins carry massive bornite and chalcopyrite with trace amounts of chalcocite and tetrahedrite. Malachite is a prominent weathering product.

Mineralization appears to be related to an intrusive body which fractured the country rock during emplacement. These fractures host the copper mineralization. Podiform mineralization of chalcopyrite and specular hematite are associated with the carbonite-rich volcaniclastics close to the intrusive. Alteration is locally present as patchy epidote in basalt with or without irregular quartz and carbonate veinlets.

DISCUSSION

The Telkwa stratigraphic succession indicates that Early Jurassic volcanism began with widespread deposition of predominantly andesitic air-fall pyroclastic material on a post-Triassic erosion surface. This volcanic cycle was followed by a short period of erosion and deposition of epiclastic beds prior to widespread eruption of basaltic lava and deposition of air-fall tuff. Eruption of lavas gradually diminished and was followed by explosive volcanism and deposition of well-bedded ash flows, air-fall tuffs, and related epiclastics with thick accumulations occurring near major vents.

Marine sedimentary strata containing late Sinemurian to early Pliensbachian fossils onlap the Early Jurassic volcanics. Biohermal reefs developed locally. These reefs may have formed chains around volcanic islands. Shallow-marine sediments are interbedded with air-fall tuffs and associated epiclastic rocks, indicating active volcanism and erosion accompanied the marine transgression.

The final volcanic-sedimentary cycle began with alternating eruptions of amygdaloidal basalt flows and deposition of red air-fall tuffs and grey felsic ash flows. The red tuff beds are onlapped by the Bajocian to Callovian marine sediments of the Smithers and Ashman formations.

Early Jurassic plutons are exhumed along the axis of the Skeena arch and in the Howson Ranges. These plutons probably formed at depth beneath major Early Jurassic eruptive centres.

SUMMARY

The major conclusions from fieldwork completed in 1989 are:

- The Telkwa Formation is divisible into five lithostratigraphic units in the map area. In ascending stratigraphic order they are andesitic pyroclastics; basalt flows and minor tuffs; siliceous pyroclastics, epiclastics, and air-fall tuffs; well-bedded shallow-marine sediments, and basalt flows and tuffs.
- Lower Jurassic volcanics of the Telkwa Range are proximal to eruptive centres.
- The Early Jurassic volcanics of the Telkwa River area are relatively undeformed and unaltered. Uplift and tilting of fault blocks rather than folding and thrust faulting are the predominate structural styles.
- Mineralization is associated with emplacement of granitic plutons.

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NOTES

GEOLOGY BETWEEN NINA LAKE AND OSILINKA RIVER, NORTH-CENTRAL BRITISH COLUMBIA (93N/15, NORTH HALF AND 94C/2, SOUTH HALF)

By Filippo Ferri and David M. Melville

KEYWORDS: Regional mapping, Germansen Landing, Omineca Belt, Intermontane Superterrane, Slide Mountain Group, Paleozoic stratigraphy, Ingenika Group, metamorphism, lead-zinc-silver-barite mineralization

INTRODUCTION

In 1989 the Manson Creek 1:50 000 mapping project encompassed the north half of the Germansen Landing map area (93N/15) and the south half of the End Lake map area (94C/2). As with previous years, the main aims of this project were: to provide a detailed geological base map of the area, to update the mineral inventory database, and to place known mineral occurrences within a geological framework.

The centre of the map area is located some 260 kilometres north-northwest of Prince George, immediately north of the settlement of Germansen Landing (Figure 1-11-1). Primary access is via all-season gravel roads from Fort St. James or Mackenzie which connect to secondary roads along Nina Creek, Nina Lake and the Osilinka drainage. A four-wheel-drive road along Nina Lake provides access to lead-zinc-silver showings northwest of Echo Lake. A major logging road along the Osilinka River services secondary logging roads providing access to the northern third of the map area. The remainder is reached on foot or by helicopter.

The northern and eastern sections of the map are bounded, respectively, by the Osilinka and Omineca rivers. At their confluence the terrain is a relatively subdued and tree covered area. This is in contrast to the Wolverine Range east of the Omineca River and a rugged, unnamed range of mountains to the southwest.

The southern part of the map area was first mapped at a 6-mile scale in the 1940s by Armstrong (1949). Gabrielse (1975) examined the northern half in the course of 1:250 000 mapping of the east half of the Fort Grahame map area. Monger (1973), and Monger and Paterson (1974) described rocks in the map area during a reconnaissance survey of Paleozoic stratigraphy. To the northwest, Roots (1954) published a 4-mile map of the Fort Graham west-half sheet. Many of the correlations made in this paper are with stratigraphy described by Gabrielse (1963, 1969), Nelson and Bradford (1987) and other workers in the Cassiar area where the miogeoclinal stratigraphy is quite similar and well known.

REGIONAL GEOLOGY

The map area lies along the western edge of the Omineca Belt, one of the five morphogeological belts of the Canadian Cordillera (Wheeler and McFeely, 1987). This area contains rocks which are part of the Intermontane Superterrane (*i.e.* accreted) and displaced North American rocks (Wheeler and McFeely, *ibid*; Figure 1-11-1). Rocks of the Foreland Belt lie

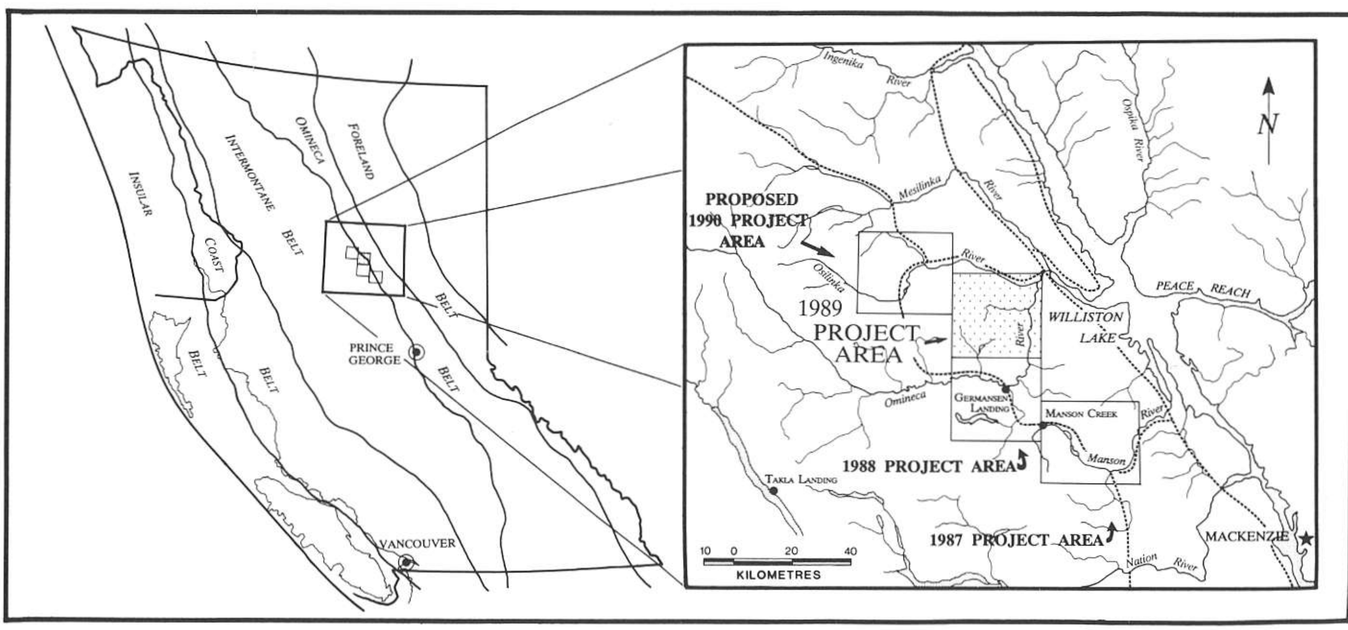


Figure 1-11-1. Location of the map area with respect to the five morphogeological provinces of the Canadian Cordillera with an expanded view in the right half of the diagram.

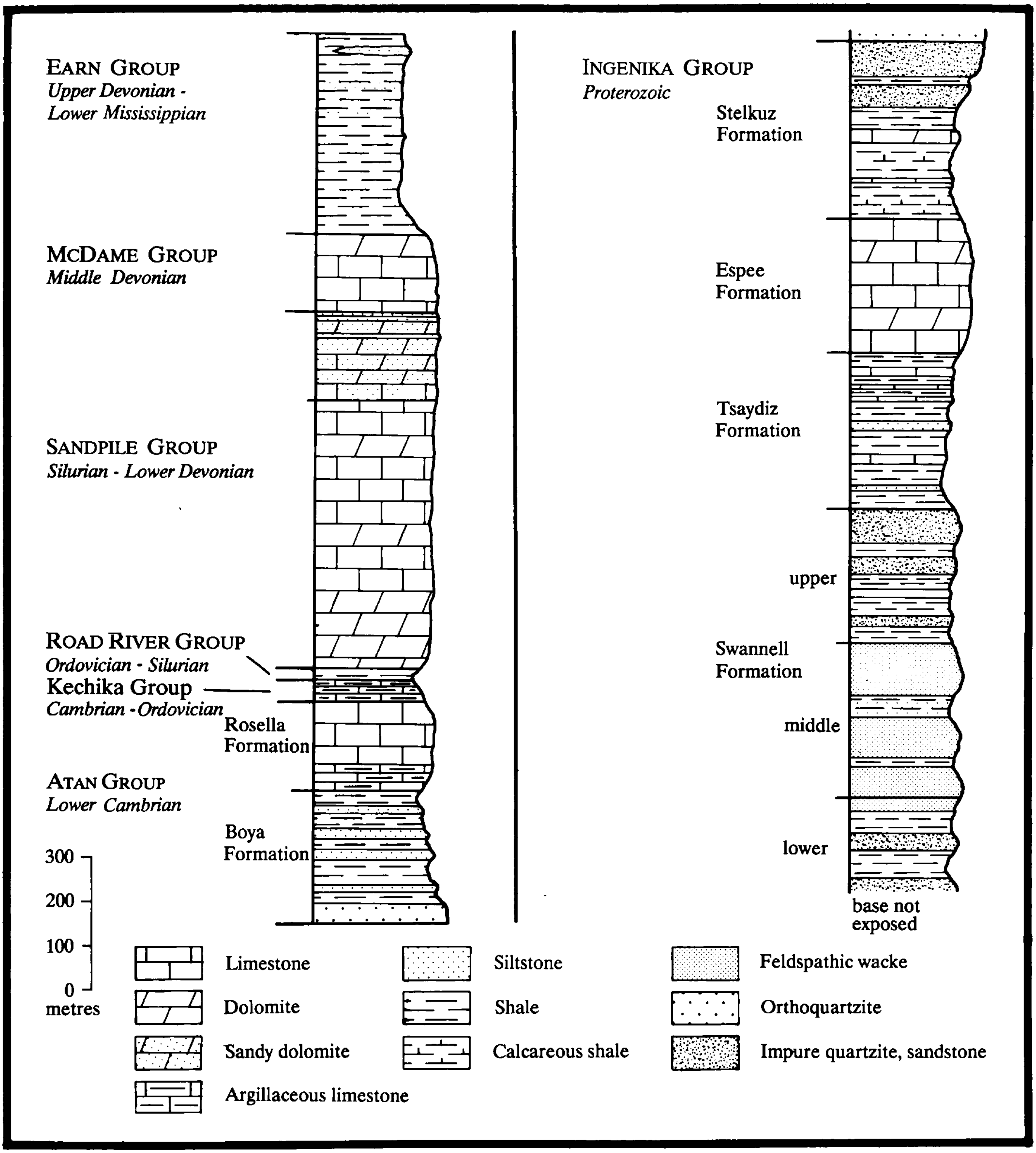
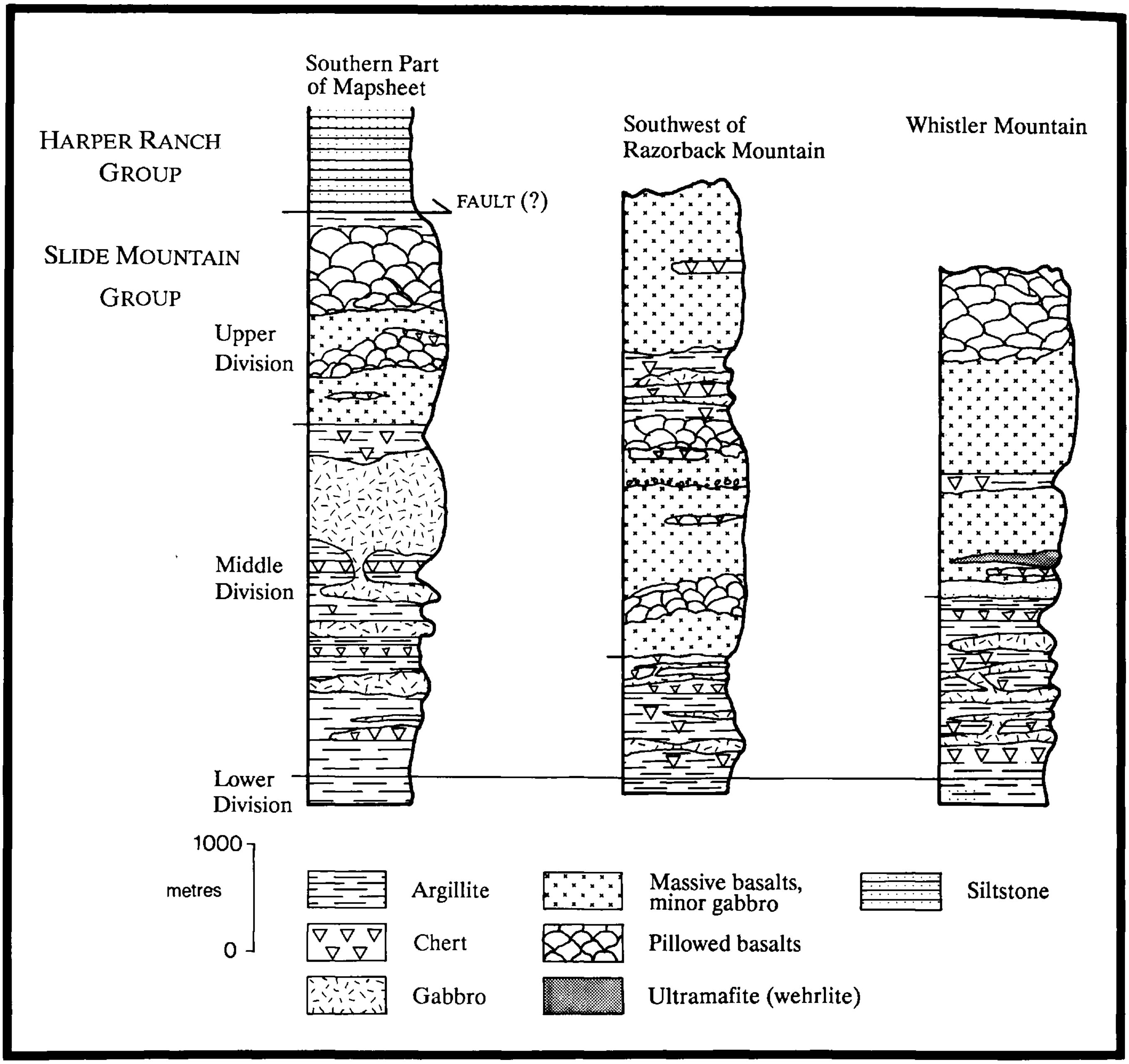


Figure 1-11-2. Generalized stratigraphic column of formations within the map area. (Includes material on facing page.)



to the east, across the Rocky Mountain Trench (now Williston Lake).

In the study area, the Intermontane Superterrane is represented by volcanic and sedimentary rocks of the Quesnel and Slide Mountain terranes. Quesnel rocks are composed of a volcanic and sedimentary assemblage of the Upper Triassic to Lower Jurassic Takla Group (Monger, 1977) and a poorly defined sedimentary and volcanic suite belonging to the Upper Paleozoic Harper Ranch Group which is basement to the Takla Group. The Slide Mountain Terrane is composed of oceanic rocks of the Upper Paleozoic Slide Mountain Group. The west side of the Quesnel Terrane is intruded by the multiphase, Triassic to Cretaceous Hogem batholith (Garnett, 1978) bounded to the west by the Pinchi fault which separates Quesnel rocks from middle Paleozoic to Triassic rocks of the Cache Creek Terrane.

Para-autochthonous rocks of North American affinity within the study area are part of a Proterozoic to Mississippian carbonate and siliciclastic miogeoclinal wedge which includes strata of the Proterozoic Ingenika Group to the Devono-Mississippian Earn Group (Figure 1-11-2). To the east, the lower parts of this sequence are highly metamorphosed (sillimanite grade) and deformed, and are incorporated within the Wolverine complex, one of several core complexes found along the length of the Omineca Belt.

The rocks above the garnet isograd roughly define a southwest-dipping package which is deformed by various generations of folds and faults. The most notable structure is

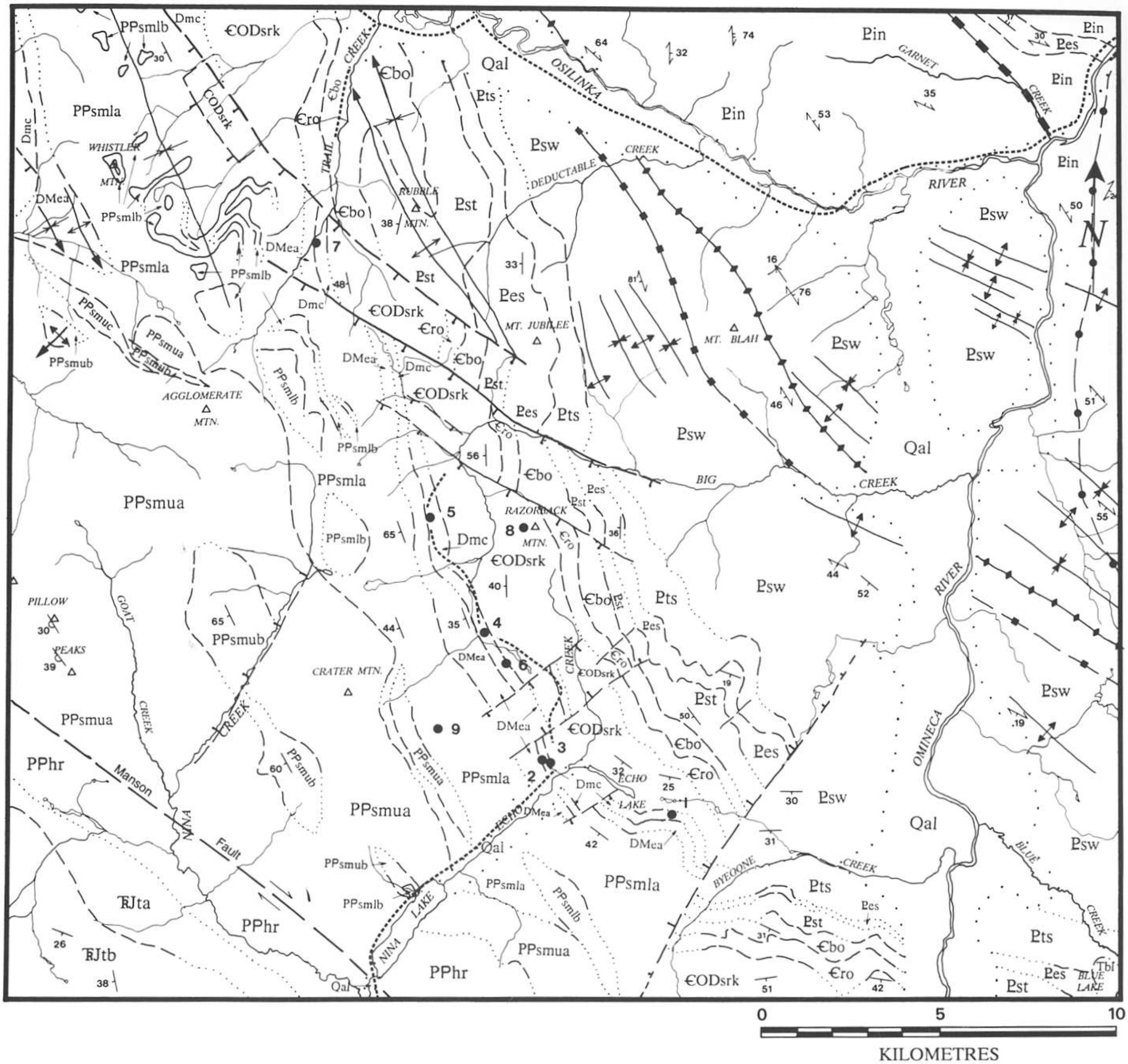


Figure 1-11-3a. Geology of the map area with location of known mineral occurrences described in Table 1-11-1.

the Manson fault zone, located along the Nina Creek valley (Figure 1-11-3a). The oldest rocks, the Proterozoic Ingenika Group, lie to the northeast. In this map area the contact between the Slide Mountain Group and North American rocks is not a west-side-down normal fault as shown by Ferri and Melville (1988) in the Manson Creek area. Its exact nature is indeterminate but, in accordance with observations made elsewhere along this contact, is assumed to be a layer-parallel thrust fault between lowermost Slide Mountain rocks and uppermost North American strata (Nelson and Bradford, 1987).

STRATIGRAPHY: NORTH AMERICAN ROCKS

INGENIKA GROUP (PROTEROZOIC)

The Ingenika Group is predominantly a clastic sequence with lesser amounts of carbonate. This package is in excess of 3.5 kilometres thick and composed of feldspathic and quartz wackes, siltstones, slates, sandstones, limestones and their higher grade metamorphic equivalents. Four subdivisions of the Ingenika Group (as defined by Mansy and Gabrielse, 1978) have been recognized in the area; these are, in ascending order, Swannell, Tsaydiz, Espee and Stelkuz formations.

LEGEND

QUATERNARY

Qal	*alluvium*

TERTIARY

Tbl	*BLUE LAKE VOLCANICS: massive basalt, volcanic breccia*

UPPER TRIASSIC TO LOWER JURASSIC

TAKLA GROUP

TJta	*argillite, siltstone, volcanic sandstone*
TJtb	*volcanic sandstone, augite porphyry basalts, agglomerate, minor siltstone, argillite*

DEVONIAN TO TRIASSIC(?)

PPhr	*HARPER RANCH GROUP: argillite, siltstone*

PENNSYLVANIAN TO PERMIAN

SLIDE MOUNTAIN GROUP

UPPER DIVISION

PPsmua	*Massive and pillowed basalt, volcanic breccia*
PPsmub	*Chert, argillite and gabbro*
PPsmuc	*Ultramafite*

LOWER AND MIDDLE DIVISION

PPsmlb	*Gabbro*
PPsmla	*Argillite at base, siliceous argillite, massive to ribbon chert, minor gabbro*

UPPER DEVONIAN TO MISSISSIPPIAN

DMea	*EARN GROUP; blue grey fissile shale, argillite, minor sandstone*

MIDDLE DEVONIAN

Dmc	*MCDAME GROUP; dark grey fetid dolomite, grey dolomite and limestone, breccia*

ORDOVICIAN TO LOWER DEVONIAN

CODsrk	*SANDPILE, ROAD RIVER AND KECHIKA GROUPS: Sandpile Group: upper part is sandy dolomite, dolomite and minor quartzite, lower massive to thickly bedded limestone to dolomitic limestone Road River Group: slate, calcareous slate Kechika Group: dark grey argillaceous dolomite*

LOWER CAMBRIAN

ATAN GROUP

Cro	*ROSELLA FORMATION: thickly bedded to platy limestone*
Cbo	*BOYA FORMATION: basal orthoquartzite succeeded by olive green siltstones and shales, minor sandstone*

PROTEROZOIC

INGENIKA GROUP

Pin	*Undifferentiated; garnet - staurolite - sillimanite schist and gneiss, marble and calcisilicate*
Pst	*STELKUZ FORMATION: basal part is shale, siltstone grading up to sandstone and minor limestone*
Pes	*ESPEE FORMATION: massive to thinly bedded limestones and recrystallized limestone, marble.*
Pts	*TSAYDIZ FORMATION: grey green slates, phyllites, minor siltstone, wackes and limestone*
Psw	*SWANNELL FORMATION: shale, phyllite, feldspathic wacke, limestone, garnitiferous schists, gneiss*

MINERAL ISOGRADS

- *BIOTITE*
- *GARNET*
- *STAUROLITE*
- *SILLIMANITE*

Figure 1-11-3b. Geological legend to accompany Figure 1-11-3a.

SWANNELL FORMATION

The Swannell Formation is the most areally extensive formation of the Ingenika Group, occupying roughly the northeastern third of the map area. The exact thickness is difficult to deduce due to polyphase deformation and lack of continuous outcrop. It is upwards of 2 kilometres thick and is tectonically thickened to the northeast. The basal part of the formation is polydeformed, metamorphosed to sillimanite grade and intruded by pegmatite and related granodiorite.

Three general subdivisions of the Swannell Formation have been recognized locally but could not be mapped out along strike. The lowest member, at least 1 kilometre thick, is composed primarily of thin to thickly bedded, very fine to medium-grained quartz and feldspathic wackes (feldspar content is typically less than 15 per cent). Subordinate to these lithologies are very fine grained impure sandstones, siltstones, grey to white marble, greenish slates and green to grey phyllite and schist. These rocks are metamorphosed to garnet, staurolite and sillimanite grade, with the metamorphic grade increasing to the northeast. Schists in the vicinity of Garnet Creek commonly contain distinct needles of metamorphic tourmaline in association with large porphyroblasts of staurolite (up to 2 centimetres long) and garnet.

Along the Wolverine Range, basal rocks of the Swannell Formation are metamorphosed to sillimanite grade (Figure 1-11-3a). They are coarsely crystalline schists, micaceous quartzites, calcsilicate, and quartz and feldspar-rich gneisses. They are injected by several generations of pegmatitic sills, dikes and related granodioritic and granitic bodies comprising over 50 per cent of the outcrop. Many of the pegmatitic sills, primarily the thinner ones, are folded or boudinaged (Plate 1-11-1), but the larger pegmatite bodies are foliated only on their margins. Later pegmatites crosscut the foliation and are undeformed.

The middle member of the Swannell Formation, some 300 to 400 metres thick, is characterized by massive beds of feldspathic wacke. These wackes are very coarse grained and approach granule conglomerates in some areas. They contain up to 30 per cent feldspar clasts and the quartz has a characteristic blue to purplish opalescence which disappears as the garnet isograd is approached. Greenish grey to silvery slate and phyllite, fine to coarse-grained quartz wackes and sandstones make up the remaining lithologies of this unit.

The upper member is approximately 300 metres thick and is characterized by massively bedded, brown-weathering, coarse-grained, impure quartzite and sandstone. These rocks may grade into or be interbedded with siltstones. Dark grey to greenish grey slates, phyllites and feldspathic wackes are also abundant.

TSAYDIZ FORMATION

The Tsaydiz Formation is typified by light greenish grey to grey crenulated slates and phyllites that are commonly interlayered with thinly bedded, buff-weathering limestone to argillaceous limestone. Lesser siltstones, quartz and feldspathic wackes and recrystallized brown-weathering, grey limestone layers 1 to 5 metres thick are also present. Limestones are more prevalent toward the base of the formation.

The formation is poorly exposed and has been inferred throughout most of the map area. The lower contact is placed at a recessive point which generally corresponds to the top of the highest resistant layer in the Swannell Formation. The thickness of the Tsaydiz Formation varies from 300 metres in the south to approximately 750 metres in the north where it may be exaggerated by tectonism.

Plate 1-11-1. Boudinaged pegmatite within lower Swannell Formation near Granite Creek. This pegmatite was intruded during either F_1 or F_2 deformation whereas other (later) pegmatites crosscut the foliation and are undeformed.

ESPEE FORMATION

The Espee Formation is a prominent cliff-forming carbonate unit which ranges from 200 to over 400 metres in thickness. It is easily traced through the entire length of the map area and is one of several marker horizons in the stratigraphic package. It is a buff, tan or grey-weathering limestone to dolomitic limestone which is moderately to thinly bedded and typically recrystallized to a coarse marble. It rarely fractures along bedding, but joints or spaced cleavage planes are common.

Its lower contact is generally not exposed but where it is seen, there is a transitional zone as interlayered slates and limestones of the Tsaydiz Formation give way to the more massive limestone of the Espee Formation. Similarly, at the upper contact with the Stelkuz Formation, buff-weathering, thickly layered Espee limestones are interlayered with green slates of the Stelkuz Formation across a zone some 50 metres thick.

STELKUZ FORMATION

Approximately 400 to 500 metres of green to grey slate and siltstone, brown to grey impure quartzite and sandstone, together with minor dolomitic limestone, make up the Stelkuz Formation. Slate and siltstone predominate in the lower part of the formation whereas sandstone, which is characteristically fine grained and planar bedded, becomes predominant toward the top. The upper part of the Stelkuz Formation contains a 100-metre-thick coarsening-upward sequence in which the amount of sandstone increases toward the top of the formation where thickly bedded, coarse-grained, impure quartzites, light grey to grey in colour, are very similar to basal Atan Group quartzites. Typically though, Stelkuz quartzites are impure and lack the glassy appearance of Atan orthoquartzites. The top of the Stelkuz Formation (and of the Ingenika Group) is placed at the base of the first thick (greater than 2 metres) sequence of white to light grey orthoquartzites. Thin layers (0.5 metre or less) of light-coloured orthoquartzites can be found below this contact, within the impure quartzites.

LOWER PALEOZOIC

Clastic and carbonate rocks exposed in the study area appear to be very similar to sedimentary rocks described in the Cassiar area by Gabrielse (1963), Fritz (1978, 1980) and Nelson and Bradford (1987). In the Cassiar area, the lower Paleozoic is represented by the Lower Cambrian Atan Group, the Cambrian to Ordovician Kechika Group, the Ordovician to Silurian Road River Group, the Silurian to Lower Devonian Sandpile Group (containing the Tapioca sandstone unit), the Middle Devonian McDame Group and the Upper Devonian to Mississippian Earn Group. We believe that a similar stratigraphy is present in the study area (as suggested by Gabrielse, 1975), but with some differences, such as the thinner nature of the Kechika and Road River groups. Where possible the formational names used in the Cassiar area are applied to units in the present map area.

Struik (1989a, b) describes a sequence of Lower to Middle Paleozoic rocks in the McLeod Lake area (east of the McLeod Lake fault) with characteristics similar to those of the study area, but containing volcanics and interlayered clastics and carbonates not encountered in our mapping.

ATAN GROUP (LOWER CAMBRIAN)

Orthoquartzites, siltstones, shales, sandstones and a thick carbonate unit above the Stelkuz Formation are assigned to the Atan Group as designated in the Cassiar area (Fritz; 1978, 1980); both the Boya and Rosella formations have been recognized.

BOYA FORMATION

The Boya Formation varies in thickness from 200 metres northeast of Echo Lake to upwards of 375 metres in the southeast. It is characterized by a white, grey, beige or maroon, massive to thickly bedded orthoquartzite, 10 to 30 metres thick, at the base of the section. It is typically fine to medium grained, but thin beds of quartz-granule conglomerate are also present. This basal unit is very distinct and therefore very useful in outlining the megascopic structures.

Thin to moderately bedded olive-green to grey siltstone, shale and beige to tan, very fine to fine-grained sandstone comprise the remainder of the formation. Typically the sandstone makes up less than 30 per cent of the sequence, although sections of sandstone and quartzite up to 10 metres thick occur in the upper part of the Boya Formation. These massive sandstones contain rare vertical *Skolithus*(?) and bedding-parallel burrows.

ROSELLA FORMATION

Uppermost Boya Formation shales are succeeded by brown-weathering nodular limestone and basal Rosella Formation limestones over a distance of approximately 5 metres. The basal part of the Rosella Formation comprises 20 to 50 metres of dark grey to grey, thin-bedded and platy, finely crystalline limestone and argillaceous limestone. This platy limestone gives way upwards to approximately 150 to 180 metres of massive, thick bedded, fine to coarsely crystalline limestone and rare dolomite. Bedding is typically outlined by thin, discontinuous to wispy argillaceous layers less than 1 metre long. Horizons of oolites, which may be silicified in the lower part of the section, are very rare.

KECHIKA GROUP (CAMBRIAN TO ORDOVICIAN)

Approximately 50 metres of argillaceous limestone is assigned to the Kechika Group. Where not exposed, this unit and the succeeding Road River Group can usually be inferred by the recessive slope between the Rosella Formation and the Sandpile Group. Correlative units thicken to the north (Nelson and Bradford, 1987).

This unit is characterized by thin-bedded, grey to black argillaceous limestone separated by thinner, tan to brown-weathering argillaceous dolomite layers.

ROAD RIVER GROUP (ORDOVICIAN TO SILURIAN)

Above the Rosella Formation is approximately 25 metres of thinly bedded grey and dark grey shale and slate together with thin-bedded, dark grey to black argillaceous limestone assigned to the Road River Group. This argillaceous limestone can be up to 10 metres thick and is found toward the top of the formation. This unit is poorly exposed in the area and is inferred over most of the map sheet. Graptolites recovered from the shales indicate a Silurian age, perhaps Llandoverian (B. Norford, personal communication, 1989).

SANDPILE GROUP (SILURIAN TO LOWER DEVONIAN)

This package of limestone, dolomite, sandy dolomite and minor quartzite has similarities to the Sandpile Group described by Gabrielse (1963) east of the Kechika fault. It is the thickest Paleozoic unit of continental affinities in the area (750 to 1000 metres) and forms prominent cliffs around Echo Lake. It is made up of two units; a lower limestone and dolomite sequence ("Echo Lake limestone", approximately 600 metres thick), and an upper sandy dolomite and quartzite sequence (100 to 200 metres thick). The upper part of this sequence resembles the Tapioca sandstone of the Cassiar area (Gabrielse, 1963; Nelson and Bradford, 1987).

The lower carbonate is characterized by sequences of light to medium grey, massively bedded limestone up to 2 metres thick. Semicontinuous quartz "vugs" and lenses (replaced algal structures), 1 to 2 centimetres thick, are found within these horizons and sometimes form an interwoven network comprising up to 20 per cent of the rock (Plate 1-11-2); rare, thin beds or lenses of grey chert or isolated bodies of polymict carbonate breccia up to 5 metres thick are also associated with these horizons. Thickly bedded limestone is interlayered with thinly bedded limestone and dolomite which commonly exhibit algal laminae and layers of silicified oolites and pisolites up to 2 centimetres in diameter.

The upper 100 to 200 metres of the Sandpile Group contains beds of sandy dolomite comprised of up to 30 per cent well-rounded, medium-grained quartz grains. Grey quartzite layers, 1 to 2 metres thick, and rare argillite or siltstone beds, together comprise about 5 per cent of the upper unit of the Sandpile Group.

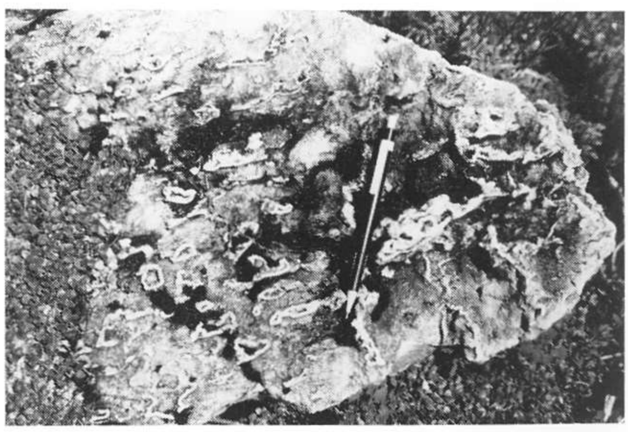


Plate 1-11-2. Quartz "vugs" and lenses within the Sandpile Group. These vugs are typically infilled with calcite and may be replaced algal features.

McDAME GROUP (MIDDLE DEVONIAN)

Some 150 to 200 metres of grey to black fetid limestone and dolomite make up the McDame Group. Exposures are poor and, away from known mineral occurrences, the map trace is tentative. The lower part of the McDame Group is characterized by thin to thick-bedded, dark grey to black fetid limestone. The limestone, though not abundantly fossiliferous, contains rugosan corals, brachiopods, gastropods, bryozoa(?), amphipora and beds of crinoid osicles, some of which exhibit twin-holed columnals. Parts of this unit are coarsely recrystallized and contain calcite and pyrobitumen-filled vugs.

The upper part of the McDame Group is slightly fetid grey to tan, finely crystalline dolomite and minor limestone which may exhibit faint bedding.

The presence of twin axial canal columnals within this formation makes it no younger than early Middle Devonian (early Eifelian; B. Norford, personal communication, 1989). Conodonts recovered from the upper part of the unit also indicate the Middle Devonian as the upper age limit on this formation (M.J. Orchard, personal communication, 1989).

EARN GROUP (UPPER DEVONIAN TO MISSISSIPPIAN)

Approximately 400 to 500 metres of blue-grey, grey or dark grey shales, argillites and minor sandstones comprise the Earn Group. The lower shales are extremely fissile, forming large, thin sheets several millimetres thick. They have a characteristic blue-grey colour typical of the Earn Group in the Cassiar area. Up-section these shales become interlayered with thicker bedded argillites and silty argillites. Uppermost Earn shales and argillites are distinguished from lowermost Slide Mountain rocks by their lack of wavy bedding.

In the area immediately south of Big Creek, thickly bedded quartz sandstones, forming a sequence up to 30 metres thick, have been assigned to the uppermost Earn Group. This lithology is not typical of the Slide Mountain Group but has been reported from the Earn Group in the Cassiar area (Nelson and Bradford, 1987).

STRATIGRAPHY: ALLOCHTHONOUS ROCKS

SLIDE MOUNTAIN GROUP (PENNSYLVANIAN TO PERMIAN)

Upwards of 7 kilometres of basalt, argillite, chert and gabbro make up the Slide Mountain Group. These rocks were first mapped by Armstrong (1949) and Roots (1954) who grouped them with the Cache Creek Group due to their very similar lithologies. Monger (1973) first recognized that these rocks belong to the Slide Mountain Group and its equivalents.

Three subdivisions are recognized in the map area and are similar to those described by Ferri and Melville (1989) to the south. The lower division is composed primarily of argillite with minor amounts of clastics and limestone. The middle division is made up of siliceous argillites, cherts and gabbro. Pillowed and massive basalt, gabbro, argillite, chert and ultramafite comprise the upper division.

In the study area, the three subdivisions of the Slide Mountain Group appear to form a continuous stratigraphic package; there is no evidence for major tectonic breaks. Elsewhere along the Cordillera, rocks equivalent to the Slide Mountain Group have been shown to be made up of repeated thrust slices, even though the package appears to be in stratigraphic continuity (Struik and Orchard, 1985; Schiarizza and Preto, 1987; Nelson and Bradford, 1987).

LOWER DIVISION

This unit is composed of 200 to 300 metres of grey to dark grey or black, rusty weathering, thin-bedded, wavy to platy argillites. The lighter coloured varieties tend to be slightly siliceous. At one locality, light coloured felsic tuff is present within this unit and appears very similar to felsic tuff seen within the Slide Mountain Group farther south (Ferri and Melville, 1989). The upper part of the unit may contain a 5 to 10-metre section of thickly bedded, interlayered buff-weathering and siliceous limestone that has yielded Lower Permian conodonts (M.J. Orchard, personal communication, 1989). Below this carbonate, and immediately above the sandstone in the uppermost Earn Group, is a 10 to 20-metre section of dark grey to black, massive to poorly bedded, chert-quartz wacke. The clasts are fine to coarse grained, predominantly chert, and make up less than 50 per cent of the rock.

The lower contact of this division, with the Earn Group, is not seen, however, Slide Mountain shales and argillites are typically more siliceous, and lack the fissility of the Earn Group rocks.

MIDDLE DIVISION

Dark argillites of the lower division become less prominent up-section and are succeeded by thin to moderately bedded grey argillites, light grey to greenish siliceous argillites, light grey, green and maroon cherts and ribbon cherts of the middle division. These rocks are intruded by gabbroic sills and dikes in the upper parts of the division. The maroon and salmon-coloured cherts are only found in the uppermost part of the middle division, either in association with the gabbro or immediately below the basalts of the upper division. Minor constituents are buff-weathering micritic limestone layers less than 0.5 metre thick and a quartz-bearing tuff, 10 to 20 metres thick, present toward the base of the unit. The latter is exposed near Whistler Mountain and is similar to quartz-bearing tuffs described within the Slide Mountain Group to the south (Ferri and Melville, 1989). Rare constituents within the lower part of the division are thin beds (less than 1 metre) of chert wackes within the argillites. The thickness of the middle division varies from approximately 2500 metres in the south to 700 metres on the western margin of the map area.

Gabbroic sills up to 1000 metres thick intrude the upper parts of the middle division and, in places, are traceable for several kilometres. These gabbros are found at various stratigraphic levels and are beautifully exposed around Whistler Mountain (Plate 1-11-3). They are fine to coarsely crystalline, with subequal pyroxene and highly altered plagioclase.

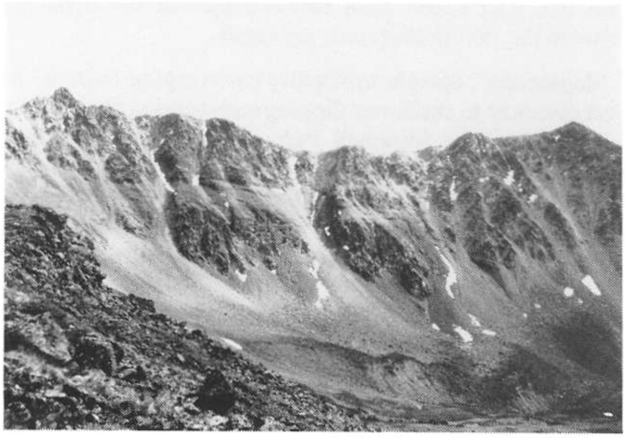


Plate 1-11-3. Gabbro sills intruding upper middle division sediments of the Slide Mountain Group near Whistler Mountain. In this photograph the sediments form the recessive sequence in the middle of the slope with gabbro sills above and below. These gabbro bodies are several hundred metres thick.

UPPER DIVISION

A thickness of at least 5 kilometres of massive and pillowed basalts, minor sediments, gabbroic and ultramafic sills are assigned to the upper division. Whether this is a true stratigraphic thickness or the result of tectonic thickening is not known. In the southern part of the map area only 2000 metres of basalt are found, overlain by siltstones of the Harper Ranch Group (*see* section on Harper Ranch). In the map area to the southwest (*see* Figure 1-11-1), these sediments were assigned to the middle division of the Slide Mountain Group by Ferri *et al*. (1989).

More than 80 per cent of the division is composed of dark green to greyish green, variolitic, pillowed to massive basalt which may contain irregular bodies of fine-grained gabbro (Plate 1-11-4). Massive and interpillow basaltic breccia is also common.

In the northwest, the basal part of the upper division contains lenticular bodies of wehrlite up to 200 metres thick, composed of undeformed clinopyroxene, olivine, serpentine (after olivine) and magnetite. They are associated with mafic gabbros and are believed to be sills.

Light to dark grey siliceous argillite, varicoloured cherts (cream, grey, green, salmon, maroon) and gabbro sills are found in sequences upward of 1 kilometre. The similarity of these sedimentary packages to middle division lithologies may indicate that they are fault slices of the upper part of the middle division. A fault contact was observed at the base of a sedimentary lens northwest of Nina Lake, but the present evidence does not allow these packages to be confidently interpreted as fault repetitions. More accurate fossil ages are required to fully resolve this problem.

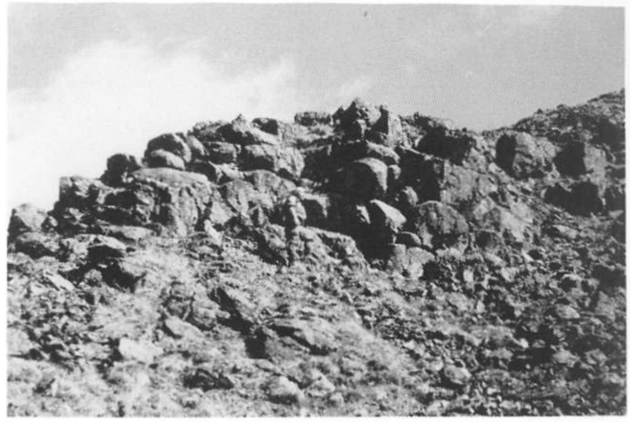


Plate 1-11-4. Large pillows within basalts of the upper division of the Slide Mountain Group on the ridge containing Pillow Peaks.

HARPER RANCH GROUP(?) (MIDDLE TO UPPER PALEOZOIC)

Light brown to greenish weathering siltstones and minor dark grey phyllites and argillites along the Nina Creek valley are assigned to the Harper Ranch Group. Similar sediments overlie basalts of the upper division on the slope southeast of Nina Lake valley and were originally mapped as middle division sediments of the Slide Mountain Group. As described above, these rocks have been reinterpreted as Harper Ranch as similar lithologies are not present in the middle division of the Slide Mountain to the north; they may, however, be a western facies of the middle division of the Slide Mountain Group.

Sediments equivalent to the Harper Ranch were designated Slide Mountain Group by the authors in map areas to the south (Ferri and Melville, 1988, 1989) where these siltstones and argillites are found within lithologies similar to middle division sediments of the Slide Mountain as described above. This season's mapping has restricted these rocks to an area south of the Manson fault, indicating that they may be separate from the Slide Mountain Group.

TAKLA GROUP (UPPER TRIASSIC TO LOWER JURASSIC)

Argillites, siltstones, lapillistone(?), volcanic sandstones and massive augite and feldspar porphyry flows assigned to the Takla Group outcrop in the southwest corner of the map area (south of Nina Creek). Two units are recognized: a lower, predominantly argillaceous unit and an upper unit composed primarily of volcanic sediments and minor flows. These are very similar to the lower two units of the Takla Group as described by Ferri and Melville (1989b) to the south.

The lower Takla Group (approximately 500 metres thick) is made up of thickly bedded green to light green siliceous argillites, grey-green to brown siltstones and fine-grained sandstones which contain volcanic and chert clasts.

The upper unit is composed predominantly of thin-bedded grey-green to brown volcanic siltstone or tuffs, thick to massively bedded volcanic sandstone with lesser dark green to green augite or feldspar porphyry flows. The volcanic sandstones are polymictic, containing volcanic clasts from several sources.

BLUE LAKE VOLCANICS (TERTIARY (?)

Immediately northwest of Blue Lake, in the southeastern corner of the map area, are massive, dark grey, augite-bearing basalts and volcanic breccia (A. Halleran, personal communication, 1989). Only a few scattered outcrops of basalt were observed and based on aeromagnetic data, it is believed that these basalts are restricted to this area.

STRUCTURE

Major tectonic elements in the study area are the left-lateral Manson fault zone and the boundary between the Slide Mountain Group and the para-autochthonous North American miogeoclinal strata. The Slide Mountain–North American contact relationship is inferred from other areas in the Canadian Cordillera (Nelson and Bradford, 1987) and is assumed to be a layer-parallel thrust fault along the contact between the two suites. Because of this, rocks on both sides of the contact appear to be in stratigraphic continuity. It should be noted that evidence for this thrust fault is not apparent in the study area; slates and argillites of the lowermost Slide Mountain Group *appear* to grade into those of the Earn Group.

The Manson fault has been placed separating Harper Ranch from Slide Mountain strata. Direct evidence for the fault (*i.e.* tectonized strata) is not present in the area due to poor exposure, but the fault trace is marked by a well-developed topographic lineament.

South of the map area the Manson fault was placed along Nina Creek due to the presence of highly deformed rocks (Ferri and Melville, 1989; Ferri *et al.*, 1989). The Manson fault is interpreted as a broad zone and clearly contains anastamosing to en echelon faults in the region between the Takla and Slide Mountain groups. It is believed the same relationship is present in the current map area.

The area can be broadly divided into two structural regions; the northeastern sector where polyphase deformation and metamorphism dominate, and the southwestern half which is typified by a west-dipping sequence where brittle deformation is more prevalent. The dividing line, or zone, between these two domains lies within the upper part of the Swannell Formation and roughly corresponds to the garnet-biotite boundary.

To the north the west-dipping package is modified by folding. In the Whistler Mountain area broad folds can be traced out within the Slide Mountain Group. Earn shales and MacDame carbonates reappear on the west side of the large syncline in this area. Across Trail Creek the Atan Group and Espee Formation delineate an upright to northeast-verging fold pair. Minor folds are rare in this area.

This west-dipping sequence is cut by two sets of normal faults; northwest and northeast trending. The northeast-trending set is found to the south; the most notable structure is the Bygone fault with a vertical displacemem on the Espee Formation of approximately 2.5 kilometres (northwest side down). To the north, the northwest-trending, west-side-down normal faults form an en echelon array cutting both the North American Paleozoic sequence and the Slide Mountain Group, with the Trail Creek fault forming part of the boundary between the two stratigraphic packages.

Megascopic, upright to slightly east-verging folds in the west give way to shallower dipping east-verging folds north-eastward into the Swannell Formation. Above the garnet isograd (*i.e.* to the northeast) schistosity or foliation becomes layer parallel as folds become tight to isoclinal. This same relationship was described by Mansy and Dodds (1976) within the Ingenika Group of the Swannell Ranges north of the map area.

At least three phases of deformation and folding (F_1 to F_3) are recognized in the map area. F_1 is characterized by tight to isoclinal folds which are synmetamorphic (*i.e.*, micas and other metamorphic minerals are axial planar to F_1 hinges). F_2 folds are axial planar with F_1 folds but are generally more open and also postmetamorphic, with micas folded and crenulated around their hinges. Along F_2 limbs, S_1 foliation has been transposed parallel to F_2 axial planes. F_2 folds are rare and only seen in the higher grade regions. In the metamorphic rocks to the northeast, compositional layering, together with the above foliations, is folded into megascopic upright to southwest-verging F_3 folds which may be associated with similarly oriented and commonly observed fold crenulations. Southwest-verging structures were observed within the high-grade metamorphic rocks of the Wolverine Range.

Parrish (1976), describes very similar fold hierarchies in the Aiken Lake area where F_2 folds appear to be more prevalent and may reflect the higher metamorphic grade (sillimanite).

In the Ingenika Range, Bellefontaine (personal communication, 1989), describes early, east-verging F_1 folds which predate main-stage metamorphism and are the same as F_1 folds in this study. Bellefontaine (1989) also describes megascopic F_2 folding that is west verging and may be related to F_3 folding in the present study area. In the Ingenika Range tight to isoclinal folds in the schistosity are rarely present and only seen in the higher grade metamorphic areas (K.A. Bellefontaine, personal communication, 1989) and are attributed to late F_1 folding.

Evenchick (1988) sees the same progression of metamorphism and deformation farther north in the Sifton Ranges.

METAMORPHISM

The southwestern two-thirds of the map area is greenschist grade or lower. Chlorite, muscovite, actinolite, clinozoisite and chloritoid are the main metamorphic minerals. To the northwest, metamorphic grade steadily increases until sillimanite is present in the Wolverine Range.

The first appearances of index metamorphic minerals (biotite, garnet, staurolite and sillimanite) is shown on Figure 1-11-3. Mineral isograds outline part of several large metamorphic highs or domes which, for the most part, follow the structural trend in the area (Gabrielse, 1975). The exception to this is the sillimanite isograd which follows the trend of the Wolverine Range (thermal event ?).

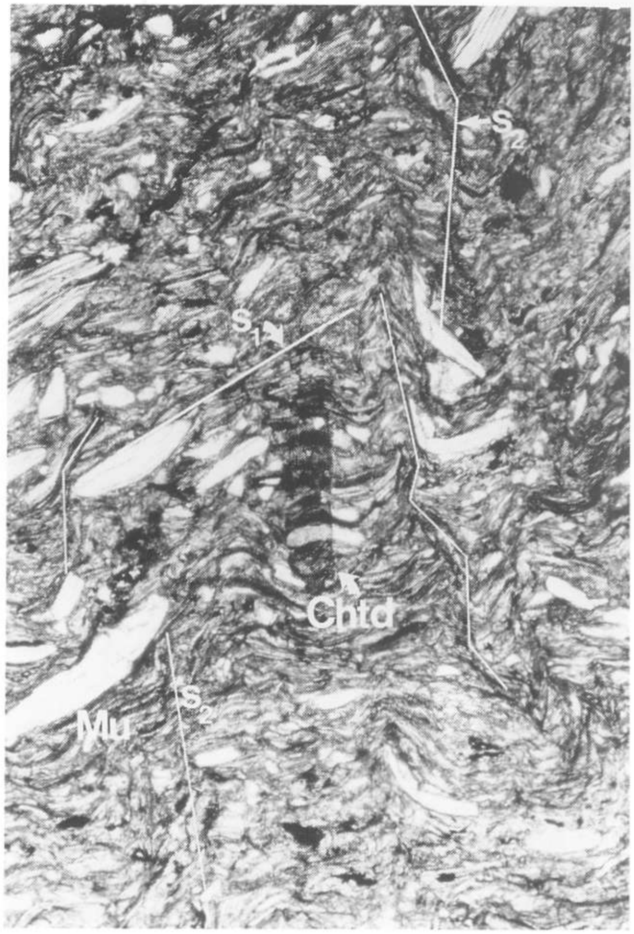
A retrograde metamorphic event has affected these higher grade metamorphic rocks and is manifested by the partial or complete pseudomorphing of garnet and biotite by chlorite. This retrogression is developed locally and is evident primarily in the lower amphibolite grade rocks.

Pegmatites and related granodiorite bodies intrude schists and gneisses of the Swannell Formation in the higher grade metamorphic areas (staurolite and above). Many of these are boudinaged or exhibit deformed margins (*see* Plate 1-11-1 and section on Swannell Formation) indicating that they were intruded during F_1 or F_2 deformation. Some pegmatites crosscut the structural fabric and are undeformed, indicating they were intruded after deformation ceased. These later pegmatites are rare.

Textural relationships between prograde metamorphic minerals and deformational fabrics (*i.e.*, S_1, S_2) are illustrated in Plate 1-11-5. In these photomicrographs an example of moderately deformed, upper greenschist grade rocks from higher in the stratigraphic package (uppermost Stelkuz Formation) is compared to isoclinally folded, amphibolite grade material from the lower Swannell Formation. In Plate 1-11-5(a) chloritoid porphyroblasts, which for the most part grew parallel or subparallel to S_1, are overgrowing the S_1 foliation outlined by layer-parallel muscovite crystals. The S_1 foliation is slightly "bowed" or flattened around porphyroblasts oriented perpendicular to foliation. A later crenulation cleavage (S_2) is superimposed on this foliation. Chloritoid porphyroblasts do not overgrow crenulation hinges or the associated cleavage, in fact this solution cleavage wraps around the chloritoid porphyroblast and is roughly parallel to the hinge planes of megascopic folds outlined by markers at this level.

Plate 1-11-5(b), from the higher grade rocks, shows porphyroblasts which can contain inclusion trails typically parallel to S_1 foliation or, less commonly, with helicitic patterns. S_1 foliation is wrapped around the porphyroblasts and is almost always parallel to S_2 foliation. Crenulations that overprint these foliations are believed to be related to a later deformation D_3, as they are upright.

From the forgoing one can confidently assume that metamorphic minerals from the different stratigraphic and deformational levels formed at the same time. Similarly S_1 foliations seen in both examples in Plate 1-11-5 are genetically equivalent. Relationships in the high-grade areas indicate that the main metamorphic event was syn-F_1. In the lower grade regions the peak of metamorphism seems to be (for the most part) late to post-F_1, unless one assumes that the layer-parallel micas were formed during an earlier metamorphic event. In higher grade rocks, later F_2 folding may have obliterated the exact timing relationships between F_1 formation and porphyroblast growth. Therefore, whether the S_2 crenulation cleavage seen in Plate 1-11-5(a) is the same as S_2 in Plate 1-11-5(b) or the later D_3 crenulation is uncertain.



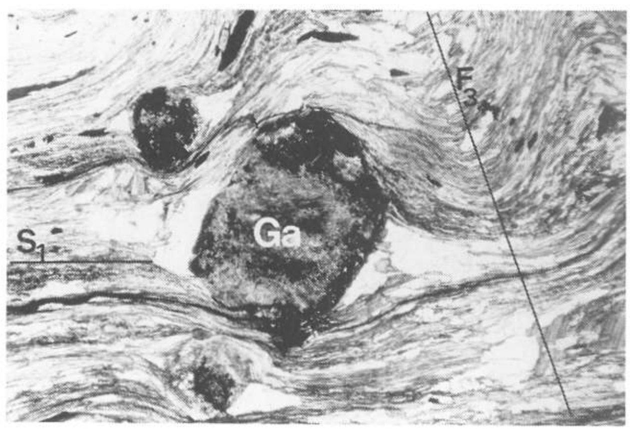


Plate 1-11-5. (a) Photomicrograph of chloritoid phyllite to schist from the upper part of the Stelkuz Formation. The large muscovite grains (Mu) form a layer-parallel fabric (S_1) which is later crenulated with an associated pressure solution cleavage (S_2). The chloritoid porphyroblasts (Chtd) overgrow S_1 but are pre-S_2. See text for details. (b) Photomicrograph of garnet-biotite schist from the Swannell Formation. In this plate the garnet (Ga) has been totally chloritized. S_1 (here parallel to S_2) is delineated by the phyllosilicate minerals which are flattened around the porphyroblasts. These are later crenulated forming the F_3 crenulations seen in the photomicrograph. See text for details. Length of each photomicrograph is 3.3 millimetres.

TABLE 1-11-1
KNOWN MINERAL OCCURRENCES
(93N/15-NORTH HALF and 94C/02-SOUTH HALF)

Map No.	Type	MINFILE Number	Name	Commodities	Geological Description
1	Statabound carbonate-hosted base metals	093N 172	Sheila	Zn, Ba, Pb, Ag	Sphalerite occurs disseminated within a fine-grained dolomite and massively with coarse galena in a barite-cemented dolomitic breccia of the McDame Group.
2	"	093N 075	W. Vernon	Zn, Pb, Ba, Ag	Sphalerite occurs as disseminated grains in fine-grained dolomite and as brecciated pods in arenaceous dolomite. Galena primarily occurs massively with barite in small localized shear zones with varying amounts of sphalerite. The hostrocks are primarily dolomites and dolomitic breccias of the McDame Group.
3	"	093N 076	Vernon	Zn, Pb, Ba, Ag	"
4	"	093N 114	Biddy	Zn, Pb, Ge, Ag	"
5	"	093N 158	Crin	Pb, Zn	"
6	"	093N 010	Jemima	Zn, Pb	Sulphide mineralization occurs in discontinuous and irregular shaped pods within arenaceous dolomites of the McDame Group.
7	"	new	new	Zn, Pb, Ba	Sphalerite, galena, barite and pyrite occur within a coarsely crystalline dolomite of the McDame Group.
8	Stockwork-hosted base metals	093N 170	Osi	Pb, Zn, Ag	A stockwork of siderite and hematite veinlets within massive limestone and dolomitic limestone in the upper unit of the Sandpile Group contains disseminated galena and sphalerite.
9	Shear-zone-hosted base and precious metals	093N 011	Nina	Au, Ag, Cu	Sulphide mineralization with varying gold, silver and base metal concentrations occurs as podiform lenses within a shear zone. The hostrocks are predominantly fine-grained gabbros or basalts(?) with lesser argillaceous cherts within the middle unit of the Slide Mountain Group.

Potassium-argon ages of metamorphism in the area consistently return dates in the 40 to 65 Ma range (unpublished analysis of metamorphic rocks by the authors from the Manson Lakes map sheet, as well as Gabrielse, 1975; Parrish, 1979) and are most likely related to the low-grade retrogression seen in the map area. Parrish (1979) presents Rb-Sr data which indicate a Middle to Late Jurassic (or earlier) age for the prograde metamorphism in the Aiken Lake area.

MINERALIZATION

Known mineral occurrences within the map area are predominantly associated with the McDame dolomitic units, except for a lead-zinc-silver showing within the Sandpile Group and a gold-silver-copper prospect in Slide Mountain rocks.

Stratabound sulphide mineralization occurs throughout an interval which extends from the Earn–McDame contact down to the uppermost Tapioca sandstone unit of the Sandpile Group. Mineralization consists primarily of argentiferous galena, brown to red sphalerite, barite and minor amounts of pyrite. In the Biddy area, germanium-bearing sphalerite is reported with the average tenor being 0.05 per cent of the sphalerite which averages about 3 to 4 per cent in mineralized areas (Leighton, 1988). Sulphide mineralization, predominantly sphalerite, is typically disseminated in arenaceous dolomites, fine-grained dolomites and sandstones. Semimassive sphalerite and galena typically favour dolomitic breccias as hostrocks and occur as matrix or carbonate-clast replacements, or a combination of both. In the McDame Group, mineralization exhibits a strong affinity with the dolomitic successions (Sonnendrucker, 1975). Remobilization has further concentrated the sulphide mineralization along shear zones where galena, sphalerite and megacrystic barite occur as irregular pods (Leighton, 1988) or as the matrix of fault breccia (Sonnendrucker, 1975).

The map area has potential for stratabound carbonate-hosted base metal deposits throughout the McDame Group. This is exemplified by the 1989 discovery of a new mineral occurrence in the northern trace of this sequence (*see* Table 1-11-1 and Figure 1-11-3a).

Fracture-hosted mineralization occurring lower in the stratigraphy is exemplified by the Osi showing where two types of occurrences have been identified. Both occur within the massive limestones and dolomitic limestones of the upper Sandpile Group. They are either a stockwork of siderite and hematite veinlets containing disseminated galena and sphalerite or larger, siliceous veins containing massive galena (Sonnendrucker, 1975).

Shear zones cutting fine-grained gabbros and argillaceous cherts in the middle unit of the Slide Mountain Group may host podiform lenses of sulphide mineralization as at the Nina prospect. Mineralization is primarily pyrite and minor chalcopyrite with varying amounts of gold and silver (Cope, 1988).

CONCLUSIONS

- Rocks of North American affinity within the study area form a continuous sequence from the Proterozoic Ingenika Group to the Devono-Mississippian Earn Group. The Paleozoic and Proterozoic stratigraphy appears very similar to that described in the Cassiar and northern Omineca Mountains. Stratigraphic nomenclature from the Cassiar area has been suggested for these rocks.
- A three-fold subdivision has been recognized within the allochthonous Slide Mountain Group; an upper basaltic sequence, a middle siliceous sediment/gabbro sequence and a lower argillite sequence. These correspond to subdivisions recognized to the south by Ferri and Melville (1988, 1989).
- The western part of the map area forms a west-dipping homoclinal succession. To the east, the lower parts of the Ingenika Group record at least three phases of deformation together with syn-F_1 prograde metamorphism (to sillimanite) and a later chlorite-grade retrograde event.
- Stratabound lead-zinc-barite-silver mineralization is found disseminated or in breccia zones within McDame Group and Tapioca sandstone equivalents. Copper-gold-silver mineralization is associated with siliceous sediments and gabbro of the middle division of the Slide Mountain Group.

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SNIPPAKER MAP AREA (104B/6E, 7W, 10W, 11E)

By J.M. Britton, B.A. Fletcher and D.J. Alldrick

KEYWORDS: Regional geology, Stikine assemblage, Stuhini Group, Hazelton Group, Texas Creek intrusive suite, Premier porphyry, Coast plutonic complex, vein, skarn, mesothermal.

INTRODUCTION

This report describes preliminary results of regional mapping near Johnny Mountain (Figure 1-12-1). It is part of the Iskut-Sulphurets project, an ongoing study of the mineral deposits of this exciting gold belt (Alldrick and Britton, 1988; Alldrick *et al.*, 1989a; Britton and Alldrick, 1988; Britton *et al.*, 1989). The project's goals are:

- to define regional stratigraphy and structure,
- to produce 1:50 000 and 1:20 000-scale geological maps,
- to document mineral occurrences, and
- to develop genetic models of mineralization.

The Snippaker map area covers about 1000 square kilometres, roughly centred on 131°00'W and 56°30'N near the headwaters of Monument Creek. Terrain is very rugged with elevations ranging from 65 metres along the Iskut River to over 2000 metres along the Alaska boundary. Below treeline at 1100 metres vegetation is a dense mixture of conifers, slide alder, devil's club, ferns and berry bushes. Permanent ice and snow cover about one third of the map area. Weather is typical of the northern rainforest. Annual precipitation exceeds 100 centimetres, much of it as snow in January and February.

Historically, river barges ascended the Iskut some 60 kilometres above its confluence with the Stikine. Present access is only by air. Fixed-wing aircraft as large as a Hercules or DC-3 can land on well-maintained gravel airstrips at Bronson Creek (elevation 90 metres) and Johnny Mountain (elevation 1075 metres). There is a disused airstrip on Snippaker Creek (elevation 530 metres). A road-access study has been completed with a view to linking the area to Highway 37, 72 kilometres east of the Bronson airstrip (Smith and Gerath, 1989).

PREVIOUS WORK

Kerr (1948) was first to map the geology and much of the topography in the 1920s. Kerr's work was incorporated into Operation Stikine (Geological Survey of Canada, 1957) with little remapping. In 1987 and 1988 the British Columbia Geological Survey Branch mapped 150 square kilometres around Bronson Creek (Lefebure and Gunning, 1989). The Geological Survey of Canada has begun a mapping program of the whole of NTS 104B (Anderson, 1989; Anderson and Bevier, 1990). About 25 per cent of the area has never been mapped at any scale.

A synthetic aperture radar (SAR) survey proposed by the province was completed in 1988 by the Canada Centre for Remote Sensing to outline major structures in the Iskut-Sulphurets gold belt. The results of a regional stream sediment sampling program conducted over this area in 1987 were released in 1988 (National Geochemical Reconnaissance, 1988), and a gold lithogeochemistry study was completed by Lefebure and Gunning (1988) in the Bronson Creek area. Other sources of information include annual reports of the Minister of Mines, assessment reports and the ministry's property file.

Fieldwork was conducted from July 8 to 13 September 1989. Traverse data were recorded on 1:15 000 airphoto enlargements and compiled on 1:20 000 and 1:50 000 base maps.

EXPLORATION HISTORY

In the late 1800s the Stikine River was a major route to the gold rushes in the Cassiar (1873) and Klondike (1896–98). At the turn of the century, prospectors heading south from the Yukon goldfields searched the Iskut valley for placer gold and staked promising looking bedrock gossans.

In 1906 custom officers F.E. Bronson and E.S. Busby, with others from Wrangell, Alaska, staked claims on lower Bronson Creek. In 1910, about a ton of high grade copper ore was shipped to a smelter in Ladysmith and assayed 2 grams per tonne gold, 1515 grams per tonne silver and 12.45 per cent copper (Minister of Mines, B.C., 1918).

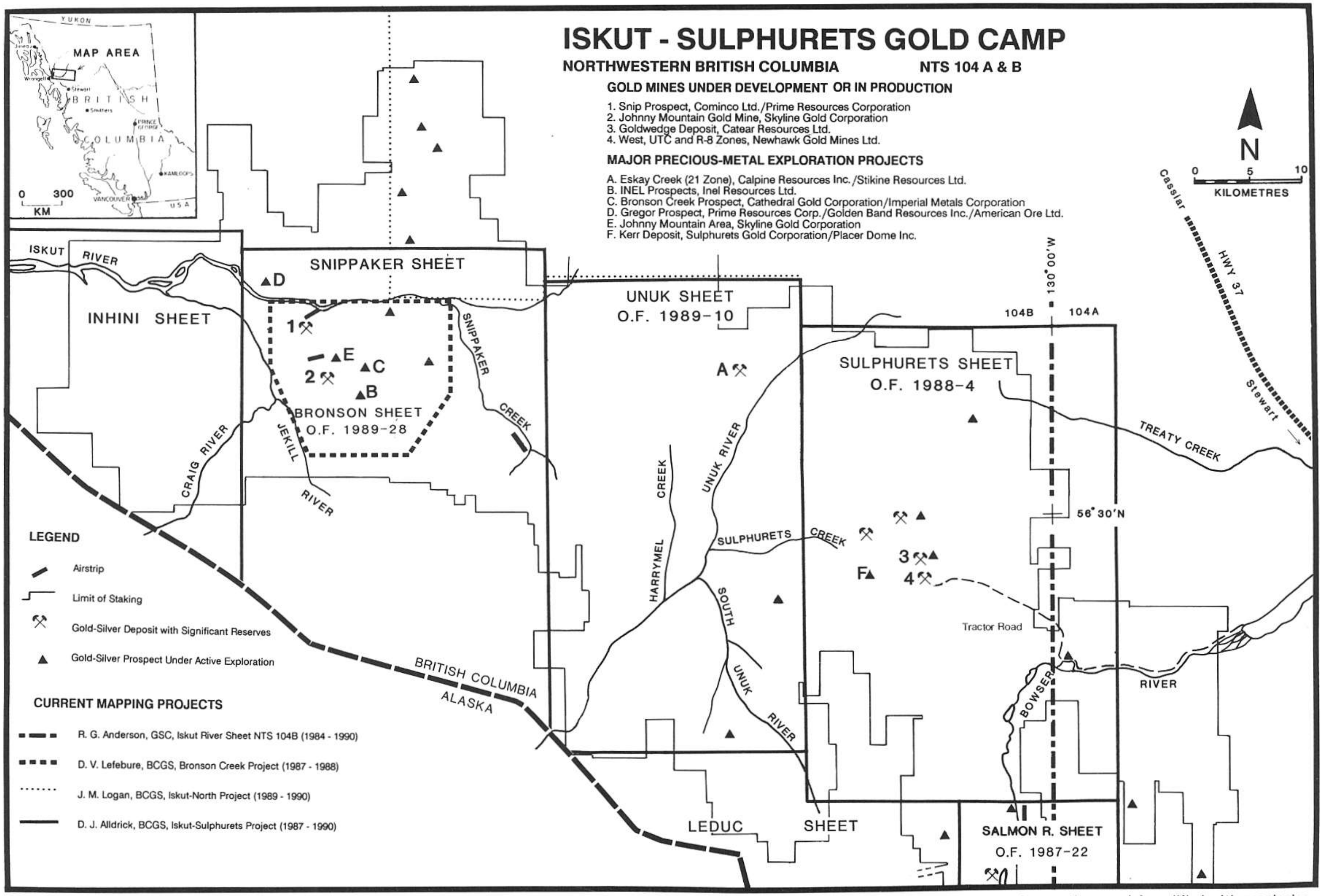
Intermittent activity continued for the next 60 years. Until the 1950s the focus of exploration was a large gossanous cliff, Red Bluff, located on lower Bronson Creek. In 1954 Hudson Bay Mining and Smelting Company discovered the Pick Axe vein on the alpine slopes of Johnny Mountain.

The porphyry copper boom in the early 1970s drew prospectors and companies into the area once more; the Inel, Ray, Shan and Tami properties were staked and explored with little success.

The new golden age began with the 1980 staking of the Reg claims by Skyline Explorations Limited. Staking rushes followed in the wake of high-grade gold discoveries on Johnny Mountain and Bronson Creek (Reg and Snip claims). The Johnny Mountain gold mine was opened in August 1988; preproduction development is underway at the Snip deposit. Surface and underground exploration continues at Inel.

GEOLOGIC SETTING

The map area is situated in the southern Boundary Ranges of the Coast Mountains physiographic belt, on the western edge of the Intermontane tectonic belt. The northern two-thirds of the area is in the Stikine Terrane; the rest is part of the Coast plutonic complex (Wheeler *et al.*, 1988).



Map courtesy of Teuton Resources Corporation, used & modified with permission.

Figure 1-12-1. Iskut-Sulphurets gold camp: precious metal deposits and current mapping projects.

Anderson (1989) proposed a stratigraphic column for the whole of NTS 104B, distinguishing four tectonostratigraphic assemblages bounded by unconformities:

- Tertiary Coast plutonic complex;
- Middle and Upper Jurassic Bowser overlap assemblage;
- Triassic-Jurassic volcanic-plutonic arc complexes; and
- Paleozoic Stikine assemblage.

Three of these assemblages are represented in the map area which is underlain by a thick (more than 5 kilometres) succession of sedimentary and volcanic rocks that range in age from late Paleozoic to Quaternary. Most strata are Upper Triassic to Lower Jurassic volcano-sedimentary arc-complex lithologies characterized by rapid facies changes. Pleistocene and Recent basaltic flows and tephra are preserved along the Iskut River, in Snippaker Creek and at Lava Lakes. Strata have been cnt by a variety of plutons representing at least four intrusive episodes spanning Late Triassic to Quaternary time. These include synvolcanic plugs, sills and stocks, minor dike swarms, isolated dikes and sills, as well as the batholithic Coast plutonic complex. The stratigraphic sequence has been folded, faulted and metamorphosed mainly during Cretaceous time, but some Paleozoic strata are polydeformed and probably record an earlier deformational event.

STRATIGRAPHY

We have divided the volcanic and sedimentary rocks of the map area into four main lithostratigraphic sequences or packages, ranging in age from Paleozoic to Quaternary (Figure 1-12-2, 3). The area is poorly fossiliferous and as yet lacks radiometric data to constrain ages of units. Stratigraphic correlations within the map area are based on lithologic similarity; relative ages are based on stratigraphic position and degree of deformation and metamorphism. Correlations with Stewart-Sulphurets stratigraphy (Alldrick, 1985; Britton and Alldrick, 1988) are complicated by a combination of facies changes and north-trending, high-angle regional faults across the Unuk River valley.

Contacts between lithostratigraphic sequences within the map area are not well exposed: commonly they are covered with moraine, disrupted by faults, or invaded by large intrusions such as the Lehto batholith and the Coast plutonic complex.

PALEOZOIC (MAP UNIT 1)

The Paleozoic Stikine assemblage (Monger, 1977), also known as the Asitka assemblage (Wheeler and McFeely, 1987; Wheeler *et al.*, 1988), has not been unambiguously identified in the map area. It is known to crop out extensively west of the Craig River (Kerr, 1948; Anderson, 1989) and northeast of Mount Verrett (Logan *et al.*, 1990, this volume). The Stikine assemblage is characterized by thick, platformal carbonate sequences, coralline reefs, and mafic to felsic volcanics. Two phases of penetrative deformation have been observed in these strata (Anderson, 1989).

Rocks tentatively assigned to the Paleozoic crop out along the Jekill and Craig River valleys, upper Olatine Creek and between Mount Zara and Mount Lewis Cass. They include abundant fine-grained, thinly layered, biotite-rich quartzofeldspathic gneiss, phyllite, metawacke, metatuff and thin recrystallized limestone (marble). The protolith of the fine-grained quartzofeldspathic gneiss is not known. Where it is interbedded with marble it is interpreted as metawacke. Almost identical rocks, with relict coarse plagioclase clusters, are considered metatuffs. Coarse volcaniclastic textures are absent. These gneisses were probably derived from tuffaceous siltstones and sandstones, with minor ash and crystal tuffs.

These rocks are the most structurally complex in the area; polyphase deformation is likely. The contact between Paleozoic rocks and overlying Mesozoic strata is probably an unconformity, based on relative states of deformation. Apart from crinoid stem debris in limestone at 1280 metres elevation south of the Johnny Mountain minesite, no macrofossils were found.

Gneissic metamorphosed sediments preserved as small screens within the Coast plutonic complex may be the oldest rocks in the map area. They form rafts a few metres thick and up to 10 metres long, entirely surrounded by Tertiary granite, consisting of very fine grained, thinly layered, felsic and mafic-rich gneisses with ductile deformation fabrics. They have been seen only northeast of Mount Lewis Cass but their distribution is unknown due to limited mapping within the Coast complex.

MESOZOIC (MAP UNITS 2 AND 3)

Most of the stratified rocks in the map area are Mesozoic. They are exposed north and south of the Iskut River, along Snippaker and Monument creeks, and from Lehua Mountain to Olatine Creek.

Mesozoic strata form a thick (3 kilometres) sequence of mixed volcanic and sedimentary rocks. Facies changes, minor unconformities and the paucity of distinctive marker horizons make stratigraphic correlation difficult. Extrusive rocks are mostly volcaniclastic: pyroclastic units with derived epiclastic facies. Plagioclase, pyroxene and hornblende are common phenocrysts; distinctive coarse potassium feldspar is minor but important. Compositions range from basalt to rhyodacite, but most are andesite to dacite. Sedimentary rocks are volcanic-derived siltstone, wacke and conglomerate with minor amounts of limestone either as relatively pure lenses or as calcareous mudstones. Limestone decreases upwards in the section and is rare in Hazelton strata.

Based on two fossil determinations from Snippaker Mountain, ages span at least Norian to Toarcian(?) time (Lefebure and Gunning, 1989). Older rocks are correlated with the Stuhini Group on the basis of age and volcanic composition. Overlying strata are correlated with the Hazelton Group based on distinctive potassium feldspar and hornblende crystal tuffs and overall similarities with Hazelton strata to the east (Hancock, 1990, this volume; MacLean, 1990, this volume; Britton and Alldrick, 1988; Britton *et al.*, 1989).

UPPER TRIASSIC (MAP UNIT 2)

Triassic strata have been divided into four main groups: sediments (Unit 2S); intermediate volcanics (Unit 2V); melanocratic basaltic tuffs (Unit 2M); and leucocratic dacitic tuffs (Unit 2L).

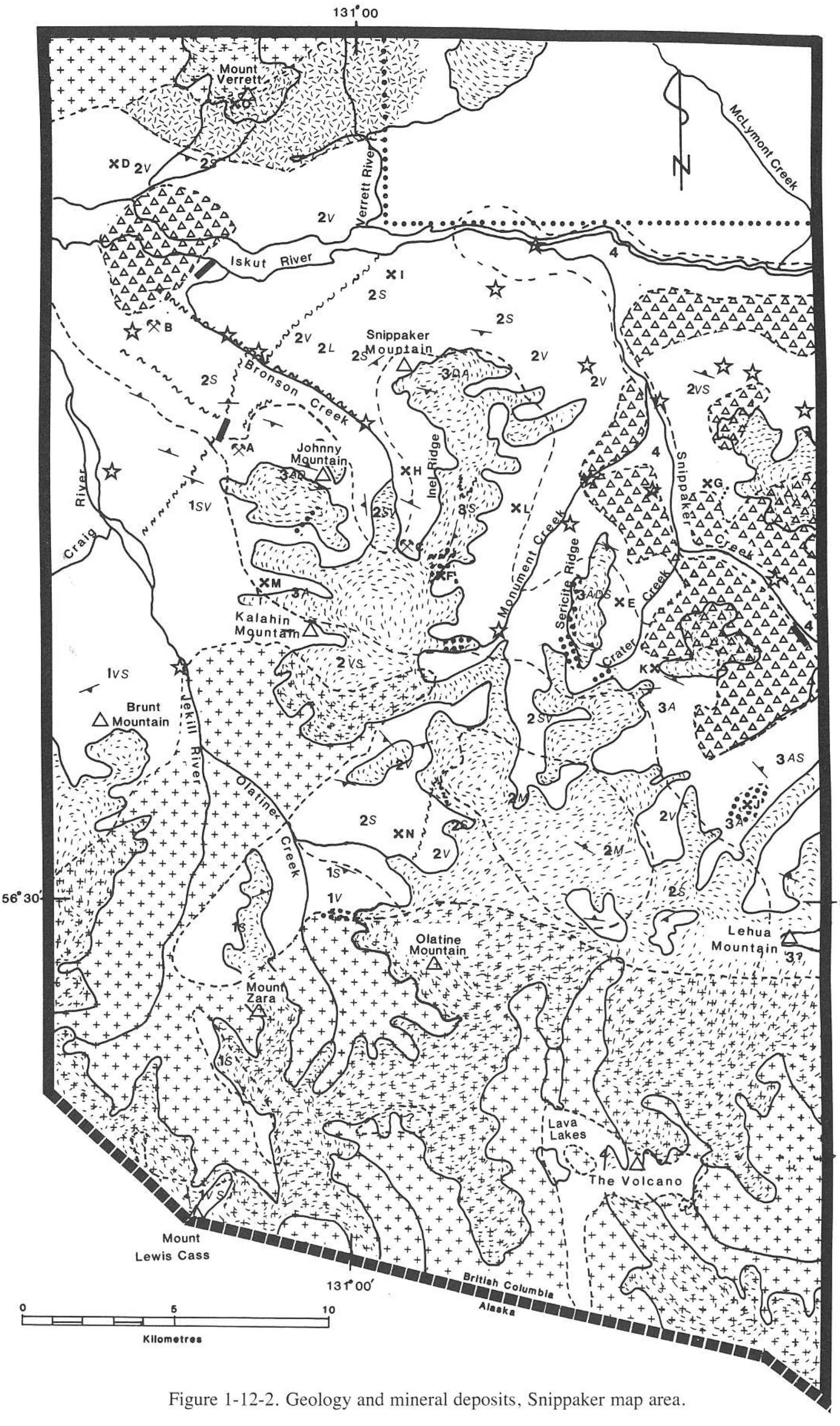


Figure 1-12-2. Geology and mineral deposits, Snippaker map area.

LEGEND

INTRUSIVE ROCKS

Tertiary		Biotite Quartz Monzonite to Granite
Early Jurassic		Lehto and Iskut River Plutons Monzonite; Quartz Diorite; K-feldspar Porphyry Hornblende Monzodiorite
Late Triassic		Hornblende Diorite

STRATIFIED ROCKS

Cenozoic	4	Basalt Flows, Tephra
Mesozoic LJr	3	Andesitic (A) to Dacitic (D) Volcanics (with <40% sedimentary strata (S))
Mesozoic UTr	2	Sediments (S) and Intermediate Volcanics (V) Leucocratic Dacitic Tuffs (L) Melanocratic Basaltic Tuffs (M)
Paleozoic	1	Metamorphosed Sediments (S) and Tuffs (V) (may include metamorphic equivalents of Triassic rocks)

SYMBOLS

Contact . . . .

Compositional layering (bedding, foliation)

Airstrip . . . .

RGS gold values > 90th percentile . . . .

Limit of mapping . . . .

Mine, developed prospect . . . . A

Prospect . . . . D

Gossan . . . .

PROSPECTS

	NAME	COMMODITY
A	Johnny Mountain	Au,Cu,Ag
B	Snip	Au,Cu,Ag,Pb,Zn
C	INEL	Au,Ag,Cu,Zn,Pb
D	Gorge/Gregor	Au,Ag
E	Sericite Ridge	Au,Cu,Fe
F	Khyber Pass/Pyramid Hill	Au,Cu,Zn
G	Josh	Au,Cu,Pb,Zn
H	Cathedral Gold	Au
I	Bug Lake	Au,Pb,Cu,Zn
J	Pins	Au,Ag,Cu,Zn,Pb
K	Lake Area	Au
L	Wolverine	Au,Cu,Pb,Zn
M	Pez-Dan	Au,Ag,Cu,Pb,Zn
N	Still	Au,Ag,Pb,Zn
O	Mount Verrett	Au

Most of the volcanic rock in the Triassic succession is basaltic to andesitic with plagioclase and pyroxene as the principal phenocrysts (Unit 2V). Pyroxene-phyric tuffs are characteristic of the Stuhini Group. Pyroclastic units are more common than flows, but many outcrops are massive and difficult to classify. For example, a thick, monotonous sequence of fine-grained, medium to dark green, feldspar porphyry andesite underlies the lower slopes of Mount Verrett and extends aross the Iskut River to Bug Lake. These rocks are moderately to completely recrystallized north of the Iskut and could be either massive crystal tuffs or flows. There are some lapilli tuffs and tuff breccias around Bug Lake, but fragmental textures are generally absent.

Unit 2S on Figure 1-12-2 indicates areas underlain mainly by sedimentary rocks. In the south these are mostly siltstone with minor fine-grained wacke. Thin rhythmic bedding is common. In the north they are interbedded mudstone, lithic wacke, feldspathic wacke, minor conglomerate and limestone lenses, with locally abundant fine-grained volcaniclastic material–ash tuff or volcanic sandstone. These rocks host the Snip deposit and other prospects uphill from Bug Lake and on lower Bronson Creek. Near the top of the section on Winslow Ridge, west of Snippaker Mountain, a limy sandstone bed contains *Epigondoella*, a middle to late Norian conodont (Lefebure and Gunning, 1989).

Unit 2M is exposed only in the south of the map area, near Olatine Mountain. It consists of over 500 metres of mafic crystal tuffs, breccias, lahars and derived wackes and siltstones. The tuffs are distinctive rocks consisting of dark green pyroxene crystal tuffs with a dark chloritized matrix. Colour index is greater than 50. Outcrops characteristically have a hackly appearance due to the resistant weathering of coarse pyroxene phenocrysts. Sediments associated with these distinctive rocks are thin-bedded, pyroxene-bearing wackes, siltstones and minor conglomerate. Some sedimentary rocks occur as large blocks (up to 10 metres across) in a melanocratic, tuffaceous matrix. These probably result from such high-energy transport mechanisms as debris flows. The restricted areal distribution of this unit suggests an unusual but perhaps long-lived eruption of magma rich in crystal cumulates. These rocks are intercalated with, and grade upwards into, more typical medium grey-green pyroxene-phyric volcaniclastics.

Unit 2L consists of light grey-green, waxy, dacitic pyroxene-plagioclase crystal and lapilli tuffs. It has been identified only on Winslow Ridge and appears to be a conformable sequence within Unit 2S. A nearby drillhole has intersected over 200 metres of underlying andesitic tuffs, lapilli tuffs and green tuffaceous siltstones.

East and west of Snippaker Creek scattered lenses and blocks of white granular recrystallized limestone are set in massive plagioclase and hornblende-phyric andesitic ash tuffs, lapilli tuffs, tuff breccias and volcanic conglomerates with minor interbedded clastic sedimentary units and thin limestone beds. Most fragments range from centimetre-sized chips to metre-sized blocks, but locally include masses 75 metres thick and 300 metres long. Internal bedding in these longer blocks is not well preserved. This sequence of probable olistostromes has been interpreted as the lowest part of the Upper Triassic succession.

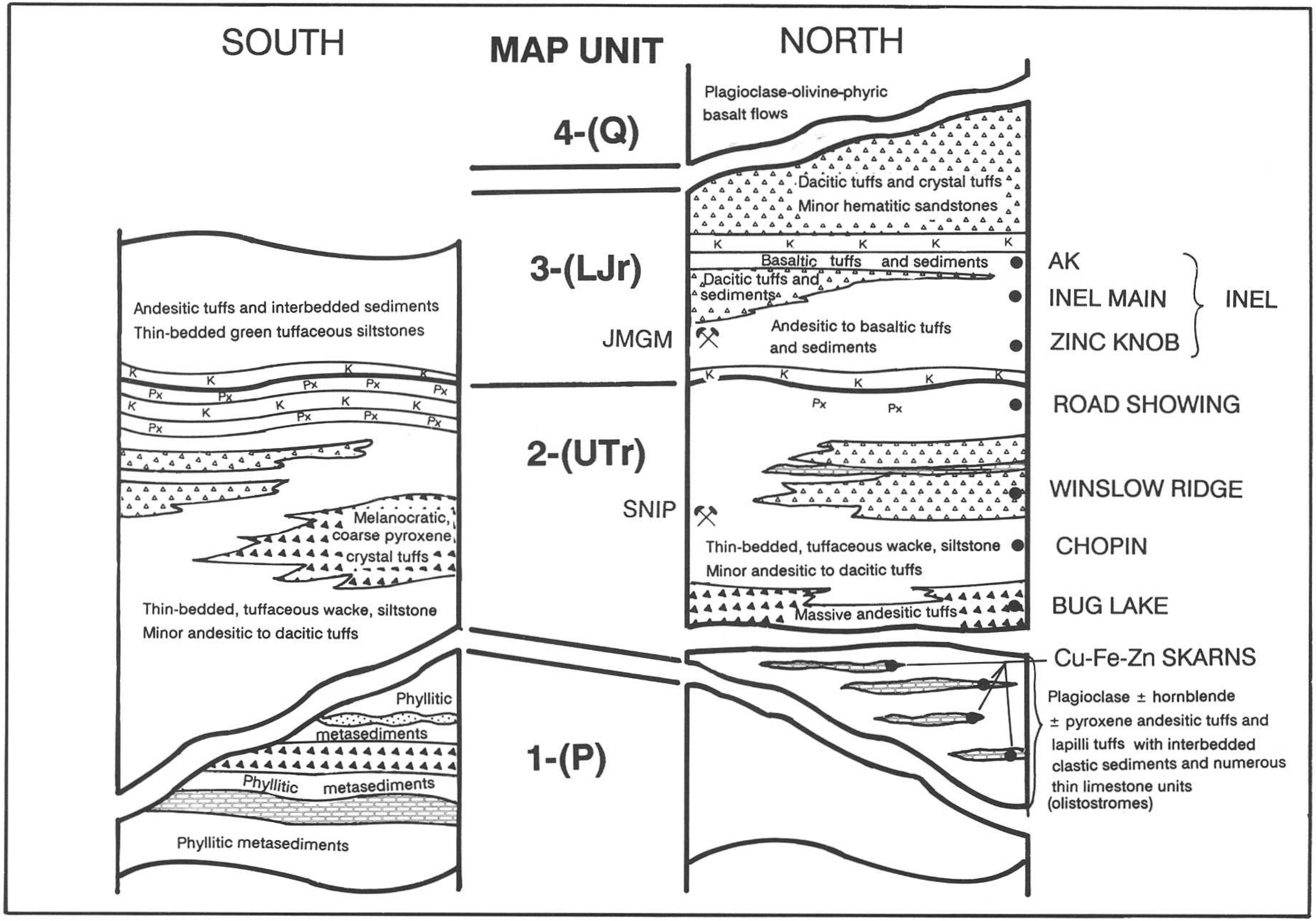


Figure 1-12-3. Schematic stratigraphic columns, Snippaker map area JMGM = Johnny Mountain Gold Mine; K = K-feldspar + plagioclase ± hornblende porphyritic andesitic tuffs; Px = pyroxene-phyric andesitic tuffs.

LOWER JURASSIC (MAP UNIT 3)

Jurassic strata are best exposed on the summit slopes of Johnny Mountain, Snippaker Mountain and Inel Ridge, but they extend southeasterly toward Pins Glacier and Lehua Mountain. Rocks are mainly andesitic to dacitic fragmental volcanics with minor basaltic tuffs and lesser amounts of siltstone, wacke and conglomerate. Marked lateral facies changes, lithologic heterogeneity and variable rock colours (grey, green, maroon and mottled combinations of these) are common.

On Johnny Mountain three distinct rock units, each at least 100 metres thick, make up the bedrock above Johnny Flats (an alpine plateau at 1000 metres elevation). The lower unit includes the "Mine Series" of the Johnny Mountain gold mine. It consists of medium to dark grey, locally greenish, plagioclase-phyric andesitie to dacitic erystal and ash tuff with some monomictic to oligomictic lapilli tuff and agglomerate. In some of these rocks plagioclase phenocrysts are rounded, suggesting they have been reworked. Zones of subrounded volcanic conglomerate also occur. This unit is mostly massive but at the Johnny Mountain mine rare compositional layering is preserved despite extensive veining, alteration and shearing. The middle unit conformably overlies the lower unit and consists of grey and tan dacitic volcanic rocks. They include flow-banded and welded ash tuffs as well as well-bedded ash and lapilli tuffs with rhyodacitic clasts. Overlying these dacites, the upper unit comprises dark grey-green, flaggy, well-foliated basaltic andesite ash tuffs with minor siltstone and wacke interbeds.

On Snippaker Mountain and extending southward, the Jurassic sequence includes at least 300 metres of matrix-supported, polymictic pebble to cobble conglomerate with minor siltstone and wacke interbeds. These are commonly orange to buff weathering due to pervasive ankeritic alteration. The unit grades laterally and upwards into green volcanic conglomerate and lithic lapilli tuff but the colour change from orange to green does not coincide with lithologic contacts. These conglomerates are locally overlain by thin-bedded, salt-and-pepper lithic arenite and siltstone with carbonized plant remains.

Light grey and green dacitic volcanics cap much of Snippaker-Inel Ridge. Some of these are similar to the middle unit on Johnny Mountain. The base of this dacitic section on the Inel property is marked by bedded potassium feldspar-hornblende-plagioclase crystal tuffs. Similar rocks also occur at the head of Crater Creek. These are texturally similar to two-feldspar or "Premier porphyry" rocks which occur at the top of the Unuk River formation and delineate the base of the Betty Creek formation in the Stewart and Sulphurets areas. Rocks east of Snippaker Mountain include mottled maroon and grey dacitic lapilli and ash tuffs overlain by hematitic siltstone with calcareous concretions. Similar interbedded dacitic volcanics and hematitic volcanic sediments crop out along Pins Glacier north of Lehua Mountain. These rocks strongly resemble the thick dacitic packages of the upper Hazelton Group on the Colagh property (MacLean, 1990, this volume) and between the Unuk River and Bruce Glacier (Britton *et al.*, 1989). These mixed dacitic pyroclastic and epiclastic sequences are characteristically hematitic, perhaps due to subaerial deposition. The Eskay Creek deposit of Calpine Resources Incorporated (Figure 1-12-1) is hosted by equivalent stratigraphy in a local paleotopographic depression indicated by minor marine turbidite units and a thick lens of pillow lavas.

Marine turbidites of the Bowser Lake Group were not found in the map area.

QUATERNARY (MAP UNIT 4)

Pleistocene and Recent basaltic lava flows, cones and tephra occupy the valleys of the Iskut River, Snippaker Creek and Lava Lakes. All flows are olivine and plagioclase phyric; many are strongly vesicular. These volcanics are part of the north-trending Stikine volcanic belt of Miocene to Quaternary eruptive centres (Souther, 1977).

The Iskut River flows vented near Palmiere Creek, just east of the map area (Alldrick *et al.*, 1989; Read *et al.*, 1989). A whole rock age for the lowermost flows is 70 000 years (Grove, 1986). Radiocarbon ages (Read *et al.*, 1989) of 8760 ± 150 years from tephra and 2610 ± 70 years in overlying sediments deposited in lava-dammed lakes bracket the youngest volcanic event.

The Volcano, a small peak near the Alaska–British Columbia border, may be the youngest eruptive centre in British Columbia. Flows from this vent dammed the Unuk River (via a tributary called Lava Forks) and formed the aptly named Lava Lakes. Carbon 14 dates on the lower flows give an age of 130 years (Grove, 1986). The youngest flows are devoid of vegetation. Glaciers nearby are dusted with black basaltic tephra from a 1904 eruption.

INTRUSIVE ROCKS

Intrusive rocks underlie about one third of the map area. At least four intrusive episodes are represented, spanning Triassic to Quaternary time. Paleozoic intrusions have not been identified.

PLUTONS

The oldest plutons are sills, dikes and plugs of hornblende diorite that are contemporaneous with Triassic hostrock volcanics. They are especially common in andesites located north of the Iskut River. Most of these are too small to be shown on Figure 1-12-2 but on Mount Verrett there is a large hornblende diorite stock of this type. The rock is texturally similar to the andesites it intrudes and consists of mesocratic medium to dark grey, fine-grained, anhedral granular diorite with fine plagioclase phenocrysts. The diorite is largely recrystallized and pervasively propylitically altered. Near its contact with the Coast batholith it has pegmatitic zones up to 50 centimetres wide by 6 metres long consisting of coarse bladed intergrowths of hornblende and plagioclase with minor biotite. Against the batholith it is migmatitic with a swirled foliated fabric in the diorite that is cut by leucogranite dikes. Contacts with andesite are indistinct and may be in part gradational.

Jurassic intrusions include synvolcanic hypabyssal stocks as well as phaneritic plutons of considerable size. Synvolcanic intrusions are thought to be comagmatic and coeval with extrusive rocks. Examples include felsite stocks on

Johnny Flats and the Inel property. These are leucocratic to holofelsic, cream to tan, porphyritic rocks with fine feldspar and quartz phenocrysts set in an aphanitic groundmass. Contacts are altered and sheared but the stocks appear to form sheet-like bodies that are crudely conformable with enclosing strata. On the Inel property the felsite stock is associated with a small felsite dike swarm.

Phaneritic intrusions of probable early Jurassic age include the Lehto batholith, the Iskut River stock and smaller plugs and dikes such as the Red Bluff porphyry. A common feature of these intrusions is the presence of coarse (up to 5 centimetres) potassium feldspar phenocrysts.

The Lehto batholith, a monzonitic to dioritic pluton first recognized in 1988 (Britton *et al.*, 1989), is the largest of these early Jurassic plutons and crops out along Snippaker Creek. Three phases have been identified. The earliest is pale green to white, medium-grained, anhedral-granular biotite-hornblende quartz diorite. It is commonly saussuritized and locally displays a weak foliation. It is intruded by mottled, pink, green, white and black medium-grained, subhedral-granular, hornblende-biotite quartz monzonite and monzodiorite. The third and youngest phase occurs as isolated masses, dikes and marginal zones in some parts of the pluton. It is hornblende monzodiorite to monzonite with coarse potassium feldspar phenocrysts set in a fine to medium-grained groundmass. The Iskut River and smaller stocks consist mainly of this third phase.

These early Jurassic plutons are texturally similar to the Texas Creek batholith near Stewart which shows a close spatial and temporal relationship with the Silbak Premier gold, silver and base metal deposit (Alldrick, 1987). Although no genetic link with mineralization has yet been established, these rocks are an important guide for exploration. Other dioritic and quartz dioritic intrusions of probable Jurassic age occur locally. All are minor, isolated stocks or plugs.

The largest intrusive mass in the map area is the Coast Mountains batholith which occupies the southern quarter and northwestern corner of the map area. It consists of medium-grained biotite and biotite-hornblende granite, granodiorite and rarely quartz diorite. Very little of it has been mapped. It is distinguished from Jurassic plutons by its fresh appearance, lack of foliation and shearing, minimal saussuritization, and abundance of quartz. Biotite is either the sole mafic mineral or else is much more common than hornblende. Near Mount Lewis Cass the margins of the batholith show very passive emplacement into gneissic country rocks. There are lit-par-lit sills of holofelsic, quartz-rich leucogranite and trains of gneissic xenoliths which show no sense of rotation. Locally, and within a few metres of the contact, some granites have a weak foliation parallel to that of the country rocks. Elsewhere the batholith crosscuts strata and its border is irregular and associated with much diking. There is little or no hydrothermal alteration or skarn developed along the intrusive contacts despite the presence of limestone units in the Paleozoic country rocks. The age of these rocks is probably middle Eocene based on potassium-argon dating near Stewart (Alldrick *et al.*, 1987) and Alice Arm (Carter, 1981).

DIKES

Isolated dikes and minor dike swarms occur locally in the map area. Strongly foliated basaltic dikes on Johnny Mountain are apparently feeders to overlying basaltic andesite tuffs. Felsite dike swarms are exposed south and west of Snippaker Mountain. Andesite, dacite and microdiorite dikes are found locally in volcanic sequences and are probably synvolcanic. Coarse potassium feldspar dikes on the Inel property may be feeders to overlying crystal tuffs. Around the margins of the Coast Mountains batholith are white, fine to coarse-grained, inequigranular, quartz-rich holofelsic leucogranite dikes that are apparently offshoots of the batholith.

Widespread biotite and hornblende lamprophyre dikes cut all other rock types including the Coast Mountains batholith. They are typically isolated and narrow (up to 2 metres wide). At the head of Monument Creek a hornblende lamprophyre dike swarm is traceable for at least 2 kilometres. It consists of a 15-metre-wide zone of up to a dozen anastomasing dikes, each less than a metre wide. The age of these dikes is probably Oligocene (Alldrick *et al.*, 1987; Carter, 1981).

Basalt dikes related to the Quaternary Stikine volcanic belt are rare, mostly occurring near eruptive centres.

STRUCTURE

FOLDS

Paleozoic rocks exhibit the strongest deformation. Folds range from crenulations through upright chevrons to recumbent isoclines with fold amplitudes of 100 metres. The largest folds plunge gently east-northeast. Crenulations and contorted open folds are also developed adjacent to faults in fine-grained sediments and tuffs of any age. These structures die out within a few metres of the fault zone.

At a regional scale the Mesozoic lithostratigraphic sequences form flat-lying packages, but Triassic and Jurassic strata show mesoscopic folds. Some of these are primary depositional features such as convolute layering in welded tuffs, flow banding and soft-sediment slumps.

FOLIATION

Many rocks, but especially fine-grained sediments, mafic tuffs and limestones, show locally intense foliation, boudinage and transposition of primary layering. Rock composition, especially mica content, largely determines the amount of foliation developed. A good example of this occurs in Johnny Mountain cirque where mafic dikes cut felsic lapilli tuffs and foliation is perpendicular to the bedding. The dikes are strongly chloritized and exhibit strong foliation and platy parting. The leucocratic dacite shows only slight stretching of lapilli and a weak flaggy cleavage.

Phyllites have formed adjacent to both flat and upright faults in rocks of appropriate composition.

FAULTS

FLAT FAULTS

There is a widespread subhorizontal cleavage in most Triassic and some Jurassic rocks. Locally this is expressed in subhorizontal faults between blocks of differing competence. An example of this is the contact between Jurassic volcaniclastic and Triassic sediments on Johnny Mountain. The underlying siltstones exhibit folding, shearing and recrystallization that decreases in intensity away from the fault. Overlying dacitic volcaniclastic rocks which act as a competent unit also show increased strain near the fault but deformation is much weaker, amounting to minor shearing and recrystallization. A similar flat fault marks the base of the thick dacitic volcanic package that caps Inel Ridge between Snippaker Mountain and the Inel camp. Interpretations proposed for these widely recognized structures include: thrust faults, extensional detachment faults, or unconformities. They are not thrust faults; they do not place older rocks on top of younger ones. Lateral displacement is not known, but is suspected to be in the order of a few hundred metres at most.

HIGH-ANGLE FAULTS

High-angle faults are common in the area and appear to postdate flat faults. Some form well-defined lineaments traceable for kilometres and visible in radar images and air photographs. Most have small displacements, in the order of tens of metres. Maximum displacement on mappable faults like those seen on Johnny and Snippaker mountains is in the order of a few hundreds of metres. Most major faults strike northeasterly or northwesterly.

METAMORPHISM

Metamorphic rank is generally low although recrystallization is complete. Local enclaves of staurolite and garnet-bearing rocks in the Paleozoic succession indicate that epidote-amphibolite facies is the highest metamorphic grade preserved in the map area. Younger rocks show typical lower greenschist facies metamorphism. Propylitic mineral assemblages are common: hornblende and pyroxene are replaced by chlorite, plagioclase is saussuritized, clay constituents form white mica.

Contact metamorphism occurs within 1 to 2 kilometres of the Coast Mountains batholith. The main effects are recrystallization with coarsening of grain size and replacement of mafic minerals by metamorphic biotite. Thin garnetite lenses occur locally in marble.

MINERAL DEPOSITS

The northern part of the Snippaker map sheet has seen intense exploration activity over the last few years. The area includes the Johnny Mountain gold mine and also the Snip and Inel properties which are in advanced stages of exploration. Detailed descriptions of the Johnny Mountain and Snip deposits are presented in Alldrick *et al.* (1989b). Several other mineral occurrences and large claim blocks are in early stages of exploration.

All of the major deposits in this area occur in quartz veins or silicified shear zones that crosscut strata. Copper-iron skarn showings are known to occur east of Snippaker Creek but have received only preliminary exploration work.

VEINS

JOHNNY MOUNTAIN GOLD MINE (STONEHOUSE; MINFILE 104B 107)

Johnny Mountain gold mine officially opened in August, 1988 and commenced commercial production of gold, silver and copper in November. Total production to the end of July, 1989 was 1050 kilograms of gold, 1850 kilograms of silver, and 430 000 kilograms of copper (George Cross News Letter No. 187, 1989). Silver and copper grades are not usually reported but mill feed runs about 27 grams per tonne silver and 0.7 per cent copper (W.L. Millar, personal communication, November 1989). Reserves (Table 1-12-1) are in six mineralized veins: Discovery (Cloutier), 16, Pickaxe, Goldrush, Zephrin and Victoria. Several other mineralized zones have recently been discovered, but are not yet fully explored.

The hostrocks (the "Mine Series") are part of a volcanic package consisting of interbedded andesitic and dacitic volcaniclastics and volcanic sediments cut by two-feldspar porphyry dikes. Andesites and dacites range from fine-grained ash tuffs to coarse lapilli tuffs. The volcanic sediments are a bedded sequence ranging from mudstones to coarse conglomerates.

Five of the main veins are sub-parallel and the Zephrin may represent a sheared and reoriented vein segment. The veins strike approximately 065°, dip steeply northwest, and are roughly parallel to the strike of the hostrocks. Where the veins cross lithological contacts, they thicken and ore grades are often higher. The veins are also enriched where they intersect cross-structures. These structures are quartz-carbonate veins with green mica (annite or chlorite) and occasional specular hematite.

The veins are surrounded by a distinctive alteration halo which is mineralogically symmetrical, but wider in the hangingwall. Outward from the the vein wall the alteration sequence is: massive potassium feldspar and ankerite alteration 1 to 2 metres from the vein; a quartz-pyrite stringer zone up to 5 metres from the vein; and disseminated pyrite up to 10 metres from the vein. Pyrite textures in quartz veins can be used as a guide to gold content (P. Metcalfe, personal communication, August 1989). Coarse cubic pyrite is invariably barren. Gold is always associated with fine grained or "milled" pyrite.

Ore consists of quartz veins, 1 to 2 metres wide, commonly with about 25 per cent pyrite, 1 to 2 per cent chalcopyrite and trace amounts of sphalerite, galena and pyrrhotite. Typical exposures on the 16 vein show a 1 to 1.5-metre vein core of massive bull quartz with patchy aggregates and disseminated sulphides, bordered by massive pyritic marginal zones 50 centimetres wide that carry the best gold values. Gold occurs with silver in electrum; coarse visible electrum is common.

Two post-ore fault sets are present within the mine workings: subvertical faults and flat "quartz faults". Subvertical faults strike north, dip about 60° either east or west and may represent conjugate sets. These faults truncate the veins. Quartz faults are low-angle brittle faults with minor offsets (<10 metres) which often contain barren ribbon quartz. Quartz faults offset the hangingwall to the north and complicate shrinkage stoping.

SNIP (TWIN ZONE; MINFILE 104B 250)

The Snip deposit occurs in the thick sedimentary sequence which underlies the volcanic package on Johnny Mountain. The main mineralized zone, the Twin zone, is a shear-hosted vein that strikes 120° and dips 45° to vertical, averaging 60° southwest.

Hostrocks are massive to bedded siltstone and feldspathic wacke. Along most of its length the vein is divided into two parts by a dark grey, fine-grained dike with fine black biotite phenocrysts known as the Biotite Spotted Unit (BSU). In the west drift the Twin zone averages 4.5 metres wide including a 2.5-metre band of barren, postore BSU dike in the core.

Three ore types have been recognized within the Twin zone:

- Massive sulphide ore with pyrite>pyrrhotite, minor sphalerite and rare arsenopyrite, galena, molybdenite and chalcopyrite;
- Crackle quartz ore consists of shattered quartz vein infilled with green mica and chlorite and disseminated sulphides;
- Streaky quartz ore consists of quartz laminae within strongly sheared and altered country rock.

Reserves are shown in Table 1-12-1.

INEL (MINFILE 104B 113)

Inel mineralization is hosted by a volcanic-sedimentary sequence consisting of sandstones, turbidites, minor cobble conglomerates and rare limestones interbedded with basalt tuffs and lapilli tuffs. Stratigraphy is cut by a swarm of mafic to felsic dikes, including distinctive potassium feldspar megacrystic dikes. Like the Johnny Mountain mine to the northwest, the area of the deposit is underlain by a large, crudely conformable felsite stock.

Mineralization occurs as numerous sulphide and quartz-sulphide veins that have a variety of orientations, but which appear to be concentrated along specific lithologic horizons that have been preferentially fractured. No single vein is sufficiently large to allow calculation of reserves, but the large number of veins and their gold grades justifies continued exploration. The Main zone hosts the highest grades obtained to date and occurs as a series of veins along a basalt-sandstone contact characterized by extensive pyritization. Recent dilling has focused on a mineralized intrusive breccia in the footwall of one of the potassium feldspar dikes (George Cross News Letter No. 199, 1989).

SKARNS

Although high-grade gold veins have been the focus of recent exploration, skarn deposits have been discovered along the margin of the Lehto batholith east of Snippaker Creek. The area coincides with a cluster of strongly anomalous gold values obtained by the Regional Geochemical Survey stream sediment survey (National Geochemical Reconnaissance, 1988). On the Josh claim, mineralization consists of replacement lenses of chalcopyrite, sphalerite and magnetite in actinolite-epidote-garnet skarn (Scott, 1983; McLeod, 1988). Grab sample assays as high as 321.6 grams per tonne silver, 24.8 per cent zinc and 32.8 per cent copper have been reported (McLeod, 1988). Gold values up to 3.7 grams per tonne have been obtained from a quartz breccia vein that cuts an epidote-quartz-garnet skarn. Hostrocks are typically calcareous siltstone or interbedded siltstone and andesitic tuffs, and, rarely, limestone.

Trace malachite staining occurs along the contact of the Coast plutonic complex in the southern part of the map area. Compared to the early Jurassic Lehto batholith, the Tertiary Coast batholith seems very weakly mineralized.

DISSEMINATED SULPHIDES

Large areas of gossanous disseminated pyrite and pervasive argillic to sericitic alteration occur between Monument and Snippaker creeks (Sericite Ridge); northwest of Lehua Mountain (Pins Ridge); and at the southern end of Inel Ridge (Khyber Pass; Pyramid Hill). These obvious exploration targets are difficult to prospect and evaluate because of their size (>5 square kilometres) and extensive surface leaching. Similar large gossans in the Sulphurets area are known to include large-tonnage low-grade gold, copper-gold and gold-molybdenum deposits such as the Snowfield, Kerr and Sulphurets prospects (Britton and Alldrick, 1988).

SUMMARY

The map area is underlain by Paleozoic Stikine assemblage, upper Triassic Stuhini Group and lower Jurassic Hazelton Group volcanic and sedimentary rocks that have been intruded by Triassic, Jurassic and Tertiary plutons. Paleozoic and Mesozoic strata are probably separated by an unconformity but relationships are obscured by faulting and intrusion. The rocks record a long and complex history of fracturing, faulting and folding which is not yet resolved.

The main targets of exploration are mesothermal quartz veins rich in gold and silver. Other potential deposit types include copper-iron-zinc skarns and large-tonnage, low-grade disseminated copper-gold deposits. So far all precious metal deposits have been found in Mesozoic strata, which may indicate a widespread but stratigraphically constrained mineralizing event associated with early Jurassic plutons.

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NOTES

GEOLOGY OF THE FORREST KERR CREEK AREA, NORTHWESTERN BRITISH COLUMBIA (104B/15)

By J.M. Logan, V.M. Koyanagi and J.R. Drobe

KEYWORDS: Regional geology, Stikine assemblage, Stuhini Group, McLymont Creek, skarns, veins, massive sulphides, gold, base metals.

INTRODUCTION

This report summarizes the results of 1:50 000-scale mapping completed during the 1989 field season in the Forrest Kerr map area (104B/15). This is the second year of a 4-year program of regional mapping and mineral deposit modelling between the Iskut and Stikine rivers in northwestern British Columbia. In 1988 mapping was centred on Galore Creek and map sheets 104G/3 and 4 (Figure 1-13-1). The 1989 program extended previous mapping southeast to the Iskut River to delimit Paleozoic Stikine assemblage and Upper Triassic Stuhini Group stratigraphy, to assess the mineral potential of the area and update the mineral inventory database (MINFILE).

The 1989 map area is located about 15 kilometres northwest of Eskay Creek, 30 kilometres southeast of the Galore Creek porphyry copper deposit and about 20 kilometres

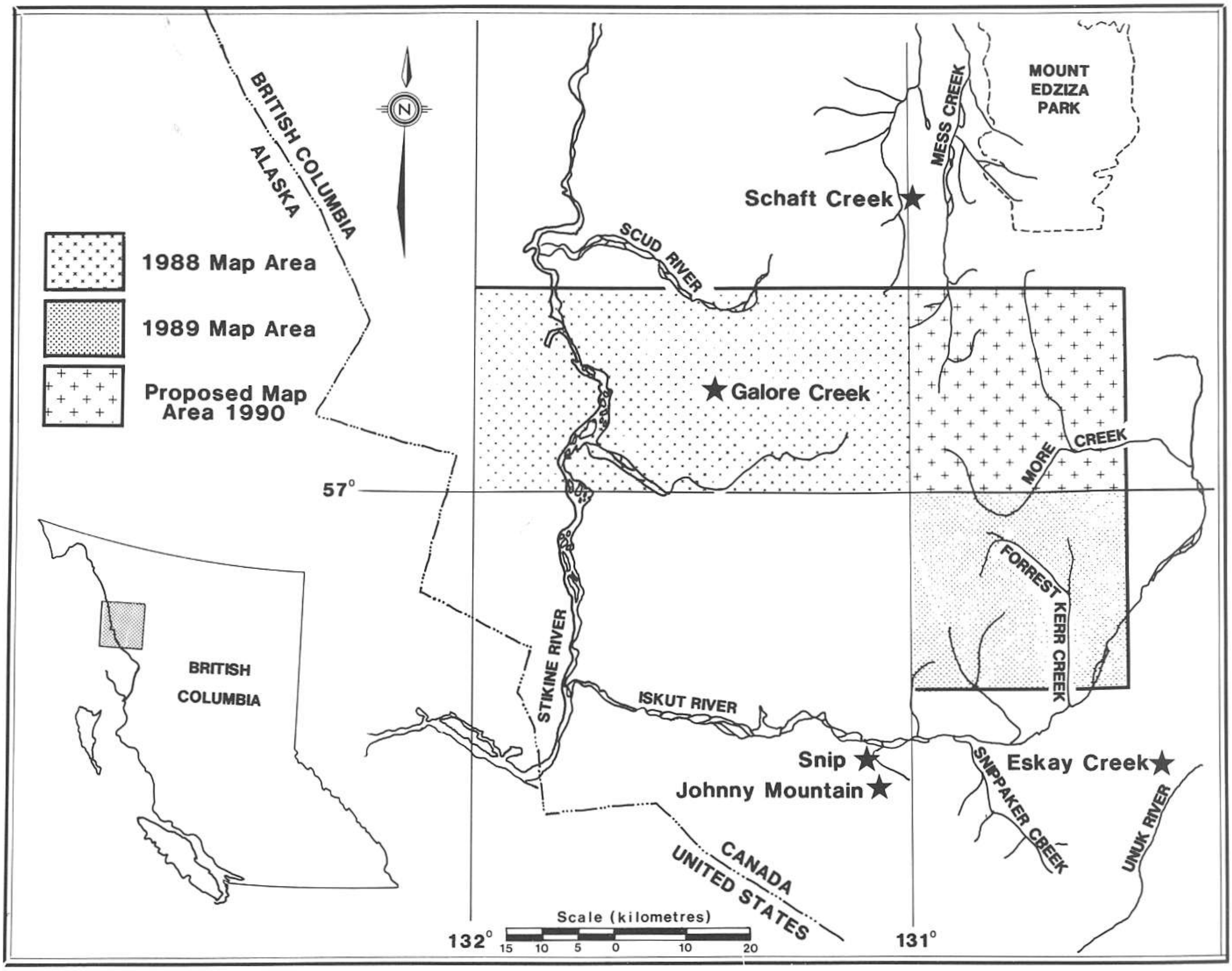

Figure 1-13-1. Location map showing the 1989 Forrest Kerr Creek map area, the 1988 Galore Creek map area and the proposed map area for 1990.

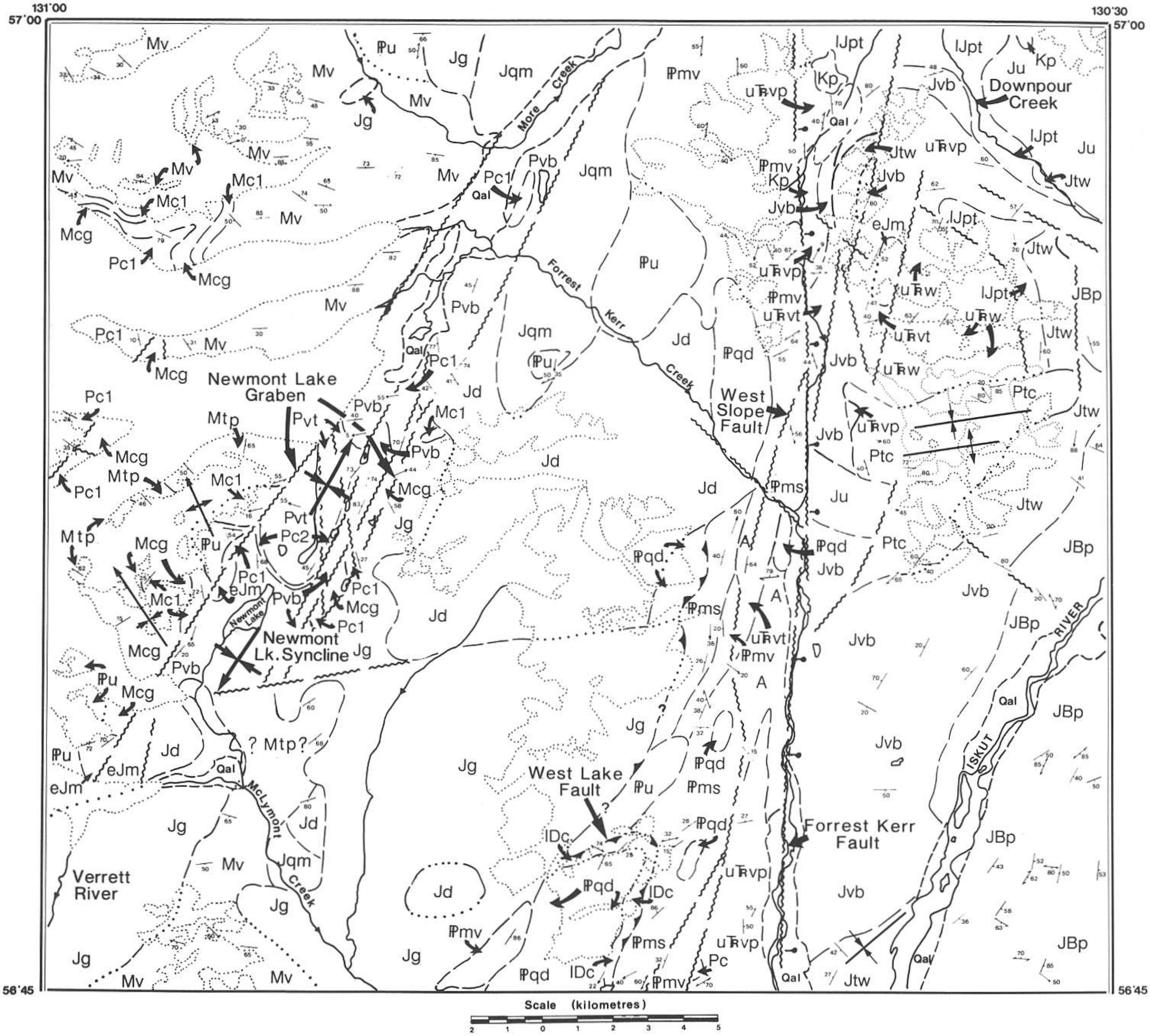

Figure 1-13-2. Geology map of the Forrest Kerr Creek map sheet 104B/15.

northeast of the Snip deposit and Johnny Mountain mine (Figure 1-13-1).

The topography of Forrest Kerr map area is characterized by deeply incised forested river valleys with an abundance of slide alder and devils club at lower elevations, and alpine areas with high peaks and ridges mantled by radiating glaciers and permanent snow fields. Permanent snow and ice cover about 20 per cent of the map area. Access is by fixed-wing aircraft from Smithers to the Forrest Kerr airstrip, or to the Bronson Creek airstrip on the Iskut River, and from there by helicopter.

REGIONAL GEOLOGY

The study area straddles the boundary between the Intermontane and Coast tectonic belts. Stikine Terrane (Stikinia) underlies the area and comprises a mid-Paleozoic to Mesozoic island arc succession which is overlapped by Middle to Late Jurassic sediments of the Bowser basin to the east.

Paleozoic volcanic and associated sedimentary rocks of the Stikine assemblage floor Stikinia. Mesozoic arc volcanism is represented in the Upper Triassic by Stuhini Group and in the Lower Jurassic by Hazelton Group volcanic and

LEGEND

QUATERNARY

Qal	TILL, ALLUVIUM

STRATIFIED ROCKS

MIDDLE TO UPPER JURASSIC BOWSER LAKE GROUP

JBp	SILTSTONE, SANDSTONE, MINOR CONGLOMERATE

JURASSIC

Ju	UNDIVIDED VOLCANICS AND SEDIMENTS
Jtw	SILICEOUS WACKE, TUFF, CONGLOMERATE
Jvb	PILLOW BASALT, BRECCIA FLOWS, SILICEOUS SEDIMENTS
lJpt	SHALE, SANDSTONE, LESSER LIMESTONE, TUFF

UPPER TRIASSIC STUHINI GROUP

uTrvt	MAROON AND GREEN EPICLASTICS, AUGITE AND PLAGIOCLASE-PHYRIC VOLCANIC BRECCIAS
uTrvp	DARK GREEN PLAGIOCLASE-PHYRIC FLOWS
uTrva	GREY-GREEN APHANITIC TUFF
uTrw	TUFFACEOUS WACKE, ARGILLITE, LIMESTONE, CONGLOMERATE WITH LIMESTONE CLASTS, PLAGIOCLASE-PORPHYRITIC ANDESITE

MIDDLE TRIASSIC

mTrs	CARBONACEOUS CALCAREOUS SILTSTONE

PALEOZOIC STIKINE ASSEMBLAGE

Pu	UNDIVIDED METAVOLCANICS AND METASEDIMENTS

WESTERN ASSEMBLAGE

PERMIAN

Pvt	FELSIC WELDED TUFF, VOLCANIC SANDSTONE AND SILTSTONE, RHYOLITE FLOWS
Pc2	THIN-LAMINATED, GREY ALGAL LIMESTONE
Pvb	INTERMEDIATE TUFF AND EPICLASTICS, MAROON LAHAR, BRECCIA FLOWS
Pc1	MEDIUM-BEDDED BIOCLASTIC LIMESTONE WITH CHERTY INTERBEDS

MISSISSIPPIAN

Mtp	SILTSTONE, SANDSTONE, TURBIDITES, LESSER LAPILLI TUFF
Mcg	POLYMICTIC VOLCANIC CONGLOMERATE
Mc1	INTERBEDDED SILICEOUS SILTSTONE AND LIMESTONE, THICK-BEDDED CRINOIDAL CALCARENITE
Mv	PILLOW BASALT, HYALOCLASTITE, ASH-FLOW FELSIC TUFF

EASTERN ASSEMBLAGE

PERMIAN

Ptc	INTERMEDIATE TO MAFIC META-TUFF, THIN-BEDDED LIMESTONE AND METASEDIMENTS
Pc	MEDIUM-BEDDED BIOCLASTIC LIMESTONE

PERMIAN AND OLDER

Pms	SILICEOUS TURBIDITES, PHYLLITES, LESSER CHERTY TUFFS
Pmv	MAFIC TO FELSIC METAVOLCANICS, METASEDIMENTS, LIMESTONE LENSES

LOWER DEVONIAN

lDc	LIMESTONE, SILICEOUS TUFF

INTRUSIVE ROCKS

CRETACEOUS AND YOUNGER (?)

Kp	PLAGIOCLASE QUARTZ PORPHYRY

JURASSIC

Jg	PINK HORNBLENDE BIOTITE GRANITE
Jqm	QUARTZ MONZONITE
Jd	HORNBLENDE DIORITE, HORNBLENDE QUARTZ DIORITE

EARLY JURASSIC

eJm	HORNBLENDE-PLAGIOCLASE-PORPHYRITIC MONZONITE, SYENITE

PALEOZOIC

Pqd	DEFORMED HORNBLENDE QUARTZ DIORITE

UNKNOWN

A	ALTERED DIORITE

sedimentary rocks. The Middle Jurassic Spatsizi and Bowser Lake groups record successor basin sedimentation. At least four episodes of plutonism affected this complex stratigraphy.

STRATIGRAPHY

PALEOZOIC STIKINE ASSEMBLAGE

The Stikine assemblage of northwestern British Columbia (Monger, 1977) includes Paleozoic rocks of Early to Middle Devonian, Mississippian and Permian age (Pitcher, 1960; Logan and Koyanagi, 1989; Anderson, 1989; Read *et al.*, 1989). Paleozoic rocks in the map area are divisible into eastern and western assemblages (Figure 1-13-2) based on structural and metamorphic disparities. Stratigraphy in general can be correlated from one division to the other (Figure 1-13-3).

EASTERN ASSEMBLAGE (lDc, Pmv, Pms, Pc, Ptc)

The eastern assemblage occupies a north-trending belt west of the Forrest Kerr fault as well as small inliers east of the fault. Rocks in this belt young southwards. In the north is a metavolcanic package (Pmv) ranging from Middle Devonian to Early Mississippian age (Read *et al.*, 1989), overlain southward by a metasedimentary (Pms) and siliciclastic package of Early Permian age, capped by Permian platformal limestones (Pc). Early Permian meta-tuffs and interbedded limestones (Ptc) outcrop east of Forrest Kerr fault. Rocks of the eastern assemblage are penetratively deformed by at least two phases of deformation.

LOWER DEVONIAN LIMESTONE (lDc)

The oldest known unit in the Paleozoic assemblage comprises at least four coralline limestone members of Early to Middle Devonian age (Anderson, 1989; Read *et al.*, 1989). Interbedded with the limestone are pebble conglomerate, siliceous and carbonaceous shales and both mafic and felsic tuffs. The maximum total thickness is 400 metres. The succession is exposed in a narrow belt, 4.5 kilometres long, north of the Iskut River between McLymont and Forrest Kerr creeks, where it is intruded by Paleozoic(?) quartz diorite (Pqd) and thrust eastward (Figure 1-13-2). Lower Devonian limestone forms the hangingwall of West Lake fault. (Read *et al.*, 1989). These are the most penetratively deformed rocks in the map area.

PALEOZOIC METAVOLCANICS (Pmv)

Paleozoic metavolcanics are variably foliated to schistose, dependent upon the protolith. Mafic to intermediate flows are weakly foliated, purple to dark green and either massive or brecciated. Amygdaloidal plagioclase ± augite-phyric flows predominate. Mottled purple and green mafic to felsic lapilli tuffs are well foliated to phyllitic. Lapilli are flattened in the plane of foliation. Fine-grained crystal tuffs and tuffaceous sediments have been metamorphosed to chlorite schists and lesser quartz-sericite schists. Interbedded with these metavolcanics are subordinate phyllites, tuffaceous pyritic argillites and recrystallized limestone.

PALEOZOIC METASEDIMENTS (Pms)

The metasedimentary unit comprises grey to light green phyllitic siltstone, graphitic argillite, siliceous phyllite and thin lenses of dark brown weathering limestone. Pale green and white siliceous turbidite couplets, comprised of 2 to 5-centimetre siltstone beds intercalated with 1-centimetre beds of fine sandstone, form a distinctive unit at least 200 metres thick. This unit is correlative with Brown and Gunning's (1989) "siliceous unit" mapped to the north at the Scud River. Apparently high(?) in the stratigraphy are vari-coloured, weakly foliated siliceous siltstones and ribbon cherts.

PERMIAN CARBONATES (Pc)

Early Permian (Artinskian to Sakmarian; Read *et al.*, 1989) limestone is exposed at the southern edge of the map area adjacent to the West Slope fault. The limestone is a massive, white to buff, sparsely crinoidal calcarenite which locally is completely recrystallized to coarse calcite.

PERMIAN META-TUFFS AND LESSER CARBONATES (Ptc)

This metavolcanic unit is exposed east of the bend in Forrest Kerr Creek. The unit is folded and foliated and comprised of interbedded chloritic tuffs, tuffaceous and siliceous siltstones and numerous thin recrystallized Early Permian limestones.

WESTERN ASSEMBLAGE (Mv, Mcl, Mcg, Mtp, Pcl, Pvb, Pc2, Pvt)

The western assemblage is well exposed north of the Iskut icefield and Newmont Lake and separated from the eastern assemblage by a Jurassic composite plutonic body. Mississippian reefal limestone (Mc) and underlying pillowed basalt (Mv) are the oldest known rocks. A polymictic volcanic conglomerate (Mcg) separates Mississippian limestone from Lower Permian (R.G. Anderson, personnel communication, 1989) limestones (Pc1). Intermediate volcanics (Pvb), a medial limestone (Pc2) and felsic tuffs (Pvt) comprise the uppermost sections. The western assemblage rocks are not penetratively deformed, not metamorphosed and the limestones are rarely recrystallized; at the time of mapping they were thought to be younger than the eastern assemblage.

MISSISSIPPIAN VOLCANICS (Mv)

Mississippian and older(?) mafic to felsic volcanics occur in the northwestern and southwestern corners of the map area and comprise a southwest-dipping homoclinal sequence of pillow lavas, flow breccia, ash-flow tuffs and stratified tuffs and epiclastics which are conservatively estimated to exceed 2000 metres in thickness. The upper contact is conformable with mid-Late Mississippian limestone (M.J. Orchard, personal communication, 1989). The lower contact was not defined. Interbedded hyaloclastite and pillowed flows are texturally distinctive and volumetrically exceed massive sheet-flows of basalt in the northwest corner of the map. Flows vary texturally from aphanitic to porphyritic and are commonly amygdaloidal. Plagioclase-phyric flows are more common than augite-plagioclase porphyries. Basaltic andesite flows are characterized by scoriaceous fragments. In detail, amygdules are distributed parallel to borders of clasts and are often concentrically zoned about the clasts. Hyaloclastite debris flows contain sparse pillows. Massive flows have brecciated tops and bottoms. Light purple to pink dacite flow breccias are spatially related to ash-flow and welded tuffs in the northwest corner of the map area. These flows and tuffs contain very angular lapilli averaging 1 to 10 centimetres across and pinkish grey porphyritic feldspar, often in clusters of three to four crystals. The irregular distribution of quartz, epidote and chlorite amygdules indicates propylitic alteration rather than regional greenschist metamorphism.

Fragmental rocks are dominated by heterolithic lapilli tuffs. Dark green to grey, angular to subrounded, densely amygdaloidal fragments are diagnostic of tuffs interbedded with pillowed flows and scoriaceous hyaloclastites. Dominantly green, orange-weathering ash-flow and welded lapilli tuffs are a distinctive unit. Lapilli are pale grey, pink and purple aphyric lithic and crystal fragments. These felsic tuffs are spatially related to dacitic and rhyolitic rocks exposed north of the headwaters of Forrest Kerr Creek. A third variety consists of thin, planar-bedded siliceous dust tuffs. These are interbedded with graded and crossbedded crystal and lapilli volcaniclastics.

This volcanic succession is interpreted to represent submarine volcanism from a seamount (pillows and hyaloclastite) with atoll-fringing reefs (reefal limestone Mc) which, over time, became emergent (ash-flow and welded ash-flow tuffs) and evolved from basalt through dacite to rhyolite.

MISSISSIPPIAN CARBONATE (Mc)

Late Mississippian reefal limestone, reef-flank carbonates and cherty siltstones overlie hyaloclastites in the northwest corner of the map area (Figure 1-13-2). West of Newmont Lake the base is faulted. At Round Lake (about 20 kilometres northwest) the limestone consists of two distinct members separated by a wedge of chert and phyllitic volcaniclastic rocks (Monger, 1970; Logan and Koyanagi, 1989). In the northwest corner of the map sheet these same lithologies are recognized, but their relationships are complicated by faulting and swarms of east-trending dikes. The lower limestone is a bioclastic calcarenite with characteristic coarse (up to 5-centimetre diameter) crinoid columns. Carbonates interpreted to be coralline reef mounds, slope-front block breccias and graded reef-flank deposits interbedded with siliceous siltstones comprise an aggregate thickness of 50 metres at most. Farther west less than 200 metres of thick-bedded, grey crinoidal calcarenite with interbeds of amorphous chert are exposed on strike. This limestone contains small crinoidal stems and fossil fragments, is locally fetid and resembles the upper limestone member exposed at Round Lake. Both limestone members are conformably overlain by a coarsening-upward sequence of siliceous siltstones and volcanic conglomerate.

MISSISSIPPIAN(?) CONGLOMERATE AND SILTSTONE (Mcg)

At least 300 metres of cherty siltstone and maroon polymictic volcanic conglomerate overly Mississippian limestone in the northwest corner of the study area and west of Newmont Lake. Dark grey and black cherty siltstones and interbedded bioclastic limestone shale-out up section, followed by thinly laminated rusty weathering cherty siltstones which, in turn, are conformably overlain by maroon conglomerates. Dark purple and green pyroxene-porphyritic and hornblende-plagioclase-porphyritic andesites, scoriaceous basalt and grey fossiliferous limestone clasts form up to 70 per cent of the rock in a plagioclase crystal rich tuffaceous matrix. In general the conglomerates are massive to thickly bedded making it difficult to determine bedding attitudes. Angular blocks of limestone up to several metres across are common. A tentative fossil identification from one of these blocks is probably Permian but possibly Mississippian (E.W. Bamber, personal communication, 1989). Near the top of the conglomerate, angular limestone clasts increase in abundance near the Permian limestone contact. From this relationship the conglomerate was formerly interpreted to be post-Permian and overturned. Mesoscopically it resembles the basal Triassic conglomerate located to the west, above Forrest Kerr Creek (Read *et al.*, 1989). Bedding-top directions and conformable contacts indicate that the conglomerate is upright and actually either Late Mississippian or Early Permian in age. It has tentatively been interpreted to represent Monger's (1977) post-Mississippian pre-Permian profound unconformity.

Mississippian Tuffs and Epilclastics (Mtp)

This unit is exposed on nunataks northwest of Newmont Lake. A second outcrop area between McLymont Creek and Newmont Lake is tentatively included. The age of these clastics and tuffs is uncertain; the interbedded tuffs and epiclastics resemble Mississippian epiclastics to the north. The unit comprises a well-bedded succession of distal(?) turbidites consisting of fine siliceous siltstone and carbonaceous siltstone, interbedded sandstone and polymictic conglomerate. Fining-upward sequences are common. Lapilli tuff is interbedded and thick accumulations of coarse breccias and lahar attest to periodic volcanic influxes. The succession southeast of Newmont Lake consists of well-indurated, commonly bleached and pyritic, finely layered siltstones and conglomerates which are hornfelsed by adjacent intrusions. Unit Mtp is thought to represent a basinward facies equivalent of the coarse conglomerate of unit Mcg.

Permian Limestone (Pc1)

Lower Permian (Artinskian; M.J. Orchard, personnel communication, 1989) limestone is best exposed on nunataks along the western edge of the map area and in fault-bound slivers on the eastern side of the Newmont Lake graben. In contrast with the greater than 1000 metres of platformal carbonates present at the Scud River (45 kilometres north) less than 200 metres are present here.

The limestone comprises primarily massive to thin-bedded grey bioclastic calcarenite and lesser buff silty dolomitic units. Thin-bedded sections are interbedded with black to yellowish buff amorphous silica beds up to 20 centimetres thick and comprising as much as 50 per cent of the outcrop. These layers are diagenetic(?) silicified fossil-rich horizons. Solitary corals, foraminifera, bryozoan, crinoids and various brachiopods are locally abundant. Limonitic and hematitic limestones are coincident with fault structures and indicate fluid flow and attendant alteration. This alteration is selective and occurs predominantly in the massive limestones and dolomitic mudstones.

Permian Volcanics and Volcanic Breccias (Pvb)

A Permian intermediate volcanic unit outcrops within the northeast-trending Newmont Lake graben. It is a distinctive purple to maroon colour and comprises thick breccia flows, lahars, tuffs and lesser mudstones and wackes. These volcanics are fresh in appearance and resemble Jurassic Hazelton Group rocks, their intercalation with limestone members is an important distinction (Anderson, 1989).

The flows are purple, plagioclase and hornblende-porphyritic andesites. They are locally amygdaloidal and generally contain 30 to 40 per cent euhedral white plagioclase and 15 per cent chloritic acicular hornblende crystals. Breccia and massive flows averaging 5 metres in thickness are interbedded with well-graded interflow epiclasties west of Newmont Lake. East of Newmont Lake are light green to pink, block and lapilli tuffs with lesser plagioclase crystal tuff. Maroon lahar and well-bedded graded conglomerates are exposed north of Forrest Kerr Creek.

This calcalkaline volcanic package, coarse "proximal facies" breccia flows, and variable oxidizing states suggest a transgressive submarine to subareal volcanic centre.

Permian Algal Limestone (Pc2)

This limestone is a good marker which can be traced around the Newmont Lake syncline. North of the lake 95 metres thickness is exposed. The limestone pinches along strike and on the east limb is only 20 metres thick. It is dark grey to black and finely laminated, weathering buff and is locally fetid. Pisolite-rich beds and cuspate stacked concave algal structures are common. The upper sections are selectively silicified.

The fine laminations are interpreted to be cryptalgal laminations (Aiken, 1967) indicative of algal mats and an intertidal zone of deposition for the limestone. The algal limestone unit is overlain conformably by well-bedded tuffaceous epiclastics and a thick package of welded tuffs.

Permian Volcanics (Pvt)

The core of the Newmont Lake syncline contains a basal package of maroon, shallow-water conglomerates, siltstones, lapilli and plagioclase crystal tuffs. Discontinuous thinly bedded siliceous limestones, up to 5 metres thick, are interspersed throughout. This basal package is overlain by more than 100 metres of brownish grey massive to thick-bedded welded ash tuffs. The tuffs exhibit good eutaxitic flow laminae and are often columnar jointed. Air-fall ash tuffs are well stratified and contain 5 to 10 per cent angular lithic lapilli. Flow-banded rhyolites and breccias occur high in the section.

Middle Triassic (mTRs)

Middle Triassic(?) rocks are restricted to small discontinuous fault-bounded exposures in the centre of the Newmont Lake syncline (Figure 1-13-2). Unit mTRs consists of black carbonaceous, locally calcareous, silty shales and argillites which contain bivalves of presumed Middle Triassic age. These rocks are similar to Middle Triassic rocks mapped in 1988 above the South Scud River (Logan and Koyanagi, 1989). North of Newmont Lake the rocks weather recessively and are less siliceous but contain characteristic rounded to elliptical concretions.

UPPER TRIASSIC STUHINI GROUP

Upper Triassic Stuhini Group rocks lie between the West Slope and Forrest Kerr faults south of Forrest Kerr Creek and east of the Kerr fault in the northeast corner of the map area (Figure 1-13-2). A generalized stratigraphy following that of Read *et al.* (1989), consists of a lowermost metasedimentary succession (uTrw), a medial metavolcanic succession (uTrva, uTrvp) and an overlying tuffaceous metasedimentary succession (uTrvt). Contacts between units are faulted or poorly exposed and, as a result, thicknesses and overall stratigraphic relationships are uncertain (Figure 1-13-3).

East of the Forrest Kerr fault Unit uTrw consists of a thick package of fine-grained volcaniclastics and sediments. Green to grey massive volcanic wackes and arenites, and interbedded black siltstones and argillites predominate, with lesser limestone and limy conglomerates. Massive to thickly bedded volcanic sandstones and poorly sorted lithic wackes contain up to 80 per cent plagioclase, the remaining 10 to 30 per cent being 2 to 4-millimetre pyroxene grains and lithic clasts of plagioclase pyroxene porphyry. Plagioclase crystal tuff, with lapilli to 5 centimetres, are intercalated with the wackes. These volcaniclastics are interbedded with thin planar-bedded to crudely crossbedded (locally) carbonaceous rusty weathering argillites. The fine-grained sediments contain fossiliferous limy horizons with abundant brachiopods intercalated with lesser limy conglomeratic beds.

At the south end of the map, east of the West Slope fault, a maroon volcanic conglomerate containing limestone clasts structurally underlies Lower Permian limestone (Figure 1-13-2). It has been interpreted by Read *et al.* (1989) to mark the base of the Stuhini Group.

Massive light grey to dark green aphanitic lapilli tuff and andesite breccia (uTrva) is a minor though distinctive unit intercalated with plagioclase porphyry flows of Unit uTrvp. The tuff is massive to stratified with monolithic scoriaceous to aphyric andesite lapilli. Read *et al.* (1989) suggested thicknesses of a few hundred to less than a thousand metres.

Green, and lesser maroon, crowded to sparsely plagioclase-porphyritic andesite breccia and flows (uTrvp) underlie the area immediately west and east of the Forrest Kerr fault. Euhedral plagioclase ranges to 5 millimetres in length and comprises 30 per cent of the rock. Alteration is characteristically hematite and chlorite with variable epidote, quartz and calcite as patches and veinlets.

Maroon to dark green tuffs and monolithic augite ± plagioclase-phyric fragmentals (uTrvt) are best exposed within the Forrest Kerr fault zone, north of the bend in Forrest Kerr Creek. Lapilli tuffs with varicoloured porphyritic volcanic and lesser scoriaceous fragments are interbedded with purple to maroon and green well-bedded, locally graded plagioclase crystal-ash tuffs and fine epiclastics. The tuffs are massive to weakly stratified. Coarse breccias and block tuffs of augite porphyry occur near the top(?) of the succession. Associated conglomerates and reworked volcaniclastics are comprised of angular to rounded, green pyroxene porphyry clasts in a pyroxene-rich matrix.

UNNAMED LOWER TO MIDDLE JURASSIC ROCKS

Unnamed Lower and Middle Jurassic rocks overlie Upper Triassic Stuhini Group rocks east of the Forrest Kerr fault. A generalized stratigraphy consists of: a sedimentary package of interbedded shales and siltstones, lesser limestones and tuffs (lJpt); followed by pillowed basalts, hyaloclastite and interflow siliceous sediments (Jvb); and finally tuffs, siliceous wackes and conglomerate (Jtw).

The inherent complications of volcanic facies changes over short distances and the lack of fossil dates precludes assigning these rocks to either the Hazelton or the Spatsizi Group.

LOWER JURASSIC SHALES AND SILTSTONES (lJpt)

At least 1000 metres of interbedded shale and siltstone outcrop in the valley floor of "Downpour" Creek and extend to More Creek north of the map area (Figure 1-13-2). The shales are fissile; siltstones and thin sandstone beds contain abundant carbonaceous wood fragments. Sandstones grade into thick-bedded granule conglomerates containing intermediate volcanic, sedimentary and limestone clasts. Fossils from interbedded limestone horizons located north of the map area indicate an Early Jurassic (late Toarcian) age (Read *et al.*, 1989). This sedimentary succession is intruded by 2 to 3-metre sills and dikes of pyroxene and plagioclase-phyric diorite (Jdi) thought to represent feeders to the overlying basalts. East of the Forrest Kerr fault, at the northern edge of the map area, Read *et al.* (1989) show felsic tuffs and rhyolite flows correlative with the Mount Dilworth formation of Alldrick *et al.* (1989). These rocks are restricted to this one locale.

JURASSIC PILLOW BASALT AND FLOW BRECCIA (Jvb)

Middle(?) Jurassic pillow and flow-breccia basalts underlie a large area between Forrest Kerr Creek and the Iskut River south of the bend in Forrest Kerr Creek (Figure 1-13-2). Smaller fault-bounded slices extend north beyond the map margin to More Creek. In the north they conformably overlie shales and siltstones of Unit lJpt.

The pillows average 30 to 100 centimetres across, are well preserved and commonly indicate facing directions. Outcrops weather dark brown to orange. Flow breccias are interbedded with the pillows and locally scour and disrupt interflow sediments. The basalt is dense, medium grey to green, locally amygdaloidal and made up of fine-grained vitreous plagioclase crystals with rare pyroxene phenocrysts and abundant disseminated pyrite. Pillows and hyaloclastite flow-breccias comprise more than 90 per cent of this unit.

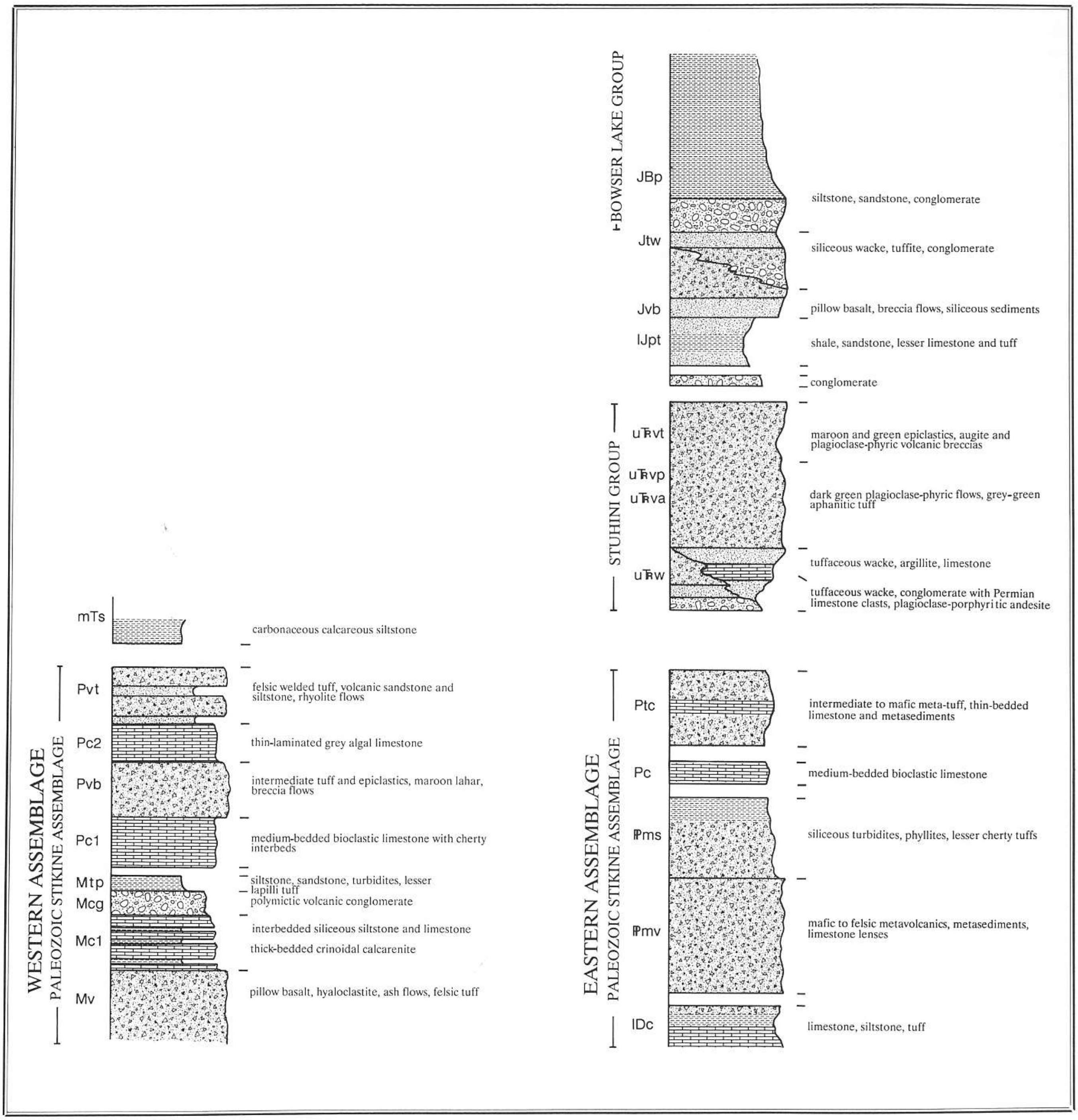

Figure 1-13-3. Schematic stratigraphic sections for the eastern and western portions in the map area.

White and grey siliceous argillites or tuffs and pyritic siltstones are interbedded with the basalts. Sills or feeders(?) to these flows are dioritic to gabbroic with characteristic euhedral felty plagioclase textures and an interstitial pyroxene-rich groundmass.

Grey and khaki siliceous siltstones, conglomerates and tuffs (Jtw) overlie and interfinger with the pillow basalts.

JURASSIC TUFF AND WACKE (Jtw)

This unit outcrops at the southern edge of the map area between Iskut River and Forrest Kerr Creek, northwest of the Iskut River 18 kilometres upstream from its confluence with Forrest Kerr Creek, and as a fault-bounded wedge east of the Forrest Kerr fault near the northern edge of the map area (Figure 1-13-2).

These rocks are characteristically drab, olive to grey in colour, unlike the maroon and dark green Upper Triassic Stuhini Group, and commonly fractured and brecciated. Dark green and grey siliceous siltstones and pyritic cherts are crackle fractured and brecciated *in situ*, forming subangular to angular centimetre-scale fragments. Tuffaceous wackes are carbonaceous and variably sheared. Interbedded with tuffaceous arenites are sedimentary conglomerates containing clasts of chert, black siltstone and intermediate to felsic volcanics. Volcaniclastics are characteristically brownish grey lapilli and crystal tuffs comprised of euhedral plagioclase and scoriaceous green and grey fragments. The contact with the Middle Jurassic Bowser Lake Group is conformable.

MIDDLE JURASSIC BOWSER LAKE GROUP (JBp)

Bowser Lake Group sediments outcrop on the eastern edge of the map area along the flanks of the Iskut River valley. South of the Iskut River, planar-bedded shale and locally crossbedded sandstone couplets are interbedded with local granule conglomerate. The conglomerate contains quartz and siltstone clasts in a limonitic sandy feldspathic matrix. Argillites have a well-developed pencil cleavage and locally host pressure solution quartz veinlets. These grey shales and siltstones are late Middle Jurassic (Callovian) age and are correlated with the Ashman Formation of the Bowser Lake Group (Read *et al.*, 1989). West of the Iskut River the lithologies consist of mainly fine to medium-grained sandstones containing 10 to 15 per cent detrital quartz, and fine siltstones.

INTRUSIVE ROCKS

Intrusive rocks underlie one third of the map area. They are chiefly restricted to a north-trending belt 10 kilometres wide, in the centre of the map. Wheeler *et. al.*, (1988) interpret this composite plutonic belt to be Jurassic. An early Jurassic U-Pb date has been obtained from a porphyritic granite south of the map area (R.G. Anderson, personal communication, 1989). A Paleozoic intrusive suite has been recognized by Read *et al.* (1989) on the eastern margin, but the belt as a whole is thought to be Jurassic or younger. A preliminary plutonic chronology based on mineralogy, textural and intrusive relationships follows. This will be refined by isotopic dating currently in progress.

PALEOZOIC(?) SUITE

HORNBLENDE QUARTZ DIORITE/TONALITE (Pqd)

This unit outcrops as discrete bodies, elongate north-south and situated west of Forrest Kerr Creek. The rock is heterogeneous due to abundant xenoliths of pyroxene and/or plagioclase-porphyritic phases and younger felsic intrusive breccias. Locally the diorite is foliated and deformed. Parts of the intrusion contain well-rounded to irregular shaped inclusions from 2 to 100 centimetres across. These are always more mafic, finer grained felty textured amphibolites and comprise up to 65 per cent (by volume) of the intrusion.

The diorite is massive, medium grained and contains about 20 per cent quartz; plagioclase occurs as glomeroporphyritic patches and makes up the bulk of the feldspar content. Hornblende also occurs as glomeroporphyritic patches to 15 per cent.

The quartz diorite intruded Paleozoic rocks in post-Early Permian to pre-Middle Triassic time (Read *et al.*, 1989).

EARLY(?) JURASSIC SUITE

PLAGIOCLASE PORPHYRITIC MONZONITE (eJm)

Dikes, sills and plugs of plagioclase-hornblende-porphyritic monzonite are restricted to the Newmont Lake graben. They are characterized by a hematitic groundmass, commonly purple to grey with pink subhedral to euhedral plagioclase (up to 50 per cent) and hornblende crystals. Trachytic textures are common.

POTASSIUM FELDSPAR MEGACRYSTIC GRANITE

A hornblende biotite potassium feldspar megacrystic granite outcrops between McLymont and Forrest Kerr creeks north of the Iskut River. A U-Pb zircon date of 191 to 195 Ma (Anderson and Bevier, 1990) indicates this body to be the same age as the Texas Creek pluton of Grove (1986).

MIDDLE JURASSIC SUITE

DIORITE (Jdi)

Coarse-grained diorite stocks and sills are spatially associated with Unit Jvb and outcrop within the Forrest Kerr fault zone. Plagioclase crystals are euhedral to subhedral acicular clots which impart a distinctive felty interlocking texture. Euhedral plagioclase to 50 per cent, pyroxene, and up to 1 per cent pyrite disseminations comprise these subvolcanic intrusions, which may represent feeders to the pillow basalts (Jvb).

JURASSIC SUITE

DIORITE (Jd), QUARTZ MONZONITE (Jqm), BIOTITE GRANITE (Jg)

Composite Jurassic intrusions comprise a batholithic body extending north from McLymont Creek to north of Forrest Kerr Creek. Three separate mappable phases have been recognized; diorite (Jd), quartz monzonite (Jqm), and biotite granite (Jg) (Figure 1-13-2).

The most mafic phase (Jd) occurs in the north. It is a heterogeneous mix of porphyritic to massive hornblende diorite with more mafic hornblendite phases. Hornblende is chloritic and comprises more than 40 per cent of the rock, quartz is less than 5 per cent, but variable to quartz diorite proportions. The remainder is subhedral plagioclase. Intrusive breccias, gneissic sills and pendants of metavolcanic and metasedimentary rocks are common. Contacts with Unit Jqm are gradational over short distances of complex intrusive mixing which commonly exhibit conflicting intrusive relationships. North of Forrest Kerr Creek a 9 square kilometre zone of metamorphosed Paleozoic rocks (Pu) separates Units Jd and Jqm.

The intermediate phase (Jqm) outcrops at the northern end of the belt on both sides of More Creek. It is composed of coarse to medium-grained hornblende quartz monzodiorite to monzonite. Hornblende averages 20 per cent, as 5 millimetre crystal laths and poikilitic clots; biotite, where present, is fine grained and less than 5 per cent. Quartz averages 10 to 15 per cent. Feldspars comprise the remainder, in a 60:40 ratio of plagioclase to potassium feldspar. The rocks are cut by flat-lying sill-like bodies and swarms of dark green diabase dikes. Contacts with the younger granite are sharp.

The youngest intrusive phase is a quartz-rich granitoid suite comprised principally of biotite granite (Jg) which occupies the southern half of the intrusive belt. A second body is centred on the Verrett River southwest of McLymont Creek. Granite intrudes all other igneous rocks in the map area as dikes and forms the anastomosing matrix of intrusive breccias. The granite is coarse to medium grained and deeply weathered producing a sandy to rubbly outcrop surface. It is pink on fresh surfaces and contains about 30 to 40 per cent quartz and 5 to 7 per cent biotite, the remainder being pink euhedral feldspars. Less commonly hornblende is the mafic mineral. The granite varies from equigranular to "quartz-eye" porphyritic. Coarse-grained quartz-rich phases (up to 50 per cent quartz) are spatially related to fault structures. These pegmatitic phases occur north of McLymont Creek at its headwaters and east of the Newmont Lake syncline. They characteristically weather hematitic.

QUARTZ FELDSPAR PORPHYRY (Kp)

Small plugs and dikes of quartz feldspar porphyry are intruded along north-trending faults in Forrest Kerr Creek valley north to the headwaters of "Downpour" Creek. The rocks contain finely disseminated pyrite and outcrops are oxidized to yellow and red colours easily visible from a distance. These rocks extend north onto the Telegraph sheet (104G) where Souther (1972) grouped them with late Cretaceous to early Tertiary felsic bodies.

ALTERED DIORITE (A)

Massive, silicified and propylitically altered intrusive rocks are exposed along the western slopes of Forrest Kerr Creek, south of the main bend. They are distributed along major north-northeasterly fault structures west of the Forrest Kerr fault. Primary textures and mineralogy have been obliterated. The rocks are aphanitic, vitreous to dull green and thoroughly fractured. Pyrite is ubiquitous as disseminations and in quartz-carbonate veinlets. The rocks are cliff formers which weather white to light green; Read *et al.* (1989) included them in Unit Jfp, Jurassic feldspar porphyry.

STRUCTURE

Five structural domains are apparent from a preliminary examination of the structural data. All five are fault bounded.

Domain I includes the area west of the Newmont Lake graben. Strata include Mississippian volcanics, limestones and coarse clastics with lesser Permian limestones (Stikine "western" assemblage) and comprise for the most part southwest-dipping and facing (?) homoclinal panels. A large-scale northwest-trending anticlinal-synclinal pair is interpreted west of Newmont Lake.

Domain II is confined to the northeast-trending Newmont Lake graben, west of the composite plutonic body, and is charaeterized by a large upright, open northeasterly trending, doubly plunging syncline in Permian volcanics and limestone. Left-lateral motion along the bounding faults has interleaved stratigraphy and resnlted in transtensional fanlting and fracturing in a north-south direction. The syncline is cored by Upper (?) Permian limestone and volcanic rocks.

Domain III is west of the Forrest Kerr fault in the "eastern" Stikine assemblage and is characterized by a moderate west-northwest-dipping schistosity and tight recumbent folds that are overprinted by a gently southwest-plunging crenulation with its axial plane dipping steeply southeast.

Domain IV is restricted to the Permian outlier where large, upright, tight, east-trending folds occur in thin-bedded tuffaceous siltstone and limestone.

Domain V, east of the Forrest Kerr fault, includes Mesozoic rocks from Stuhini Group to Bowser Lake Group. Large-scale folds are generally open, upright, northwesterly trending and isoclinal in volcanic rocks, to more tightly chevron folded in sediments.

FAULTS

Regional faults cross the map area and control the distribution of lithostratigraphic packages. Fault trends are mainly northeast to north; northwesterly faults are less significant structurally, but are important controls for mineralization.

The northeast-trending Newmont Lake graben is 3 kilometres-wide. The eastern side is comprised of a network of parallel fault structures which separate various Jurassic intrusive phases from Permian limestones, volcaniclastics and clastics. The western fault is a single, strong, 040°-trending structure separating Mississippian from Permian strata. Overall apparent movement across the graben is left lateral. North and northeasterly trending faults crosscut this structure. Both left and right lateral senses of motion are evident along these steep faults, with offsets of no more than tens of metres.

Read *et al.* (1989) have recognized folded regional-scale faults. Deformation of the West Lake, West Slope and Kerr Bend (not shown) faults occurred during the Middle to Late Jurassic and is characterized by low-angle easterly directed movement.

The West Lake fault dips moderately to the west. Hangingwall rocks, Unit lDc in the south and Unit Pqd in the north, have been thrust eastward with respect to the Early Permian footwall.

The West Slope fault is moderately steep and easterly dipping. Upper Triassic Stuhini Group volcanics and sediments are exposed in the down-dropped footwall with Paleozoic metasediments and metavolcanics occupying the hangingwall. The fault is marked by a wide limonitic alteration zone along its length.

The Forrest Kerr fault is a northeasterly trending, vertical to steep easterly dipping normal fault. It separates metamorphosed and deformed Paleozoic strata on the west from Lower to Middle Jurassic rocks on the east. Read *et al.*

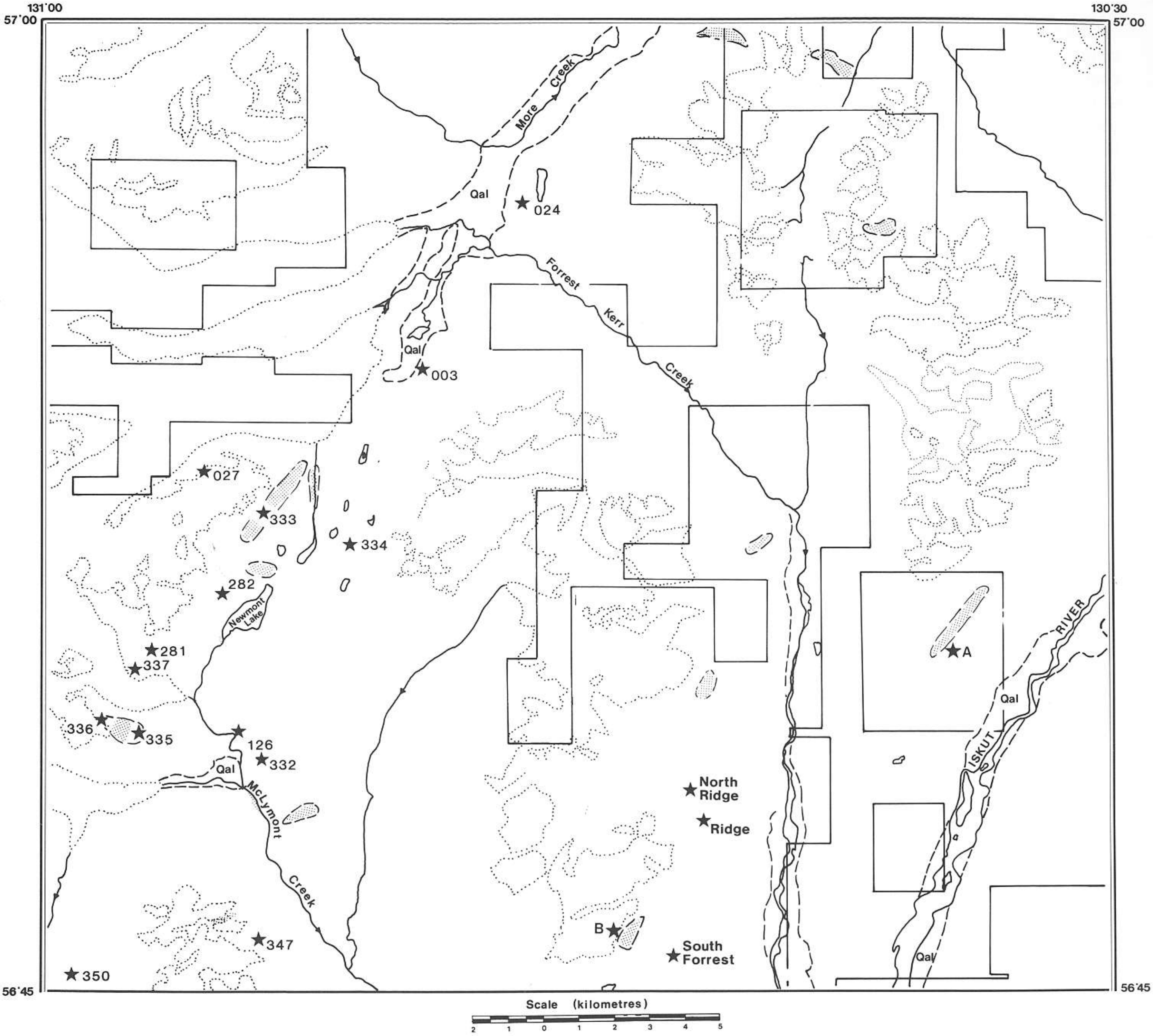

Figure 1-13-4. MINFILE locations, mineral occurrences and gossanous zones (shaded) in the Forrest Kerr Creek map area 104B/15 (*see* Table 1-13-1 for descriptions of MINFILE occurrences).

(1989) suggests a minimum of 2 kilometres east-side-down and 2.5 kilometres of left-lateral oblique-slip motion. Movement along these major structures has continued into the Miocene (Souther, 1972).

EXPLORATION ACTIVITY

The map area is located at the northwest end of the Stewart–Sulphurets–Iskut gold belt, 15 kilometres northwest of Eskay Creek. It is completely staked, including sections of the Iskut icefield. Exploration activity was accelerated after the release of assay results from Eskay Creek.

Gulf International Minerals Limited completed diamond drilling (approximately 7000 metres), geophysics and surface mapping on its Northwest skarn mineral zone. Kestrel Resources Limited carried out soil and rock geochemical sampling, prospecting and trenching on vein mineralization on its Tic, Arc and Mon claims. Pamicon Development Limited cut lines, mapped and sampled its Ridge and South Forrest mineral zones on the Forrest claims. Cominco Ltd. carried out prospecting, boulder tracking, and a geophysical program (UTEM and magnetometer) to locate the source of more than 200 massive sulphide boulders on its Fore More claims, located 10 kilometres north of the headwaters of Forrest Kerr Creek on NTS sheet 104G/2.

MINERAL PROSPECTS

MINFILE lists 14 mineral occurrences in the map area (Table 1-13-1). Figure 1-13-4 shows their distribution, including two new mineral showings located during our work. A direct relationship is evident between epigenetic mineral occurrences and northeast-trending regional structures. Mineralization can be divided into the following categories:

- gold-copper-iron skarns;
- mesothermal copper-gold-silver-bearing quartz veins;
- stratabound massive sulphides.

GOLD-COPPER-IRON SKARNS

Newmont Mining Corporation of Canada Limited first assessed the copper potential of these skarns in the early 1970s but it was not until 1987 that the potential for gold enrichment was discovered by Gulf International Minerals Limited. On its McLymont Creek property, the Northwest zone (MINFILE 104B 281) contains stratabound skarn mineralization hosted by up to 200 metres of thin to medium-bedded siltstones, chert, sandstone, marble and minor conglomerate of Mississippian age (Unit Mtp). Mineralization is developed in marble horizons and along contacts between tuffaceous sandstones and marble where faults and fractures have provided permeability for the ore solutions. Several mineralized zones are semiconformable with bedding and extend outward from a central northeast-trending structure (the 040° western-bounding fault of the Newmont Lake graben).

Sulphides include pyrite, chalcopyrite, sphalerite and galena with a gangue of barite, calcite, gypsum, coarse-grained magnetite and specular hematite. The best gold mineralization is associated with coarse euhedral pyrite (E.W. Grove, personnel communication, 1989). The 1989 drilling has tested the zone for 300 metres along strike and 200 metres below the surface.

Strong structure, proximity to intrusive bodies and chemically reactive hostrocks have all contributed to localizing this deposit.

MESOTHERMAL COPPER-GOLD-SILVER-QUARTZ VEINS

Early exploration on the McLymont property (MINFILE 104B 126) focused on testing precious metal veins. Two types of veins were recognized. The first are an early quartz-pyrite-chalcopyrite vein set that trends 120° to 140° (Camp zone). The veins and attendent potassic alteration selvages fill fractures in quartz-rich granite. Mineralization comprises minor sphalerite, galena and free gold (Grove, 1987). The second vein type consists of northwest or northeast-trending en echelon vein swarms that postdate the earlier quartz veins. The veins are ankerite-quartz-pyrite replacement veins which contain sparse chalcopyrite and local gold values.

The Forrest claims extend approximately 10 kilometres north from the Iskut River along the west side of Forrest Kerr Creek. They cover 11 mineral occurrences concentrated in three areas; South Forrest, Ridge and North Ridge. Mineralization consists of quartz stockworks and veins which trend either 135/70NE or 360/90, hosted by Stikine assemblage rocks (Eastern assemblage) close to the contact with Jurassic granitic rocks. Plagioclase-porphyritic diorite intrusions (Pqd) are spatially related to veining in the Ridge area. Mineralization consists of gold and silver-bearing quartz-chalcopyrite veins with or without malachite, azurite, arsenopyrite, galena, bornite and hematite.

STRATABOUND MASSIVE SULPHIDES

Stratabound pyritic horizons, associated with dacitic(?) pyroclastics or altered hyaloclastite horizons outcrop discontinuously within cherty siltstones and black carbonaceous argillites in a thick succession of basic pillow and breccia flows north of the Iskut River, 12 kilometres upstream from the mouth of Forrest Kerr Creek (A, Figure 1-13-4). Massive fine-grained pyrite and pyrrhotite occur as bedding-parallel layers several centimetres thick and as disseminations. Rusty limonitic gossans and the white-weathering felsic rocks can be traced along the ridge for 1500 metres. Samples have been submitted for assay and geochemical analysis. The stratigraphic similarities to Eskay Creek warrant follow-up.

Foliation-parallel massive pyrite zones occur in pyritic felsic volcaniclastic rocks 6 kilometres northwest of the confluence of Forrest Kerr Creek and the Iskut River (B, Figure 1-13-4). Mineralization lies stratigraphically below Lower Devonian limestone within a package of tuffaceous sediments and pyritic felsic sills. Preliminary analytical data show no elevated base or precious metal values.

Polymetallic massive sulphide float occurs on Cominco's Fore More claims. The claims are situated outside the map area, on the south tributary of More Creek (NTS 104G/2) about 10 kilometres north of the headwaters of Forrest Kerr Creek. More than 200 massive sulphide boulders have been found in glacial outwash gravels. They are chiefly fine-grained massive sulphides containing, in order of abundance, pyrite, sphalerite, galena, barite, chalcopyrite and, locally, silver minerals. Texturally the mineralization varies from massive to layered. Less common are limestone boulders with crosscutting massive pyrite mineralization. A thin bedded fossilliferous limestone boulder shows preferential replacement by massive sulphides along bedding planes. This limestone contains probable algal laminations or stromatoporoid *Favosites* sp. of Late Ordovician to Middle Devonian age (B.S. Norford, personal communication, 1989). Stream sediment sampling (RGS, 1988) detected no geochemical anomalies in streams draining the area containing these massive sulphide boulders.

MINERAL POTENTIAL

Deposit models applicable to the geological setting and known mineral occurrences in the map area indicate favourable potential for the following:

(1) **Precious metal bearing skarn mineralization** occurs at the McLymont Creek prospect and Ken showing. Mississippian and Permian limestones are abundant in the western part of the map area but significant mineralization is confined to thin-bedded limy horizons and only where these rocks are cut by northeasterly trending structures.

TABLE 1-13-1. MINFILE OCCURRENCES IN THE FORREST KERR CREEK MAP AREA 104B/15.

MINFILE No.	NAME	COMMODITY	DEPOSIT TYPE	DESCRIPTION
104B 003	Don, Don 12, Don 40	Cu, Ag, Au	SKARN	Skarn mineralization occurs near the contact between Permian limestone and a Jurassic(?) diorite intrusion; mineralization includes disseminated pyrite, chalcopyrite and tetrahedrite.
104B 024	Mag	Fe, Cu	SKARN	Skarn mineralization occurs near the contact between Permian limestone and a Jurassic diorite intrusion; mineralization consists of massive magnetite with minor pyrite and chalcopyrite.
104B 027	Ken, Dirk, Glacier, Rope	Au, Cu	SKARN	Limy beds within a Permian volcanic/sedimentary package of rocks contains gold-bearing skarn mineralization consisting of garnet, epidote, magnetite, chalcopyrite, pyrite and possibly tetrahedrite; grab samples returned assay values to 33.7 grams per tonne gold.
104B 126	McLymont, Camp	Au, Ag, Cu	VEIN	Northwest-trending pyrite-chalcopyrite-quartz-carbonate-barite veins hosted by a quartz-rich granite.
104B 281	Northwest, Warrior 4	Au, Ag, Cu, Ba	SKARN	Mineralized zones consisting of barite, calcite, gypsum, coarse-grained magnetite, hematite, pyrite, chalcopyrite, sphalerite and galena occur as stratabound bands within chert layers and along bedding contacts between marble and chert; hostrocks are of probable Mississippian or Permian age.
104B 282	Gab 9, Gab 7, Warrior 7	Au, Ag, Cu	SHEAR/FAULT	A northeast-trending fault structure hosts gold-bearing sulphide and oxide material; 1988 diamond drilling reports indicate 900 metres in five holes.
104B 332	NE McLymont	Au, Ag	VEIN	Narrow auriferous quartz-pyrite veins cut a granitic intrusion; narrow rusty limonitic dikes of granite cut Permian and Mississippian stratified rocks; hornfelsed siltstone pendants within the granite body contain massive and disseminated pyrite and some sphalerite.
104B 333	Gab NW	Au	SKARN	Mineralization consisting of tetrahedrite, malachite and barite occurs along a northeast-trending fault thought to be an extension of the McLymont Northwest zone (MINFILE 104B 281); ten diamond-drill holes totalling about 900 metres returned best assay value of 4.78 grams per tonne gold over an interval of 2.5 metres.
104B 334	Cuba, Gab 8	Ag, Pb, Zn, Ba, Cu	SKARN	Mineralization consisting of silver, sphalerite, galena and barite is found within a northeast-trending fault that cuts Permian limestone; seven best samples from two copper-barite areas returned values averaging 1023.0 grams per tonne silver.
104B 335	Gab 12, SW	Ag, Zn, Pb, Cu, As, Au	VEIN	Mineralization consisting of magnetite, chalcopyrite, galena and sphalerite is associated with a northeast-trending fault crosscutting sediments of probable Permian and Mississippian age; gold mineralization occurs within iron-carbonate veins and pods mineralized with pyrite and coarse-grained magnetite; assay samples from a 1988 diamond-drill hole intersection returned values grading 77.14 grams per tonne gold over 60.0 centimetres.
104B 336	Gab 11, SE	Au, As, Fe, Cu, Ag	SHEAR/FAULT	Massive fine-grained pyrite occurs within sedimentary rocks of Permian and Mississippian age; a grab sample from a pyritized zone returned values of 23.5 grams per tonne gold and 116.9 grams per tonne silver.
104B 337	Gab 12 NE	Au	SHEAR/FAULT	A northeast-trending fault within Permian and Mississippian sedimentary rocks contains massive and disseminated magnetite, chalcopyrite, sphalerite and galena; grab samples from a gossanous zone contain values as high as 26.7 grams per tonne gold.
104B 347	Egg, Verjoy, Ret 7	Au	VEIN	Quartz-barite veins infilling shear or fracture systems are mineralized with pyrite, chalcopyrite, galena, hematite, magnetite and malachite; gold values are low but silver, cobalt and copper values are above background; massive pyrite veins to 5.0 centimetres width contain chalcopyrite, hematite and magnetite.
104B 350	Adrian	Au, Ag, Cu, Pb, Zn	VEIN	Quartz veins hosted by quartz-rich granite returned gold values ranging from 2.7 to 30.0 grams per tonne.
	North Ridge, Ridge, South Forrest	Au, Ag, Cu, Pb, Zn	VEIN	Mineralization consists of gold and silver-bearing chalcopyrite-quartz stockwork veins and shears with one or more of the following: malachite/azurite, arsenopyrite, galena, bornite, and hematite; these mineralized veins are hosted within Eastern assemblage rocks of the Paleozoic Stikine assemblage.

(2) **Stratabound mineralization** occurs in Paleozoic and Jurassic strata as the following:
   (a) Polymetallic massive sulphide mineralization occurs north of the map area on the Fore More claims and pyritic volcanic horizons are intercalated within the Lower Devonian limestone section. The Paleozoic Stikine assemblage rocks are a potential exploration target for volcanogenic massive sulphide deposits. The Devonian mafic and felsic metavolcanic package appears to have the highest potential.
   (b) Middle(?) Jurassic basaltic and dacitic submarine volcanic strata underlie the eastern and southeastern parts of the map. This stratigraphy resembles the geological setting of the Eskay Creek gold prospect. Mineral potential is high.

(3) **Mesothermal quartz veins** and stockworks located adjacent to the Newmont Lake graben and West Slope fault warrant further exploration.

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NOTES

GEOLOGY OF THE STIKINE RIVER – YEHINIKO LAKE AREA, NORTHWESTERN BRITISH COLUMBIA (104G/11W AND 12E)

By Derek A. Brown and Charles J. Greig

KEYWORDS: Regional geology, Stikine River, Yehiniko Lake, Stikine assemblage, Stuhini Group, Hazelton Group, Sustut Group, Sloko Group, Nightout pluton, Yehiniko pluton.

INTRODUCTION

The second summer of 1:50 000 geological mapping for the Stikine project was completed in the Yehiniko Lake (104G/11W) and Chutine River map areas (104G/12E), to adjoin mapping of the Scud River (Brown and Gunning, 1989a, b) and Galore Creek (Logan and Koyanagi, 1989a, b) areas that were completed in 1988. The present project area, approximately 30 kilometres southwest of Telegraph Creek, lies within a northwest-trending mineral-rich belt that includes important precious and base metal deposits such as Premier, Sulphurets, Eskay Creek, Johnny Mountain, Snip, Galore Creek and Golden Bear (Figure 1-14-1). A gravel road provides four-wheel-drive access to the northwest corner of the project area and the central part can be reached by boat along the Stikine River, but practical access to much of the area is by helicopter from Telegraph Creek or Dease Lake.

Fieldwork in 1989 focused on Mesozoic stratigraphy. Preliminary results include recognition of: (1) weakly deformed Lower to Middle Jurassic, Upper Cretaceous to Eocene (?) and Eocene successions preserved in the central part of the map area and resting unconformably on more highly deformed Upper Triassic rocks; (2) four episodes of plutonism; (3) moderate mineral potential; narrow, discontinuous gold-bearing quartz veins and auriferous porphyry copper mineralization are possible exploration targets; and (4) an extensive pyritic alteration zone that trends northeast across the map area is related to a Middle Jurassic (?) dike swarm.

The project area straddles the boundary between the Coast and Intermontane belts, and is underlain by rocks of the Stikine Terrane, an integral part of the Intermontane Belt. Previous regional studies and the regional geologic framework were summarized in Brown and Gunning (1989a).

STRATIGRAPHY

The stratigraphic succession consists of: pre-Permian arc-volcanic and sedimentary rocks and Permian platformal limestone of the Stikine assemblage; bimodal, dominantly marine arc-volcanic and related sedimentary rocks of the Upper Triassic Stuhini Group; subaerial and subordinate marine Lower to Middle Jurassic volcanic and sedimentary rocks, equivalent to the Hazelton Group; nonmarine clastic rocks of the Upper Cretaceous to Eocene (?) Sustut Group; and felsic to mafic volcanic rocks of the Eocene Sloko Group (Figures 1-14-2 and 3).

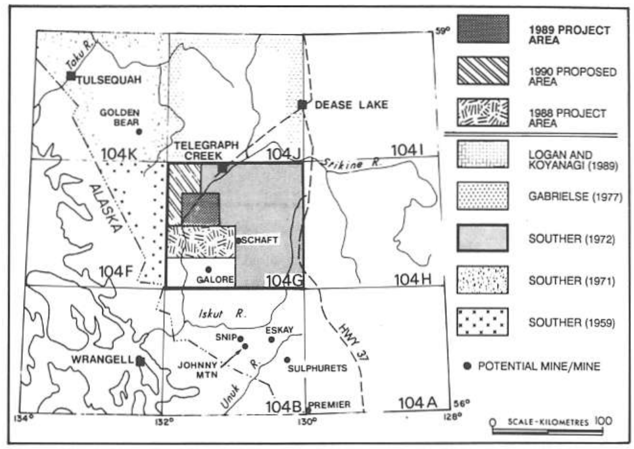


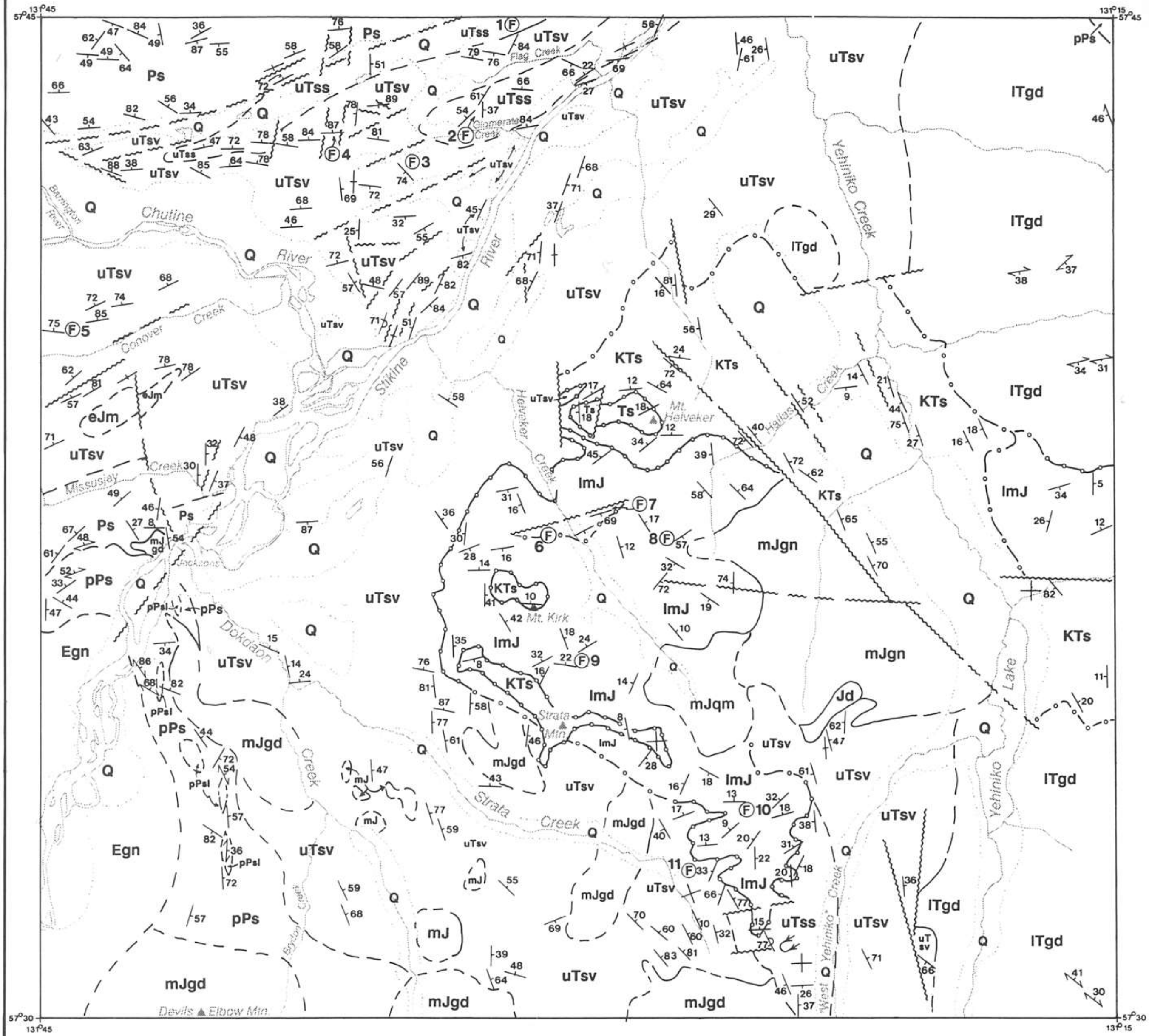
Figure 1-14-1. Location map for Stikine project with areas of previous work indicated.

PALEOZOIC STIKINE ASSEMBLAGE

(UNITS Ps & pPs)

Paleozoic Stikine assemblage rocks underlie three small areas in the western half of the map area, each distinguished by different structural trends and the presence or absence of thick Permian limestone. Comparative studies of these and more complete sections in the Scud River area are the subject of a masters thesis by M.H. Gunning being undertaken at the University of Western Ontario. Detailed discussion of several Scud River sections is found in Gunning (1990, this volume).

The oldest rocks in the map area, discontinuous Mississippian limestone lenses within pyritic metasiltstone and argillite (Kerr, 1948), are exposed in the Devils Elbow area. In contrast to Triassic and younger units, these rocks display a penetrative foliation that is commonly bedding parallel, north to northwest-trending, and moderate to steeply east-dipping. Thick and complexly deformed Permian limestone, which is missing in the Devils Elbow area, dominates exposures near Missusjay Creek and the Barrington River. In the Missusjay Creek area Permian limestone is structurally underlain by silicic metasedimentary rocks, similar to those mapped by Brown and Gunning to the south (Units A, B and C; 1989a). In the Barrington River area, the section includes minor maroon and green tuffaceous mudstone, silicic tuff and chert. In contrast to the Devils Elbow area, major structures trend northeast (Missusjay Creek) and east-west (Barrington River). The vergence, age and origin of these structures are poorly understood. An inferred north-dipping reverse fault in the Barrington River area, which juxtaposes Stikine assemblage rocks on the north with Stuhini Group volcanic rocks on the south, suggests that at least some of the deformation occurred in post-Late Triassic time.



STRATIFIED ROCKS

QUATERNARY

Q	UNCONSOLIDATED DEPOSITS

TERTIARY - SLOKO GROUP

Ts	DACITE, TRACHYANDESITE FLOWS, PYROCLASTIC ROCKS

UPPER CRETACEOUS TO TERTIARY - SUSTUT GROUP

KTs	CONGLOMERATE, WACKE, RHYOLITE TUFF

LOWER TO MIDDLE JURASSIC

lmJ	ANDESITE, BASALT, RHYODACITE TUFFS & FLOWS, WACKE

UPPER TRIASSIC - STUHINI GROUP

uTs	BASALT, ANDESITE TUFFS & FLOWS (**uTsv**); SILTSTONE, WACKE, LIMESTONE (**uTss**)

PERMIAN - STIKINE ASSEMBLAGE

Ps	LIMESTONE, RARE TUFF

PERMIAN OR OLDER - STIKINE ASSEMBLAGE

pPs	FOLIATED ARGILLITE, SILTSTONE, ANDESITE FLOWS, MINOR TUFF, LIMESTONE (**pPsl**)

INTRUSIVE ROCKS

EOCENE

Egn	**SAWBACK PLUTON** -- BIOTITE GRANITE

MIDDLE JURASSIC

mJ	**YEHINIKO PLUTON** -- PINK HORNBLENDE BIOTITE GRANITE (**mJgn**); QUARTZ MONZODIORITE (**mJqm**)
mJgd	GRANODIORITE, QUARTZ MONZODIORITE

LATE TRIASSIC TO MIDDLE JURASSIC

Jd	HORNBLENDE DIORITE

EARLY JURASSIC

eJm	**CONOVER PLUTON** -- HORNBLENDE MONZONITE, MONZODIORITE

MIDDLE TO LATE TRIASSIC

lTgd	**NIGHTOUT PLUTON** -- FOLIATED, BIOTITE HORNBLENDE GRANODIORITE

KILOMETRES

0 1 2 3 4 5

SYMBOLS

Geological boundary (defined, approximate or inferred) ....

Unconformity (defined, approximate or inferred) ....

High-angle fault (defined, approximate or inferred) ....

Overturned synform ....

Fossil location.. (F) 2

Figure 1-14-2. Simplified geology of the Stikine River – Yehiniko Lake area (104G/11W and 12E).

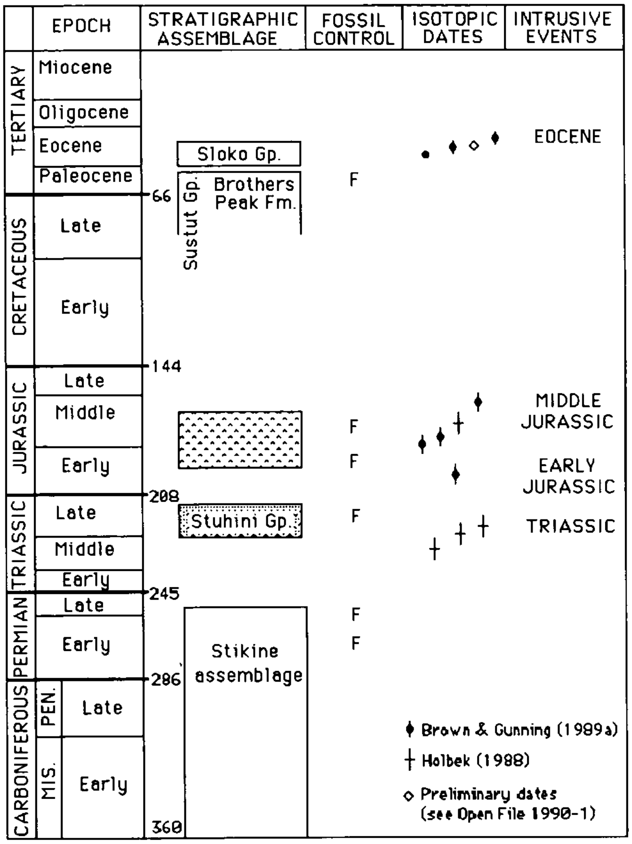


Figure 1-14-3. Schematic column of stratified rocks in the map area, with intrusive events shown.

UPPER TRIASSIC STUHINI GROUP

(UNIT uTs)

The Upper Triassic Stuhini Group is the most abundant map unit within the project area. The best exposures occur in the south near the headwaters of Dokdaon, Strata and "West Yehiniko" creeks, and in the northwest on the ridges west and northeast of the confluence of the Stikine and Chutine rivers. Much of the area underlain by Stuhini Group rocks, such as Stikine River valley, is at lower elevations and covered with unconsolidated deposits and thick vegetation. Poor exposure in these areas, a lack of marker units, and the similarity in colour and weathering characteristics of constituent lithologies, hinder both subdivision of the Stuhini Group and elucidation of its structural features.

Fossil control for the Stuhini Group in the project area is relatively good. Six new localities with Upper Triassic (probable Norian) fossils were discovered during mapping for this project (Fossils at Localitles 3, 4, 5, 7, 11 identified by H.W. Tipper, personal communicatlon, 1989; Locality 6 by T.P. Poulton, personal communication, 1989 as shown in Figure 1-14-2). In addition, two localities with fossils interpreted as Lower Jurassic by Kerr (1948) near Glomerate and Flag creeks (Localities 1, 2; Figure 1-14-2) were recollected, and also yielded Upper Triassic (probable Norian) fossils (T.P. Poulton, personal communication, 1989). This suggests that Lower Jurassic rocks mapped by Kerr (1948) and Souther (1972) in the Stikine valley are probably Upper Triassic.

Volcanic rocks of the Stuhini Group, although similar in some respects to overlying Lower to Middle Jurassic volcanic rocks, are more mafic and contain a larger component of subaqueous deposits. In addition, Upper Triassic flows, sills and beds typically dip at high angles, whereas Lower to Middle Jurassic rocks are typically gently dipping. Augite-bearing volcanic rocks, regarded as a hallmark of the Upper Triassic in British Columbia, occur in both the Upper Triassic and Lower to Middle Jurassic packages, but are more abundant and augite-rich in the Stuhini Group.

The most abundant Stuhini Group lithologies are mafic crystal-lithic lapilli tuff, ash tuff and lapilli tuff-breccia. They are typically dark green, massive and rich in pyroxene and plagioclase. Layering in tuff units is uncommon and attitudes must generally be measured on interbedded finer grained pyroclastic or volcaniclastic units.

Dark grey-green pyroxene and plagioclase-phyric (amygdaloidal) basalt or basaltic andesite flows containing blocky pyroxene and lath-shaped plagioclase phenocrysts appear to be characteristic of the Stuhini Group in the project area. Less distinctive but perhaps more abnndant Stuhini Group flows include pyroxene-phyric and plagioclase-phyric basalt or basaltic andesite. Locally, such as in the hills northeast of the confluence of the Stikine and Chutine rivers, pillowed basalt flows were recognized. On the ridges immediately south of Strata Creek, more mafic volcanic rocks are subordinate, and plagioclase-rich andesite flow-breccia and tuff-breccia dominate.

Felsic volcanic rocks are an important and widespread, yet volumetrically insignificant component of the Stuhini Group. Pale green to dark grey, pyritic, laminated subaqueous siliceous dust to ash tuff are most common, but siliceous lithic lapilli tuff and welded to unwelded ignimbrite are present locally. In places, mafic crystal-lithic tuff and tuff-breccia contain felsic clasts. The presence of this felsic component amid the predominantly mafic volcanic pile indicates that Stuhini Group volcanism was bimodal.

Though the bulk of the Stuhini Group consists of resistant volcanic rocks, recessive sedimentary rocks are commonly part of the section and may underlie a greater area than is apparent in outcrop (Unit uTss; Figure 1-14-2). Olive coloured mafic and lesser brown-weathering arkosic wacke, siltstone and shale, minor tuffaceous(?) and sandy limestone, limestone conglomerate/breccia, granitoid-bearing polymictic conglomerate, and black ribbon chert occur in belts that trend east-northeast from the confluence of the Barrington and Chutine rivers. In a narrow belt that trends roughly east-west across Helveker Creek, large lenses of steeply south-dipping, thick-bedded to massive limestone and minor thin bedded, light grey chert grade upward into well-bedded, dark grey siltstone and green sandstone (Plate 1-14-1). Between Strata and "West Yehiniko" creeks a generally conformable but somewhat disrupted sedimentary package includes well-bedded tuffaceous rocks. Several horizons of pale grey weathering, thick-bedded to massive, micritic and bioclastic limestone layers (up to 20 metres thick) occur at the transition with massive flows and breccias to the southwest. Dark grey to black siltstone, green and pale grey arkose and black shale overlie the limestone. Maroon, mauve and brick-red massive lapilli tuff, tuffaceous mudstone and lesser laminated limy ash tuff lie above the sediments. The distribution of these "sedimentary belts," and the orientation of their beds relative to layered rocks in adjacent volcanic-rich parts of the Stuhini Group, suggest that the sedimentary rocks are younger and that Stuhini Group volcanism in the project area was largely pre-Norian.

Stuhini Group rocks are characterized by steep dips. Folds are recognized only within well-bedded, dominantly sedimentary sequences; more massive volcanic rocks appear unfolded and are rarely foliated. This suggests that structures are high-level and that lithology controlled structural style. A high-level structural setting is corroborated by preliminary petrographic study of metamorphic mineral assemblages, which indicates that the Stuhini Group rocks are metamorphosed only to zeolite grade. The unconformable contact between folded (steeply-dipping, northeast-verging overturned beds) and faulted Upper Triassic rocks and overlying, gently dipping Lower Jurassic flows and tuffs (Plate 1-14-2) indicates that much of this deformation occurred between Late Triassic and Early Jurassic time.

The nature of the contact between the Stuhini Group and rocks of the Stikine assemblage in the project area is uncertain. Contacts are poorly exposed and are interpreted as structural. However, near the headwaters of Conover Creek, the presence of abundant clasts typical of the Stikine assemblage within limy crystal-lithic lapilli tuff of the Stuhini Group suggests that the contact may originally have been an unconformity (M.H. Gunning, personal communication, 1989).

Lower to Middle Jurassic Rocks

(UNITS lMj)

In the central part of the map area, Lower to Middle Jurassic volcanic and minor sedimentary rocks underlie much of the drainage of Helveker Creek (Figure 1-14-2). They also outcrop northeast of Yehiniko Lake along the east margin of the map area, where they were mapped as Stuhini Group by Souther (1972). Ammonite fragments, belemnites, brachiopods and scarce bivalves from three new fossil localities (Figure 1-14-2, Localities 8, 9 and 10) have been tentatively assigned Toarcian and Bajocian ages (Lower to Middle Jurassic; H.W. Tipper, personal communication, 1989). Prior to this study, fossils from one locality near Mount Kirk were interpreted to be Jurassic (Locality 28 of Kerr, 1948). Potassium-argon dating of several unaltered flows from the Jurassic map unit is in progress (Open File 1990-1).

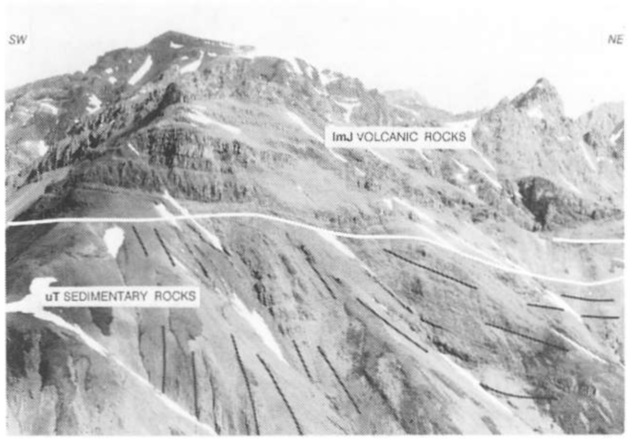


Plate 1-14-2. View northwest to folded and faulted Upper Triassic siltsone and sandstone unconformably overlain by gently dipping Lower to Middle Jurassic volcaniclastic rocks and flows; ridge between Strata and "West Yehiniko" creeks.

Plate 1-14-1. View northeast to steeply south-dipping Upper Triassic sedimentary rocks unconformably overlain by gently dipping Lower to Middle Jurassic volcanic rocks; northeast side of Helveker Creek.

The Lower to Middle Jurassic section is characterized by the presence of gently dipping, massive and thick (up to tens of metres), plagioclase-rich andesite flows and tuffs with maroon or purple hues. Tuffs and flows occur in subequal quantities. Crystal-lithic andesite tuff-breccia and lapilli tuff are commonly maroon to brick-red and less typically grey-green to mottled maroon and green. They contain angular volcanic country rock fragments up to 1 metre in diameter. Thin to thick-bedded units of poorly indurated, flaggy, maroon, mauve and pale green ash and fine-grained lapilli tuff are commonly interbedded with coarse-grained tuffaceous rocks. Rare pale grey grit and maroon tuffaceous grit, thought to be derived from underlying plutonic rocks, occur northeast of Yehiniko Lake and on Mount Kirk.

The most common flows are pyroxene and plagioclase-phyric andesite. They are typically grey-green with a faint dark purple or maroon hue, and are locally flow banded and amygdaloidal (zeolite, calcite, chlorite, epidote, quartz, pyrite). Subordinate flows include: buff to rusty weathering, flow-banded aphanitic rhyolite (up to 60 metres thick) that occurs at the base of the section in the southwesternmost exposures (Plate 1-14-3); pink-weathering hematitic flow-banded and locally flow-folded and autobrecciated rhyolitic ignimbrite on the southwest flank of Mount Kirk (traced over 2 kilometres along strike); and dark green, chlorite-amygdaloidal olivine basalt exposed 2 kilometres southeast of Mount Helveker and 3 kilometres east and southeast of Strata Mountain.

Facies changes within the package suggest that subaerial and marine deposition were contemporaneous. For example, on the ridge 1.5 kilometres north of Strata Mountain, a southwest to northeast facies transition occurs between maroon (subaerial) andesite flows and tuff-breccia, and pale green (subaqueous) andesite ash and dust tuff, green andesite flows with interbedded dark green siltstone and arkosic wacke containing belemnites, bivalves and rare ammonites. In turn, the marine rocks grade vertically and laterally (farther to the northeast) into maroon tuff-breccia. This suggests that contemporaneous subaerial and marine deposition occurred in an emergent island setting. Overall, subaerial rocks appear to dominate the Lower to Middle Jurassic package, but other marine rocks are exposed southeast of Strata Mountain (argillite, siltstone, fossiliferous wacke, carbonate-cemented amygdaloidal basalt pillow-breccia and rare bioclastic limestone lenses), north of Mount Kirk (mauve lapilli tuff with limestone lenses) and northwest of

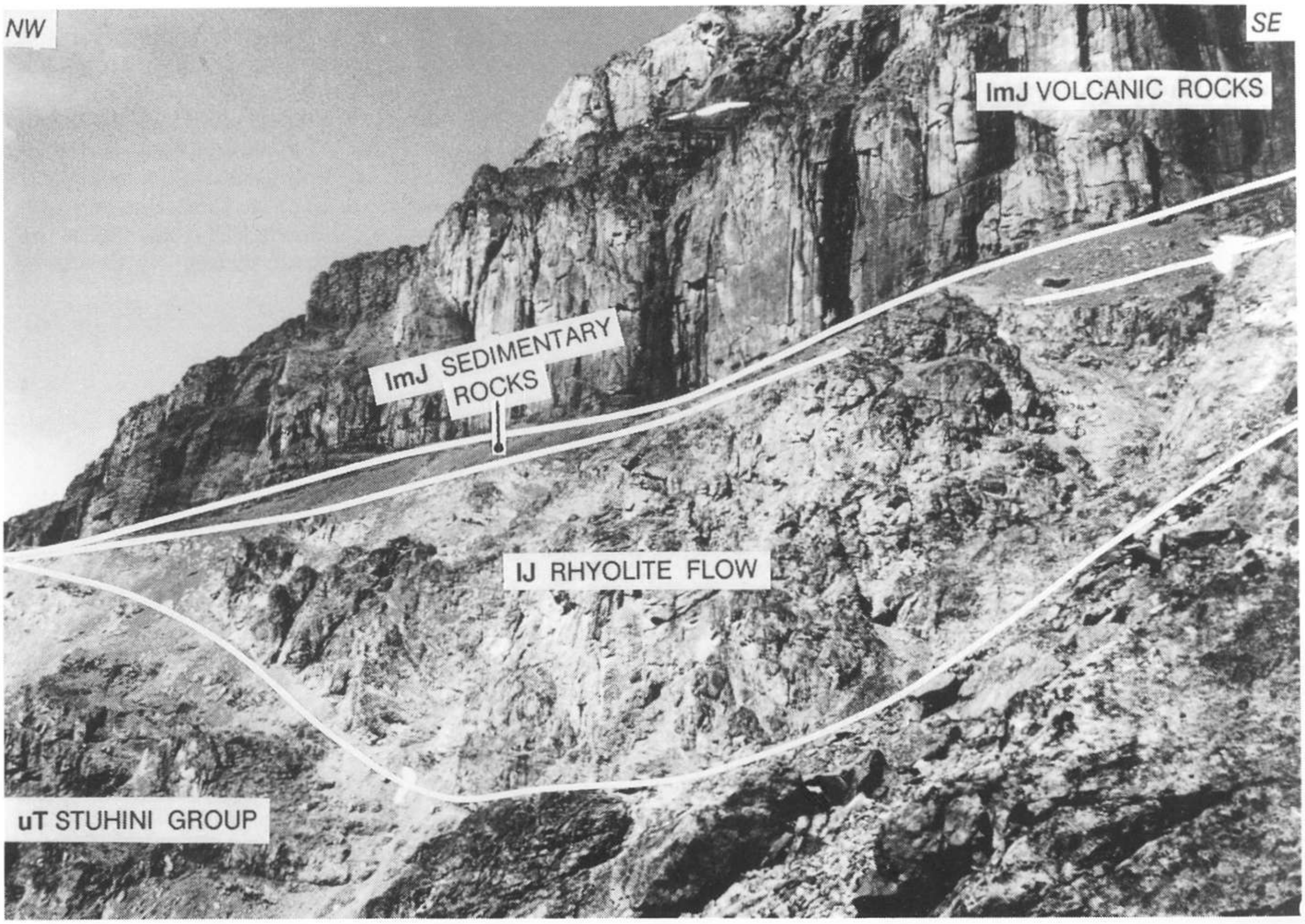


Plate 1-14-3. View northeast to Stuhini Group rocks overlain unconformably by pale-weathering rhyolite flow which is in turn conformably overlain by a relatively thin layer of gently dipping Lower Jurassic (Toarcian ?) sedimentary rocks and by prominent cliffs of columnar-jointed Lower to Middle Jurassic basaltic andesite flows; northeast of Strata Creek.

Mount Kirk (green, carbonate-cemented arenaceous sandstone, siltstone and polymictic cobble conglomerate, and pale grey massive limestone).

In the central part of the project area, Lower to Middle Jurassic rocks rest unconformably on rocks of the Stuhini Group except on their east margin, where they are intruded by the Yehiniko pluton. The unconformity is marked locally by maroon polymictic cobble to boulder conglomerate, but in most places there is an angular discordance of up to 90° (Plate 1-14-2) between Lower to Middle Jurassic lithologies and the underlying Stuhini Group. The paleosurface is commonly highly irregular (Plate 1-14-3). Relief on the unconformity is variable, but locally appears to be as much as 500 metres and quite abrupt (*e.g.* Helveker Creek). Northeast of Yehiniko Lake, maroon tuff and minor tuffaceous grit rest unconformably on the Nightout pluton.

The Jurassic package appears much less deformed relative to the underlying Stuhini Group, except for several high-angle brittle faults with minor vertical offsets. Preliminary petrographic study of the Jurassic volcanic rocks suggests that they, like the Stuhini Group rocks, have undergone zeolite facies metamorphism. Locally the zeolite assemblage is overprinted by an albite-epidote assemblage (calcite, chlorite, epidote, albite, quartz, pyrite, rare actinolite) that is spatially related to plutons and dike swarms.

UPPER CRETACEOUS TO EOCENE (?)

SUSTUT GROUP (Unit uKs)

Sedimentary and subordinate volcanic rocks correlated on a lithologic basis with the Brothers Peak Formation of the Sustut Group (Souther, 1972; Eisbacher, 1974) occur in a belt that trends northwest from the eastern boundary of the map to Mount Helveker and as outliers southwest of the belt at Mount Kirk and Strata Mountain (Figure 1-14-2; Plate 1-14-4).

Extremely poorly indurated polymictic cobble conglomerate, in places resembling Quaternary glaciofluvial deposits, characterize the Sustut Group in the project area. Sandstone and siltstone and conspicuous pale-coloured rhyolite tuff are subordinate. Basalt flows are rare. Brick-red, brown and grey conglomerates are massive to moderately well bedded and locally contain crossbeds and foreset beds. They are generally moderately sorted and clasts are well rounded, except in Yehiniko Creek valley, where debris flows or fanglomerates are common and clasts, locally up to 3 metres in diameter, vary from well rounded to subangular. Clasts are derived from all older lithologies and on Mount Kirk include distinctive pink flow-banded rhyolite derived from immediately underlying Lower to Middle Jurassic volcanic rocks. Only conspicuous white bull quartz clasts (up to 10 per cent) have an unknown provenance. Coaly plant stems, leaves and wood fragments occur locally. Sandstone lenses and beds are buff, brown, pale green, grey, maroon and olive, and locally contain fresh biotite. Thin but prominent, white, mauve and pale green, locally welded, biotite quartz-eye rhyolite to rhyodacite ash to lapilli tuff horizons, up to 10 metres thick, are sometimes interbedded with conglomerate (Plate 1-14-5).

The Sustut Group lies with angular unconformity on Lower Jurassic volcanic rocks on Mount Kirk and Mount Helveker and on Upper Triassic volcanic rocks on Strata Mountain (Plate 1-14-4). The basal contact of the Sustut Group is 350 metres higher in elevation at Strata Mountain than at Mount Helveker, suggesting either differential uplift in Tertiary time or variable relief on the Cretaceous paleosurface. East of Yehiniko Lake subhorizontal conglomerate lies unconformably on the Nightout pluton. Northwest of

Plate 1-14-4. View northwest to Strata Mountain which is underlain by Sustut Group conglomerate and sandstone lying unconformably on altered Lower to Middle Jurassic volcanic rocks cut by Middle Jurassic (?) dikes.

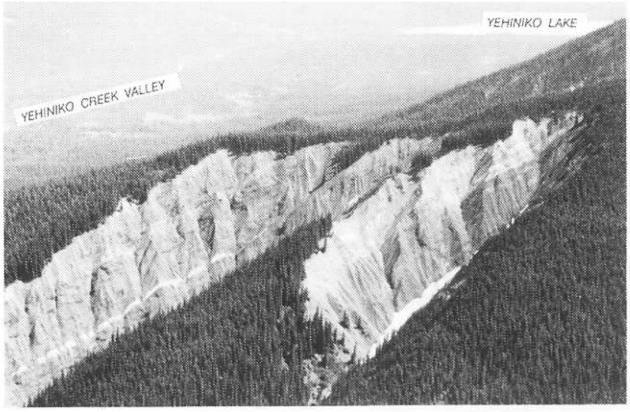


Plate 1-14-5. View southeast to bedded Sustut Group conglomerate with a prominent horizon of white rhyolite tuff.

Yehiniko Lake steeply northeast-dipping strata of the Sustut Group are faulted against the Yehiniko pluton. In the Mount Helveker area, basalt and felsite dikes, believed to be Sloko Group feeders, intrude the sediments.

The deformational style of the Sustut Group in Yehiniko Creek valley is characterized by abrupt changes in dip. Lack of evidence for folds, except for a possible dragfold on the eastern flank of Mount Helveker (R.T. Bell, written communication, 1989), and the poorly sorted nature of much of the conglomerate, suggest that dip changes may be in part related to syndepositional block faulting.

Palynomorphs from a sample collected 4 kilometres east of the map boundary indicate an Early Paleocene age, which is younger than the age of the Brothers Peak Formation in its type area (A.R. Sweet, personal communication, 1989). Fossil deciduous leaves and coaly plant stems have been submitted for identification, and biotites from rhyolite tuff and tuffaceous sandstone are being dated.

A high-energy fluvial paleoenvironment with alluvial fans feeding a fault-controlled basin that received intermittent pyroclastic deposits and rare flows is envisioned for the Sustut Group. Imbricated clasts and foreset beds at several localities suggest north to northeastward-directed paleocurrents.

EOCENE SLOKO GROUP

(UNIT Ts)

Sloko Group volcanic rocks, the youngest stratified rocks in the map area, cap weathered, limonitic Sustut Group conglomerate on Mount Helveker. The exposed section is about 330 metres thick and comprises flat-lying tuff, breccia, and dacite and trachyandesite flows.

The base of the section (at 1740-metre elevation) consists of pale green to white, laminated welded tuff with prominent eutaxitic textures. On the west side of Mount Helveker, welded tuff is overlain by brick-red volcanic breccia and well-bedded lithic lapilli tuff (pyroclastic surge and flow deposits) which is in turn overlain by light grey weathering hornblende and plagioclase-phyric dacite flows and sills. Lithologically similar dikes cut Sustut Group rocks on Mount Kirk, 6 kilometres to the southwest. Overlying the dacite flows near the top of the south side of Mount Helveker are resistant, dark brown weathering, columnar-jointed plagioclase-phyric trachyandesite flows 2 to 3 metres thick. The top of Mount Helveker is capped by poorly indurated, light grey weathering, andesitic volcanic breccia and hornblende crystal lithic lapilli tuff. Tuff contains fresh hornblende in grey to olive lithic fragments and as crystal fragments. Tuff was collected for K-Ar age determination.

The Sloko Group sequence is somewhat disrupted by north-trending, high-angle normal faults with displacements of at least 30 metres. It is also cut by northeast to east-trending columnar-jointed basalt dikes.

Overall there is a subtle angular discordance ($<10°$) between the Sustut and Sloko groups, however, beds are locally conformable. The contact marks an abrupt change from largely high-energy fluvial deposition with intermittent volcanism, to mainly subaerial felsic to intermediate volcanism.

INTRUSIVE ROCKS

Intrusive rocks underlie roughly 25 per cent of the project area. This is in sharp contrast to the adjacent Scud River area, of which approximately 75 per cent is underlain by intrusive rocks. Preservation of Lower Jurassic, Upper Cretaceous and Eocene stratified rocks in the current map area and their absence in the Scud River area suggests that uplift and erosion were greater to the south. Age assignments in the following section are based on field relationships or lithologic similarity with nearby dated intrusions. Potassium-argon dating is in progress for several of the suites. Compositions of intrusive rocks were determined from slabbed and stained hand specimens utilizing the classification scheme of Streckeisen (1976). Four plutonic episodes are tentatively defined: Middle to Late Triassic, Early Jurassic, Middle Jurassic and Eocene.

MIDDLE TO LATE TRIASSIC

NIGHTOUT PLUTON (UNIT lTgd)

The Middle to Late Triassic (Holbek, 1988) Nightout pluton, consisting of foliated to massive medium-grained biotite hornblende granodiorite that grades locally to tonalite, quartz monzonite, monzodiorite and diorite, underlies much of the eastern margin of the study area. It is continuous with exposures of Nightout pluton to the south, described by Brown and Gunning (1989a). In the present study area the pluton is characterized by a widespread magmatic foliation and by coarse-grained (up to 2 centimetres) poikilitic potassium feldspar grains.

In the northeast corner of the project area, the Nightout pluton intrudes Stuhini Group mafic volcanic rocks on its west margin, and mylonitic Stikine assemblage rocks on its northeast margin. Foliation in the pluton parallels the contact and the foliation in the country rocks. Along the east-central margin of the study area, the Nightout pluton is overlain unconformably by both Lower to Middle Jurassic volcanic rocks and by Upper Cretaceous to Eocene (?) conglomerate of the Sustut Group. A sample of biotite hornblende granodiorite collected near the pre-Lower Jurassic unconformity is being processed for K-Ar dating.

In the southeast corner of the study area, along the southwest side of Yehiniko Lake and farther to the south, unfoliated, medium-grained equigranular (biotite) hornblende tonalite, possibly a marginal phase of the Nightout pluton, intrudes and hornfelses mafic volcanic and volcaniclastic rocks of the Stuhini Group.

Early Jurassic (?)

CONOVER PLUTON (UNIT eJm)

West of the confluence of the Stikine and Chutine rivers, Stuhini Group rocks are intruded by distinctive seriate to plagioclase-porphyritic ("crowded") hornblende monzonite to monzodiorite. The intrusions, including the Conover pluton and numerous associated sills, though texturally heterogeneous, typically contain blocky to lath-shaped, locally trachytic plagioclase phenocrysts 3 to 5 millimetres long, in a groundmass of hornblende, potassium feldspar, plagioclase and quartz. Based on compositional and textural similarities with alkaline plutons to the south, the Conover intrusions are assigned an Early Jurassic age. The alkaline plutons to the south, however, are potassium feldspar megacrystic. Hornblende from the Conover pluton is being processed for K-Ar dating.

Late Triassic to Pre-Middle Jurassic

DIORITE AND QUARTZ MONZODIORITE (UNIT Jd)

Roughly 5 kilometres west of Yehiniko Lake, along the southern margin of the Yehiniko pluton, fine to medium-grained, unfoliated hornblende diorite intrudes and hornfelses Stuhini Group rocks and is itself intruded by granite and quartz monzonite apophyses of the Yehiniko pluton.

Southeast of Strata Mountain, on both sides of Strata Creek, Stuhini Group rocks are intruded by texturally heterogeneous hornblende diorite and subordinate leucodiorite. On the south side of Strata Creek, this intrusion varies from fine to coarse grained, is locally foliated and, in places, weak gneissic banding is developed. On the steep slopes north of Strata Creek, it is cut by numerous dikes, is intensely altered and, as a result, contact relationships with Lower to Middle Jurassic rocks to the north are uncertain. Pink, fine to medium-grained (biotite) hornblende granite to syenite dikes, thought to be comagmatic with the Yehiniko pluton, intrude the diorite.

On the northeast side of Strata Creek, west of Strata Mountain, medium-grained biotite-hornblende quartz monzodiorite intrudes rocks of the Stuhini Group. As with the pluton southeast of Strata Mountain, contact relationships with Lower to Middle Jurassic volcanic rocks to the north are uncertain.

Middle Jurassic

YEHINIKO PLUTON (UNIT mJgn)

West of Yehiniko Lake, distinctive pink, medium and locally fine-grained hornblende biotite granite to quartz monzonite and subordinate pale grey quartz monzodiorite and quartz monzonite comprise the Yehiniko pluton. It has a somewhat more limited distribution than shown by Souther (1972) and is more compositionally and texturally heterogeneous than Middle Jurassic plutons to the south, with which it has been correlated (Unit 9 of Brown and Gunning, 1989a, b).

On its northeast margin, the Yehiniko pluton is interpreted to be in fault contact with conglomerate of the Sustut Group. Contacts with Upper Triassic volcanic rocks and post-Late Triassic hornblende diorite on the south, and with Lower to Middle Jurassic volcanic rocks on the west, are intrusive and country rocks are strongly hornfelsed.

Near the west and southwest contacts of the Yehiniko pluton, abundant and distinctive pink dikes of biotite-hornblende-plagioclase (rarely potassium feldspar) porphyritic (quartz) syenite, granite and (quartz) monzonite intrude Lower to Middle Jurassic and Upper Triassic country rocks within an irregular north-northeast to northeast-trending zone of intense diking and alteration. This suggests that the dikes, and in part, the alteration, are genetically related to the Yehiniko pluton, and together with the prominent hornfels zone, suggest that the pluton is epizonal. Farther southwest along the trend of this zone, on the ridge between Dokdaon and Strata creeks, hornblende granite to quartz monzonite stocks and hornblende-plagioclase-porphyritic (quartz) syenite and (quartz) monzonite dikes have been tentatively correlated with the Yehiniko pluton. The stocks, which are associated with mineralization on the Dok claims (Ulrich, 1971), were formerly assigned to the Early Jurassic (Souther, 1972).

Granodiorite and Quartz Monzodiorite

(UNIT mJgd)

Along the southern edge of the project area, from the headwaters of Strata Creek west to Devils Elbow, five intrusions of unfoliated, medium-grained (biotite) hornblende granodiorite to quartz monzodiorite intrude rocks of the Stuhini Group or Stikine assemblage. The three southernmost plutons are continuous to the south with intrusions assigned to the Middle Jurassic suite by Brown and Gunning (1989a, b). A pluton continuous with that exposed along Dokdaon Creek, near the southern boundary of the present project area, yielded a Middle to Late Jurassic K-Ar hornblende date (158 ± 6 Ma; Brown and Gunning, 1989b). Other dates for their Middle Jurassic suite are somewhat older (hornblende: 182 ± 7 Ma; biotite: 163 ± 6 Ma; Brown and Gunning, 1989b).

Eocene

SAWBACK PLUTON (UNIT Egn)

In the southwestern corner of the study area, massive (hornblende) biotite granite, characterized by its medium to coarse grain size, well-developed but widely spaced joints and relatively unaltered nature, comprises the Sawback pluton. Brown and Gunning (1989b) obtained a Middle Eocene K-Ar date (48.0 ± 1.7 Ma; biotite) on a sample collected approximately 8 kilometres southwest of the present study area.

MINERAL OCCURRENCES

The 13 mineral occurrences recorded in MINFILE for the area can be subdivided into veins, volcanic-hosted porphyry copper occurrences and skarns (Table 1-14-1, Figure 1-14-4). None of them have defined reserves and most are apparently small, but exploration to date has been limited.

The Dok claims (MINFILE 104G 038, 43, 74) cover large limonitic and pyritic alteration zones which contain minor malachite, azurite and rare chalcopyrite and chalcocite. Irregular mineralized veins are associated with northeast-trending, pink granite to (quartz) syenite dikes that intrude Stuhini Group volcanic rocks.

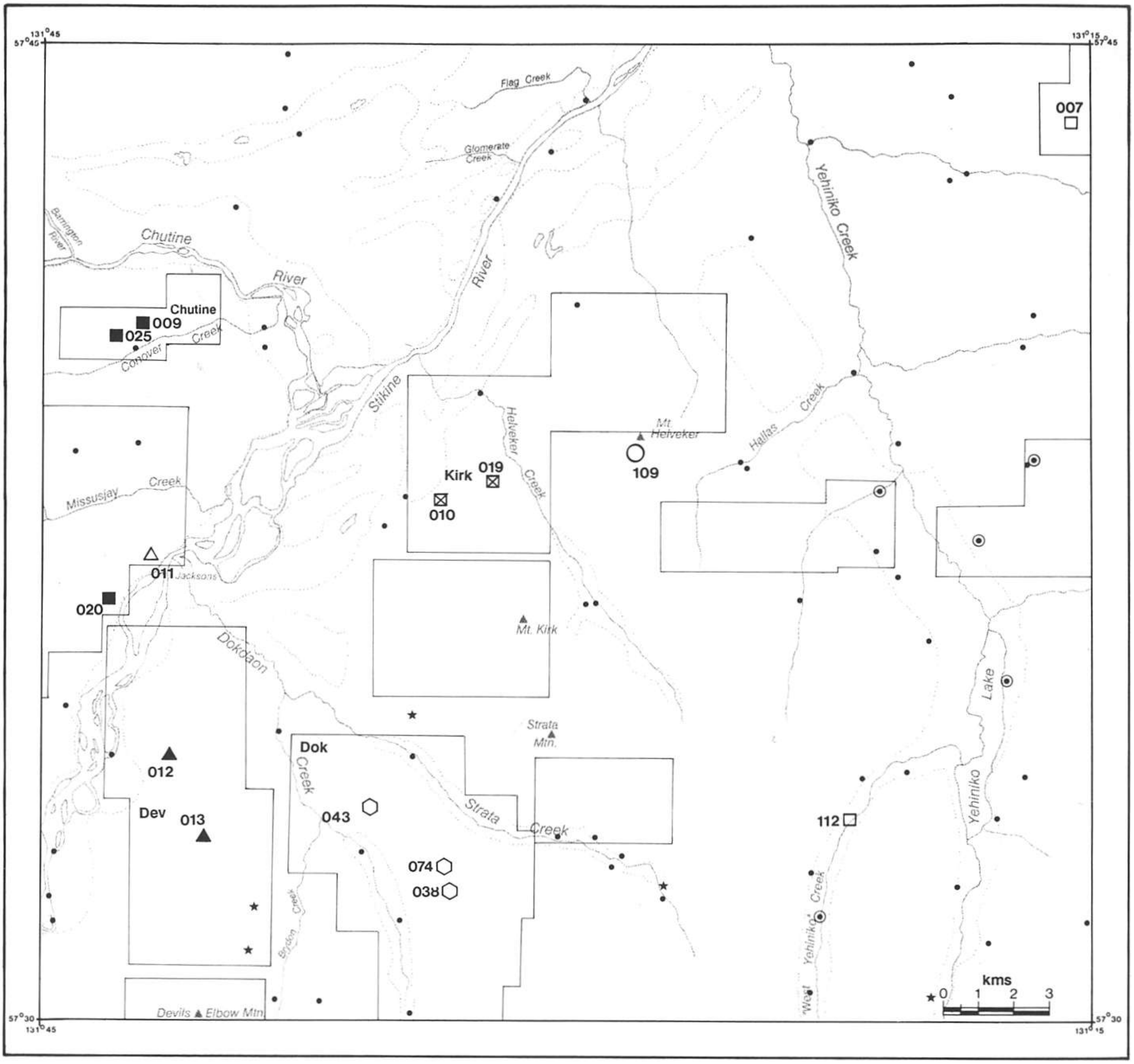


Figure 1-14-4. MINFILE occurrence localities, current claim locations (October, 1989), and RGS sample locations in the Stikine River – Yehiniko Lake area. Solid circles denote RGS silt sample locations, stars indicate multi-element anomaly sites and circled dots are single element gold anomaly sites. MINFILE occurrences are grouped according to Table 1-14-1: solid rectangles = Au-Ag base metal veins; crossed-rectangle = Au-Cu quartz veins; open rectangle = base metal veins; open hexagon = volcanic-hosted porphyry Cu; solid triangles = Ag-Au base metal skarn; open triangle = base metal skarn; open circle = stratabound U.

TABLE 1-14-1
SUMMARY OF MINERAL OCCURRENCES (104G/11W AND 12E)

Minfile (104g)	Name	Host	Economic Minerals	Alteration	Orientation	Work Completed	Possible Age	Reference
Au-Ag base metal quartz veins								
009	Jackson, BlK,	Stuhini vol.	cpy,gln,sph,py	Sil, py	NE to NW trending	trenching	Post-Tri	1,#591, 14,216
025	Lady Jane	Stuhini vol.	cpy,gln,sph,py	Sil, py	180°/60° W	trenching	Post-Tri	1,#591, 14,216
020	Lucky Strike,	Stikine assem.	Au,gln,cpy,sph,pyr	—	NW-trending	?	?	
Au-Cu quartz veins								
010	August	Stuhini vol.	cpy,bo,py	—	Steep SE-dipping	adit	Post-Tri	1,#13,662
019	Mt. Goat, Kirk Cu	Stuhini/Lower Jur?	cpy,bo	—	Flat pods	trenching	Post-Tri	1,#13,662
Base metal veins								
007	Callbreath	Nightout pluton	cpy,bo	ep, ch	240°/steep	trenching	Late Tri?	#4717
112	Yehiniko West	Stuhini vol.	cpy?	—	0°/steep E	none	Post-Tri	Souther (1972)
Volcanic-hosted porphyry Cu								
038	LLK	Stuhini vol.	cpy,cc,mo	propylitic		m,ss,tr	Mid Jur?	#3029
043	Dok	Stuhini vol.	cpy	propylitic		m,ss,tr,dr	Mid Jur?	#3029,3238
074	PR	Stuhini vol.	cpy	propylitic		m,ss,tr	Mid Jur?	#3029,#3846
Cu skarn								
011	Drapich	Stikine assem.	cpy,spy,gln,mag,py	calcsilicates	steep, irregular	none	Jur?	1
Ag-Au base metal-W skarn								
012	Devils Elbow	Stikine assem.	mag,pyr,gln,cpy,sph	calcsilicates	variable	3 adits	Mid Jur?	1,#11,262
013	Apex	Stikine assem.	gln,spy,mag,cpy,sph	calcsilicates	irregular	unknown	Mid Jur?	1
Stratabound U								
109	Hel	Sustut Group	Saleeite,torbernite	none	—	m,ss,tr	Tertiary	Bell (1981), #7708

Abbreviations: Au = native gold; bo = bornite; cc = chalcocite; ch = chlorite; cpy = chalcopyrite; ep = epidote; gln = galena; m = mapping; mo = molybdenite; mag = magnetite; py = pyrite; pyr = pyrrhotite; sch = scheelite; sil = silicification; sph = sphalerite; ss = soil sampling; tr = trenching; 1. = Kerr (1948); # = Assessment Report number.

Showings on the Kirk claims (MINFILE 104G 010, 19) are narrow, discontinuous, northeast-trending auriferous quartz veins with minor chalcopyrite and bornite that are hosted in sheared Stuhini Group volcanic rocks (Kerr, 1948). The Chutine claims (MINFILE 104G 009, 025) cover a 500-metre-long pyritic, siliceous alteration zone within Stuhini Group volcanic rocks and include the Lady Jane and Jackson quartz veins, which contain minor chalcopyrite, galena and sphalerite (Kerr, 1948). The Dev claims and claims at Jacksons (MINFILE 104G 011, 12, 13) cover precious and base metal skarn showings hosted in the Stikine assemblage.

Uranium mineralization on the Hel claims (MINFILE 104G 109) is hosted in Sustut Group conglomerate and sandstone. It occurs as secondary minerals in conglomerate, as radioactive coaly fragments in sandstone, and as radioactive Sloko Group trachyte talus (Bell, 1981). The source of the uranium is thought to be Sloko Group volcanic rocks which lie 20 to 30 metres above the main showing. Uranium may have been leached from the Sloko Group by migrating groundwater and precipitated on organic-rich material as secondary uranium minerals (Bell, 1981; Salat and Noakes, 1979).

NEW SHOWINGS

New showings are concentrated in four parts of the map area (Figure 1-14-4). Most are narrow (<5 cm) and discontinuous veins. Veins with massive pyrite, and chalcopyrite-bornite quartz-carbonate veins, occur in the headwaters of Strata Creek. Quartz-carbonate veins containing chalcopyrite, galena and pyrite outcrop northeast of Strata Creek and southwest of Strata Mountain. The fault zone southwest of Yehiniko Lake is erratically mineralized with pyrite. Sample locations and geochemical results for 90 grab samples collected during the study are available in Open File 1990-1.

GEOCHEMISTRY

Regional stream sediment sampling data (Regional Geochemistry Survey; B.C. RGS 20) for the Telegraph (104G) map sheet were released in July 1988 and included analyses of 75 stream sediment and water samples collected from the study area (Figure 1-14-4; Open File 1990-1). Numerous sample sites yielded anomalous geochemical results (*i.e.* 90th percentile based on the entire 104G population).

Multi-element anomalies occur in several places. Anomalies west of Brydon Creek are probably related to skarn mineralization along Devils Elbow ridge (MINFILE 104G 013). North and east of Strata Creek anomalies occur near galena-bearing quartz veins. Anomalies south of Yehiniko Lake occur close to north-trending faults that are erratically mineralized with pyrite. Coincident tungsten, molybdenum, uranium and fluorine anomalies at the southwest edge of the map area correlate with biotite granite of the Sawback pluton and may be derived from local greisen zones. Single element gold anomalies in Yehiniko Creek valley have no obvious source. Nickel and cobalt anomalies are scattered throughout the map area and appear to be associated with mafic Stuhini Group volcanic rocks.

EXPLORATION ACTIVITY

Exploration activity was moderate in 1989. Preliminary property work and regional prospecting were done by Cominco Exploration Limited, Homestake Mineral Development Company, Equity Engineering Limited and Coast Mountain Geological Services Limited.

MINERAL POTENTIAL

Several areas of RGS anomalies and new small showings remain unstaked and deserve further exploration and evaluation. The syenite dike swarm and associated limonitic alteration zones hold potential for disseminated and vein mineralization. Prominent rusty pyritic alteration zones north of Devils Elbow Mountain warrant more evaluation. The local areas of Stuhini Group felsic submarine volcanism are an appropriate setting for massive sulphide deposits. Epithermal, structurally controlled alteration and mineralization associated with Eocene felsic dikes and volcanism are potential targets, however, no mineralization has yet been identified.

ACKNOWLEDGMENTS

Thanks go to Mike Gunning who provided excellent support in the field for a second season and contributed a great deal to the project. Eric Maeckelburg provided cheerful and capable field assistance. Fossils were promptly identified by H.W. Tipper (GSC, Vancouver), T.P. Poulton and A.R. Sweet (ISPG, Calgary). The cooperation of I.A. Paterson and S.B. Noakes of Cominco Limited was greatly appreciated, as were the services of Northern Mountain Helicopters Inc., Trans North Air Ltd. and Tel Air Ltd. Special thanks are extended to the MacGregor family for their hospitality.

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NOTES

STRATIGRAPHY OF THE STIKINE ASSEMBLAGE, SCUD RIVER AREA, NORTHWEST BRITISH COLUMBIA (104G/5, 6)

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KEYWORDS: Stratigraphy, lithologies, Stikine assemblage, Scud River, Permian, limestone, fossils, chert nodules, jasper, mafic flows, sericitic tuffs.

INTRODUCTION

This is a preliminary report on the stratigraphy of the Late Paleozoic Stikine assemblage in the Scud River area of northwest British Columbia. Fieldwork was done in the summer of 1989 in conjunction with the Stikine Project (Brown and Greig, 1990, this volume). Relevant data collected last year for the Stikine Project (Brown and Gunning, 1989 a, b) is included. Additional fieldwork will be conducted next summer in conjunction with a graduate thesis.

The study area is near the confluence of the Scud and Stikine rivers in the rugged Coast Mountains (Figure 1-15-1). Strata examined occur in a northwest-trending belt exposed between the Scud River and Butterfly Lake. Fieldwork for the study was serviced from Telegraph Creek about 50 kilometres to the north.

The objective of this report is to provide detailed lithologic descriptions of the Stikine assemblage and to develop lithostratigraphic subdivisions in context with the structural setting of the sequence. It is based upon preliminary data from surface sections measured through Permian and "Permian or older" strata using hip chain, altimeter, Brunton compass and airphoto techniques.

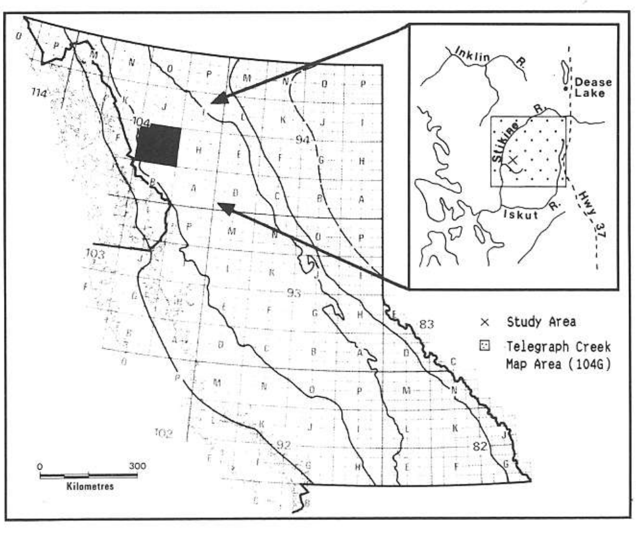

Figure 1-15-1. Location map for study area.

PREVIOUS WORK

The Stikine assemblage (Monger, 1977) refers to all Late Paleozoic rocks in northwest British Columbia which occur west of the Bowser basin and south of the Cache Creek Terrane. Rocks are Early to Middle Devonian to late Early Permian in age (Anderson, 1988). Thick successions of very fossiliferous Permian limestone characterize the assemblage and permit regional correlation between isolated and discontinuous exposures which occur along the western margin of the Stikine Terrane from Terrace to the Tulsequah River.

Regional mapping projects in the lower Stikine River region include the work of Dawson (1889), Kerr (1948) and Souther (1972). Several British Columbia Ministry of Energy, Mines and Petroleum Resources 1:50 000-scale mapping projects are in progress (Brown and Gunning, 1989, a,b; Brown and Greig, 1990, this volume; Logan and Koyanagi, 1989, a,b; Logan *et al.*, 1990, this volume). Detailed studies of the stratigraphy of the Stikine assemblage have been done by the Pan American Petroleum Company (Fitzgerald, 1960; Rigby, 1961), Monger (1970, 1977) and Holbek (1988). Paleontologic and paleostratigraphic studies have been completed by Pitcher (1960), Monger and Ross (1971), Mamet (1976), Rycerski (1985) and Stevens and Rycerski (1989). The work of Brown, Logan and their colleagues has greatly increased the paleontologic data base for the Stikine assemblage.

STRATIGRAPHY

Paleozoic strata in the Scud River area comprise four main subdivisions. From the bottom up, they are very deformed, pervasively foliated dacitic tuffs and phyllites (Unit A) overlain by lower greenschist, intermediate to mafic volcanic flows and lesser crystal tuffs (Unit B). Thinly bedded sericitic dacitic lithic tuffs and siliceous siltstones (Unit C) conformably overlie the metavolcanic rocks and are in turn unconformably overlain by a thick Permian succession of fossiliferous limestone and subordinate varicoloured siliceous sedimentary rocks (Units D, E, F, G).

The age of Units A, B and C is unknown and the base of the assemblage is not exposed. At the toe of the Scud Glacier, the assemblage is conformably overlain by pyroxene-bearing greywackes of probable Late Triassic age.

The nature of the contact between Permian and "Permian or older" units varies but all contacts between units within the Permian succession are conformable. Lateral continuity of the units is also variable. True stratigraphic thickness is often indeterminate due to structural disruptions, rapid facies changes and inaccessibility of exposures.

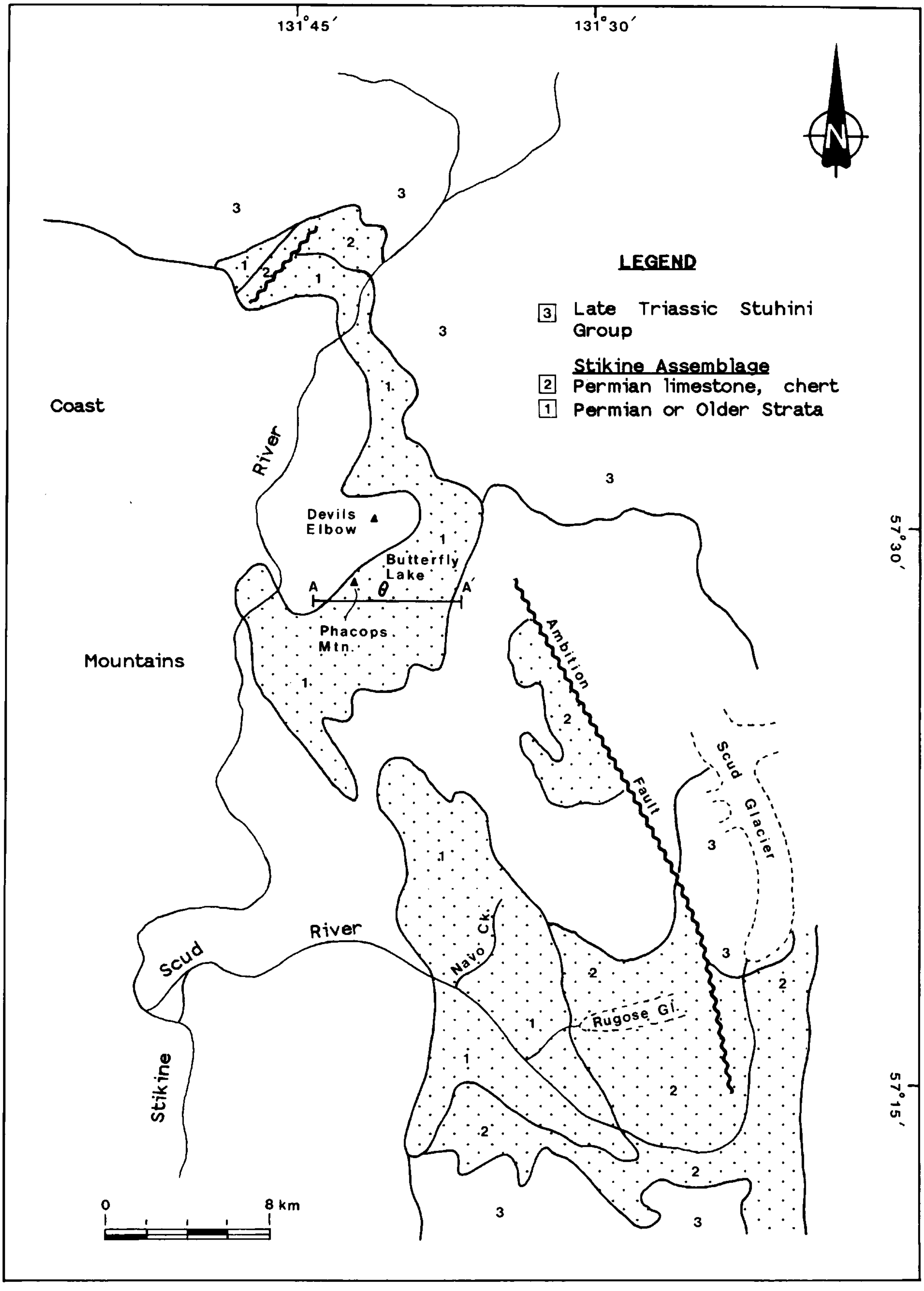

Figure 1-15-2. Simplified geology map showing the distribution of the Stikine assemblage in the Scud River area (modified from Brown and Gunning, 1989b).

Macrofauna and microfauna examined by E.W. Bamber of the Institute of Sedimentary and Petroleum Geology and M.J. Orchard of the Geological Survey of Canada in Vancouver respectively, indicate a probable Early Permian age (Artinskian) for the limestone succession although it could be as old as late Middle Pennsylvanian (Brown and Gunning, 1989b). There are no age constraints for strata which are structurally and probably stratigraphically below the Permian limestone.

Permian or Older Strata

Rocks below known Permian strata are well exposed between Navo Creek and Butterfly Lake. They are north to northwest-trending metavolcanic rocks (Units A, B) conformably overlain by fine-grained siliceous tuffs (Unit C).

There is no area with continous, accessible exposures of all three units. Very deformed dacitic tuffs and phyllites of Unit A were examined west of Butterfly Lake. Variably chloritic andesitic flows with subordinate felsic tuffs and volcanic conglomerate of Unit B were examined east of Butterfly Lake (Basement Unit, Volcanic Facies; Brown and Gunning, 1989a, b). Well-bedded siliceous tuffs of Unit C were examined just north of Navo Creek where the underlying strata of pelites, foliated greywackes, and limy siliceous siltstones (Basement Unit, Sedimentary Facies; Brown and Gunning, 1989b) indicate there is a facies change within Unit B. The transition from dominantly volcanic to dominantly sedimentary rocks in Unit B may be about 2 kilometres north of Navo Creek where chloritic andesite flows and possibly sills are intercalated with foliated greywacke and argillite.

"Permian or older" metavolcanic and metasedimentary strata are continuous both north and south of the study area. Pelitic rocks north of Devils Elbow includes fossiliferous Mississippian(?) limestone (Kerr, 1948) and may be correlative with Unit B.

UNIT A

This unit is well exposed along the south flank of Phacops Mountain west of Butterfly Lake (Figure 1-15-1, 1-15-2) and consists of dacitic to rhyolitic lithic tuffs with subordinate intermediate to mafic flows. Relict bedding is generally overprinted by a parallel sericitic foliation. These rocks are more intensely deformed and more metamorphosed than those of Units B or C and may be the oldest in the region.

Poorly preserved load casts at one locality indicate stratigraphy is right-way-up. Beds strike to the southeast and dip steeply to the southwest. This orientation is consistent and strongly discordant to other "Permian or older" strata (Units B, C). Crenulation fabrics are common throughout this unit; vergence is variable.

The upper contact is obscured by a small, heterogeneous potassium feldspar megacrystic stock that crops out around Butterfly Lake. It intrudes Units A and B and is crosscut by a major north-striking vertical fault. The base of the unit is not exposed; it is intruded by a large body of middle Jurassic granodiorite southwest of Phacops Mountain.

Pale green sericite imparts a pervasive foliation throughout the unit. The degree of foliaton ranges from sericitic bedding planes to the formation of phyllites and sericite schists. Sericite content increases westward from Butterfly Lake at stratigraphically deeper levels.

Where preserved, bedding is defined by alternating bands of light and dark grey, fine-grained to aphanitic dacitic tuff less than 20 centimetres thick. White to pale grey weathering surfaces common in dacitic tuff beds in Units B and C are not present in this unit. Cloudy white felsic fragments and relict crystals are deformed. The fragments, up to 15 centimetres long, make up 10 to 15 percent of the rock and are commonly flat with depth to length to width ratios of up to 5:20:1.

A 3 to 4-metre bed of laminated siltstones and thinly bedded dacitic to rhyolitic lapilli tuffs is exposed about 1100 metres west of Butterfly Lake. The tuffs are deep green on fresh surfaces with well-developed pale green sericite on bedding planes. Bedding is defined by 15 to 25-centimetre zones made up of 60 to 80 per cent spherical to elliptical siliceous grains set in a milky white groundmass. Spherical grains are 1 to 3 millimetres in diameter; elliptical grains can be up to 15 millimetres long and are generally oriented parallel to bedding. The grains are cloudy grey and amorphous with a narrow (less than 2 millimetres) light grey rim. The rims may have formed during growth of the grains or they may be an alteration halo. Load casts and possible grading in the lapilli tuffs indicate that strata are right-way-up.

Variably chloritic, massive to poorly foliated dark green andesitic flows and possibly sills are intercalated throughout the dacitic tuffs but comprise much less than 1 per cent of the unit.

UNIT B

Unit B is well exposed east of Butterfly Lake and consists of variably chloritic intermediate to mafic volcanic flows with subordinate siliceous tuffs and volcanic conglomerate (Figures 1-15-2, 1-15-3). True stratigraphic thickness is uncertain because of structural shortening but the unit is less than 2500 metres thick.

The lower contact with Unit A is obscured by the Butterfly Lake alkalic stock. The upper contact with overlying siliceous tuffs of Unit C is conformable and gradational. The same contact relationship between Unit C and the sedimentary facies of this unit occurs just north of Navo Creek.

Attitudes of bedding and parallel chloritic and sericitic foliations are variable. Strikes vary within 20° of north. Dips vary from east to west and are generally less than 50° in the upper half of the sequence, and steep to vertical in the lower half. Correlative strata exposed both to the north and south of Butterfly Lake have a more consistent northwest strike and moderate to steep northeast dip. Load casts, flame structures, and graded bedding indicate that stratigraphy is right-way-up.

The unit consists mainly of massive to faintly foliated, hornblende and plagioclase-porphyritic to equigranular, fine-grained to aphanitic andesite to basalt flows. The flows are dark green on weathered and fresh surfaces. Chlorite content is variable but fracture surfaces are almost always chloritic. Phenocrysts are less than 15 per cent. Plagioclase is in cloudy white subhedral crystals less than 3 millimetres across. Euhedral, lath-shaped hornblende phenocrysts are up to 4 millimetres long. Traces of fine-grained biotite are also present.

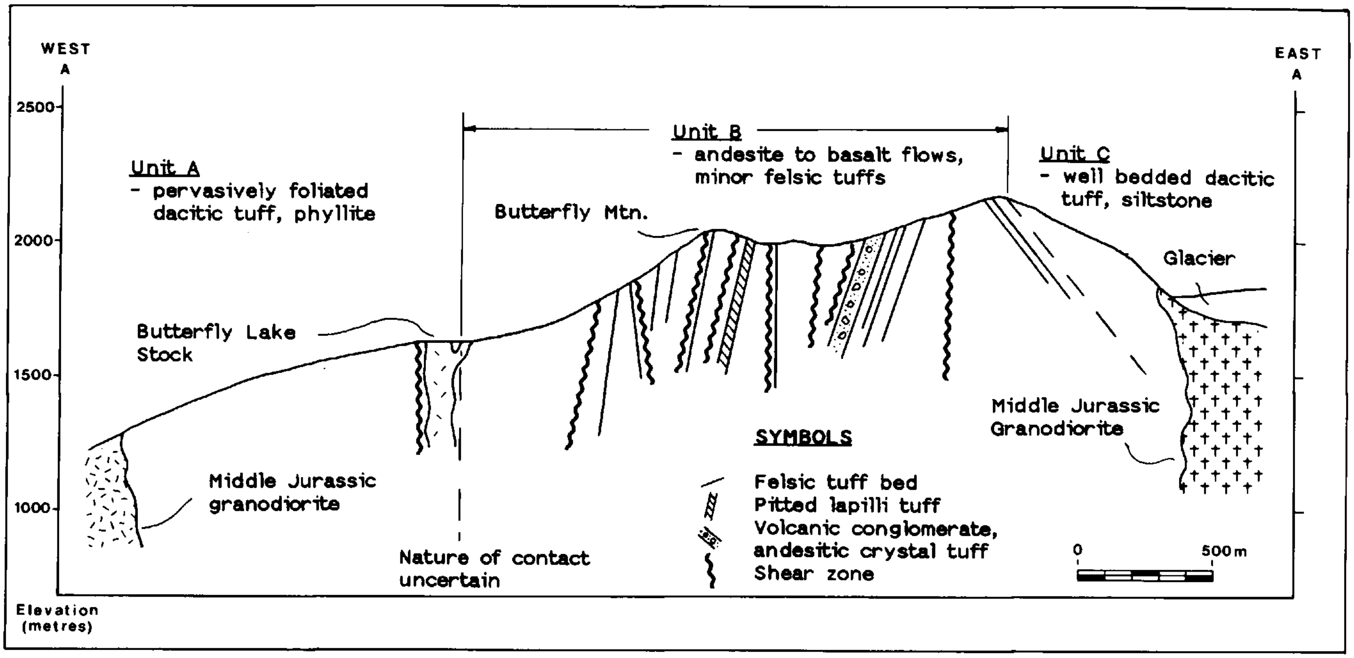

Figure 1-15-3. Schematic cross-section of "Permian or older" strata at Butterfly Lake.

Felsic crystal tuffs and volcanic conglomerates comprise about 10 per cent of the unit. There are at least 13 beds of fine-grained to aphanitic, pale grey to green, white-weathering siliceous (dacitic) tuffs in the unit. The tuffs are thinly bedded, well laminated, variably sericitic on bedding planes, and contain less than 5 per cent flat felsic fragments. The tuff beds are generally less than 2 metres thick and appear to accommodate much of the strain in the unit.

Approximately 2 kilometres east of Butterfly Lake, there is a zone 210 metres thick of interbedded andesitic crystal-lithic tuff and volcanic conglomerate within massive andesitic flows. The upper 70 metres consists of brown-weathering, pitted, limy lapilli tuff with 10 to 20 per cent subangular, fine-grained siliceous tuff fragments. Pitted textures are from weathered-out lapilli less than 5 centimetres across. The tuffs are interbedded with brown-weathering, matrix-supported monomictic volcanic conglomerate of similar composition. Clasts are well rounded, up to 40 centimetres across, and make up less than 40 per cent of the rock. The cobbles have not been deformed.

The lower 140 metres of this zone consists of white-weathering, andesitic feldspar-crystal tuff and polymictic volcanic conglomerate. The conglomerate is clast supported, made up of 70 to 90 per cent subangular cobbles. The matrix is aphanitic, dark grey and andesitic. Clasts are mainly light grey, fine-grained siliceous tuffs and lesser equigranular salt-and-pepper granodiorite and grey recrystallized limestone. Clasts average 5 to 10 centimetres across but can be up to 70 centimetres across. There are no known Paleozoic intrusions in the area.

This unit is tightly folded and cut by vertical shear zones. Fold axes plunge gently north with steep axial planes. Two fold cores were identified and eight more inferred from measurements of bedding and cleavage-bedding intersections. Nine zones of pervasive chlorite and sericite foliation are mapped and may be faults. The zones are generally less than 5 metres wide with gradational contacts; they commonly occur along siliceous tuff interbeds within massive andesitic flows where the tuffs grade into phyllites and the flows into chlorite schists. At least four of the zones appear to represent shearing along fold cores. Well-developed crenulation fabrics were observed adjacent to three of the structures with vergence variable from north to northwest to southwest.

UNIT C

This unit consists of fine-grained dacitic crystal-lithic tuffs (Siliceous Unit; Brown and Gunning, 1989a, b). It is well exposed north of Navo Creek (Figure 1-15-2) where it is approximately 1350 metres thick, although stratigraphic thickness varies greatly along strike. The unit is continuous through the study area, but like the Permian limestone, is absent north of Butterfly Lake where Late Triassic volcanic strata overlie "Permian or older" pelitic rocks. Continuity south of the Scud River is uncertain.

The section examined begins approximately 120 metres stratigraphically below the top of the unit. The tuffs are lithologically and texturally homogeneous and conformably overlie dark grey to black, well-foliated pelites and metagreywacke. The tuffs are well bedded, strike northwest and dip moderately to the northeast. Beds are well laminated, wavy, and generally less than 20 centimetres thick; pale green sericite is almost always present on bedding planes. Load casts and flame structures occur throughout and indicate strata are right-way-up.

The tuffs are white to dark grey or green on weathered surfaces. The lighter coloured tuff is generally very well laminated. It has less than 1 per cent poorly defined cloudy white lithic fragments less than 3 millimetres across, hosted in a light green to grey, amorphous, siliceous groundmass. Less than 1 per cent disseminated pyrite euhedra are present. The dark grey to green, more andesitic tuffs contain trace mafic crystals and 2 to 3 per cent cloudy white feldspar crystals and lithic fragments up to 3 millimetres across.

Black, well-laminated, amorphous siliceous siltstone beds occur throughout the unit and contain from 1 to 2 per cent disseminated pyrite. Thin, discontinuous layers of micrite and argillite comprise less than 1 per cent of the upper third of the unit. The beds are lensoidal, less than 20 centimetres thick, usually less than 50 metres long, and generally cleaved. The micrite is dark to light grey wlth no macrofossils.

PERMIAN STRATA

A thick succession of late Early Permian limestone and subordinate chert is well exposed between Rugose Glacier and the toe of the Scud Glacier (Figure 1-5-2). Four lithologically and faunally distinct units make up the sequence (Units D, E, F, G; Figure 1-15-4) and are described below from the base up. Limestone descriptions are according to Folk (1962).

The sequence is well bedded with north to northwest strikes and moderate to shallow east dips. Folds are open and plunge gently to the south and southeast. The sequence is over 2 kilometres thick although structural disruptions within the suceession, in combination with rapid lateral facies changes and inaccessibility of some exposures, make determination of true stratigraphic thicknesses difficult. The contact between Permian and "Permian or older" strata varies from conformable (disconformity ?) to a 30° angular unconformity. It is structurally disrupted between Rugose Glacier and Navo Creek. The upper contact with Late Trlassic volcanic and sedimentary rocks is conformable.

Ten kilometres to the north of Rugose Glacier, Late Triassic volcanic rocks overlie "Permian or older" metasedimentary rocks; the entire Permian limestone succession is absent. The limestone is more continuous to the south, exposed over 20 kilometres away in the Sphaler Creek area (Logan and Koyanagi, 1989a, b), and 50 kilometres to the south in the Iskut River district (Anderson, 1988).

UNIT D

This is a discontinuous, rusty weathering argillite that is best exposed at the toe of Rugose Glacier where it is approximately 85 metres thick. Thickness changes dramatically along strike and the argillite pinches out 2 kilometres south of the glacier. Continuity to the north is uncertain because of inaccessibility of exposures. The upper contact with Unit C is gradational.

The lower half of the unit is massive to well-laminated rusty weathering argillite with 2 to 10 per cent fine-grained pyrite and pyrrhotite occuring as disseminations and layer-parallel stringers. The upper strata have less pyrite and are more limy, grading into a black argillaceous micrite with rare solitary rugose corals up to 6 centimetres across.

UNIT E

This is the thickest unit in the Permian sequence. It is well exposed along the south side of Rugose Glacier where thinly bedded bioclastic limestone is over 1500 metres thick with no lithologic or paleontologic evidence for stratigraphic repetition. The same section of limestone exposed on the north side of the glacier is complexly folded with at least two layer-parallel structural disruptions. The unit appears to be folded inhomogeneously as only two small southwest-verging z-folds occur in the section examined.

The limestone is very fossiliferous and much of the paleontologic data for the Stikine assemblage comes from strata correlative with this unit. Fossil collections have been made from the base to the top of the section exposed at Rugose Glacier and there is no discernible age transgression.

The lower third of the unit (550 metres) is interbedded light and dark grey bioclastic micrite. The lower 30 metres of the unit is argillaceous and contains ubiquitous solitary rugose corals up to 5 centimetres across. Beds are generally planar and from 10 to 70 centimetres thick. The light grey biomicrite consists of up to 50 per cent white crinoid fragments and lesser bryozoa and other fossil debris. The fossil pieces are generally less than 7 millimetres across and are in a light grey micritic matrix with little sparry cement. The dark grey biomicrite is very bioclastic with up to 80 per cent branching and fenestrate bryozoa and well-preserved fusulinids. The fossils are both whole and fragmented and generally lie parallel to bedding in a dark grey to brown, fine-grained micritic matrix. Fenestrate bryozoa mats may be up to 15 centimetres long and 10 centimetres wide.

Approximately 200 metres from the base of the unit, there is a 3-metre bed of light grey biomicrite that contains abundant syringipora corals, large brachiopods, gastropods, branching bryozoa and fusulinids. About 350 metres from the base of Unit E, there is a bed of dark brown biomiclutite over 30 metres thick that consists of over 70 per cent delicately preserved branching and fenestrate bryozoa, fusulinids, gastropods and ubiquitous large solitary horn corals up to 40 centimetres long that are commonly well exposed on weathered bedding planes.

The middle 700 metres of the unit is thinly bedded light grey biomicarenite and varicoloured amorphous chert. The chert is pale yellow and rarely black. It occurs as nodules and planar to wavy beds from 10 to 80 centimetres thick (Plate 1-15-1). The upper 100 metres of this section has less than 5 per cent chert beds and contains abundant solitary rugose corals. One 30-centimetre bed has 60 to 80 per cent small solitary rugose corals which generally lie parallel to bedding. There are also thin (less than 20 centimentres) beds of dark grey biomicrite composed almost entirely of layer-parallel, black, recrystallized, round to elliptical fusulinids and oblong bryozoa. The fusulinids are generally less than 1 centimetre long and the bryozoa can be up to 3 centimetres long.

The upper third of Unit E (510 metres) is thickly bedded grey to buff biomicrite and biosparite. Beds are 30 centimetres to greater than 1 metre thick and there are no chert beds.

About 400 metres from the top of the unit there is 30 to 40 metres of very bioclastic dark grey biomicrite that contains

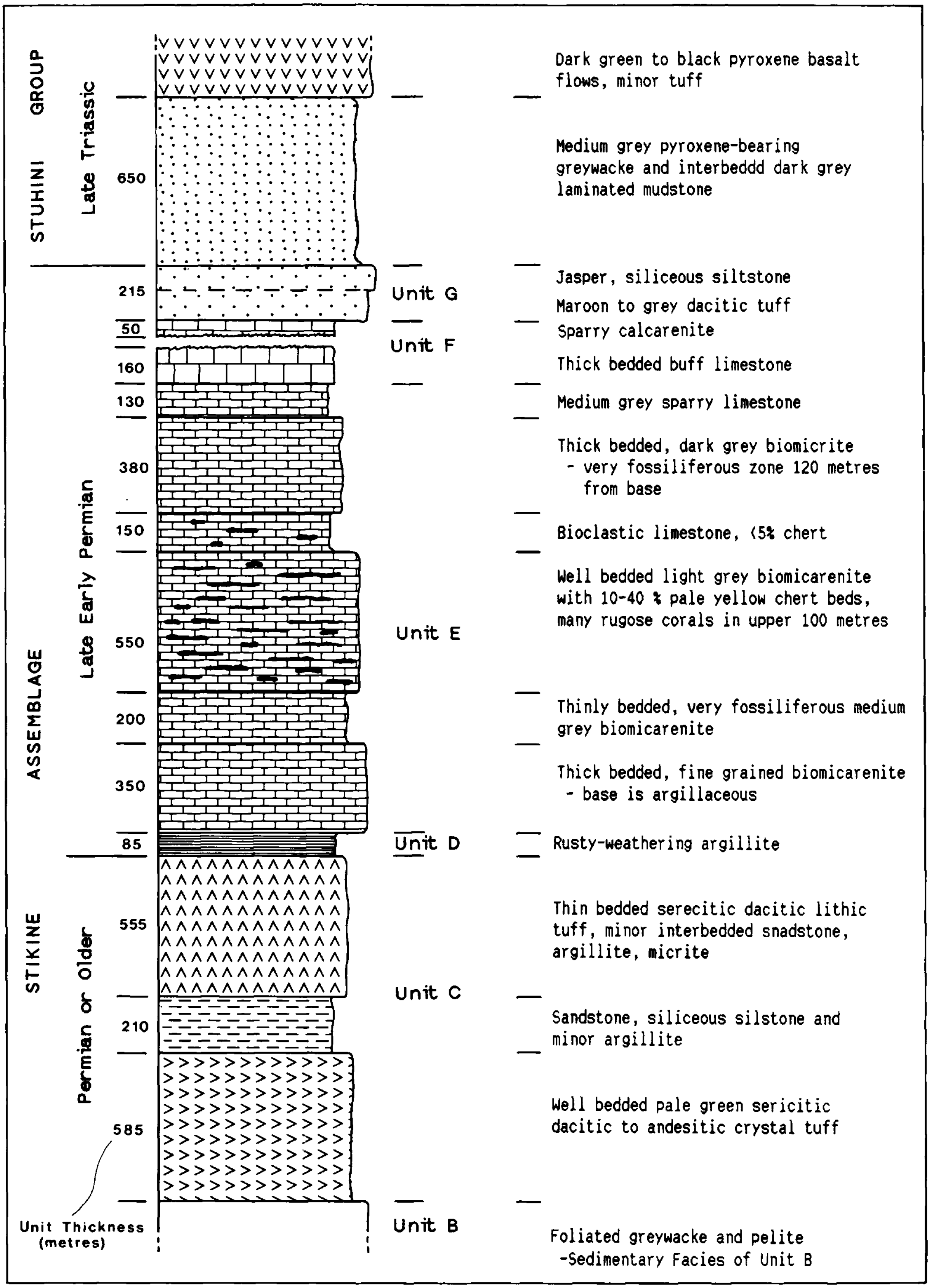

Figure 1-15-4. Schematic stratigraphic column of the Stikine assemblage in the Navo Creek, Rugose Glacier and Scud Glacier areas.

Plate 1-15-1. Bedded to nodular pale yellow chert within light grey biomicarenite in the mid-portion if Unit E.

abundant colonial corals and branching bryozoa up to 25 centimetres long. There are large solitary corals (*heritschioides* sp.) over 15 centimetres across and lesser fenestrate bryozoa, small rhynchonellid brachiopods, fusulinids and gastropods. Beds of dark grey biomicrite with fusulinids and branching bryozoa occur throughout the upper third of Unit E.

UNIT F

Unit F forms most of the peaks between Rugose and Scud glaciers. Stratigraphic thickness is uncertain due to structural disruption along the Ambition fault but the unit has a minimum thickness of 180 metres.

The lower part is mostly inaccessible and consists of massively bedded, white to buff-weathering sparry calcarenite with few macrofossils and no chert beds. The upper part is exposed on the east side of the Scud Glacier valley, east of the Ambition fault. It consists of approximately 50 metres of massively bedded buff to white-weathering sparry calcarenite with thin discontinous beds of argillite and maroon crystal-lithic lapilli tuff. The tuffs contain 7 to 10 per cent quartz and feldspar crystals, and less than 7 per cent black, subangular to flat lithic fragments less than 2 centimetres long. Some of the tuff beds appear to be structurally disrupted.

The upper 15 metres of Unit F is thinly bedded biocalcarenite and sparry calcarenite. The beds are 15 to 40 centimetres thick and load casts indicate that stratigraphy is right-way-up. The biocalcarenite beds have 60 to 70 per cent recyrstallized, white, elliptical to oblong fusulinids up to 3 centimetres long that are parallel to bedding. There are also small, poorly preserved solitary corals less than 4 centimetres across. Green to maroon crystal tuff and maroon tuffaceous micrite are interbedded with the biocalcarenite.

Conodonts from the fusulinid-rich limestone are Permian but have significantly different morphologies from those in the thick sequence of thinly bedded bioclastic micrite of Unit E (M.J. Orchard, personal communication, 1989).

UNIT G

Varicoloured siliceous sedimentary and volcanic rocks of Unit G comprise about 210 metres of strata well exposed east of the toe of the Scud Glacier. The lower contact with the tuffaceous limestone of Unit F is gradational. The unit is conformably overlain by pyroxene-bearing greywackes correlative with the Late Triassic Stuhini Group. Beds are right-way-up and strike northwest with moderate east dips. These rocks were previously mapped as Middle Triassic.

Fine-grained siliceous tuffs comprise the lower 130 metres of the unit. The tuffs are maroon to dull grey, fine grained to aphanitic, and massive to faintly foliated, with 5 to 15 per cent lithic fragments less than 5 millimetres across. Mafic crystals less than 3 millimetres across occur in the maroon tuffs. Exposures of the tuffs are generally rounded with poorly defined bedding. Orientation of the tuffaceous strata is commonly defined by rare interbeds of laminated, dark grey, very poorly sorted, coarse-grained volcanic greywacke.

Overlying the tuffaceous strata are 85 metres of fine-grained siliceous sedimentary rocks. The lower 15 metres consist of thinly bedded, wavy, pale green amorphous siliceous siltstone. Above the siltstone is a very distinctive horizon of bright red, thinly bedded radiolaria-bearing jasper. Beds are wavy, from 5 to 10 centimetres thick, and have dark grey mudstone lamellae. Thin interbeds of amorphous, green siliceous siltstone and black argillite comprise less than 5 per cent of the jasper sequence.

The jasper is overlain by rusty weathering, thin-bedded, pyritic green siliceous siltstone and a distinctive horizon of black ribbon chert. Permian conodonts were retrieved from the ribbon chert (M.J. Orchard, personal communication, 1989) which consists of 20 to 40-centimetre, black, amorphous chert beds separated by beds of graphitic argillite less than 10 centimetres thick. The upper 15 metres of Unit G is amorphous, thin-bedded green siliceous siltstone with less than 1 per cent disseminated pyrite. Well-laminated beds are wavy and from 1 to 30 centimetres thick.

LATE TRIASSIC STRATA

At the toe of the Scud Glacier, thin-bedded, well-laminated, dark to light grey mudstone, siltstone and greywacke conformably overlie the Permian siliceous sedimentary rocks. The beds contain many sedimentary structures including load casts, mudstone rip-ups, flame structures, graded bedding, and cut-and-fill channels which indicate stratigraphy is right-way-up. Narrow 1 to 2-metre zones of well-laminated, dark grey to brown mudstone and siltstone occur within much thicker zones of medium to light grey greywacke.

The thickness of this unit is about 650 metres; true stratigraphic thickness is uncertain because of cover by the Scud Glacier. On the west side of the glacier, the laminated greywacke grades upward into coarse-grained andesitic tuffs and polymictic volcanic breccia overlain by a tremendous thickness of dark green to black pyroxene basalt and minor intercalated maroon volcaniclastics. These rocks are correlative with the Upper Triassic Stuhini Group.

SUMMARY

Paleozoic strata in the Scud River area consist of a thick succession of late Lower Permian limestone and subordinate chert which unconformably overlies more deformed and more metamorphosed felsic to mafic volcanic rocks and metasedimentary strata. Exposures at the toe of the Scud Glacier suggest that the Upper Triassic Stuhini Group conformably overlies Permian strata of the Stikine assemblage. Developing a better understanding of the Permo-Triassic boundary, and of the nature of the pre-Permian unconformity, will be important aspects of this study.

There are seven distinct lithostratigraphic subdivisions which comprise over 4 kilometres of strata in the Scud River area. True stratigraphic thickness is difficult to determine, as is the lateral continuity of the various units; many of the units will not be useful for regional correlations. Continued fieldwork, in conjunction with a graduate thesis by the author, will further develop detailed lithologic descriptions and stratigraphic analyses. These data will be examined in context with the structural setting of the area to produce a better understanding of the paleoenvironment and tectonic evolution of the Stikine assemblage.

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NOTES

GEOLOGY AND MINERALIZATION, TATSAMENIE LAKE DISTRICT, NORTHWESTERN BRITISH COLUMBIA (104/K)

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KEYWORDS: Economic geology, gold, structural geology, stratigraphy, Stikine Terrane, ophiolite, mid-Triassic unconformity.

INTRODUCTION

This report presents the preliminary results of a second year of study of the geology and mineral occurrences of the Tatsamenie–Muddy Lake district, northwestern British Columbia. The project area is located approximately 140 kilometres west of Dease Lake (Figure 1-16-1). One producing gold mine, the Golden Bear, and several other base and precious metal occurrences are hosted dominantly by the Upper Paleozoic and Mesozoic rocks of this region.

During the 1989 field season, 1:10 000-scale geological mapping was completed along the north side of Tatsamenie Lake and south towards the Golden Bear deposit. Mapping focused on the three major structural features of the area: the Tatsamenie antiform, the Tatsamenie alteration zone, and the West Wall fault and associated ultramafic rocks south of Tatsamenie Lake.

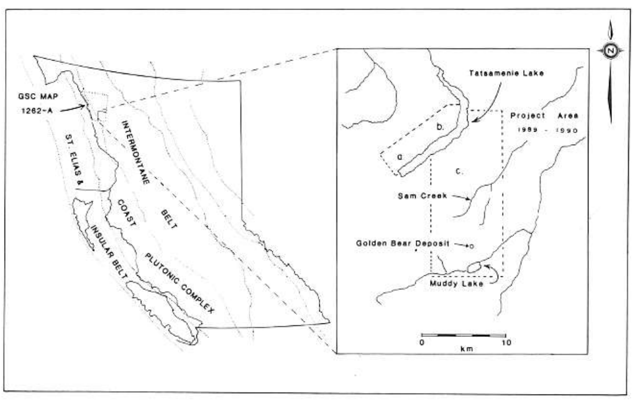

Figure 1-16-1. Location of the Tatsamenie Lake map area within the tectonic framework of the Canadian Cordillera. The location of major structural features, the Tatsamenie antiform (a), The Tatsamenie alteration zone (b), and West Wall fault (c), are also shown.

PREVIOUS WORK

The geology of the Tatsamenie Lake area was examined in a field program conducted by Souther between 1958 and 1960 (Souther 1971). His report and accompanying map established the general stratigraphic and structural relationships of the Upper Paleozoic and Mesozoic rocks of this area. The Paleozoic section exposed along the shores of Tatsamenie Lake was used by Monger (1970) in his development of an island arc tectonic model for Stikine assemblage rocks.

Selected mineral occurrences in this area have been briefly described by Schroeter (1986). The area was the focus of a broad precious metal reconnaissance program conducted by Chevron Minerals Ltd. in the early 1980s. The results of its field programs are documented in a series of assessment reports (Brown and Walton, 1983; Brauset, 1984; Shannon, 1982; Shaw 1984; and others). Parts of this region, principally the main Tatsamenie alteration zone, were explored for their base metal potential, prior to 1980. In a B.Sc. thesis, Hewgill (1980) described the alteration and dated a large albite dike exposed above treeline south of Tatsamenie Lake.

The geology and structural features of rocks south of Sam Creek, and the Golden Bear deposit, have been described by Oliver and Hodgson (1989).

REGIONAL GEOLOGY

Much of the project area is underlain by an 800-metre-thick sequence of thick-bedded limestones and lesser interbedded cherts, shales and sandstones of Permian age (Souther, 1971). The Permian succession is conformably overlain by a sequence of Lower Triassic mafic pyroclastics and lesser flows, several hundreds of metres thick (Figure 1-16-2).

Unconformably lying on Lower Triassic rocks are weakly metamorphosed and deformed Middle Triassic volcanic rocks of the Stuhuni and King Salmon Groups; a pre-Middle Triassic unconformity is found throughout the North American Cordillera (Read and Okulitch, 1977, Siberling, 1973).

In the Stikine Terrane, late Triassic volcanic rocks are dominantly fragmental and commonly alkalic. Based on bulk rock chemistry, both extensional, back-arc, and fore-arc tectonic settings have been suggested for these rocks (Mortimer, 1986; Souther, 1977). It is probable that both fore-arc and back-arc lithologies may have evolved to form the larger composite that is the Stikine Terrane.

Within the project area, supracrustal rocks are flanked, to west and south, by intrusions which form the core of the Stikine arc. These intrusions trend northeast across the north-northwest grain of the Cordillera. They are typically Triassic to early Jurassic in age and are believed to be coeval with late Triassic volcanic rocks. Late Triassic felsic plutonic rocks have an extremely limited distribution in the Cordillera, occurring only in the northwestern Stikine arc (Anderson, 1984; Armstrong, 1988).

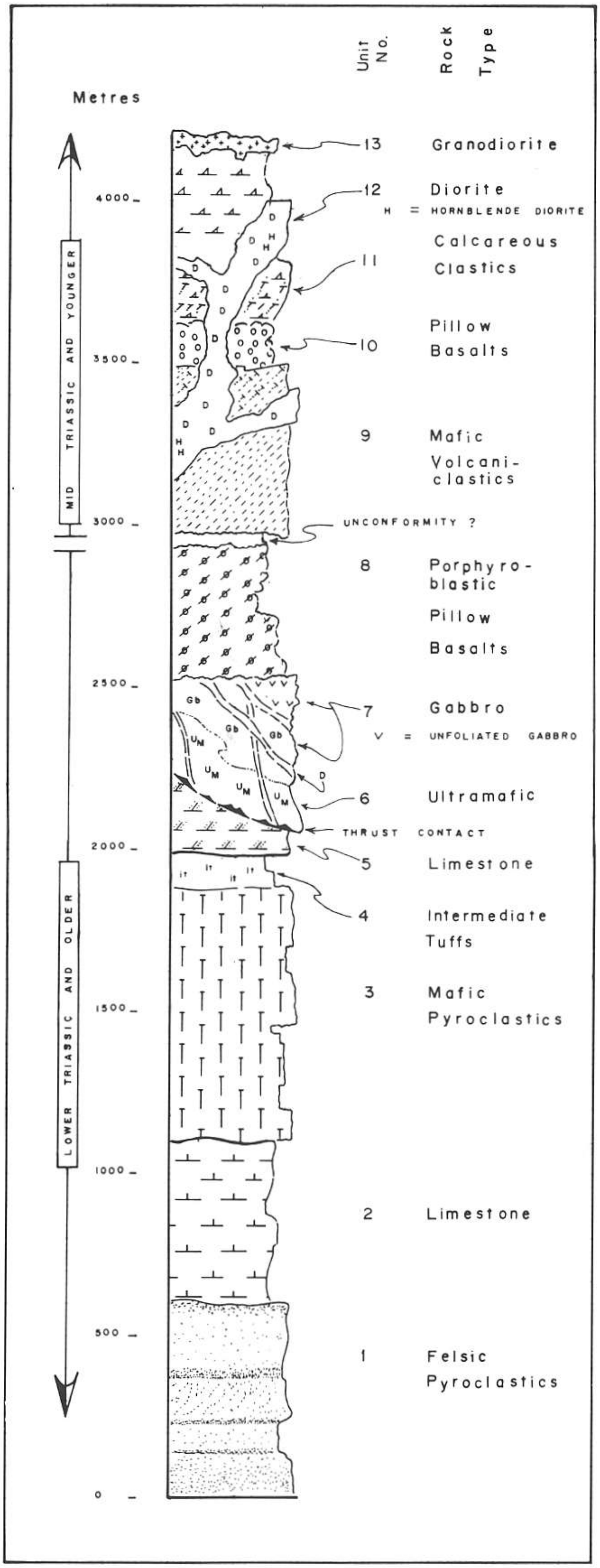

Figure 1-16-2. Generalized stratigraphic column, Tatsamenie Lake area. Additional subdivisions and modifiers of these major rock units are detailed and expanded in the text.

Three major deformations have influenced the form and distribution of supracrustal rocks in this area. The youngest structures are Eocene extensional faults which cut all rock types and locally are intruded by Tertiary diabase and felsic dikes. Volumous basaltic volcanism associated with Eocene extension, produced lavas of the Level Mountain Group; the youngest rocks in the project area. The late extension in the region was related to a transtensional tectonic environment caused by a change in plate convergence direction at 54 Ma (Engerbretson *et al.*, 1985).

The second deformation was related to a mid-Jurassic accretionary event. The deformation involved the formation of southwest-verging thrust faults, for example the King Salmon fault, coupled with the development of broad, open fold structures, and weak penetrative deformation. Many of the older southwest-verging tight to overturned folds in the penetratively deformed rocks below the mid-Triassic unconformity were refolded in the second deformation. The marked change in deformational style between post and pre-Middle Triassic rocks has been used as a field criterion for locating the Middle Triassic unconformity in the area (Souther, 1971; Kerr 1948).

SUPRACRUSTAL ROCKS

PERMIAN AND OLDER

UNIT 1: FELSIC TUFFS AND LESSER FELSIC FLOWS

Pale cream to pink-buff-weathering, fine-grained felsic tuffs and lesser flows occupy the core of the Tatsamenie antiform. This unit is intensely foliated and usually displays well-developed bedding-cleavage intersection lineations. The thickness of the combined pyroclastic and flow sequence may exceed 500 metres.

In the flows, primary feldspar phenocrysts occur in a matrix composed of greater than 40 per cent potassic feldspar. In tuffs, well-developed compositional layering is defined by variations in the ratio and size of quartz and feldspar. Layers average 0.5 to 2.5 centimetres in width. Unabraded primary zircon phenocrysts are abundant in thin section and establish the primary pyroclastic origin of this unit.

UNIT 2: LIMESTONE AND DOLOMITIC LIMESTONE

Massive to weakly bedded pale cream to dark grey limestone conformably overlies the felsic tuff unit in the core of the Tatsamenie antiform. This rock is moderately to strongly recrystallized and weakly foliated with the foliation being defined by the alignment of muscovite porphyroblasts. Quartz grains, some of detrital origin, average 15 to 20 per cent of the rock volume. Quartz content appears to increase towards the top of the section, as does the abundance of thin argillite beds. Stromatolitic mounds are developed at one locale near the upper contact of the limestone with the overlying tuffaceous units.

The carbonate sequence is approximately 450 to 500 metres in true thickness. Midway through the section, a mafic tuff bed, 20 to 40 metres thick, forms a useful internal marker horizon.

PERMIAN TO LOWER TRIASSIC

UNIT 3: BEDDED MAFIC TUFFS

A 600 to 650-metre sequence of thin bedded mafic tuffs, with minor calcareous interbeds, conformably overlies Unit 2. These rocks are characterized by 30 to 40 per cent porphyroblasts of actinolite and lesser hornblende, and may show well-developed compositional layers of actinolite-calcite-quartz and feldspar, 2 centimetres or less in thickness. They may also carry up to 15 per cent disseminated pyrite and magnetite. The rocks weather buff to grey-green, and locally are friable due to a well-developed foliation.

Tuffaceous beds range in thickness from 2 to less than 20 centimetres thickness. Plagioclase crystal tuff horizons can be recognized at the outcrop scale, but appear to have little lateral continuity and do not form mappable units. Flows have not been recognized in this unit.

UNIT 4: INTERMEDIATE TUFFS

This unit has only been recognized above 2000 metres elevation south of Tatsamenie Lake. It occurs as lenticular beds with little lateral continuity. The unit is well bedded with discrete compositional layering visible on all scales, and is blue to grey-green weathering. Highly strained quartz and potassic feldspar fragments are usually less than 1 centimetre on the long axis and may show asymmetric quartz and feldspar pressure shadows. Clasts composed of quartz and feldspar are strongly recrystallized and often deformed into isoclinal microfolds. Porphyroblastic biotite, and lesser chlorite, are present in the matrix and replace primary mafic minerals. The rock contains 15 per cent magnetite and pyrite, which is commonly rimmed by porphyroblastic biotite.

UNIT 5: LIMESTONES, INTERBEDDED ARGILLITES AND MINOR CALCAREOUS TUFFS

This unit consists of thin interbeds of chemical and clastic sedimentary rocks, and directly overlies either Unit 3 or Unit 4 (Figure 1-16-2). It does not exceed 150 metres in true thickness. A clean white limestone unit, 50 to 60 metres thick, forms the base of the section and is in turn overlain by a mixed sequence of argillites, minor siliceous phyllites, and extremely calcareous mafic tuffs. Calcareous mafic tuff members are very thin bedded, highly crenulated and disharmonically folded. The entire rock package appears to have reacted to deformation in a ductile fashion.

UNIT 8: PORPHYROBLASTIC PILLOW BASALTS

Diagnostic field criterion for this flow sequence are abundant, prominent, medium-grained actinolite porphyroblasts and highly strained chloritic pillow remnants. Morphologically the pillows are consistently elongate and of moderate to small size, seldom exceeding 75 centimetres along the long axis. Radial fracture patterns are present but strongly rotated. Estimates of thickness are difficult to make due to the high strain, but the unit appears to be approximately 350 metres in true thickness. Massive flows, lacking pillows, may form about 50 per cent of the section.

MIDDLE TRIASSIC AND YOUNGER

UNIT 9: MAFIC VOLCANICLASTICS

The map unit consists of 10 to 15 per cent fine-grained black clastic sedimentary rocks, 20 per cent carbonate-rich sedimentary rocks, 30 per cent mafic pyroclastics and flows, and 20 per cent volcanic wacke. This sequence is exposed along the north shore of Tatsamenie Lake. In this area, rocks are extensively hydrothermally altered and the original protolith is usually difficult to determine. The rocks are not strongly porphyroblastic, actinolite is absent and penetrative deformation is weak. Soft-sediment deformational features and fault-related drag-folds are the principle structures identified.

UNIT 10: PILLOW BASALTS

Unlike Unit 8, flows in this 100 to 150-metre-thick unit are moderately altered and unfoliated. Pillows have about 3 per cent vesicles by volume. The vesicles tend to be relatively large (2 to 4 millimetres) and are commonly filled with epidote with lesser analcime. The rock matrix is composed of 60 per cent fine-grained, unaligned plagioclase microliths, in a carbonatized pyroxene and lesser olivine matrix. The rock mineralogy is consistent with a basaltic rather than andesitic origin. The mineralogy also indicates that these rocks have only experienced very low-grade metamorphism, zeolite facies or burial metamorphism. The unit has the internal stratigraphy characteristic of subaqueous basaltic flows. The stratigraphic column consists of massive flows at the base, overlain by pillow flows and topped by well-defined pillow breccias.

UNIT 11: CALCAREOUS CLASTIC ROCKS, MINOR LITHIC WACKES

This unit forms the upper parts of the steep cliffs in the Tatsamenie alteration zone. The unit as a whole is red-buff weathering and is locally strongly altered. It consists of varying proportions of volcanic clasts, quartz grains and carbonate-rich mud. Angular, poorly sorted lithic fragments weather in relief against the buff-orange carbonate-rich matrix. Shaly limestone interbeds are present locally and provide excellent markers. Soft-sediment deformational features and large-scale slump structures are common. The unit may be slightly more tuffaceous at the base, grading upwards into a more clastic rock.

INTRUSIVE ROCK UNITS

UNIT 6: ULTRAMAFIC ROCKS

Ultramafic intrusive rocks are exposed at two localities south of Tatsamenie Lake. At both locations, they appear to be spatially associated with, and likely bounded by, large faults. In their unaltered form, the intrusions are coarsely phenocrystic with hypidiomorphic olivine and lesser pyroxene cut by small serpentine veinlets. Magnetite and other oxide phases average 10 per cent of the rocks by volume. Within 50 metres of large faults, the rocks become increasing serpentinized and carbonatized with steatite visible in both hand specimen and in thin section. Closer to the main faults significant grain-size reduction develops, producing well-

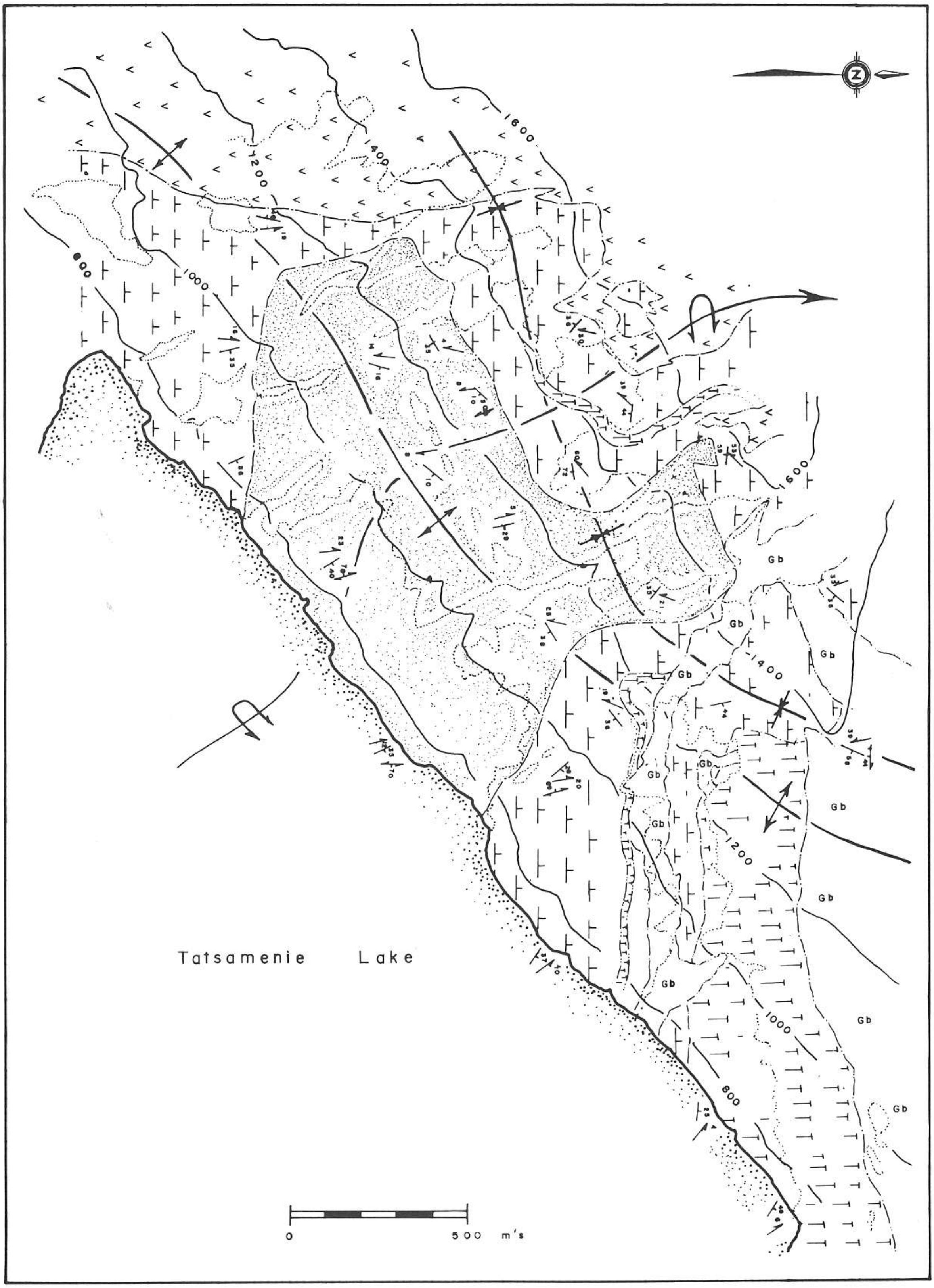

Figure 1-16-3. Geological plan of the Tatsamenie antiform. Lithologic patterns correspond to those of the stratigraphic column, Figure 1-16-2.

defined deformation bands. No contact aureoles or compositional zonation of the ultramafic rocks were mappable at these locations.

UNIT 7: GABBRO

Gabbroic intrusions are abundant on both the north and south sides of Tatsamenie Lake, and appear to have been emplaced in at least two intrusive events. The earliest gabbros are moderately to strongly foliated and partially carbonatized. Contact relationships between these intrusions and the overlying mafic flows and pyroclastic rocks suggest that the gabbros may in part be subvolcanic and coeval with the mafic volcanic rocks of Units 3 and 4 (Figure 1-16-2). Younger gabbros are unfoliated and essentially unaltered plagioclase-pyroxene-hornblende phenocrystic rocks which intrude all lower Triassic and Permian strata.

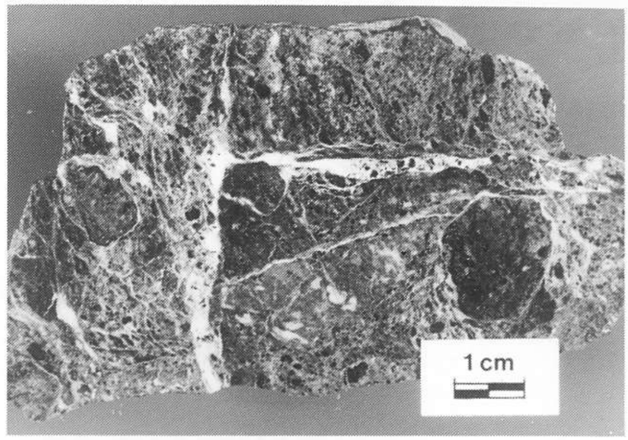

Plate 1-16-1. Example of an intrusion breccia from the Tatsamenie alteration zone. Veins are dominantly carbonate and the lighter flecks within fragments are green mica.

UNIT 12: DIORITE

Rocks of dioritic composition predominate among the intrusions in the area. They range from coarse-grained potassic hornblende diorites, to strongly plagioclase-porphyritic diorites (Map Unit PD), to well-foliated and weakly compositionally layered dioritic gneisses. Hornblende diorites and dioritic gneisses may have a weak igneous fabric, defined by aligned hornblende, but it seldom coincides with the regional penetrative foliation in the supracrustal rocks. Diorites form the core and flank the main alteration zone at Tatsamenie Lake. In this zone, they are extensively sheared and chloritized (Map Unit Dc). The margins of the intrusions are commonly sericitized (Map Unit Ds). Intrusive breccias, with fragments displaying well-developed hydrothermal reaction rims, and potassic alteration occur in this zone (Plate 1-16-1). Fine-grained plagioclase-phyric diorites, with well-defined chilled margins occur as numerous dikes which invade ultramafic intrusions. Larger dioritic bodies show clear crosscutting relationships with ultramafic and gabbroic intrusions.

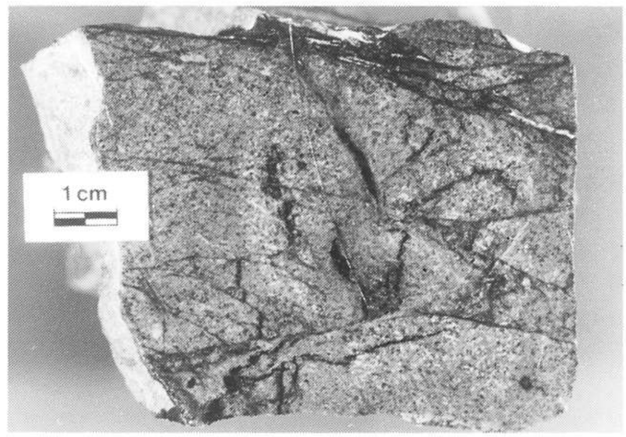

Plate 1-16-2. The sample is of a quartz-free, strongly carbonatized potassic intrusion from the Tatsamenie alteration zone. Small veinlets are chlorite-pyrite. Bleached halos adjacent to the veinlets are carbonate.

UNIT 13: GRANODIORITE

Small felsic intrusions occur as dikes and minor sills across the property. They are orange-buff weathering, typically strongly plagioclase porphyritic, and locally albitized or extensively carbonatized. Chlorite-pyrite veinlets and microveinlets, sometimes with chalcopyrite, occur locally (Plate 1-16-2). Alteration is most intense within the Tatsamenie alteration zone. In thin section, dikes of felsic appearance, and other types of intrusions in the alteration zone, appear quartz deficient; compositionally some of them may be monzonites to syenomonzites.

MAJOR STRUCTURAL FEATURES

TATSAMENIE ANTIFORM

The structure of the Tatsamenie antiform is shown on Figures 1-16-3 and 1-16-4. This large antiform is west verging, slightly overturned and plunges north at 10° to 15° with felsic tuffs and flows (Unit 1) in the core. A north-trending asymmetric lobe in the core of the structure is the result of upright cross-folding which trends east across the northerly trend of the earlier overturned antiform. Early bedding-cleavage intersection lineations are strongly rotated in the hinge regions of these east-west folds. The core of the antiform is cut by a single north-trending, east-side-down normal fault with an apparent offset of 50 to 75 metres. On the east limb of the antiform, the limestones of Unit 2 are strongly disrupted by numerous small-scale faults, and by early gabbroic intrusions. Most of these faults strike 045° and dip subvertically; many of them host iron-stained quartz-ankerite veins, locally well brecciated and up to 1.5 metres wide. Disseminated malachite is found in limestone talus near the gabbro contacts. Minor amounts of copper staining were noted on the steep cliffs which form the east limb of the antiform.

Two pale cream chert horizons, 20 to 30 metres thick, occur in felsic rocks of Unit 1 in the core of the antiform. These rocks are extensively sheared along sericitic foliation surfaces, and display the characteristic yellow-brown weathering product of arsenopyrite, scorodite. They have

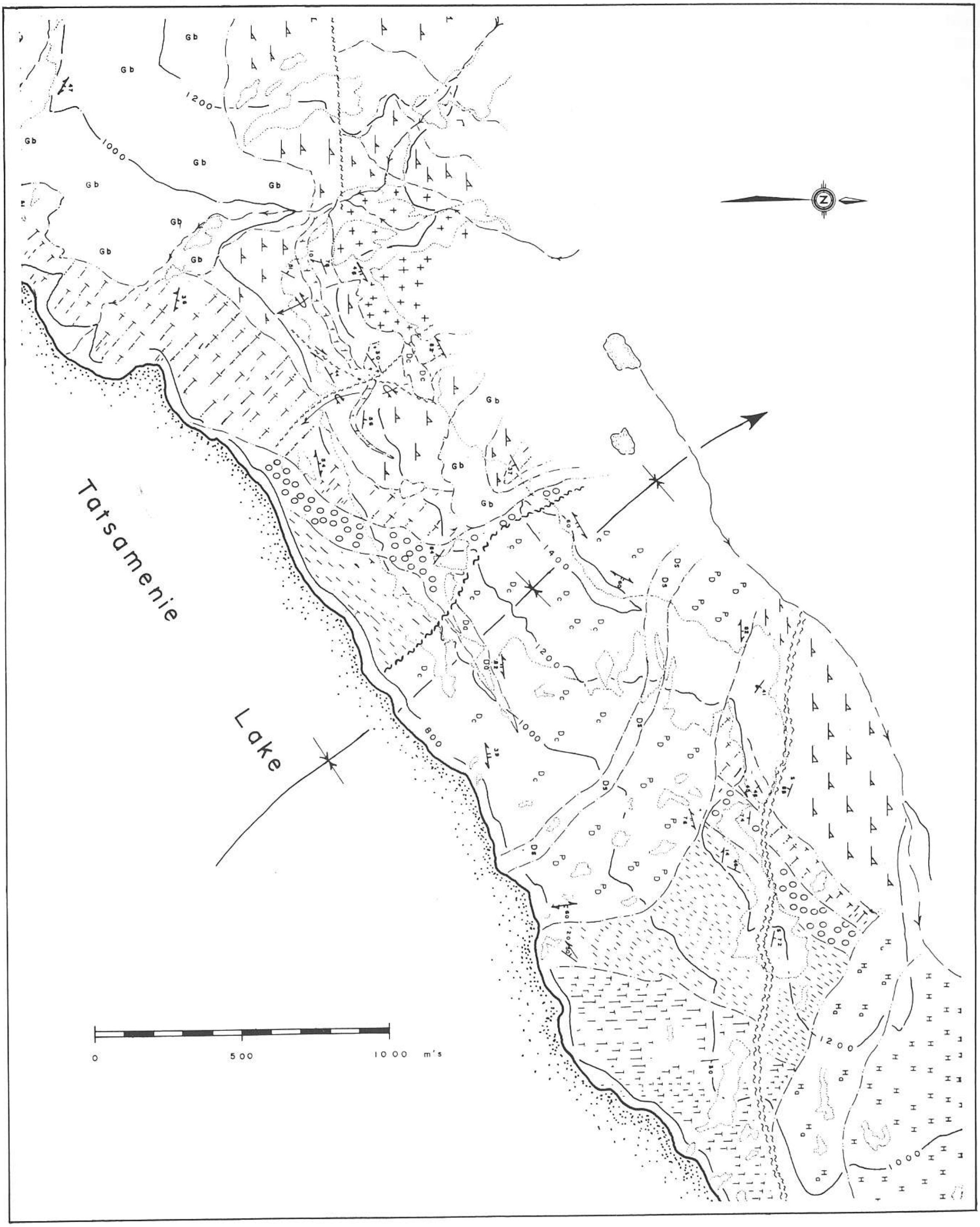

Figure 1-16-5. Geological plan of the Tatsamenie alteration zone. Lithologic patterns correspond to those of the stratigraphic column, Figure 1-16-2.

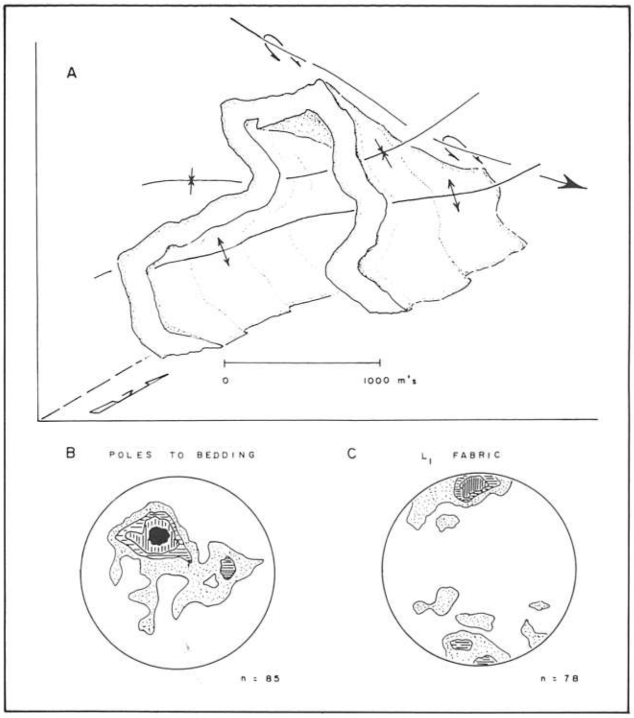

Figure 1-16-4. Isometric down-plunge projection through the Tatsamenie antiform (4A), sterographic equal-area contoured poles to bedding (4B) and contoured linear fabric (4c). Sterographic contour intervals are in area increments of 2%.

many of the field characteristics of cherts which occur distally to volcanogenic massive sulphide deposits. The felsic tuffs and flows of Unit 1 are cut by many white quartz veins. These veins do not have well-developed selvedges and are not sulphide bearing; they may have been formed by the migration of metamorphic fluids during deformation.

Limestones of Unit 2 on the western limb of the antiform are intruded by a massive to weakly foliated gabbro (Unit 7). Calcsilicate-magnetite alteration occurs on the northwestern contacts of this and other smaller intrusions. No copper skarn mineralization was identified on the gabbro contacts.

TATSAMENIE ALTERATION ZONE

The Tatsamenie alteration zone is an area of bright orange-buff weathering, carbonate and potassium alteration exposed on the north side of Tatsamenie Lake and exceeds 14 square kilometres in area. Primary textures in supracrustal rocks are only occasionally preserved. The principal structural features are shown on Figure 1-16-5.

The alteration zone is both cored and frequently dissected by several intrusive phases. An intensely chloritized and highly sheared dioritic intrusion (Unit Dc) is exposed in the central part of the zone. This early intrusion is in fault contact with supracrustal rocks on its western margin. Movement on this fault displaces rusty weathering alteration zones 50 to 75 metres down to the west. A younger porphyritic diorite stock (Unit PD) lies to the east of the earlier diorite. Strongly sericitic alteration zones may be developed (Unit Ds) over distances of 40 to 60 metres along this contact.

Heavily carbonatized plagioclase-porphyritic granodiorites to monzonites (Unit 13) are best exposed in the northwestern parts of the alteration zone. They do not contain significant free quartz and some may be syenitic. Many of them carry discrete chlorite-pyrite and potassium feldspar veinlets. Quartz veins are also present, but much less common. Extensive replacement by carbonate minerals has obliterated many of the textures in the igneous rocks.

Several faults cut the main alteration zone. Strong silicification, development of green micas (fuchsite?) and multistage veins are common within and marginal to these faults (Plate 1-16-3). The largest fault dips subvertically and is exposed for over a kilometre on the northeastern margin of the alteration zone, (Figure 1-16-5). It is marked by a zone of alteration and deformation at least 50 metres wide. Alteration consists of abundant graphite, extensive silicification and fuchsite-sericite alteration assemblages in mafic rocks (Plate 1-16-4). Large-scale overturned drag-folds are developed in a zone up to 100 metres wide either side of this fault. Stratigraphic offsets and minor structures suggest that the fault is normal and the movement is south side down. This fault and related structures offset and flatten the limbs of a broad, open synform trending through the core of the alteration zone.

THE WEST WALL FAULT AND ULTRAMAFICS SOUTH OF TATSAMENIE LAKE

Two ultramafic bodies are exposed on the higher plateaus south of Tatsamenie Lake; the geology and structural relationships of the more westerly of the two are shown on Figure 1-16-6. The rocks range from anorthositic gabbro to peridotites. The larger of the two crops out 2.5 kilometres east of the ultramafic body shown on Figure 1-16-6. At this location, a coarse-grained, pyroxene-rich peridotite is truncated and invaded by younger dioritic intrusions (Unit 12). The intrusion is fault bounded to the east, where a prominent zone of orange carbonatization defines the contact. The fault is a relatively old structure and is truncated by younger dioritic

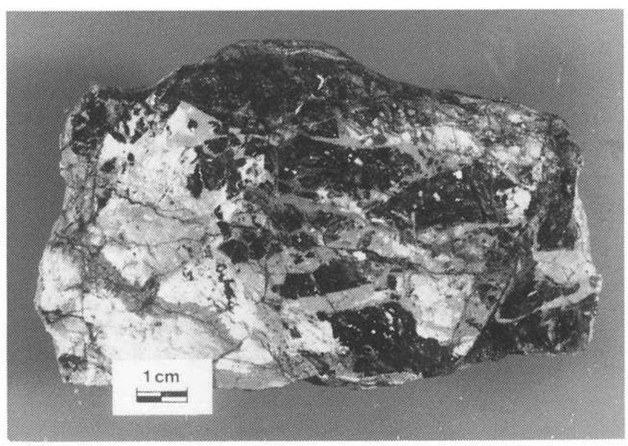

Plate 1-16-3. The hand specimen is taken from the fault breccias exposed in a major east-west fault bordering the Tatsamenie alteration zone. Graphitic sediments are intensely disrupted by carbonate (white) and potassium silicate veinlets (light grey).

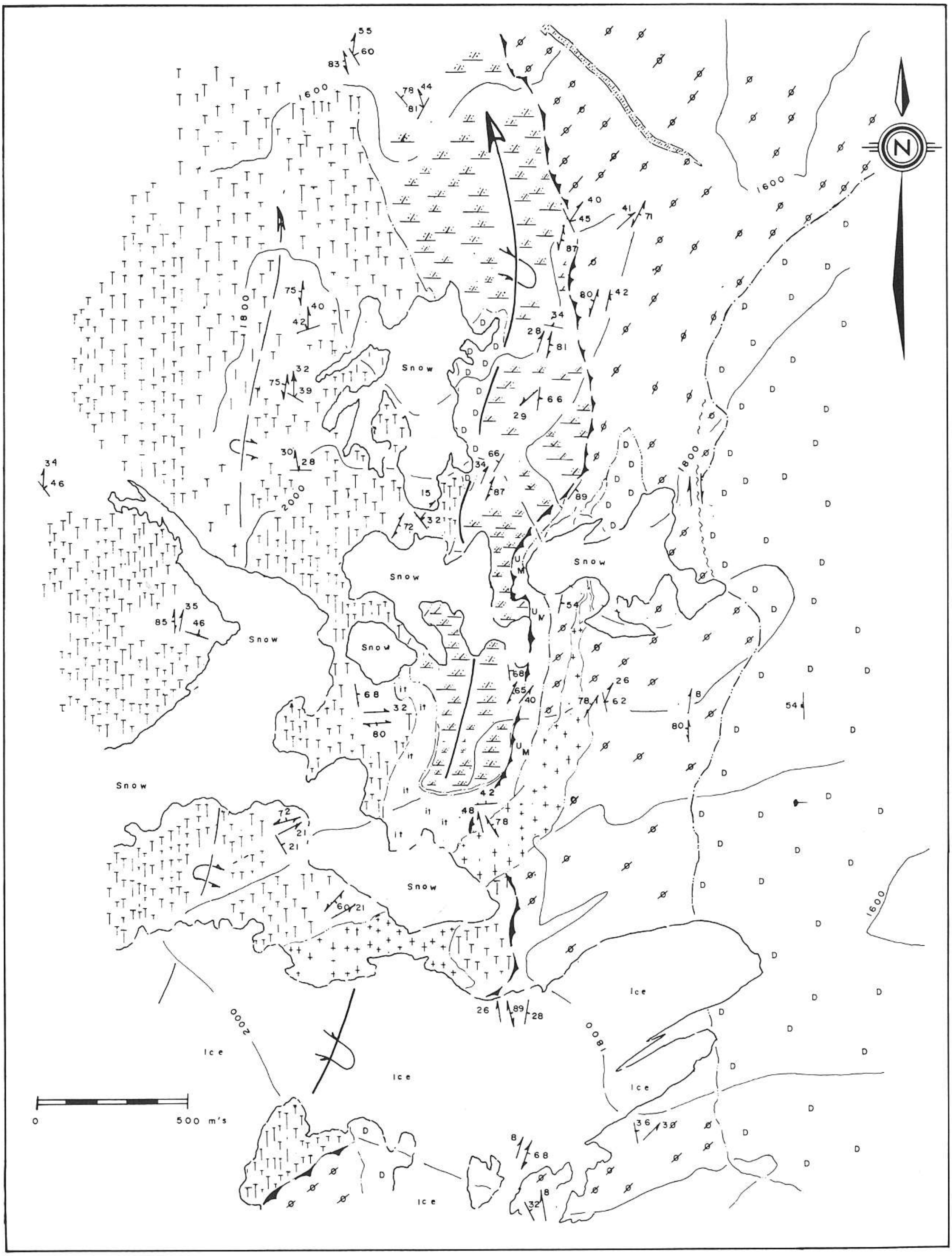

Figure 1-16-6. Geological plan of the West Wall fault and ultramafic zone. Lithologic patterns correspond to those of the stratigraphic column, Figure 1-16-2.

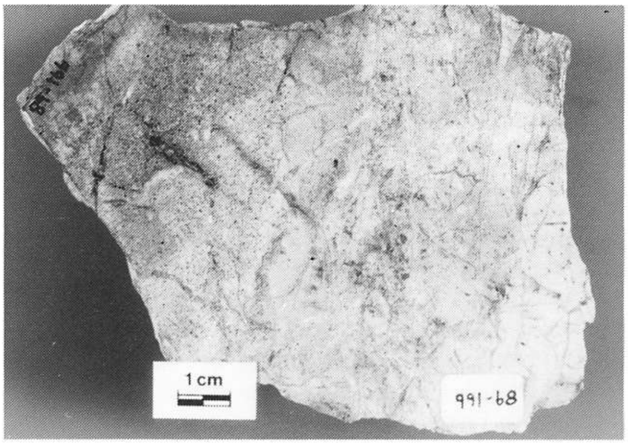

Plate 1-16-4. The protolith of this sample is a fine-grained mafic pyroclastic from the Tatsamenie alteration zone. The sample is pervasively carbonatized; the darker grey flecks are pyrite and green mica.

intrusions to the north and south. This ultramafic has been partially serpentinized, but primary igneous textures are easily discernible.

The ultramafic shown on Figure 1-16-6 ranges from a completely serpentinized peridotite to coarse-grained anorthositic gabbro. This rock is generally extensively tectonized including the development of mlyonitic bands within 25 metres of the bounding fault. Grain size reduction occurs not only in extensively serpentinized ultramafics, but also in extremely coarse grained anorthositic gabbros.

The distribution of the more westerly ultramafic body, closely corresponds to the surface expression of a major structure, the West Wall fault, first identified by geologists with Chevron Minerals Ltd. (Wober and Shannon, 1985). Across this fault, intensely porphyroblastic, highly strained pillow flows in the hangingwall are in contact with fine-grained mafic and intermediate tuffaceous rocks in the footwall. The strike of argillite marker beds in the hangingwall flows appears discordant with the strike of the fault and with bedding in the footwall. Stratigraphic and structural relationships, and the presence of gabbroic to ultramafic bodies aligned along the fault, all indicate that this fault is a west-verging thrust.

Closely associated with the fault, and principally developed on the footwall side, is a north-plunging, west-verging overturned synform, mapped for over 3 kilometres of strike. The intensely deformed, overturned limb appears to be truncated by the West Wall fault. Close to the highly altered ultramafic body splays from the West Wall fault are mineralized (10 to 20 grams per tonne gold) over relatively narrow (0.6 metre) widths (Barr, 1989). A zone in which penetrative linear fabrics show a marked rotation coincides with a dilational zone and a gold showing named the "Two Ounce Notch" (Schroeter, 1986). The rotation of these lineations may be due to superposition of a later fold across the overturned synform, or by fault deflections.

DISCUSSION

REGIONAL GEOLOGY

The geological relationships described in this preliminary report have significant implication for interpretation of the regional geology. Several geological features must be explained:

(1) There is a marked decrease in the state of strain in the rocks which form the Tatsamenie alteration zone relative to the surrounding rocks. Within the alteration zone, the rocks lack a strong penetrative fabric, and contain porphyroblastic zeolites, consistent only with very low grade (burial) metamorphism. Pillows are relatively undeformed and the quartz and carbonate components of the clastic sediments are not recrystalized. Rocks to the west and south of the alteration zone are, in contrast, strongly porphyroblastic, contain the metamorphic mineral assemblage amphibole (actinolite)-albite-biotite and are strongly recrystalized. Small-scale kinematic indicators suggest these rocks record more than one high-strain event. It appears likely that rocks within the alteration zone lie above the mid-Triassic unconformity and belong to the Stuhuni group. The large extensional faults mapped in the area of the alteration zone may be part of a major graben.

(2) Rocks south of Tatsamenie Lake cannot be correlated with those exposed in the Tatsamenie alteration zone. This discontinuty of rock types and structural style is best explained by inferring an unconformity between the two rock packages. Stratigraphy south of Tatsamenie Lake, and west of the alteration zone, appears to be Lower Triassic and older.

(3) The distribution of ultramafic rocks south of Tatsamenie Lake suggests that these rocks are part of a dismembered and poorly preserved ophiolite assemblage. This interpretation is based on the following evidence:
   (a) The ultramafic rocks form long, linear bodies, confined virtually exclusively to the hangingwall of the West Wall fault. They are found in only one instance to the west of this fault, and at this exposure which is less than 10 square metres in area, they may not be in place.
   (b) These rocks are directly overlain by a highly strained mafic pillowed-flow sequence.
   (c) The ultramafic rocks are peridotites which are texturally and mineralogically consistent with alpine-type ultramafic bodies.
   (d) The ultramafic intrusions lack signifcant contact aureoles and field data strongly suggest that the contacts are tectonic rather than intrusive.
   (e) One of the intrusions is cut by numerous mafic dikes. However, the dikes have chilled margins and appear to be dioritic as opposed to gabbroic in composition. The diorites may have been emplaced in a mid-Mesozic intrusive event, rather than feeders to the pillowed flows.
   (f) The orientation and indicated sense of movement on the northern parts of the West Wall fault is consistent

with the orientation of other larger west-verging ultramafic bounded faults, for example the Nahlin fault.

Implications For Exploration

The data presented in this report suggest that the following type of exploration targets may occur in the Tatsamenie Lake area:

- The felsic tuffs and flows in the core of the Tatsamenie antiform contain chert horizons which carry disseminated arsenopyrite and pyrite. They may be the lateral equivalents of massive sulphide horizons, similar to those mined at the Tulsequah Chief. The outcrop area of the felsic sequence should increase to the south of Tatsamenie Lake and north of Sam Creek, up the plunge of the Tatsamenie anticline. Mineralized cobbles and boulders carrying disseminated lead and zinc sulphides have long been documented in the upper drainage of Sam Creek (Souther, 1971).
- The Tatsamenie alteration zone, and other similar alteration zones in the area, are an impressive target for precious metals exploration. The large-scale extension faults, the presence of numerous quartz-bearing and quartz-free intrusions, the development of intrusive and fault breccias, widespread carbonatization, fuchsitic and potassic alteration, are all suggestive of a highly favorable precious metal environment.
- The association of intensely altered ultramafic rocks and large-scale low-angle faults, south of Tatsamenie Lake, favours the formation of shear-related gold veins. Zones of dilatancy within these linear faults are prime targets and these will generally occur where second-order fault structures intersect the major faults; where changes in rock competency occur at depth along the fault trace; and in areas where different fold systems are superimposed near the faults. Careful geological mapping and interpretation is the best way of outlining these targets.

ACKNOWLEDGMENTS

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NOTES

GEOLOGIC AND ISOTOPIC ANALYSIS OF THE NISLING – NORTHERN STIKINE TERRANE BOUNDARY NEAR ATLIN, BRITISH COLUMBIA (104M/8)

By Jay L. Jackson, George E. Gehrels and P. Jonathan Patchett
The University of Arizona

KEYWORDS: Regional geology, Nisling Terrane, Stikine Terrane, Stuhini Group, Laberge Group, neodymium isotopes.

INTRODUCTION

Three tectonic assemblages underlie the eastern flank of the Coast Range batholith in northern British Columbia: from west to east, the Nisling, northern Stikine and northern Cache Creek terranes (Figure 1-17-1). Each assemblage preserves a unique stratigraphy and geologic history, and the boundaries between these assemblages are marked by faults along much of their lengths. One of the goals for understanding the accretionary history of western North America is to document the time at which adjacent crustal fragments first came together. This study examines the age of juxtaposition of the Nisling and northern Stikine assemblages, and is part of a larger project focused on assessing the early Mesozoic tectonic relationships between the Nisling, northern Stikine and northern Cache Creek terranes.

The Nisling Terrane comprises metamorphosed sedimentary and volcanic rocks interpreted to belong to a Proterozoic to Paleozoic pericratonal basinal assemblage (Wheeler and McFeely, 1987). Nisling rocks lie west of volcanic and sedimentary strata of the Upper Triassic Stuhini Group, part of the northern Stikine Terrane. In northern British Columbia and southern Yukon, these two assemblages are separated by the north-northwest-striking Llewellyn fault zone (Bultman, 1979; Mihalynuk and Rouse, 1988; Mihalynuk *et al.*, 1989) and Tally Ho shear zone (Doherty and Hart, 1988; Hart and Pelletier, 1989; see Figure 1-17-1).

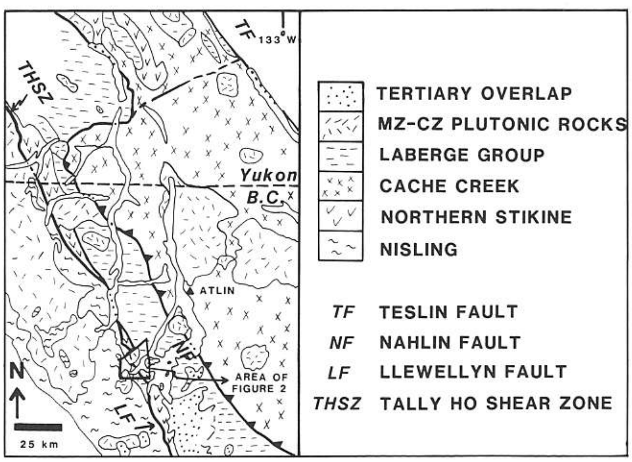

Figure 1-17-1. Generalized geology of the area around Atlin, British Columbia (modified after Wheeler and McFeely, 1987). Refer to the text for discussion of map units and structures. Note location of the Willison Bay study area.

Although northern Stikine lithologies are nowhere preserved in demonstrable primary depositional contact with Nisling rocks, several authors have discussed evidence pointing toward a Late Triassic link. Bultman (1979) noted: (1) the presence within Stuhini conglomerates of metamorphic clasts that resemble Nisling lithologies and of porphyritic granodiorite clasts that resemble a Late Triassic plutonic suite that intrudes Nisling rocks; (2) the presence within Nisling rocks of augite porphyry dikes that are similar to augite porphyry flows within the Stuhini succession; and (3) a zone of weathering, interpreted to be pre-Stuhini, within the porphyritic granodiorite near its contact with Stuhini strata. Werner (1978) also discussed the possibility of an original unconformity separating Nisling and Stuhini rocks, on the basis of similar chemical composition of pyroxenes in Stikine augite porphyry flows and in an augite porphyry dike that intrudes the Nisling assemblage. Mihalynuk and Rouse (1988) believe that an unconformity may be preserved in the Tutshi Lake map area (104M/15) due to a lack of deformational features within strata adjacent to the contact.

The goal of this research project is to evaluate geologic relationships noted previously along the Nisling-Stuhini contact, and to provide more quantitative constraints on the nature of the contact through analysis of the neodymium and strontium isotopic signatures of rocks near the contact. We chose to begin the study in Willison Bay at the south end of Atlin Lake (104M/8) where excellent exposures of Nisling and Stuhini rocks are preserved. In this report, we first describe the geology of the Willison Bay area, then discuss the isotopic studies in progress at the University of Arizona.

GEOLOGY OF THE WILLISON BAY AREA

Figure 1-17-2 shows a geologic map of the Willison Bay shoreline, illustrating the major rock units found along the Nisling-northern Stikine contact. General descriptions of these rocks are given below, as are relevant structural observations and interpretations of relationships between units. All descriptions are based on field observations only, as thin section analysis and quantitative petrography are in progress. This geologic framework was assembled as a basis for collecting and interpreting isotopic samples. For a more detailed account of the geology of the region, the reader is referred to Mihalynuk and Mountjoy. (1990, this volume).

LAYERED ROCKS

NISLING ASSEMBLAGE (PPZn)

The western end of the study area is underlain by Nisling assemblage rocks composed of biotite-quartz-feldspar

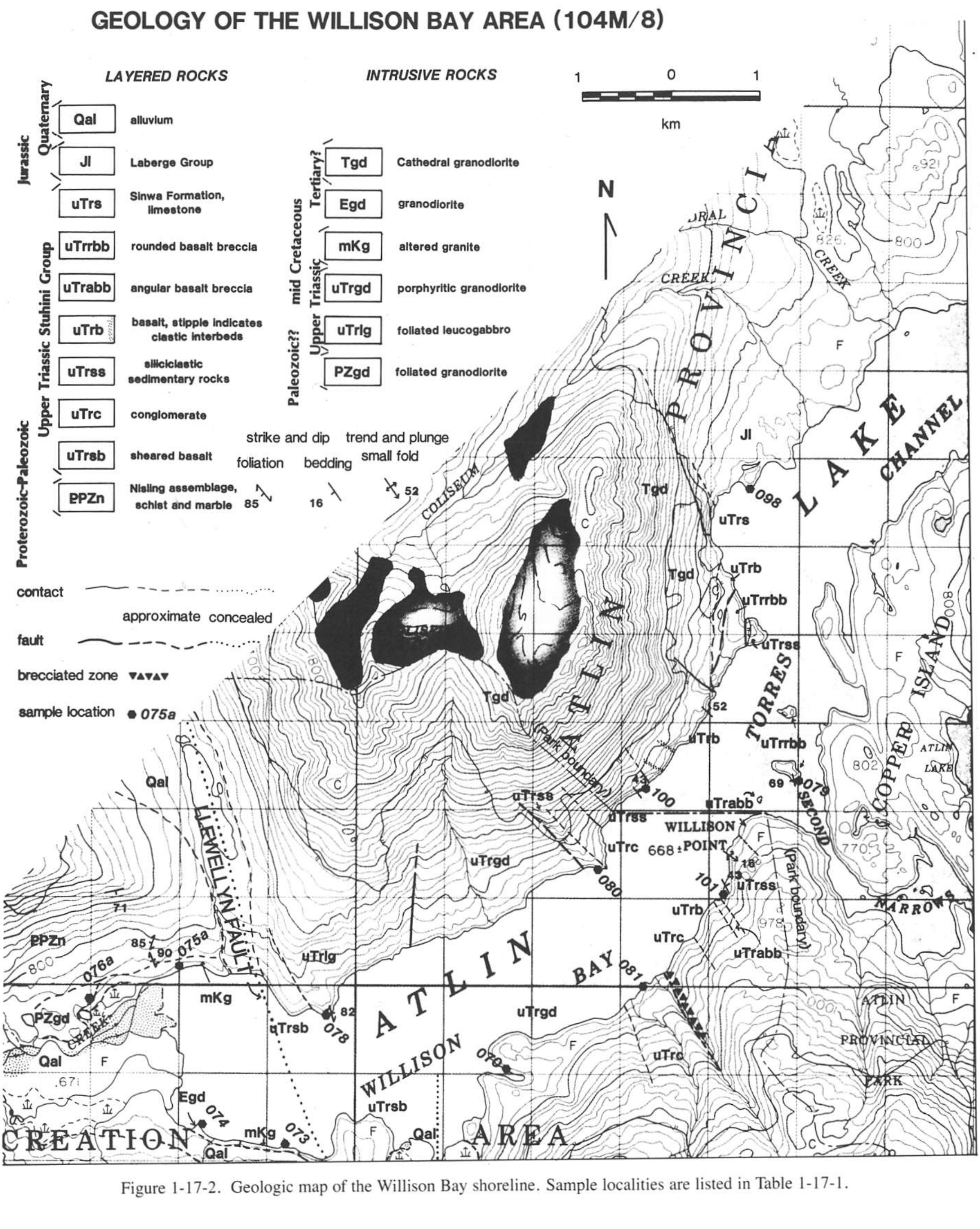

Figure 1-17-2. Geologic map of the Willison Bay shoreline. Sample localities are listed in Table 1-17-1.

schist, and marble units ranging from coarsely crystalline, white calcite marble to light grey, fine-grained well-foliated marble. Quartzofeldspathic schist grades into calcsilicate schist near contacts with the marble layers. Folding is apparent at outcrop scale and Mihalynuk *et al.* (1989) have interpreted at least four phases of deformation within the Nisling assemblage. Nisling rocks are intruded by the foliated granodiorite (PZgd) and the undeformed mid-Cretaceous granite (mKg) and Eocene(?) granodiorite (Egd). The Llewellyn fault forms the eastern extent of Nisling exposures.

UPPER TRIASSIC STUHINI GROUP

SHEARED BASALT (uTrsb)

Just east of the Llewellyn fault zone, the lowermost Stuhini unit consists of medium to dark green massive basalt and pyroxene-phyric basalt that have been strongly sheared, brecciated and locally altered by calcite veining. The fabric in these rocks is not penetrative, nor does it have a strong preferred planar orientation. Most of the brittle shear surfaces are, however, subvertical. Aside from the deformation and alteration, this unit resembles other Upper Triassic basalt layers along Willison Bay. The foliated leucogabbro (uTrlg) described below intrudes the eastern margin of this package.

CONGLOMERATE (uTrc)

A poorly sorted, matrix-supported, cobble to boulder conglomerate overlies the Late Triassic granodiorite (uTrgd). Alluvium conceals the contact along the shoreline, but higher on the slopes of the Cathedral (south of Coliseum Glacier, Figure 1-17-2) Mihalynuk (personal communication, 1989) has documented an unconformity preserved beneath the conglomerates. Clasts in the conglomerate consist mainly of felsic intrusive rocks, dominated by a porphyritic granodiorite identical to the underlying lithology. Locally, up to 5 per cent of the clasts are crenulated chlorite-quartz schists that resemble some lithologies in the Nisling assemblage. Other clast lithologies include intermediate to mafic volcanic rocks, argillite and minor limestone. All clasts, except the argillite fragments, are extremely well rounded, regardless of size. They are supported by a matrix of fine to medium-grained, dark green volcanic lithic arkose similar to the overlying sandstone. The siliciclastic strata (nTrss) conformably overlie this conglomerate.

SILICICLASTIC STRATA (uTrss)

This unit contains a variety of lithologies. On the north side of the bay, the dominant rock type is a green, pebbly to fine-grained, graded, volcanic lithic arkosic sandstone in beds from 50 to 200 centimetres thick. Most of the sandstone layers are slightly calcareous. Along the south shoreline, medium grey, calcareous greywacke, siltstone, argillite and mudstone characterize this unit, and the latter locally contains up to 1 per cent pyrite. These rocks contain abundant evidence of soft-sediment deformation, including contorted beds, small folds, rip-up clasts of sandstone within argillite, and dewatering structures. Coarse to medium-grained sandstone, similar to that along the north shore, are also interbedded with this sequence. Rare intraformational isoclinal folds plunge gently southeast and indicate deformation in the sedimentary rocks along the south shore (Figure 1-17-2).

BASALT (uTrb)

Along the south shore, Stuhini basalt is predominantly massive, medium green in color (due to low-grade metamorphism?) and contains small (<1 millimetre) veinlets of calcite. Along the north shore, a prominent, well-exposed section contains relict pillows about 70 centimetres in average length and 30 centimetres in average height. Zones with demonstrable pillows typically display pronounced carbonate coatings and minor calcite veining. Within this section are interbedded grey siltstone and fine-grained sandstone beds that have a platy character and weather to a brown colour.

ANGULAR BASALT BRECCIA (uTrabb)

Pyroxene porphyry basalt is found locally as monolithologic angular blocks within a basaltic matrix. Breccia blocks range in size from several centimetres to 50 centimetres in diameter. These deposits are poorly sorted in general, and contain rare quartz ± epidote veins that strike northwest, dip moderately northeast and display minor folds.

ROUNDED BASALT BRECCIA (uTrrbb)

Rounded fragments (bombs?) of pyroxene-phyric basalt up to 1 metre in diameter in a fragmental basaltic matrix comprise this unique lithology. Mihalynuk *et al.* (1989; unpublished field guide) interpret this unit as a phreatomagmatic breccia erupted in a shallow submarine setting.

SINWA FORMATION (uTrs)

Bultman (1979) correlated Norian limestone at the top of the Stuhini Group with the Sinwa Formation in the Tulsequah (104K) map area (Souther, 1971). Along the north shore of Willison Bay, Sinwa rocks consist of light to dark grey, well-layered, steeply dipping limestone and interlayered brown-weathering argillite sequences up to 30 metres thick.

LOWER TO MIDDLE JURASSIC LABERGE GROUP (Jl)

Laberge Group strata gradationally overlie the Stuhini Group and consist of interbedded argillite, siltstone and greywacke that are deformed into open, and locally tight to isoclinal, northwest-trending folds. Where sampled along the north shore of Willison Bay, Laberge rocks are dark grey, platy, homogeneous siltstone and argillite that weather to a light brown colour.

INTRUSIVE ROCKS

FOLIATED GRANODIORITE (PZgd)

At the northwest corner of Willison Bay, a well-foliated, coarse-grained hornblendebiotite granodiorite intrudes the Nisling assemblage and is in turn intruded by the mid-Cretaceous granite (mKg). The degree of fabric development in this rock suggests that it is older than the Late Triassic granodiorite (uTrgd) that shows only a local, moderately developed foliation.

FOLIATED LEUCOGABBRO (uTrlg)

Within the Stuhini Group are strongly to weakly foliated hornblendepyroxene leucogabbro (plagioclase content to 50 per cent) and gabbro bodies that are probably related to Stuhini basaltic volcanism, The Late Triassic granodiorite (uTrgd) intrudes the eastern margin of this unit.

LATE TRIASSIC GRANODIORITE (uTrgd)

A potassium feldspar porphyritic, coarse to medinm-grained hornblende-biotite (15% and 5% respectively) granodiorite (60% feldspar, 20% quartz) underlies a large portion of the centre of the field area. Stuhini conglomerate unconformably overlies or is faulted against the eastern contact of this pluton, while the western contact is an intrusive and faulted contact with foliated leucogabbro (uTrlg). Bultman (1979) obtained a K-Ar hornblende age of 215 ± 35 Ma on this body. This granodiorite is similar to Late Triassic plutons that intrude Nisling rocks west of the Llewellyn fault and Tally Ho shear zone along much of their lengths (Mihalynuk *et al.*, 1989; Hart and Pelletier, 1989).

MID-CRETACEOUS GRANITE (mKg)

A large biotite granite (biotite 15%, quartz 35%, feldspar 50%) body intrudes Nisling rocks and older intrusive bodies west of the Llewellyn fault. Epidote alteration has affected this granite and biotite grains have a greenish tinge.

EOCENE GRANODIORITE (Egd)

A small (400 square metres) outcrop of very fresh, white, biotite granodiorite (biotite 20%, quartz 20%, feldspar 60%) is exposed within the mid-Cretaceous granite. The lack of alteration and difference in composition suggest that this is a distinct and probably younger plutonic body. No published radiometric ages constrain this unit.

TERTIARY CATHEDRAL GRANODIORITE (Tgd)

The Cathedral granodiorite caps Cathedral Mountain in the centre of Figure 1-17-2 and received only cursory attention in this study. Where observed, the intrusive body is a white to grey-weathering biotite granodiorite to quartz diorite.

NEODYMIUM ISOTOPIC STUDIES

The presence of metamorphic clasts and porphyritic granodiorite clasts within the Stuhini conglomerate strongly suggests that the Nisling assemblage and related intrusive rocks served as a source area for detritus deposited into basins associated with the northern Stikine assemblage. One way to quantitatively test this hypothesis is to characterize the neodymium isotopic signature of the sedimentary rocks in the northern Stikine Terrane. Previous isotopic studies show that much of the Nisling assemblage is composed of detritus from older source regions that were at least in part Proterozoic in age. L. Werner (in Monger and Berg, 1987) reported a 900 Ma rubidium-strontium isochron from Nisling lithologies. Samson *et al.* (1989a) interpreted an early Proterozoic neodymium mantle separation age for material in the Nisling Rocks. Gehrels *et al.* (in press) report an early Proterozoic source for detrital zircons in the Nisling assemblage based on uranium-lead geochronologic analyses.

If detritus from the Nisling Terrane is present in Stuhini sedimentary rocks and has been incorporated into plutons that intrude these strata, neodymium isotope data should show a distinct mixing of Proterozoic material with younger, mantle-derived material from the Stikine volcanic rocks. To determine if this is the case, we have collected 12 samples in the Willison Bay area (list of sample localities given in Table 1-17-1) and 65 other samples from northern British Columbia and southern Yukon. This sample suite will provide provenance signatures for the northern Stikine Terrane and overlapping sedimentary strata of the Lower to Middle Jurassic Laberge Group and Upper Jurassic to Lower Cretaceous Tantalus Group.

TABLE 1-17-1

Sample Number	Unit	Location (UTM)
89-AT-070	Late Triassic granodiorite	517-691
89-AT-073	mid-Cretaceous granite	492-682
89-AT-074	Eocene(?) granodiorite	483-684
89-AT-075a	mid-Cretaceous granite	480-702
89-AT-076a	foliated granodiorite	470-698
89-AT-078	Late Triassic foliated leucogabbro	498-697
89-AT-080	Upper Triassic pyroxene-phyric basalt	549-724
89-AT-081	chlorite schist clast in Late Triassic conglomerate	532-700
89-AT-098	Jurassic Laberge siltstone	545-757
89-AT-100	Upper Triassic fine-grained sandstone	533-723
89-AT-101	Upper Triassic grey argillite	542-711

Locations of neodymium-strontium isotopic samples collected in the Willison Bay area. Sample sites are shown in Figure 1-17-2 (note that only the last three digits are given on the map). Locations are UTM grid references; all are within NTS sheet 104M/8, Edgar Lake.

REGIONAL TECTONIC SIGNIFICANCE

If Upper Triassic Stuhini Group strata of the northern Stikine Terrane unconformably overlie, or received detritus from Nisling assemblage rocks, then this northern portion of the Stikine Terrane may be somewhat different from its southern counterpart. In the south, Anderson (1989) interprets Upper Triassic Stuhini rocks to lie unconformably on the Stikine assemblage, a succession of Devonian to Permian limestones and arc-type volcanic rocks. Samson *et al.* (1989b) studied the neodymium and strontium isotopic signature of southern Stikine rocks and concluded that much of this assemblage has a primitive isotopic signature, with rocks consisting primarily of new additions from the mantle rather than recycled continental material. This study will serve as a basis for comparison with previous isotopic work in the southern part of the Stikine Terrane and will provide more rigorous, quantitative constraints for relationships between units in the northern part of the terrane.

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Geological Survey of Canada in Vancouver provided additional logistical support. J. Roska served as Jackson's field assistant. This work has been funded by U.S. National Science Foundation grant EAR-8903764 to Gehrels and Patchett, research grants to Jackson from the Geological Society of America, Sigma Xi Grants-In-Aid of Research, British Columbia Geoscience Research Grant Program (grant number RG89-13) Homestake Mineral Development Company of Canada, Shell Oil Company (U.S.) and Standard Oil Production Company (U.S.) and a fellowship to Jackson from Chevron U.S.A.

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NOTES

GEOLOGY OF THE TAGISH LAKE AREA (104M/8, 9E)

By M.G. Mihalynuk and K.J. Mountjoy

KEYWORDS: Regional geology, Boundary Ranges metamorphic suite, Nisling assemblage, Stuhini Group, Laberge Group, Tagish volcanic suite, Coast Belt, Llewellyn fault, Engineer mine, gold veins.

INTRODUCTION

A third season of regional 1:25 000-scale geological mapping (compiled at 1:50 000) was completed in the Tagish Lake area between Skagway, Alaska and Atlin, British Columbia during 1989 (Figure 1-18-1). Tagish Project mapping began in 1987 at the division between the Coast and Intermontane belts on the British Columbia–Yukon border and has continued southeast on contiguous half map sheets, finishing this year near the south end of Atlin Lake, adjacent to the Atlin Provincial Park. Between early June and mid-September, approximately 1100 square kilometres were mapped in the Edgar Lake (104M/8) and Fantail Lake (104M/9E) areas to complement the 1700 square kilometres covered in the previous two seasons (Mihalynuk and Rouse, 1988a, b; Mihalynuk *et al.*, 1989a,b). A belt of metamorphic rocks in central 104M/8 was mapped by Lisel Currie (1990, this volume) as part of her doctoral thesis at Carleton University. Parts of this belt were also mapped as a component of this study.

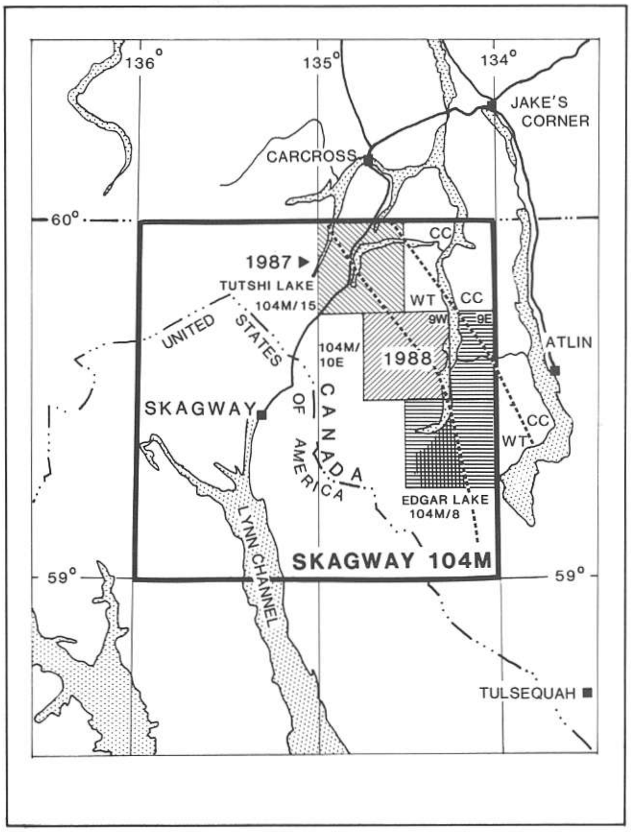


Figure 1-18-1. Location map showing the course of Tagish project mapping. The Whitehorse trough is outlined west of Atlin Lake by the dotted lines and is labelled WT; CC denotes Cache Creek Terrane. The double hatched region within 104M/8 outlines the study area of Currie (1990, this volume).

As in past seasons, a regional geochemical moss-mat survey was conducted in concert with geological mapping; sample density is approximately one per 12 square kilometres of drainage basin. Some ninety mineralized samples were collected for analysis to aid in evaluating the mineral potential of the area.

The area is part of an anomalous arsenic-antimony (and sporadic gold) province extending into the Yukon (Schroeter, 1986) and has high economic mineral potential. Mapping and sampling have been conducted to evaluate the involvement of depositional and late deformational processes in the formation of ore deposits. Potential hostrocks for mineralization range from Proterozoic(?)-Paleozoic metamorphics to Tertiary intrusives and extrusives.

ACCESS AND PHYSIOGRAPHY

Large lakes at elevations ranging from 656 to 844 metres provide boat access to about 30 per cent of the map area. These lakes are most easily reached by floatplane or helicopter, both stationed in Atlin some 40 kilometres to the east. During high water (early July through to late September) Tagish Lake can also be reached from Atlin by powerboat via the Atlin River. At highest water, this route is perilous and recommended only for experienced boaters.

Treeline varies from 1000 to 1400 metres elevation with major peaks extending to over 1900 metres. Rapidly receding glaciers cover about 20 per cent of the 104M/8 alpine area and both permanent and fresh snow covers many north-facing slopes for most of the summer months.

REGIONAL GEOLOGICAL SETTING

Extensive regional geological mapping was previously conducted in the area by Christie (1957) and Bultman (1979). Both have provided excellent guidance, and the generalized geological picture of this study does not vary greatly from that established by Bultman for strata within the Whitehorse trough (Figure 1-18-1).

NOTES ON THE TERRANE ARCHITECTURE

The map area covers a short segment of the north-northwest-trending boundary between the Coast and Intermontane geomorphological belts. The structural grain of the

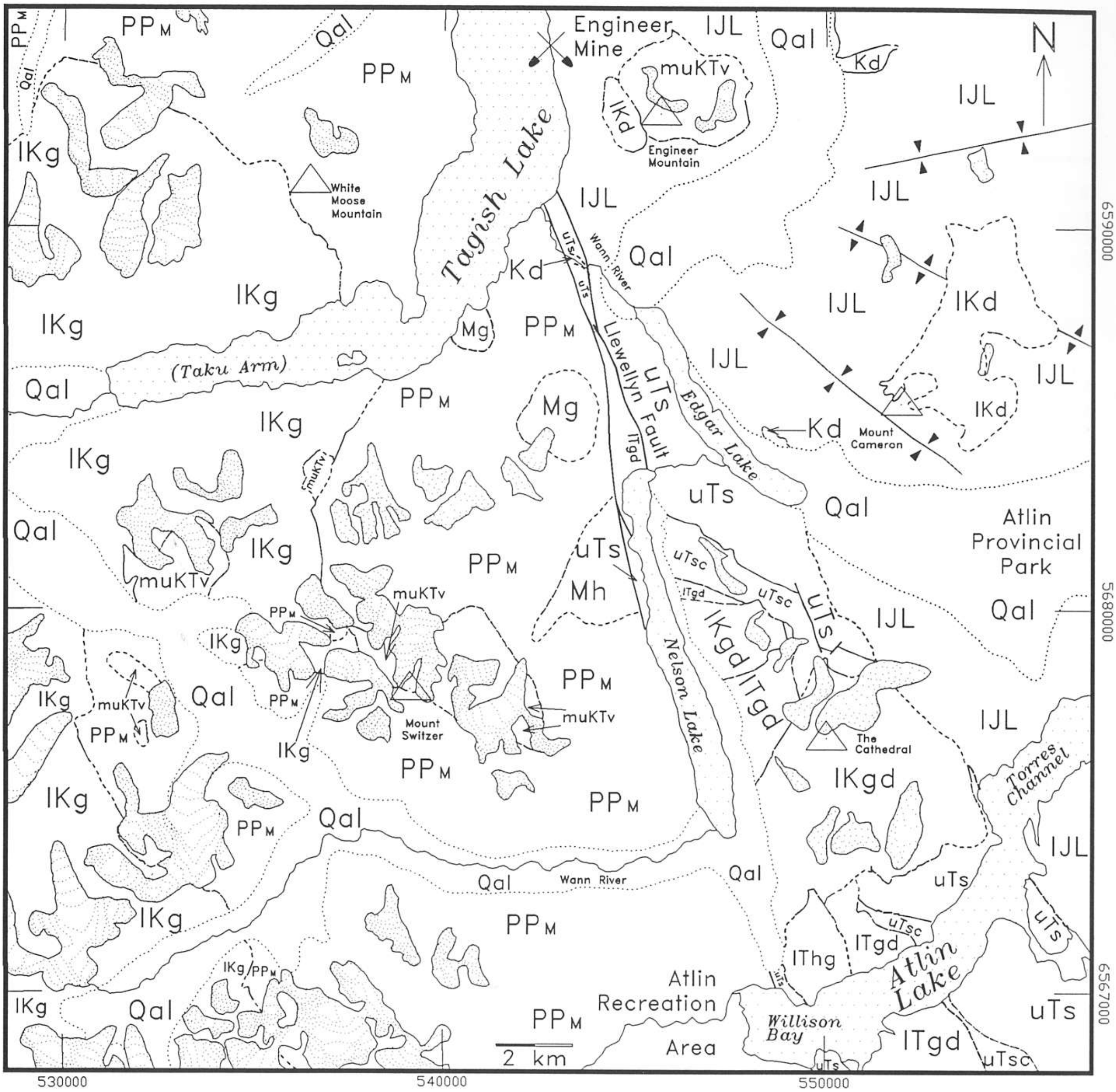
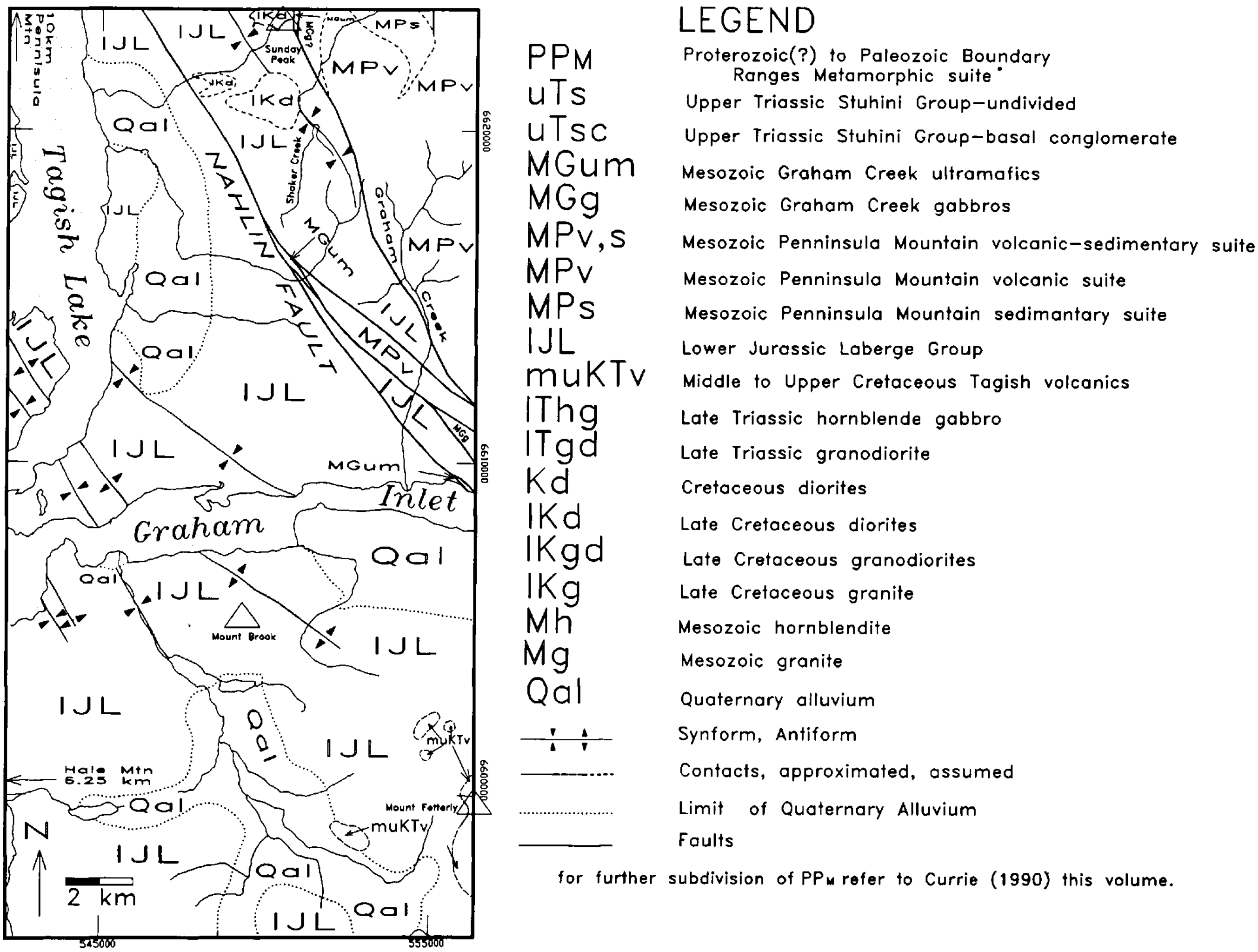


Figure 1-18-2. Simplified geology of 104M/8, *see* Currie (1990, this volume) for a subdivision of Unit PPM; the area mapped by Currie is shown by the double hatched pattern on Figure 1-18-1

entire area is subparallel to this boundary. This same boundary is roughly coincident with what has historically been considered the contact zone between undivided "central gneiss" (Tipper *et al.*, 1981), more recently termed the Nisling Terrane (Wheeler *et al.*, 1988) [This naming convention is followed by Mihalynuk *et al.* (1989), however, since the Nisling rocks have yet to be demonstrated as unique terrane, a name such as the "Nisling assemblage" is more suitable (D. Brew, personal communication, 1989)], and strata of the Whitehorse trough. In contact with the eastern margin of Whitehorse trough is the Cache Creek Terrane. Trough sediments are dominated by Upper Triassic and younger arc volcanics and clastics which are thought to overlap, and therefore link, the Proterozoic to Paleozoic metamorphosed and displaced Nisling continental margin assemblage to the west with low-grade oceanic rocks of the Cache Creek Terrane to the east. Thus, strata in the Whitehorse trough have come to be known as the Inklin overlap



assemblage (Wheeler *et al.*, 1988), yet in few places is an unconformable contact unequivocal. One such location is in the west-central Tutshi Lake area where probable Toarcian (late Lower Jurassic), belemnite-bearing conglomerates contain pebbles and boulders derived from the metamorphic rocks and rest with angular unconformity on the metamorphic suite. Even at this locality the contact has been masked by post-depositional shearing. In north-central Tutshi Lake area volcanic strata of presumed Late Triassic age also appear to rest unconformably on, as well as in fault contact with, the metamorphic suite. The eastern contact relationships are potentially more ambiguous because of lithologic similarities between rocks in the Whitehorse trough and the youngest sediments of the Cache Creek Terrane (J. Jackson, personal communication, 1989).

In south-central British Columbia, well-documented relationships show Cache Creek clasts in the Upper Triassic volcanic succession, and rocks with lithology identical to Upper Triassic volcanics within subduction-related mélange of the Cache Creek (Monger, 1984) demonstrating their proximity by Late Triassic time. Although no such clear cut relationship has been demonstrated in northwestern British Columbia, Bloodgood and Bellefontaine (1990, this volume) observe Norian and younger strata above Cache Creek rocks. They have not seen a stratigraphic contact, but one is expected. Their overling strata are gritty limestones of Norian age and a conglomerate containing clasts of this limestone, as well as boulders of intrusive rock and clasts derived from the underlying Cache Creek strata. Both lithologies are similar to rocks of the Stuhini Group in the Edgar Lake area (with the exception of Cache Creek Group clasts in the conglomerate). It therefore seems that conglomerates deposited on the eastern and western margins of the Whitehorse trough were derived respectively from the underlying Cache Creek Group and Nisling Terrane. This comprises an overlap assemblage consistent with the regional tectonic syntheses of Wheeler and McFeely (1987) and Wheeler *et al.* (1989).

LAYERED ROCKS

PROTEROZOIC(?) TO PALEOZOIC METAMORPHICS (PPm)

This belt of rocks, generally less than 15 kilometres wide, extends from the British Columbia–Yukon border to south of Atlin Lake. Within the Florence Range (south-central

104M/8) metamorphism attains upper amphibolite grade, but rapidly decreases eastward to transitional greenschist-amphibolite and greenschist adjacent to Nelson Lake. A similar, but more gradual, south to north decrease in metamorphic grade is also observed, with a low-grade culmination near the British Columbia–Yukon border. For a detailed description of these rocks within 104M/8 the reader is referred to Currie (1990, this volume). In brief, Currie recognizes four major lithologic subdivisions (roughly east to west; Figure 1-18-2): the Boundary Ranges metamorphic suite, Hale Mountain granodiorite, Wann River gneiss and Florence Range metamorphic suite. Boundary Ranges metamorphic rocks are predominantly chlorite-actinolite schists with lesser chlorite schist, thin marble, quartzite and orthogneiss (*see* also Mihalynuk *et al.*, 1989a, b). The foliated Hale Mountain granodiorite is medium grained, hornblende-biotite rich with plagioclase and less abundant but larger potassium feldspar augen. Epidote, evenly distributed as fine grains, is diagnostic. Wann River gneisses are well layered on a millimetre to decimetre scale, containing 20 to 60 per cent hornblende in layers alternating with quartzofeldspathic layers. Florence Range metamorphic rocks are pelites and semipelites (mostly without graphite), marbles, amphibolites, calcsilicates and minor quartzite. Currie has tentatively interpreted the contacts between these units as shear zones.

STUHINI GROUP (uTs)

Within southern 104M/8 a continuous section of Stuhini Group sediments and volcanic rocks displays units that can be correlated for tens of kilometres. The entire package represents a transition from coarse terrigenous sediments to submarine mafic volcanics. As the volcanic piles built, they became more felsic and less voluminous. After the end of volcanism, epiclastic and reefal carbonate deposition covered and preserved the volcanic piles. The culmination of carbonate deposition marks the top of Stuhini strata (Figure 1-18-3).

BASAL CONGLOMERATE (uTSc)

A basal conglomerate can be correlated intermittently from at least the south end of Atlin Lake to the British Columbia–Yukon border. Bultman (1979) termed these rocks Unit A of the Stuhini Group and recognized a similarity to the basal King Salmon Formation in the Tulsequah area (Souther, 1971). Where best developed it attains a thickness of at least 800 metres although its thickness is probably variable due to paleotopographic effects. Identifiable clasts range in size up to 2 metres, but are typically in the 2 to 20-centimetre range and invariably well rounded. The concentration of different clast types varies locally with intrusive, volcanic and metamorphic clasts prevalent.

No fossils have been identified in these rocks so their age is not precisely known, however, they sit unconformably atop potassium feldspar megacrystic hornblende granodiorites that elsewhere have yielded K-Ar (hornblende) and U-Pb isotopic dates of 212 to 220 Ma (BCGSB unpublished data; Bultman, 1979; Hart and Pelletier, 1989). Metamorphic clasts, in order of abundance, include muscovite-biotite schists (no aluminosilicates were observed) and phyllites, chlorite-muscovite schists, amphibolitic gneisses (Wann River gneiss) and rare marble.

Foliated intrusive clasts such as Hale Mountain hornblende granodiorite are locally abundant. Strongly foliated hornblende gabbro and diorite clasts derived from an older body are sparse but conspicuous. Potassium feldspar megacrystic hornblende granodiorite clasts are very evident due to their light colour (Plate 1-18-1a), but close to the parent body they may comprise almost 100 per cent of the outcrop, making it difficult to discern the sediment-intrusive contact. Foliated to unfoliated leucogranitic clasts are probably derived, to a large extent, from pegmatite and aplite dikes within Wann River gneiss and Hale Mountain hornblende granodiorite.

Volcanic clasts include pyroxene and feldspar-porphyritic varieties. Source terrains of two ages are probable as clasts derived from epidote-chlorite-actinolite-altered volcanic conglomerate indicate a previously eroded and hydrothermally metamorphosed volcaniclastic succession (*i.e.*, second generation conglomerates, Plate 1-18-1b). On the north

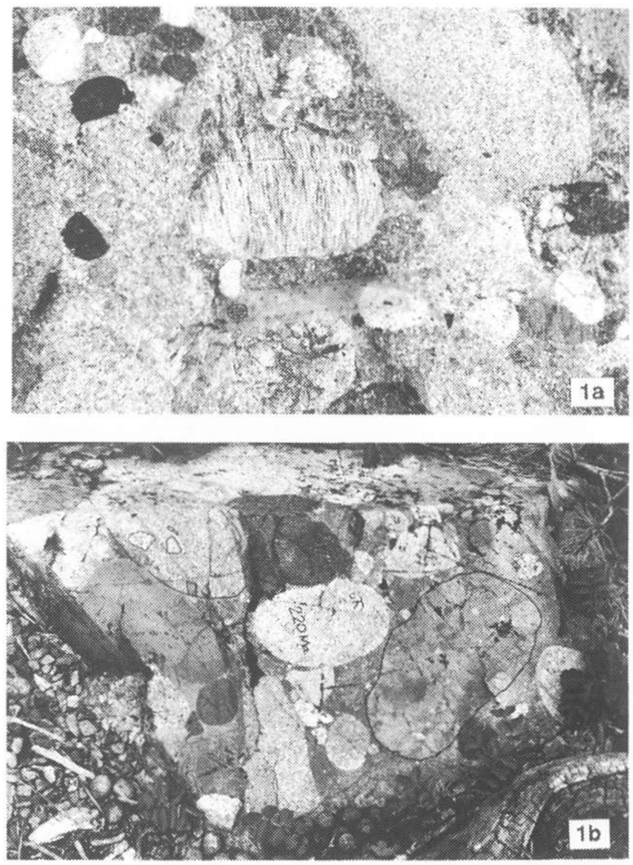


Plate 1-18-1. Upper Triassic basal Stuhini Group conglomerates: (a) clasts are composed of Hale Mountain intrusive, Wann River gneiss (well banded, centre), Late Triassic granodiorite, pegmatite and pyroxene-phyric volcanics in a medium to coarse-grained volcanic matrix. Large clast in centre is about 12 centimetres in maximum dimension; (b) a wave-washed boulder containing second generation cobble conglomerate (outlined) and epidote-altered lapilli tuffs (angular fragments outlined), as well as Late Triassic potassium feldspar megacrystic granodiorite.

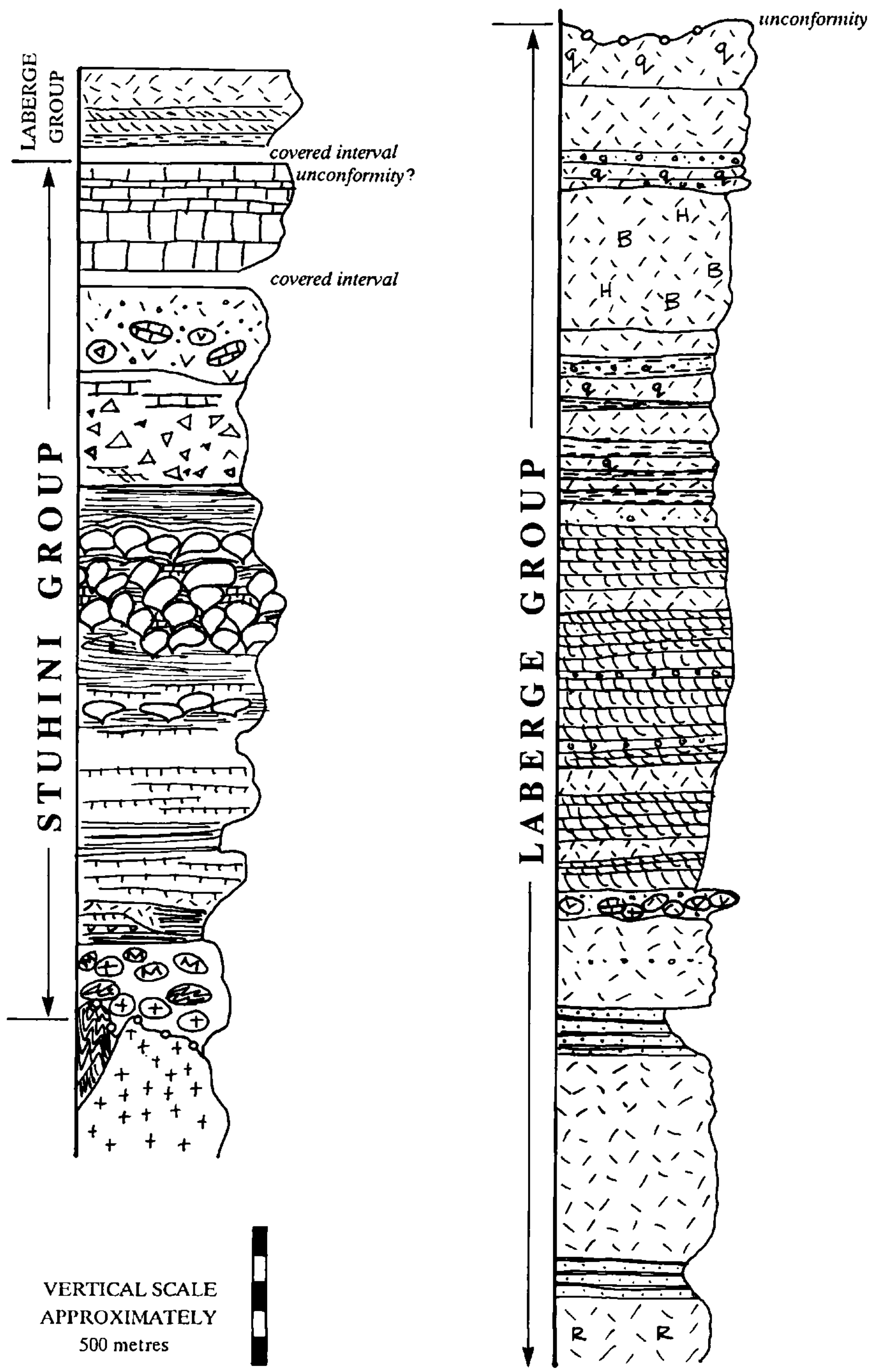


Figure 1-18-3. Generalized stratigraphic columns for the Stuhini and Laberge groups.

shore of Willison Bay a conglomerate unit 500 metres thick is perceptibly zoned; volcanic and intrusive clasts dominate the lower portions whereas metamorphic fragments become more abundant upwards.

Both matrix and clast-supported conglomerates are common. They are normally massive, comprised of lensoid or sheet-like subunits.

SUBAQUEOUS PYROXENE-PHYRIC FLOWS AND INTERLAYERED SEDIMENTS (uTSpx)

Sediments transitional between the conglomerates and this volcanic-dominated unit are generally disrupted, fine-grained, grey-green cherty wackes. They may contain crude layers of andesitic or basaltic blocks (pillow breccia?). At several localities these rocks give way to a pyritic sharpstone conglomerate comprised dominantly of bleached, pebble-sized, angular metamorphic clasts and mica-rich matrix.

Pyroxene-phyric flows (mapped as 'Unit B' by Bultman, 1979) are either massive or pillowed as exposed in resistant, dark grey to black outcrops. Massive flows may be over 20 metres thick, but 2 to 10 metres is more typical. Pillows are generally 0.3 to 2 metres in diameter and commonly have interpillow laminated micrites. Vesicular pillow and flow interiors are plagioclase and pyroxene porphyritic. Locally these rocks are crosscut by gabbroic dikes.

Interlayered argillaceous siltstones drape pillowed flow units; their upper surfaces are disrupted by succeeding volcanic units. Finely ribbed bivalves (*Halobia*?) and carbonized plant material are abundant. These units mark brief episodes of local volcanic inactivity. They are generally less than 3 metres thick and finely parallel bedded (displaying rare ripple cross-stratification). Outcrops are typically rusty and recessive even though the sediments are well indurated with a subconchoidal fracture.

PHREATOMAGMATIC PYROXENE-PHYRIC BRECCIA (uTSpb)

Dusty green, poorly lithified, monolithic breccias form a conspicuous unit above the flows. The change is abrupt, although the actual contact is not exposed. Black pyroxene-rich blocks, set in a matrix of dusty green crystal-ash tuff, range in size up to 0.5 metre, but are generally less than 20 centimetres. These breccias are both clast and matrix supported. These rocks have the same general bulk composition as the underlying unit (uTSpx) and in combination are suggestive of a volcanic pile building to within about 300 metres of the surface where steam generated eruptions occur (Tanakadote, 1935). The abrupt contact between flows and breccias is, however, not supportive of a slowly changing physical condition such as decreasing pressure (water depth) as an explanation for the change in rock type.

Within the map area this lithology is recognized only in the Willison Bay area and corresponds to 'Unit C' of Bultman (1979).

VOLCANICLASTIC UNIT (uTSvc)

Quartz-rich volcanic sandstones crop out at several localities above the pyroxene-phyric breccias, but are not continuous and probably reflect small disconnected basins of deposition. Thickest sections are in central 104M/8 where these rocks attain a thickness of about 800 metres. Planar bedding or shallow, large-scale (several metres) trough crossbedding are common. On southern Copper Island their lower contact is with Unit uTSpb where blocks of the breccia are redeposited with the epiclastics. In upper parts of the unit, carbonate pebbles to large boulders are common, as are crowded feldspar and hornblende porphyry clasts which are lithologically identical to the finer grained parts of the megacrystic hornblende granodiorite. Disrupted, volcaniclast-rich, bioclastic carbonate layers, common within the volcaniclastics, mark the transition to dominantly carbonate sedimentation. Fossils include corallites, colonial corals, bryozoa and bivalves.

SINWA LIMESTONE (uTSs)

These fossil-poor light grey, massive to less commonly well-bedded and argillaceous carbonates have been dated as Late Triassic on the basis of micro and macrofossils. Extensive calcite veining and internal deformation are common, especially in massive beds. This carbonate-rich horizon occurs discontinuously for over 320 kilometres; from a major reef buildup in the Tulsequah map area (type locality; Souther, 1971), through the Tagish Lake area (Bultman, 1979), to patch reefs in Yukon (*e.g.*, Lime Peak; Reid and Tempelman-Kluit, 1987).

The contact between the Upper Triassic Stuhini Group and the Lower Jurassic Laberge Group is most closely constrained on the southwest corner of Copper Island in south Atlin Lake and immediately to the north along the southeast face of the Cathedral. Here greywackes and argillites of the Lower Jurassic Laberge are stratigraphically separated from the Upper Triassic Sinwa carbonate by what is interpreted as a disrupted erosional unconformity. A hiatus is evidenced by disparate ages obtained from fossils in the Sinwa Formation and the overlying Laberge greywackes. The youngest age determination of the Sinwa Formation is uppermost Norian (conodonts from carbonates within 104M/15, identified by M.J. Orchard, Geological Survey of Canada). The oldest fossils collected from overlying Laberge rocks in the area are ammonites of Sinemurian age (identified by H.W. Tipper, Geological Survey of Canada).

GRAHAM CREEK IGNEOUS SUITE AND PENINSULA MOUNTAIN VOLCANIC-SEDIMENTARY SUITE (MGum, MGg; MPv, s)

These rock packages of uncertain age and affiliation crop out between Graham Creek and Sunday Peak in 104M/9. The Graham Creek igneous package includes tectonized harzburgites and gabbros that have a close recurrent spatial relationship with cherts, pillow basalts and succeeding felsic volcanics and epiclastics of the Peninsula Mountain suite. One interpretation of these rock associations is that they represent a dismembered ophiolitic suite, possibly a part of the Cache Creek Terrane.

"Peninsula Mountain" is the name used by Bultman (1979) to describe "a pre-Laberge sequence" of grey-green tuff and tuff breccia and minor, locally interbedded chert and siltstone that crops out along the northeastern edge of the Whitehorse trough and is well exposed on Peninsula Mountain (10 kilometres northwest of Sunday Peak). Since this description closely resembles what is seen in the map area, this terminology has been retained.

Graham Creek igneous rocks have tectonized harzburgite at their base (Plate 1-18-2a) or serpentinite that crops out on the north shore of Graham Inlet 2.5 kilometres east of the mouth of Graham Creek and again along strike in the headwaters of Shaker Creek. Another structurally unrelated sliver occurs on the southeast flank of Sunday Peak. They are bright orange and hackley weathering, strongly sheared and quartz-carbonate-mariposite altered. Maximum thickness is about 30 metres including serpentinized equivalents. Altered gabbros crop out on the southwest flank of Table Mountain structurally above, but not in contact with, the harzburgite.

The Peninsula Mountain volcanic-sedimentary suite sits structurally above the gabbroic rocks. Strongly deformed, well-bedded wackes and cherts are assumed to have a conformable contact with a thick pillow basalt succession since, at two localities, the basalts contain black inter-pillow cherts.

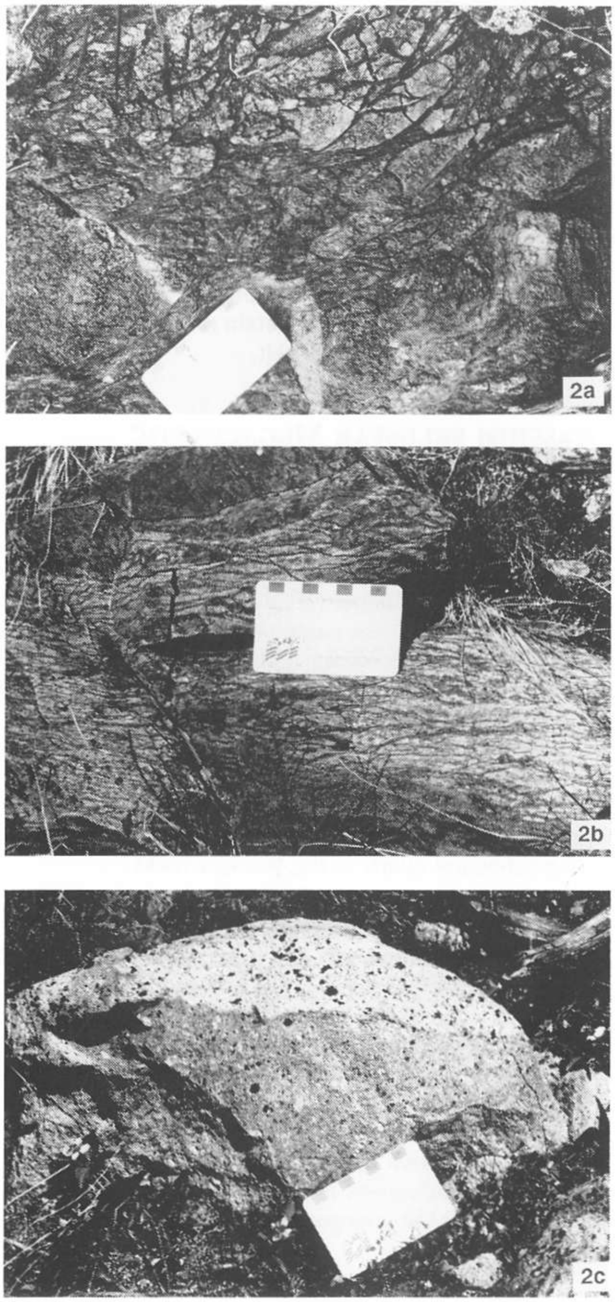


Plate 1-18-2. Mylonitization and admixture of harzburgites shown in (a) with a coarse quartz feldspar porphyry dacitic plug pictured in (c) to produce cataclasite (b).

Contact relationships with the overlying felsic volcanic suite are not easily determined, but intermediate hyaloclastic rocks at the base of the felsic unit suggest initial deposition under subaqueous conditions and felsic volcanism may therefore have been continuous with basalt deposition. The upper volcanics are mainly intermediate to rhyolitic, medium-grained feldspar porphyry breccias and subordinate flows (Plate 1-18-3). Sparse feldspar-biotite ash flows with or without quartz are volumetrically minor but distinct.

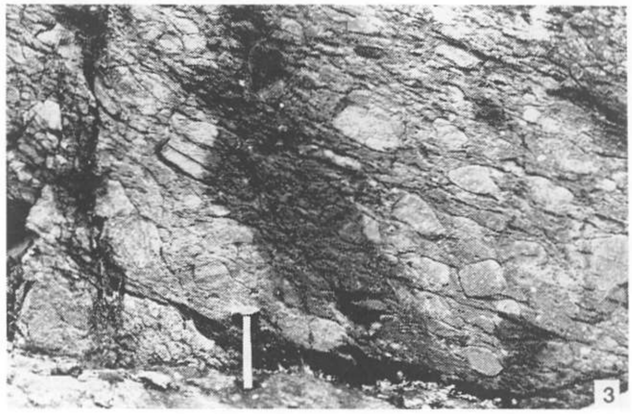


Plate 1-18-3. Coarse, flow-banded rhyolitic feldspar porphyry breccias of the Peninsula Mountain volcanic suite.

LABERGE GROUP (lJL)

Many of the units within the Laberge Group sediments have a limited facies-dependent distribution, which is interpreted to be one of coalescing subaqueous turbidite fans (Bultman, 1979). A ministry project is currently in progress to determine facies and subfacies within these sediments.

ARGILLITES (lJL$_a$)

Laberge Group argillites are divided into two major types. **Irregular, thinly interbedded** brown to black argillites of variable thickness (most commonly millimetres to decimetres thick) generally occur as sets within wacke-dominated successions. **Rhythmically bedded** argillites form successions low in the Laberge stratigraphy that are 10 to 100 metres or more thick. Beds are 2 to 5 centimetres thick and grade from silty, light-coloured bases to dark, argillaceous tops.

GREYWACKES (lJL$_g$)

Greywackes are the dominant rock type within the Laberge Group. A major focus of mapping was to divide wackes into feldspathic, quartz-rich and lithic-rich varieties with further subdivisions based on the type and abundance of rock fragments and mafic minerals present.

Feldspathic wackes are the most abundant type and typically occur in massive or well-bedded units with well-defined normal grading. Individual beds vary considerably in thickness, ranging from 5 centimetres to 10 metres or more.

Where bedding is distinguishable, it typically consists of a base of granule conglomerate to very coarse grained wacke with 75 to 90 per cent feldspar grains (the remainder consists of variable amounts of quartz and rock fragments, dominantly altered biotite or hornblende grains, and a muddy matrix) grading upward into a fine to very fine grained feldspathic wacke which rarely exhibits trough cross-stratification and ripples. The weathering colour of feldspathic wackes is a greenish grey or orange and they are resistant compared to adjacent argillite beds which are rusty brown to black.

Lithic-rich wackes are subordinate to feldspathic wackes. They usually occur as beds 10 to 100 centimetres thick, interbedded with argillites (*see* irregularly and thinly bedded argillites above). Overall, lithic-rich wackes are finer and less variable in grain size than feldspathic or quartz-rich wackes.

For a more thorough description of Laberge Group lithologies the reader is referred to Mihalynuk *et al.* (1989a).

TAGISH VOLCANICS (MOUNT SWITZER AND ENGINEER MOUNTAIN SUITES, $muKT_v$)

Rhyolitic through basaltic flows and pyroclastic sequences exist as erosional remnants above metamorphic and trough strata. Contacts are both unconformable and fault bounded. Within 104M/8 these rocks are best exposed in the Mount Switzer and Engineer Mountain areas and, prior to Quaternary erosion, probably blanketed the intervening areas as well. Correlation between the these areas is based upon the common occurrence of two widespread units. **A light green heterolithic lapilli tuff** containing conspicuous white rhyolite fragments up to 15 per cent and variegated aphyric to medium-grained feldspar porphyry clasts. It is recessive and platey weathering. Isolated basalt "blocks" up to 10 metres wide are common and probably represent cross-sections through channelled flows. **Black monolithologic feldspar porphyry breccias and tuffs** are possibly the more widespread unit. They are resistant, weathering into large blocks. Feldspars are medium grained and tan-grey on weathered surfaces.

These rocks are older than the Engineer stock which crosscuts them. This stock is presumed coeval with compositionally similar intrusions (for example, just to the north on Bee Peak) which have yielded a date of 80.3 ± 2.4 Ma, suggesting that the volcanics are older than Late Cretaceous. South of western Tagish Lake, granites that crosscut the Mount Switzer volcanics have been dated as 64.1 ± 1.3 Ma (both dates are K-Ar determinations on hornblende separates by Bultman, 1979).

HUTSHI/MOUNT FETTERLY VOLCANICS ($muKT_v$?)

These rocks cap Mount Fetterly and the low mountains to the west. They are probably correlative with the Engineer suite, based upon basaltic flow units that display identical textural and weathering characteristics. At their base, well-developed conglomerates mainly composed of underlying Laberge sediments crop out in paleotopographic lows. Conglomerates grade upwards to olive-green, well-bedded epiclastic rocks rich in plant fossils. Above these are white to rusty weathering, massive, aphyric to quartz and feldspar-phyric rhyolites (at one locality an ignimbrite marks this horizon) that may be several hundred metres thick. Not observed in contact, but presumably overlying, are fractured, brown-weathering, feldspar-porphyritic to aphyric basalts that break to a black, conchoidal surface.

INTRUSIVE ROCKS (oldest to youngest)

VARIABLY FOLIATED HORNBLENDE-RICH GABBRO (lThg)

A strongly foliated to unfoliated diorite to hornblende gabbro crops out on the southwest flank of the Cathedral. In the strongly foliated zones it is reduced to a chlorite schist, but normally surviving igneous textures identify it as a composite intrusive. Abrupt variations in the plagioclase content, on an outcrop scale, and the crosscutting of earlier, more highly foliated zones by less foliated material, indicate emplacement within a structurally active zone at moderate to shallow depth. Intrusives with strikingly similar features occur within the Hogem (Garnett, 1978), Iron Mask (Kwong, 1987) and Copper Mountain (Preto, 1972) bodies; all in the Nicola-Takla-Stuhini belt.

POTASSIUM FELDSPAR MEGACRYSTIC HORNBLENDE GRANODIORITE (lTgd)

Grey-weathering granodiorites, white, pink or tan on fresh surfaces, occur in a linear belt on the eastern side of the Llewellyn fault. These are equivalent to the Bennett Range granite of Hart and Pelletier (1989) which lies to the west of the fault. A weak or, less commonly, moderate foliation is locally developed. Potassium feldspar megacrysts (up to 4 centimetres) may be weakly perthitic and commonly contain concentric zones of plagioclase and hornblende poikilocrysts. According to Hart and Pelletier these rocks can be differentiated from younger granodiorites sharing many of the same characteristics on the basis of the lack of foliation, more perthitic and more lightly coloured megacrysts, and smokey subhedral quartz in the younger rocks.

At its southeastern margin near Splinter Peak, this body loses its holocrystalline texture for a hypabyssal texture with 10 to 15 per cent megacrysts in a matrix of 60 per cent 0.5-centimetre plagioclase phenocrysts and an aphanitic grey to pink groundmass. Volcaniclastic fragments almost identical to these border granodiorites are commonly observed within Unit uTSvc, raising the question of whether the intrusive had extrusive equivalents. If so, then how did these extrusive clasts become entrained high in a stratigraphic pile that sits unconformably above the potassium feldspar megacrystic hornblende granodiorites? Presumably compositionally and texturally similar but younger intrusive bodies had volcanic equivalents.

TECTONIZED WANN RIVER HORNBLENDE DIORITE (Kd)

In the Wann River valley an elongate foliated diorite body crops out for about 500 metres. It is dark green when fresh and white or red on weathered surfaces, with local brittly deformed zones that may be pervasively crosscut by quartz veinlets comprising up to 3 per cent of the rock. It is considered a syntectonic body emplaced along the Llewellyn fault, probably in the Cretaceous, and is perhaps coeval with the diorites described below.

ZONED GRANODIORITE – DIORITE BODIES (lKgd, lKd; CATHEDRAL MOUNTAIN, ENGINEER MOUNTAIN AND SUNDAY PEAK)

A zoned intrusive body, 2 kilometres long, crops out on the southwest flank of Engineer Mountain and a second body, 10 kilometres long, underlies much of the Cathedral. Smaller related dioritic bodies probably include those on the east flank of Engineer Mountain, the south ridge of Bee Peak, Mount Cameron, the east side of southern Edgar Lake and Sunday Peak. A K-Ar (hornblende) age of 80 ± 3 Ma for tonalite from Bee Peak has been determined by Bultman (1979) who included these rocks with intrusives in the Whitehorse trough.

The largest of these bodies contains at least two of the following four phases: an orange-weathering, olive-brown to greasy grey, fresh, medium to coarse-grained hornblende biotite diorite to "anorthositic" biotite diorite, most commonly occurs as a border phase; tan to salmon, platey weathering, sparse potassium feldspar and rare quartz-phyric rhyolite occurs as irregular zones or dike-like bodies with sharp or digested margins; varitextured fine to medium-grained hornblende granodiorite irregularly admixed with the second variety and locally containing abundant biotite-hornblende-rich mafic xenoliths; and white to pink, fine to medium-grained hornblende granodiorite to tonalite with subhedral plagioclase and hornblende (1-5 millimetres and 2-7 millimetres respectively) with mainly interstitial quartz and potassium feldspar. Evenly distributed patches of finely intergrown acicular hornblende and plagioclase 0.5 to 2 centimetres in diameter are characteristic of the Cathedral body.

COAST INTRUSIONS (lKg)

Coast intrusions crop out along the western margin of 104M/8 and were probably emplaced as two separate plutons, one underlying the south end of Tagish Lake and the other in the southwest corner of 104M/8. The south Tagish body is pink to white, medium to coarse grained and may contain up to 5 per cent potassium feldspar megacrysts up to 4 centimetres long. A xenolith-rich chilled margin is observed at the contact with Mount Switzer volcanics. Elsewhere the contact zone lacks xenoliths, but may be peraluminous, bearing 1 to 2 per cent fine garnets. The southwestern pluton is very homogeneous with only the slightest internal variations. It is tan to pink, blocky weathering, medium-grained biotite granite generally containing several per cent perthitic potassium feldspar 1 to 2 centimetres long. At its southeast margin it is in chilled contact with hornblende biotite diorite that forms the northern extension of a large body to the south.

DEFORMATION

FAULTS

The structure of the area is dominated by the northwest-trending Llewellyn and Nahlin faults; both involve basement rocks and crop out within the map area as disaggregated brittly deformed zones containing various lithologies and their sheared equivalents.

Several lines of evidence suggest that within the map area the Llewellyn fault is a long-lived, dextral, west-side-up transcurrent structure. It probably acted as a basin-bounding fault to the Whitehorse trough. Uplift on this structure most likely began at least as early as the Triassic, as the metamorphic and igneous source-terrain to the west was exumed to supply detritus to conglomerates of Unit uTSc. A fabric was locally imparted in synkinimatic intrusives such as the variably foliated hornblende-rich gabbros (*see* INTRUSIVE ROCKS) that crop out adjacent to the Llewellyn fault. Latest motion is approximately mid-Cretaceous as it cuts volcanics overlying Lower Jurassic Laberge sediments in the Tutshi Lake area, but is plugged north of the British Columbia–Yukon border by the Pennington pluton (Hart and Pelletier, 1989; Hart, written communication, 1989) which is interpreted to be of Late Cretaceous age.

Metamorphic rocks west of the fault display top-to-the-south, partly annealed, plastic deformation (Plate 1-18-4a, b). A later dextral and east-side-down brittle deformation also affects the rock (Plate 1-18-4c). This late event is probably related to motion on the Llewellyn fault, but its age is not known. An undeformed pegmatite is both crosscut and invades these minor fault planes.

Three surface exposures of highly deformed rocks, including tectonized harzburgite, are interpreted to lie along the trend of the Nahlin fault within 104M/9E (except at Sunday Peak). Fabrics within this zone are all near vertical and strike 120°. Well-developed shear bands in a mylonite zone (Plate 1-18-2b) indicate local intense cataclasis. Shear sense, however, could not be unequivocally determined because of the lack of a consistent lineation. This fault crosscuts the Laberge and Graham Creek rocks and is plugged by the Birch Mountain pluton which has yielded K-Ar ages of 56 ± 6 and 48 ± 3 Ma (on biotite and hornblende respectively; Bultman, 1979).

FOLDS

Folding affects all pre-Cretaceous rocks in the map area. Megascopic folds in all major lithologies, except the chlorite-actinolite schists of the Boundary Ranges metamorphic suite, are dominated by upright to overturned, open to closed and gently plunging fold styles. These generally have northwest-trending axial surfaces (Figure 1-18-4). Mainly east-trending, broad folds affecting the Laberge Group rocks are thought to postdate the northwest folds, but these structures are poorly known.

METAMORPHISM

Nine samples containing garnet and biotite were selected from the low-grade zone within the 1988 map area for the

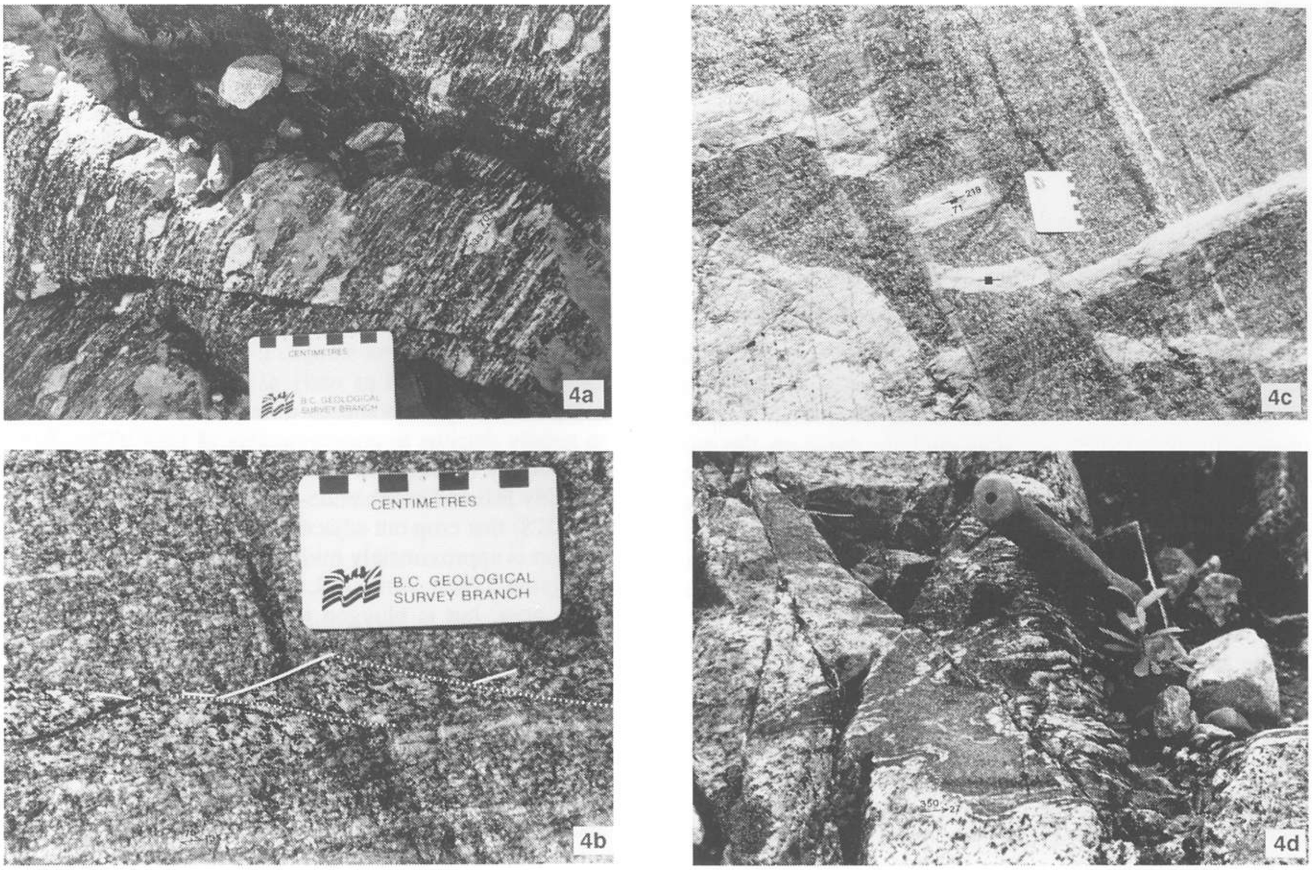


Plate 1-18-4. Deformation within the Hale Mountain hornblende granodiorite: (a) sigma-type augen indicating top-to-the-left (south) motion. A well-developed mineral lineation outlined by hornblende is approximately parallel to the outcrop surface (view is to the west, the foliation is 135/70); (b) synthetic (dashed) and antithetic (dotted) fault surfaces in coarse Hale Mountain intrusive phase are consistent with a sinistral shear sense; (c) pervasive brittle faults show east-side-down and dextral displacement; (d) mafic-rich layers within the Hale Mountain intrusive locally display tight to isoclinal folds with axes parallel to the regionally developed mineral lineation (here 350/27).

purpose of Fe-Mg exchange geothermometry. Thin-section studies show that four of these samples are too retrograded to be worth analyzing. Microprobe analyses were conducted on the remaining five samples (Table 1-18-1), but one of them yielded biotites with low K_2O and high SiO_2; clearly altered and unlikely to produce meaningful temperatures (not listed in Table 1-18-1). In all cases, the samples displayed some degree of replacement of biotite by chlorite. One sample (MM 27-6) contained mainly unaltered biotites. In another sample (MM 11-1), the analysis of garnet and biotite in textural equilibrium was not possible. Calculated temperatures based on the calibration of Ferry and Spear (1978) and Thompson (1976) and the average composition of garnet-biotite pairs, yields a systematic core-to-rim temperature decrease from 592°C (576°C) to 557°C (547°C) for MM 27-6. In sample MM 11-1, analyses of garnet rims and isolated interstitial biotite grains ('INTER' on Table 1-18-1) yielded temperatures of 562°C (551°C). Both LC 11-1A and MM 14-2B yield similar temperatures. However, these temperatures should only be considered approximate as the titanium-aluminum content of the biotites and the calcium-manganese content of the garnets fall outside the compositional limits that are considered to be a reliable uncorrected biotite-garnet geothermometer. Actual temperatures probably fall within the ±75°C error limits of the Ganguley and Saxena (1984) 'corrected' geothermometer which, although not tabulated in Table 1-18-1, were calculated and found to be 20° to 70°C higher than the temperatures obtained from the Thompson calibration (in parentheses). In any case, such temperatures indicate amphibolite-grade metamorphism; they are incompatible with the greenschist to transitional greenschist-amphibolite grades implied by the present mineral assemblage.

Low-grade metamorphic minerals, particularly chlorite and actinolite, display both syn and predeformation textures in that they may be aligned or bent by the last phases of folding. Late folds (F_3 of Mihalynuk *et al.*, 1988b, 1989a) within the metamorphic rocks are approximately coplanar with the main phase of folding within the Laberge Group strata of the Whitehorse trough (Figure 1-18-4). Similarly, near the British Columbia–Yukon border, metamorphic rocks and unconformably overlying Laberge sediments are folded together about an axis parallel to the Whitehorse trough structural grain. Unless these folds have been transposed, F_3 could be of Middle Jurassic to Early Cretaceous age. [Volcanics interpreted to rest disconformably atop Toarcian sediments are folded, but younger Montana Mountain volcanics (*circa* 90 Ma, Hart and Pelletier, 1989) apparently

TABLE 1-18-1
COMPOSITIONS OF (A) BIOTITE AND (B) GARNET FROM BOUNDARY RANGES METAMORPHIC SUITE RETROGRADED GARNET-BIOTITE-MUSCOVITE-CHLORITE-ACTINOLITE SCHISTS.

Table 1-18-1a **MICROPROBE ANALYSES: BIOTITE**

SAMPLE		LC11-1A		MM11-1	MM14-2B		MM27-6		
N		5	6	5	10	4	4	4	4
place		RIM	INTER	INTER	RIM	INTER	CORE	MIDDLE	RIM
WEIGHT PER CENT OXIDES									
Oxide	DL								
SiO2	0.06	34.75	34.50	36.02	35.47	36.51	35.87	35.91	35.91
TiO2	0.03	1.39	1.28	1.72	1.41	1.42	2.79	2.78	1.69
Al2O3	0.07	17.43	17.28	17.16	18.11	18.09	17.04	17.00	17.03
FeO*	0.04	23.31	23.05	21.47	23.22	22.50	22.57	21.98	20.98
MnO	0.05	0.10	0.10	0.07	0.12	0.10	0.34	0.25	0.28
MgO	0.03	10.50	10.87	10.42	8.98	8.84	8.35	9.23	9.94
CaO	0.01	0.07	0.06	0.02	0.04	0.18	0.10	0.05	0.05
Na2O	0.02	0.21	0.11	0.16	0.03	0.28	0.03	0.03	0.05
K2O	0.01	8.00	7.99	9.19	8.76	8.98	9.43	9.47	9.64
BaO	0.11	0.00	0.00	0.20	0.00	0.00	0.39	0.35	0.36
Cl	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
F	0.05	0.18	0.20	0.21	0.22	0.23	0.19	0.19	0.21
Total (anhydrous)		95.93	95.44	96.64	96.36	97.13	97.10	97.24	96.14
NUMBER OF ATOMS BASED ON 12 O + (OH+F+Cl)									
Si		5.40	5.39	5.54	5.49	5.59	5.54	5.52	5.57
Ti		0.16	0.15	0.20	0.16	0.16	0.32	0.32	0.20
AlIV		2.60	2.61	2.46	2.51	2.41	2.46	2.48	2.43
AlIII		0.59	0.57	0.65	0.80	0.85	0.64	0.60	0.68
Fe*		3.03	3.01	2.76	3.01	2.88	2.92	2.83	2.72
Mn		0.01	0.01	0.01	0.02	0.01	0.04	0.03	0.04
Mg		2.43	2.53	2.39	2.07	2.02	1.92	2.12	2.30
Ca		0.01	0.01	0.00	0.01	0.03	0.02	0.01	0.01
Na		0.06	0.03	0.05	0.01	0.08	0.01	0.01	0.02
K		1.59	1.59	1.80	1.73	1.75	1.86	1.86	1.91
Ba		0.00	0.00	0.01	0.00	0.00	0.02	0.02	0.02
OH		3.63	3.65	3.60	3.62	3.58	3.61	3.59	3.62

Biotite-garnet Fe-Mg exchange thermometer results are listed at the bottom of (b).

Table 1-18-1b **MICROPROBE ANALYSES: GARNET**

SAMPLE		LC11-1A		MM11-1			MM14-2B		MM27-6		
N		6	8	4	4	5	2	8	4	4	4
place		MIDDLE	RIM	CORE	MIDDLE	RIM	MIDDLE	RIM	CORE	MIDDLE	RIM
WEIGHT PER CENT OXIDES											
Oxide	DL										
SiO2	0.06	36.95	37.15	37.66	37.56	37.13	36.98	37.02	37.23	37.41	37.31
TiO2	0.03	0.08	0.06	0.10	0.06	0.04	0.16	0.08	0.25	0.21	0.13
Al2O3	0.08	20.76	20.97	20.83	20.91	20.62	21.03	21.17	20.65	20.42	21.08
FeO	0.04	31.77	31.96	29.99	31.17	31.90	30.43	31.96	27.49	27.94	28.66
MnO	0.06	3.04	2.35	4.60	3.44	2.81	3.93	2.12	5.27	4.28	2.98
MgO	0.02	2.37	2.60	2.40	2.62	2.61	1.94	2.37	2.00	2.10	2.27
CaO	0.01	4.68	4.57	5.15	4.96	4.54	5.47	4.97	6.81	7.16	7.43
Total		99.64	99.66	100.73	100.72	99.65	99.92	99.69	99.70	99.52	99.86
NUMBER OF ATOMS BASED ON 12 O											
Si		2.99	2.99	3.01	3.00	3.00	2.98	2.98	3.00	3.01	2.99
Al		1.98	1.99	1.96	1.97	1.96	2.00	2.01	1.96	1.94	1.99
Ti		0.00	0.00	0.01	0.00	0.00	0.01	0.00	0.02	0.01	0.01
Fe2		2.15	2.15	2.00	2.08	2.15	2.05	2.15	1.85	1.88	1.92
Mg		0.28	0.31	0.29	0.31	0.31	0.23	0.28	0.24	0.25	0.27
Mn		0.21	0.16	0.31	0.23	0.19	0.27	0.14	0.36	0.29	0.20
Ca		0.41	0.39	0.44	0.42	0.39	0.47	0.43	0.59	0.62	0.64
Fe3+		0.02	0.01	0.04	0.03	0.04	0.00	-0.01	0.04	0.06	0.01
MOLE FRACTION END MEMBERS											
Xalm		0.70	0.71	0.66	0.68	0.70	0.68	0.72	0.61	0.61	0.63
Xprp		0.09	0.10	0.09	0.10	0.10	0.08	0.09	0.08	0.08	0.09
Xsps		0.07	0.05	0.10	0.08	0.06	0.09	0.05	0.12	0.10	0.07
Xgrs		0.13	0.13	0.13	0.13	0.12	0.16	0.15	0.18	0.19	0.21
Xadr		0.01	0.00	0.01	0.01	0.01	0.00	0.00	0.01	0.02	0.00

GEOTHERMOMETRY RESULTS

lnKD	-1.87	-1.71	-1.79	-1.74	-1.77	-1.74	-1.66	-1.67	-1.70	-1.78
Thompson**	526*	565	545*	557*	551*	558*	578	576	568	547
Ferry and Spear	529*	582	548*	570*	562*	572*	599	592	585	557

N = NUMBER OF ANALYSES (AVERAGED); DL = DETECTION LIMIT

* Temperature in degrees celsius resulting from mineral pairs that DID NOT grow in textural equillibrium (i.e. biotite is interstitial (INTER) and isolated from the garnet)

** Thompson calibration with pressure correction of 4kilobars used

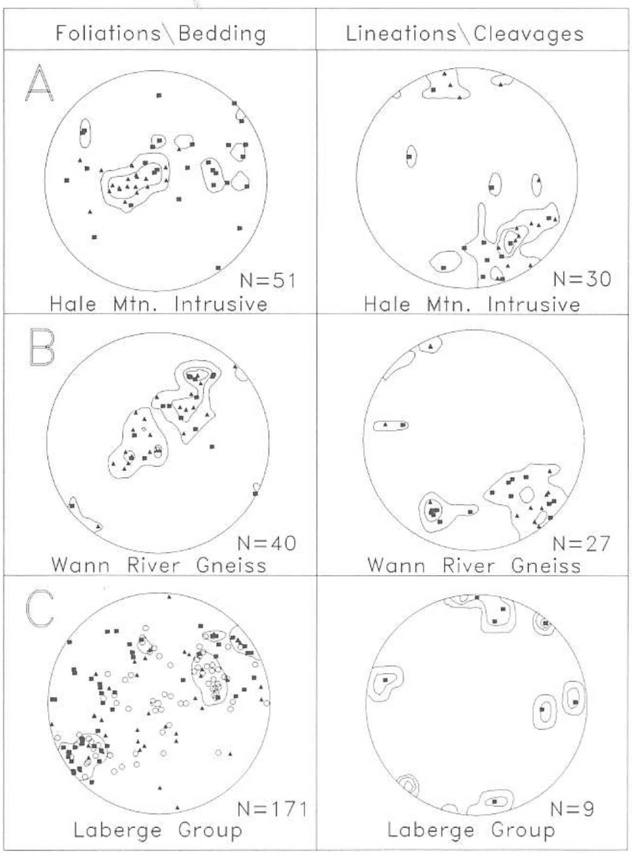


Figure 1-18-4. Lower hemisphere stereo plots showing foliations and mineral lineations within (a) the Hale Mountain granodiorite and (b) the Wann River gneiss and (c) poles to bedding and cleavages within the Laberge Group. Contour interval is five per cent per one per cent area.

are not. F_1 and F_2 must be pre-Late Triassic because foliated metamorphic clasts are found within conglomerates of that age.] F_3 deformation may affect rocks in the chlorite schist package more strongly due to the proximity of the Llewellyn fault, which in combination with extensive plutonism to the west, probably acted as a conduit supplying enough heat-driven fluids to retrograde the region and to facilitate deformation. It is doubtful whether all low-grade metamorphic rocks are a result of retrogression, at least not a retrograde event of post-Middle Jurassic age, since chlorite-muscovite-quartz schists occur as clasts within Upper Triassic conglomerates in the Llewellyn Inlet area.

MINERALIZATION

Exploration activity has increased dramatically in the Tagish Lake area in 1989 (Figure 1-18-5).

Six distinct types of mineralization are recognized within the study area. Occurrences include past producers such as the Engineer mine, a sulphide-poor gold and gold-tellurium-silver-bearing quartz vein and the Ben-My-Chree sulphide-rich gold-silver-bearing quartz vein. Others include gold $\pm$ silver quartz veins with no sulphides, as at the Sweepstake showing; sulphide-bearing calcite and/or quartz veins, and massive sulphide pods like the Anyox-Rodeo showing. The association of carbonatized ultramafic rocks and gold, characteristic of the Atlin camp, may also apply to the Graham Creek area. Quartz-carbonate-mariposite-altered ultramafics crop out within this drainage, which is the westernmost placer stream of the Atlin camp.

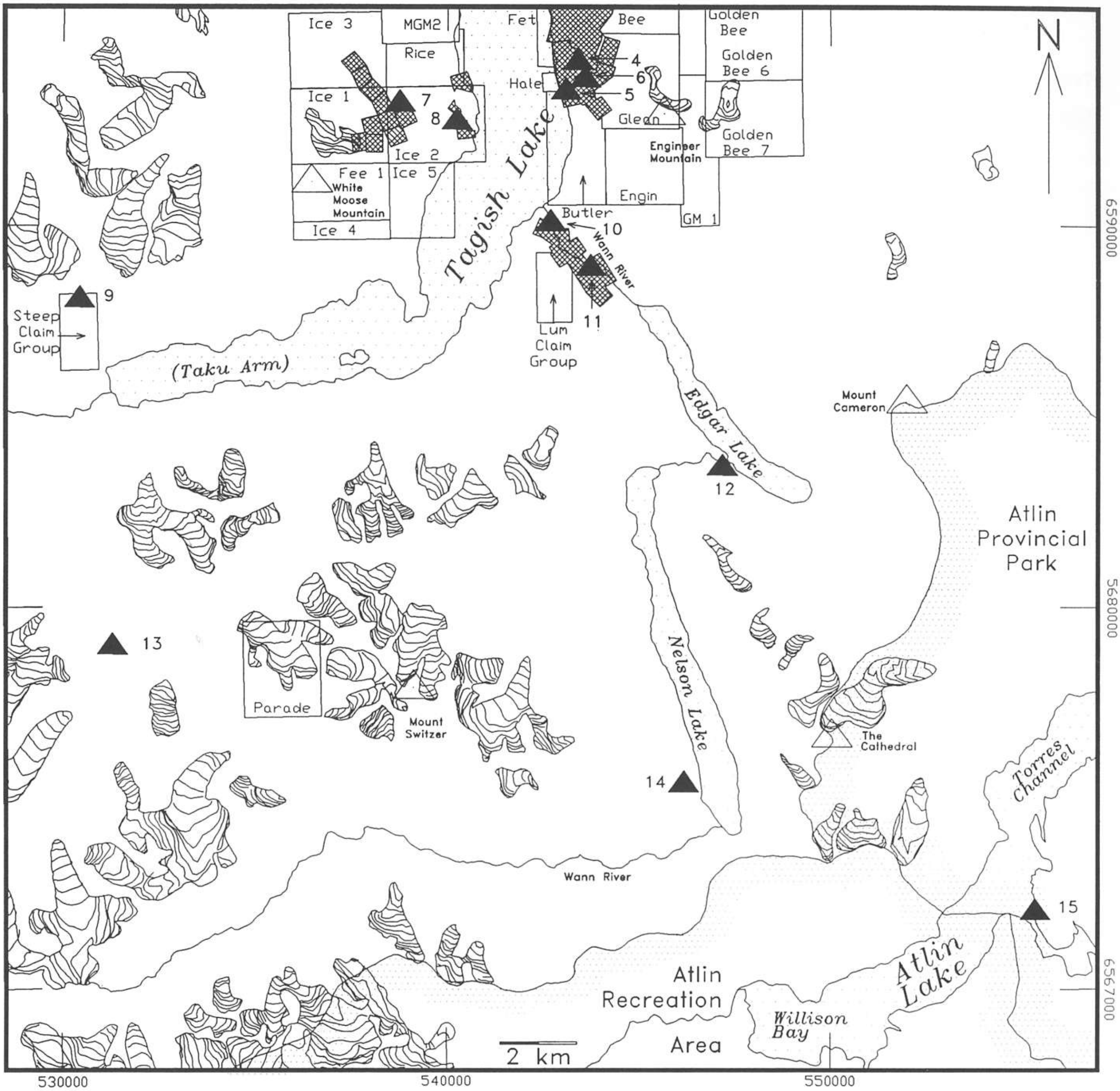


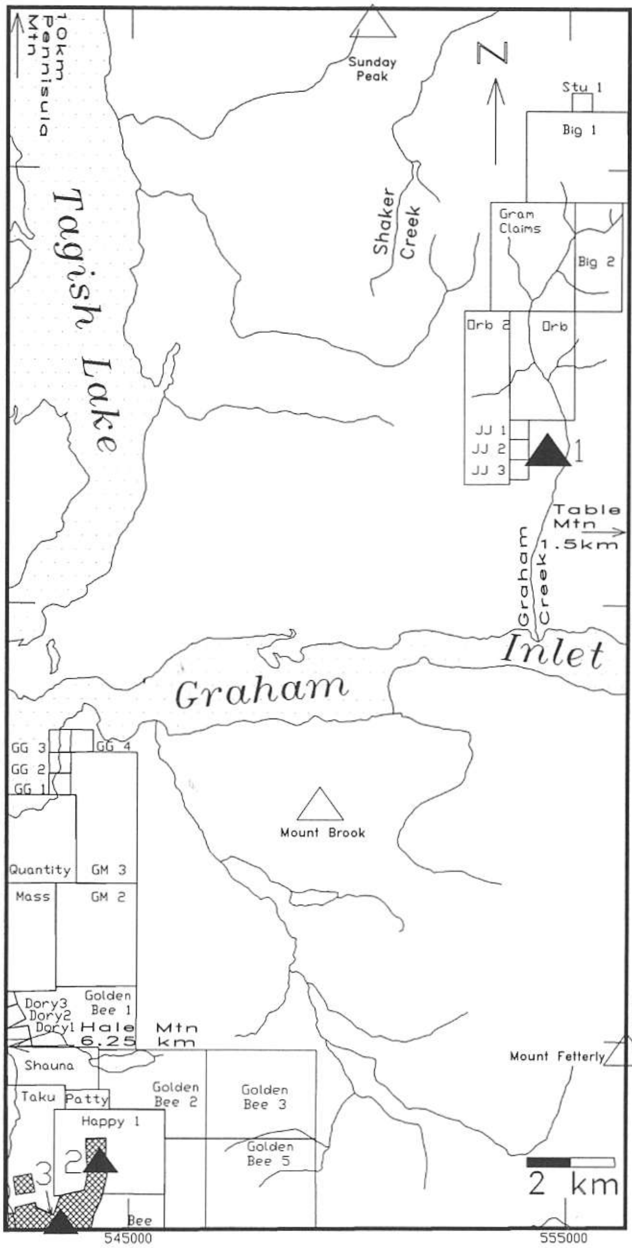
Figure 1-18-5. Claim boundaries as shown on September, 1989 mineral claims map. MINFILE occurrence locations denoted by solid triangles, numbers correspond to those in Table 1-18-2. Crown grants are denoted by the cross-hatched pattern.

Gold and Gold-Tellurium-Bearing Quartz Veins with Trace Base Metals

Historically the Engineer mine is the largest lode gold producer in the area, yielding approximately 560 kilograms of gold with an average recovered grade of 36 grams per tonne between 1913 and 1952. Native gold, tellurides, pyrite and traces of allemontite (SbS), arsenopyrite, mariposite and berthierite needles ($FeS\text{-}Sb_2S_3$), identified by x-ray analysis, occur in quartz-calcite veins. Good comb structures as well as banding and vugs characterize these veins which crosscut Lower Jurassic Laberge argillites and greywackes (Schroeter, 1986).

The Gleaner, Lumsden, Myosotis, Lake View and Taku Chief gold vein occurrences are similar mineralogically and geologically to the Engineer mine and are included as part of the Engineer system (Table 1-18-2).

The Happy Sullivan prospect has a similar mineralogical and geologic setting to that of the Engineer mine, however, arsenopyrite is locally up to 20 per cent and dendritic crystals



of native gold have been found (Table 1-18-2; Schroeter, 1986).

Gold-Silver Quartz Veins and/or Brecciated Zones with Base Metals

Although showings of this type have received significant exploration and development work, historically they have been uneconomic. This mineral occurrence type comprises about one half of all mineral occurrences in the 104M/8 and M/9 map areas (Table 1-18-2) and is restricted to areas adjacent the Llewellyn fault zone (Figure 1-18-5). These occurrences include veins on the White Moose, Rupert and Steep claim groups and the Kim and Nelson Lake showings. Significant associated minerals are galena, chalcopyrite, malachite, pyrite (with or without bornite, tetrahedrite, azurite and pyrrhotite) with minor native gold and silver. Veins on the White Moose and Rupert claim groups crosscut Boundary Ranges metapelites whereas those on the Steep claim group (with up to 450 grams per tonne gold) are hosted by Late Cretaceous granites, as at the Nelson Lake showing.

Gold($\pm$ Silver) Quartz Veins without Base Metals

Veins at the Kirtland and Jersey Lily showings are of this type. They differ from the Engineer veins in that they contain gold and silver but lack tellurides and base metals. Veins at the Kirkland occurrence are just south of the Engineer vein system. The Sweepstake occurrence also shows similar mineralogical and geological characteristics, but differs in that the veins contain no silver.

Base Metal Veins within the Llewellyn Fault Zone

The Brown, Jack Pine, Wann Fraction and Anyox-Rodeo showings are base metal occurrences within the Llewellyn fault zone. An adit 10 metres long, driven along the Brown vein, exposes an anastomosing network of irregular quartz veins and veinlets, ranging from less than 1 to 35 centimetres wide, parallel to the country rock foliation (which is oblique to the main trace of the fault). Gouge material ranges from Boundary Ranges chlorite-actinolite schists to altered granodiorite and Stuhini volcanics. Mineralization consists of tetrahedrite, chalcopyrite, malachite, azurite, molybdenite, pyrite, sphalerite and galena.

Massive Sulphides within the Llewellyn Fault Zone

The Anyox-Rodeo showing is a copper-nickel-platinum-palladium massive sulphide occurrence hosted in Boundary Ranges chlorite-actinolite schist near its contact with probable Upper Triassic Stuhini volcanics. Fractured actinolite porphyroblasts up to 3 centimetres long are accompanied by interstitial or fracture-filling pentlandite, pyrrhotite, chalcopyrite and pyrite. Anomalous platinum and palladium analyses have been reported but have not been confirmed.

Placer

The Graham Creek placer claims are the westernmost known occurrence of placer gold in the Atlin and Tagish Lakes area. Property exploration and development to date have not yielded substantial returns. Extensive geochemical analysis of water samples collected from the Graham Creek drainage basin has consistently yielded anomalous values in gold (1 to 7 parts per trillion). Furthermore, nickel tellurides, chromite and gold grains (with electrum) have been recorded from heavy mineral separates (B. Ballantyne, Geological Survey of Canada, personal communication, 1989; *see* also Hall *et al.*, 1986). Silt samples may yield anomalous arsenic values, but not necessarily with accompanying high gold. Soil samples are generally anomalous with respect to lead. According to Ballantyne, these geochemical observations

TABLE 1-18-2
MINFILE OCCURRENCES

MAP. NO	NAME (MINFILE NUMBER)	COM-MODITY	ASSAY (Reference)	SAMPLE WIDTH	DESCRIPTION
1	Graham Creek (104M 023)	Au			Westernmost placer gold occurrence of the Atlin gold camp.
2	Happy Sullivan (104M 013)	Au Ag	323.6 g/t 226.2 g/t (AR 10511)	Grab	Mineralized quartz veins, up to 0.9 m wide, are located within a shear zone about 24 m wide and 3 km long in Lower Jurassic Laberge greywackes and argillites.
3	Sweepstake (104M 025)	Au	n.a.		
4	Engineer Mine (104M 014)	Au Ag Sb Te	past producer (1913-1952) 15 564 tonnes grading 36.0 g/t Au 17.9 g/t Ag		Sporadic mineralization, mainly gold, occurs in discordant quartz-calcite veins less than 2 m wide. Veins are hosted in Laberge greywacke and argillite and are oriented oblique to local and regional northwest-trending folds.
5	Engineer gold camp Kirkland Group (104M 015)	Au Ag	n.a.		Veins on the Kirkland group are a southerly extension of the Engineer vein system. The main vein, the Jersey Lily, is about 60 cm wide, composed of quartz and minor gold. It is hosted in Laberge sediments.
6	Engineer gold camp Gleaner Group (104M 016)	Au Ag Te	n.a.		Gleaner veins are situated on the northeast side of a major northwest-trending shear zone. Pyrite and gold are hosted in brecciated quartz stringers up to 1.2 m wide.
7	Rupert claim group (104M 008, 35, 36 and 37)	Au Ag Pb Cu Zn	7.6 g/t 925 g/t 74 % 0.28 % 860 g/t (KMO89-47-1)	Grab from 'G' occurrence	Mineralized quartz veins occur near the contact between Hale Mountain meta-intrusive and the metavolcano-sedimentary Wann River gneiss suite. Veins are frequently subparallel to, or bounded by, felsite dykes.

can be explained by the occurrence of a hydrothermal system (such as that producing the silica-flooded rhyolite breccia in Graham Creek) rooted in mafic and ultramafic lithologies such as those exposed both east and west of Graham Creek.

MINERAL POTENTIAL

A number of geologic environments indicate good potential for precious metals. Several of these have been stressed in Mihalynuk *et al.* (1989a), however, subsequent regional mapping has refined the earlier analysis and indicated other favorable geologic environments, as follows:

(1) Veins hosted by Laberge Group strata and associated with splays of the Llewellyn fault zone and/or adjacent dioritic intrusions and volcanics as at the old Engineer mine. Such veins may core folds in the Laberge strata. Further prospecting in the area should place emphasis on the relationships between these features.

(2) Both concordant and discordant (with respect to the foliation) quartz veins occur in the Paleozoic-Proterozoic Boundary Ranges metamorphic rocks. Exploration for occurrences of this type requires a careful focus on late crosscutting metal-bearing veins as barren, concordant quartz sweats are abundant.

(3) Sheared and altered (broadly silicified) or quartz veined rocks within and adjacent to the Llewellyn fault zone are known to be anomalous in gold (Mihalynuk and Rouse, 1988a, b; Mihalynuk *et al.* 1989b).

(4) Brecciated contacts between an en echelon belt of Late Cretaceous to Early Tertiary(?) volcanics and Paelozoic-Proterozoic Boundary Ranges metamorphic rocks including the Teepee Peak and Mount Switzer volcanics.

(5) Mafic and ultramafic rocks occurring adjacent to major fault structures or capped by volcanics have precious metal potential.

(6) Calcsilicate rocks within the Florence Range metamorphic suite locally display sulphide potential. Loose boulders of calcsilicate containing sphalerite and galena (up to 5 per cent) were traced to the base of steep outcrops.

8	White Moose claim group (104M/009, 10, 12)	Au Ag Pb Cu Zn	2.06 g/t 27.4 g/t 2.45 % 0.01 % (AR 8384)	Grab	Mineralized quartz veins occur in Proterozoic to Paleozoic Boundary Ranges greenschist, Hale Mountain meta-intrusive and Wann River gneiss. Veins range from 0.4 to 1.2 m wide.
9	Steep claim group (104M 011)	Au Ag Pb Cu Zn	450.0 g/t 11.0 g/t 4.2 % 0.1 % (AR 9133)	Grab (best assay)	Mineralized quartz-calcite veins are hosted in Cretaceous granite of the Coast plutonic suite. Ore shipped in 1911 yielded 31 103 grams of silver and 93 grams of gold.
10	Brown, Jackpine, Wann Fraction (104M 026)	Au Ag Pb Cu Zn	17.9 g/t 347 g/t 2.62 % 0.56 % 1.00 % (MMI89-59-2A)	Grab (best assay)	A series of subparallel to parallel anastomosing mineralized quartz veins up to 35 cm thick occur within a major splay of the Llewellyn fault zone over more than 70 m. Host rocks are meta-intrusive and metasedimentary lenses.
11	Anyox-Rodeo (104M 017)	Cu Ni Pt Pd Co	0.15 % 0.60 % n.a. n.a. 0.12 % (KMO89-26-3)	Grab	A massive copper-nickel sulfide occurrence is hosted in chloritic schists of probable Proterozoic to Paleozoic age, which lie within major splay of the Llewellyn fault zone.
12	Edgar Lake (104M 018)	Cu	n.a.		Native copper occurs in a number of calcite veins and as disseminations in pyroxene-phyric lapilli tuffs of the Stuhini Group.
13	Kim (104M 063)	Ag Au Cu	109.7 g/t 0.7 g/t 4.0 % (EMPR property file)	Grab	Mineralization occurs in one of several poorly exposed shear zones in a granodiorite host.
14	Nelson Lake (104M 019)	Au Ag Pb Cu	4.6 g/t 198 g/t 3.9 % 1.25 % (MMI89-62-1)	Grab	Highly deformed pelitic schists and marbles host these sulphide-rich veins.
15	Copper Island (104M 020)	Cu	n.a.		Native copper and associated oxides occur in calcite veins.

Occurrence locations are shown on Figure 1-18-5.

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METAMORPHIC ROCKS IN THE FLORENCE RANGE, COAST MOUNTAINS, NORTHWESTERN BRITISH COLUMBIA (104M/8)

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KEYWORDS: Regional geology, Nisling Terrane, Boundary Ranges, Florence Range, metamorphic suite, Hale Mountain granodiorite, Mount Switzer volcanics, Wann River gneiss.

INTRODUCTION

Fieldwork completed during 1989 focused on 1:20 000-scale mapping of the previously undivided metamorphic rocks in the Florence Range of the Coast Mountains of northwestern British Columbia (Figure 1-19-1). These rocks have been included in the Nisling Terrane, which is interpreted as a displaced continental margin assemblage of unknown origin (Wheeler and McFeely, 1987). Nisling Terrane rocks resemble continental margin rocks of western North America (*e.g.* Windermere Supergroup). However, east of the Florence Range there are oceanic rocks of early Paleozoic to early Mesozoic age that separate the Nisling Terrane from ancestral North America. Either the Nisling Terrane is a rifted fragment of North America or a fragment of another continent. In addition to its origin, the apparent pre-Upper Triassic metamorphism of the terrane is unusual in the Cordillera and is presently unexplained. These are subjects that will be addressed by this study.

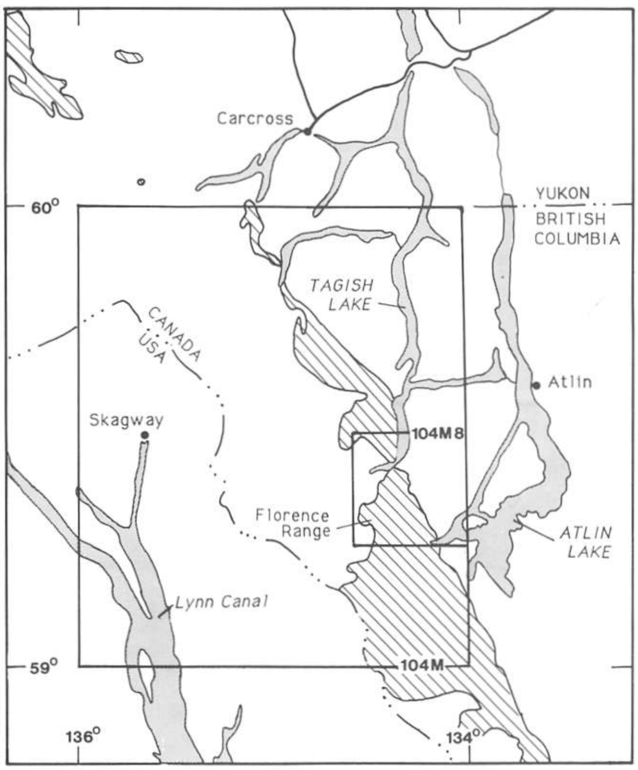

Figure 1-19-1. Location map for the Florence Range. Lined areas indicate the location of metamorphic rocks of the Nisling Terrane (modified from Wheeler and McFeely, 1987).

GENERAL GEOLOGY AND PREVIOUS RESEARCH

The Florence Range lies between the south ends of Tagish and Atlin lakes (Figure 1-19-1). Metamorphic rocks are bounded on the west by undeformed, probably late Mesozoic, granitic and granodioritic intrusives of the Coast plutonic-metamorphic complex. To the east, the Llewellyn fault separates metamorphic rocks from Upper Triassic volcanic rocks of the Stuhini Group (Stikine Terrane) and undeformed Mesozoic plutonic rocks (Christie, 1957; Bultman, 1979; Werner, 1978; Figure 1-19-2).

Mapping at a scale of 1:250 000 by Christie (1957) outlined the regional extent of exposed metamorphic rocks in map area 104M. He divided the metamorphic rocks into (1a) micaceous quartzite, hornblende-quartz-feldspar gneiss, amphibolite, schist and limestone, and (1b) chlorite schist, feldspar-chlorite gneiss, amphibole gneiss and limestone. Other than mapping the locations of large carbonate layers in the latter package, Units 1a and 1b were not subdivided.

More detailed mapping was conducted in the Florence Range by Bultman (1979), who mapped the eastern margin of the range at reconnaissance scale, and Werner (1977, 1978), who mapped the metamorphic rocks exposed south of the Wann River and south of Willison Creek at 1:30 000 scale (note that throughout this paper 'the Wann River' will refer to the Wann River above Nelson Lake). Neither study subdivided the metamorphic rocks, although Werner (unpublished mapping) mapped carbonate bands and major axial-surface traces (Werner, 1978).

Metamorphic rocks continuous with those in the Florence Range extend to the northwest to the British Columbia – Yukon border (Figure 1-19-1). They have been mapped at 1:50 000-scale as the Boundary Ranges metamorphic suite (Mihalynuk and Rouse, 1988a, 1988b; Mihalynuk *et al.*, 1989; Mihalynuk, 1989). Lithologies include chlorite-actinolite schist, biotite-plagioclase-quartz schist, chlorite schist, and graphitic schist, with minor marble, pyroxene-plagioclase schist, impure metaquartzite and orthogneiss. The orthogneiss bodies include altered and deformed leucogranite and quartz diorite, Bighorn granite, and Hale Mountain hornblende-biotite granodiorite (Mihalynuk and Rouse, 1988b; Mihalynuk *et al.*, 1989).

North of the Yukon border metamorphic rocks no longer form a continuous belt but are exposed as isolated pendants in Mesozoic plutons (Wheeler, 1961; Doherty and Hart, 1988).

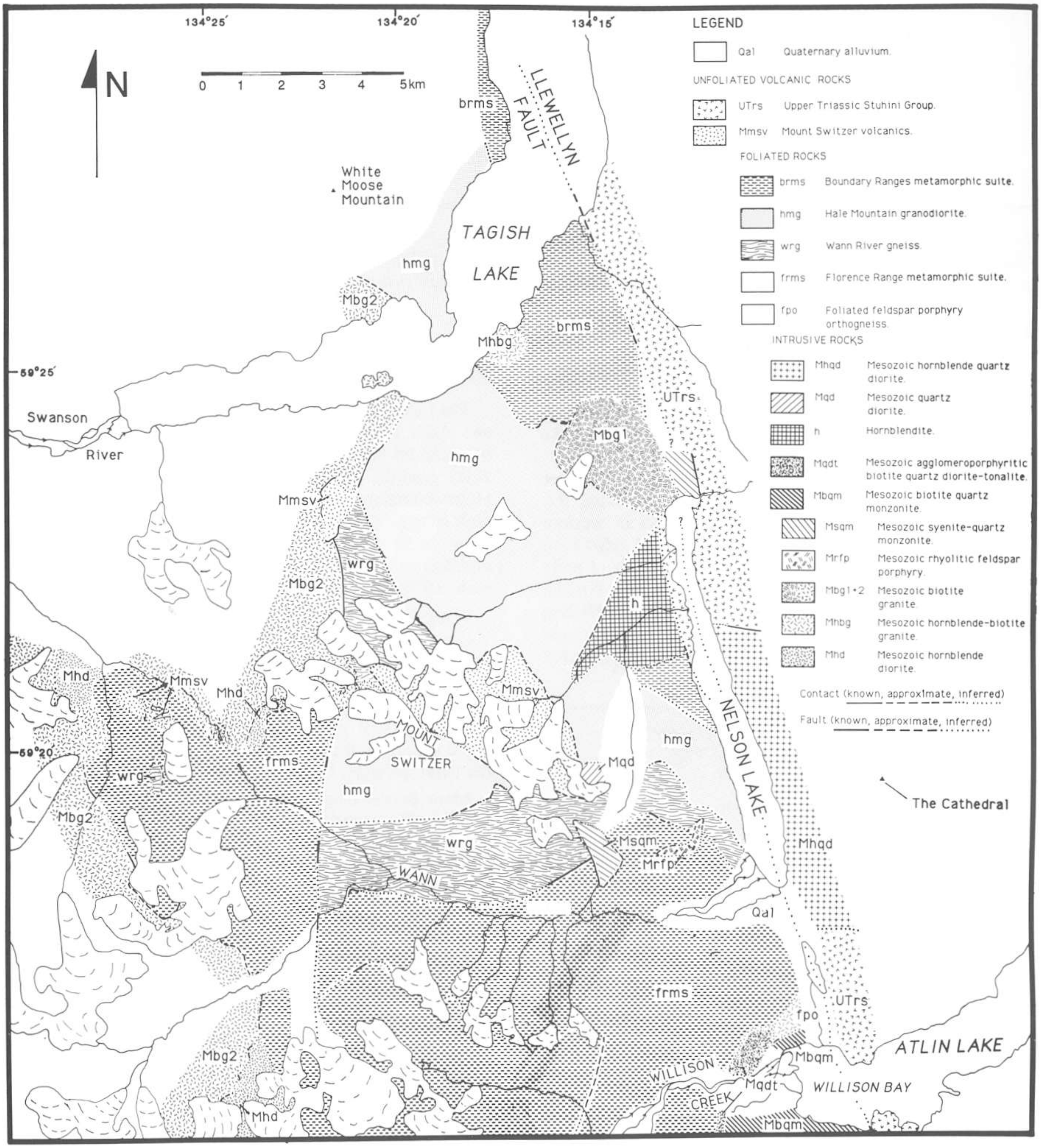

Figure 1-19-2. Geologic map of the Florence Range. The locations of some contacts were generously provided by M. Mihalynuk of the British Columbia Geological Survey Branch, who mapped in the 104M/8 and 104M/9E map areas during the 1989 field season.

There they comprise biotite-muscovite-quartz-feldspar schist, chlorite-rich biotite-granite gneiss, quartzite and minor quartz-mica schist with rare amphibolite bands, and foliated hornblende and hornblende-biotite granodiorite, quartz diorite and quartz monzonite (Doherty and Hart, 1988).

The nature of the basement that the metasedimentary rocks of map area 104M/8 were deposited on is uncertain. The ages of deposition, deformation and metamorphism are inferred to be pre-Late Triassic because clasts similar to metamorphic rocks of the Florence Range occur in the Upper Triassic Stuhini Group, at the south end of Atlin Lake in Willison Bay (Bultman, 1979). The Boundary Ranges suite is interpreted to have been deposited, deformed and metamorphosed before the Early Jurassic because sediments containing Early Jurassic fossils have been mapped as unconformably overlying it north of the Florence Range (Mihalynuk and Rouse, 1988a and 1988b).

LITHOLOGIC SUBDIVISIONS

Four lithologic subdivisions of metamorphic rocks are recognized in the Florence Range (Figure 1-19-2). They are the Boundary Ranges metamorphic suite, Hale Mountain granodiorite, Wann River gneiss and Florence Range metamorphic suite. They have been intruded by undeformed plutonic rocks and are overlain by undeformed volcanic rocks.

BOUNDARY RANGES METAMORPHIC SUITE (brms)

The Boundary Ranges metamorphic suite (Mihalynuk and Rouse, 1988) is exposed along the west and south shores of Tagish Lake (Figure 1-19-2). It is primarily composed of chlorite schist and chlorite-actinolite schist, with minor chlorite-pyroxene schist, discontinuous carbonate layers, quartzite, layer-parallel felsic layers 1 to 20 centimetres thick, and orthogneiss. The orthogneisses vary in composition from granitic to dioritic.

Possible protoliths for the chlorite-bearing schists include pelitic, calcareous marine sedimentary, volcanic and reworked volcanic rocks. Felsic layers may have been deposited as tuffs or felsic flows and the quartzite may be metamorphosed sandstone or chert. A possible tectonic environment for the deposition of these sediments is a volcanic arc; if the protolith for the quartzite is a sandstone, then the arc may have formed on or near continental crust.

The contact between the Boundary Ranges metamorphic suite and the structurally overlying Hale Mountain granodiorite is characterized by interlayering of both units on a decimetre-scale, which may be a result of shearing along the contact. This contact is tentatively interpreted to be a shear zone.

HALE MOUNTAIN (HORNBLENDE-BIOTITE) GRANODIORITE (hmg)

The Hale Mountain hornblende biotite granodiorite (Mihalynuk *et al.*, 1989; Mihalynuk, 1989) is exposed north of the Wann River in the Florence Range, and on White Moose and Hale mountains to the north (Figure 1-19-2; *see* Mihalynuk, 1989). It is characterized by medium-grained plagioclase phenocrysts in a fine-grained groundmass of hornblende with minor biotite, chlorite and epidote. Metre-scale compositional layering is ubiquitous. More felsic layers (0.5 to 3 metres thick) are coarse grained (Plate 1-19-1) and occur in only the structurally highest exposures.

The granodiorite is variably foliated. In some areas a strong foliation and lineation have been developed, but in others fabrics have not been observed in the field, however, phenocrysts are recognisable throughout. Undeformed potassium feldspar pegmatites that crosscut the Hale Mountain granodiorite are characteristic of this unit.

WANN RIVER GNEISS (wrg)

The Wann River gneiss is exposed along the Wann River valley (above Nelson Lake), structurally above the Hale Mountain granodiorite and below the Florence Range metamorphic suite (Figure 1-19-2), and as small outcrops in thrust slices of the Florence Range metamorphic suite. It exhibits a distinctive millimetre to decimetre-scale compositional layering. The composition varies from dioritic (20 per cent hornblende) to gabbroic (50 per cent hornblende), with minor biotite and epidote. The small scale and the gradational

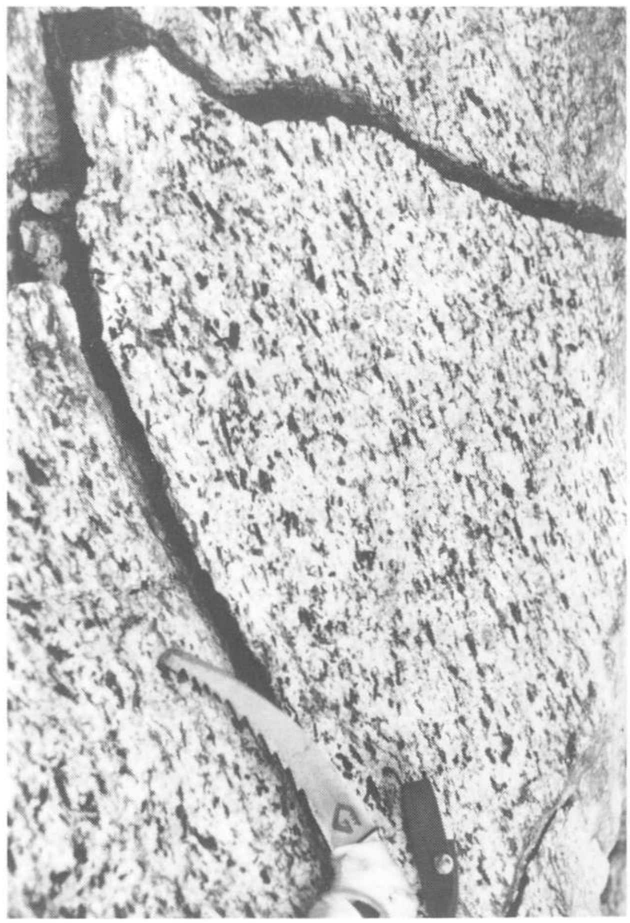

Plate 1-19-1. Coarse-grained felsic layer in the Hale Mountain granodiorite.

nature of compositional layering suggest that it is a primary fabric and therefore the gneisses are interpreted to be metavolcanic rocks of broadly intermediate composition. The Wann River gneiss differs from the Hale Mountain granodiorite in that it does not contain plagioclase phenocrysts, felsic lithologies are medium grained, compositional layering occurs on a smaller scale and the gneisses are everywhere strongly foliated and crosscut by plagioclase-bearing pegmatites.

The contact zone between the Hale Mountain granodiorite and the Wann River gneiss may be abrupt; an interlayered boundary that is continuous over at least 100 metres; or a brecciated and pegmatite-flooded zone about 200 metres wide (M. Mihalynuk, personal communication, 1989). The foliation becomes more strongly developed, within both units, toward the contact, indicating that this contact is sheared. On the ridge west of Nelson Lake a succession of metasedimentary rocks 20 metres thick is preserved at the contact between the Hale Mountain granodiorite and the Wann River gneisses.

FLORENCE RANGE METAMORPHIC SUITE (frms)

The Florence Range metamorphic suite crops out on the slopes north of the Wann River and continues to the south, beyond the south edge of map area 104M/8 (Souther, 1971; Werner, 1977, 1978; Figures 1-19-1 and 2). It structurally overlies the Wann River gneiss and where the contact has been seen it corresponds to an abrupt lithologic change in a zone of strained, well-layered gneiss. This zone is tentatively interpreted as a shear zone.

The Florence Range metamorphic suite is composed of semipelitic, pelitic, carbonate, amphibolitic and calcsilicate rocks, with minor quartzite and graphite-bearing pelitic and semipelitic rocks. Semipelitic and pelitic layers (biotite, quartz $\pm$ plagioclase, muscovite, garnet, kyanite, sillimanite, graphite) are from 0.1 to 30 metres thick. Quartzite layers are often impure ($\pm$ biotite) and may be up to 3 metres thick. Amphibolite (hornblende $\pm$ plagioclase, garnet) is associated with carbonate rocks. These two distinctive rock types, which may be used to define a stratigraphy, form layers 0.1 to 20 metres thick (Plates 1-19-2 and 3) that can be continuous over several kilometres and are thicker and increasingly abundant toward the west, whereas semipelitic layers become more common toward the east. Calcsilicate rocks form layers 0.02 to 1 metre thick and are composed of calcite with or without tremolite, diopside, actinolite, grossular garnet and anorthite. A feldspar porphyry orthogneiss (fpo) crops out north of the mouth of Willison Creek (Figure 1-19-2).

Lithologies within the Florence Range suite, such as extensive quartzites and carbonates, are indicative of a continental margin setting. Amphibolitic rocks are interpreted to be metamorphosed flows, tuffs and reworked tuffs.

MOUNT SWITZER VOLCANICS (Mmsv)

The Florence Range suite is unconformably overlain by the undeformed Mount Switzer volcanics, which are preserved on Mount Switzer (Mihalynuk and Mountjoy, 1990, this volume; Figure 1-19-2). Clasts of granite, biotite schist, foliated amphibolite, and quartzite are preserved in the conglomerate overlying the unconformity.

PLUTONIC ROCKS

Two hornblendite bodies (h) have been mapped in the Florence Range. Hornblendite also occurs in map-area 104M/10 (Mihalynuk, 1989). Although undeformed, they are not necessarily entirely post-tectonic, as the lack of foliation within the hornblendite bodies may be due to their extreme competence. South of the Wann River, garnet-

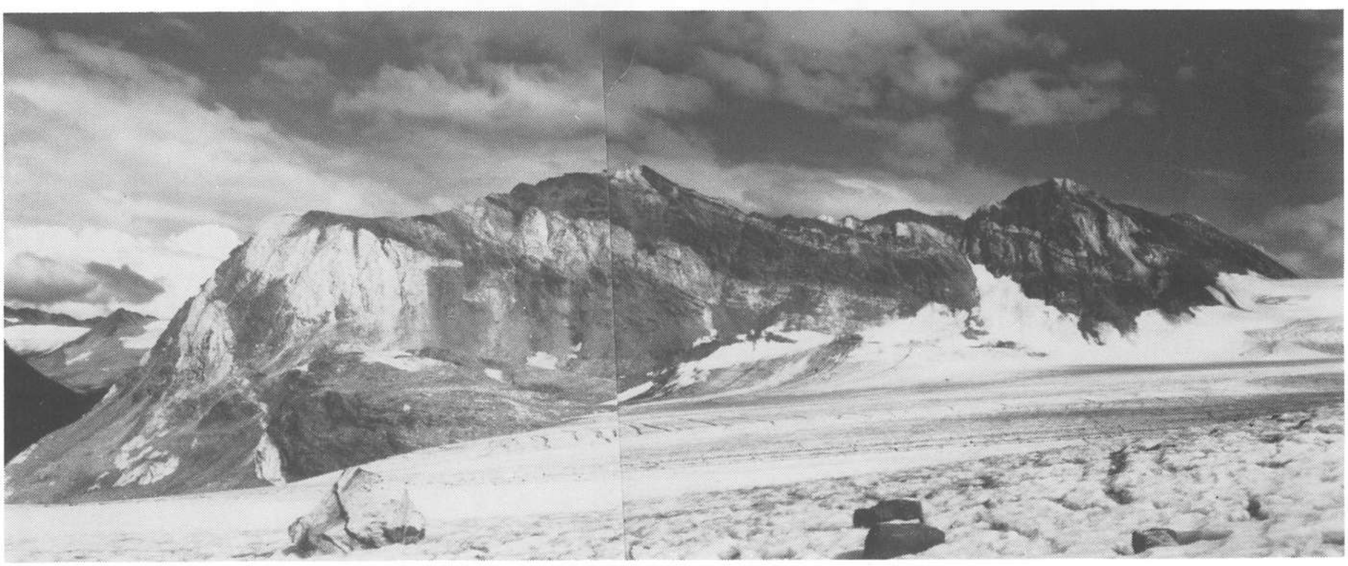

Plate 1-19-2. Northeast-verging folds in the Florence Range metamorphic suite are outlined by carbonate (c), pelitic rocks (p), and amphibolite (a); viewed looking southeast (see Figure 1-19-3).

Plate 1-19-3. Northeast-verging thrust in the Florence Range metamorphic suite (see Figure 1-19-4).

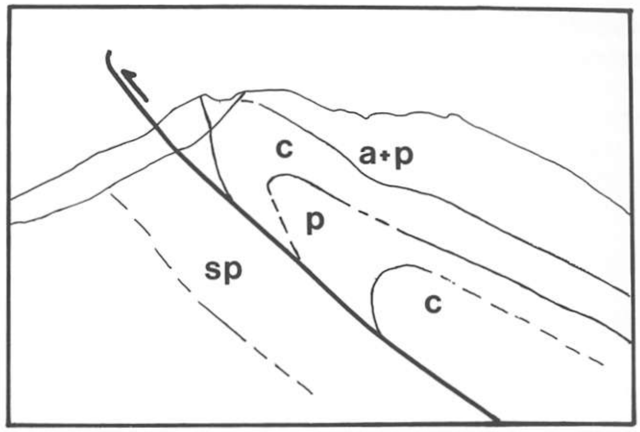

Figure 1-19-4. Line drawing of Plate 1-19-3. (a = amphibolite, c = carbonate, p = pelitic rocks, sp = semipelitic rocks).

bearing hornblendite crosscuts layering and foliation within the Florence Ranges metamorphic suite, indicating that the hornblendite must have been intruded after some deformation had occurred, and then metamorphosed. North of the Wann River, foliation in the Hale Mountain granodiorite wraps around xenoliths of hornblendite.

Cathedral Mountain is primarily underlain by hornblende quartz diorite to granodiorite (Mhqd). The biotite quartz monzonite (Mbqm) that crops out north and south of Willison Creek is not part of the Cathedral Mountain batholith as previously reported by Bultman (1979). It does, however, intrude the variably deformed, medium-grained agglomeroporphyritic biotite quartz diorite (Mbqd) that occurs north and south of the mouth of Willison Creek. The absence of a penetrative foliation indicates that this intrusion postdates deformation that produced a penetrative fabric in the Florence Ranges metamorphic suite. The fabrics that have been developed may be related to later faulting.

Syenite to quartz monzonite (Msqm) intrudes the contact between the Wann River gneiss and the Florence Range metamorphic suite west of the south end of Nelson Lake. The pluton is medium grained and may be trachytic. A rhyolitic feldspar porphyry (Mrfp) that intrudes the Florence Range metamorphic suite and the Wann River gneiss is also exposed west of south end of Nelson Lake.

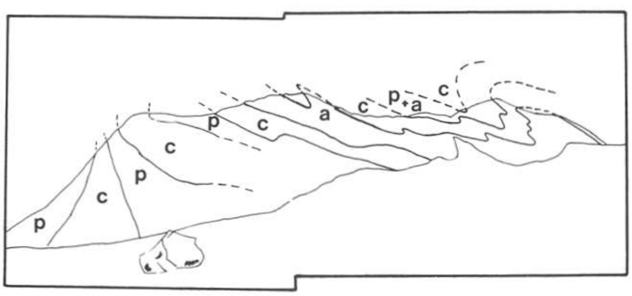

Figure 1-19-3. Line drawing of Plate 1-19-2.

Medium to coarse-grained biotite granite crops out on the north shore of Nelson Lake and the ridge to the northwest. The intrusive contact between the biotite granite (Mbg1) and the Florence Range metamorphic suite has been mapped on the ridge, but no contacts have been seen in the valley below. Where the granite is exposed on Nelson Lake, it lies on the east side of a splay of the Llewellyn fault and is therefore thought to intrude the Llewellyn fault in part. It may also intrude the Upper Triassic Stuhini Group, that is exposed to the east.

Hornblende biotite granite (Mhbg) exposed on the south shore of Tagish Lake is significantly different from the granite that crops out west of Edgar Lake. No contacts with the country rock have been seen.

The hornblende diorite (with minor biotite; Mhd) that crops out in the southwest quarter of the map area is intruded by granites that are exposed to the west and intrudes the Florence Range metamorphic suite. It commonly contains inclusions of the country rock.

Medium to coarse-grained biotite granites of the Coast plutonic-metamorphic complex (Mbg2) intrude the western margin of the Florence Range metamorphic suite, the Wann River gneiss, the Hale Mountain granodiorite, the Boundary Ranges metamorphic suite and the hornblende diorite.

STRUCTURE

FOLDS

Mesoscopic folds are observed in the Boundary Ranges and Florence Range metamorphic suites, which are strongly layered lithologic assemblages, but not in the Hale Mountain granodiorite or the Wann River gneiss. This may be due to a lack of variation in competence in the gneiss and granodiorite, however, they are folded by macroscopic open folds.

The Boundary Ranges metamorphic suite has experienced at least three phases of folding (*see* Mihalynuk *et al.*, 1989). A pre-existing metamorphic fabric is commonly isoclinally folded by mesoscopic folds and carbonate layers outline refolded folds. Areas of planar layering are interpreted to be on the limbs of large-scale folds. However, the lack of marker horizons, such as carbonate layers, makes the mapping of large scale structures in this suite difficult.

Within the Florence Range metamorphic suite, a layer-parallel metamorphic foliation is deformed by northeast-verging folds that range in size from crenulations (0.5 to 3 centimetre scale) to megascopic folds (Plate 1-19-2, Figure 1-19-3). Variations in the vergence of these folds are attributed to later faulting or folding, some examples of which have been seen in outcrop.

FAULTS AND SHEAR ZONES

The contacts between the Boundary Ranges metamorphic suite, the Hale Mountain granodiorite, the Wann River gneisses and the Florence Range metamorphic suite are interpreted to be faults or shear zones. They are offset by later, steeply-dipping, northerly striking faults and truncated by the Llewellyn fault (Figure 1-19-2).

Faults within the Florence Range metamorphic suite are interpreted as northeast-verging thrusts (Plates 1-19-3 and Figures 1-19-2 and 4) that place carbonate-rich thrust sheets over thrust sheets dominated by semipelitic rocks. The Wann River gneiss occurs at the base of some of the thrust sheets. The original relationship between the Florence Range metamorphic suite and the Wann River gneiss is not known.

The contacts between the Hale Mountain granodiorite and both the Wann River gneiss and Boundary Ranges metamorphic suite are characterized by strongly developed ductile fabrics. They are in part interpreted to be shear zones.

Steep, northerly striking faults are recognised in the southern half of the field area where different lithologic units are juxtaposed (Figure 1-19-2). Although similar faults may be present to the north, they were not recognised because lithologies are not as varied.

The Llewellyn fault (Bultman, 1979) truncates all penetrative fabrics and structures in the metamorphic rocks of the Florence Range; the sense and amount of displacement on it are uncertain. However, the juxtaposition of metamorphic rocks on the west and unmetamorphosed rocks on the east is indicative of east-side-down relative movement, strike-slip movement or possibly a combination of both.

METAMORPHISM

The metamorphic grade in the Florence Range varies from greenschist to transitional greenschist-amphibolite facies in the Boundary Ranges metamorphic suite, to upper amphibolite facies (sillimanite/fibrolite present) in the Florence Range suite. Garnet occurs sporadically in schist and amphibolite, and kyanite (with muscovite alteration) occurs rarely in schist. Garnets are commonly chloritized.

CONCLUSIONS

The previously undivided metamorphic rocks of the Florence Range are grouped into four fault-bounded, lithologically distinct subdivisions: the Boundary Ranges metamorphic suite, Hale Mountain granodiorite, Wann River gneiss and the Florence Range metamorphic suite.

Rock types that make up the Boundary Ranges suite correspond with Christie's (1957) Unit 1b and are continuous with a northwest-trending belt of metamorphic rocks that extends as far north as the Yukon border. They are interpreted to have formed in a volcanic arc setting with possible continental influence. They lack abundant quartzose and carbonate rocks that are more typical of the Nisling Terrane. However, the Florence Range metamorphic suite comprises rock types typical of a continental margin. It more closely resembles Christie's (1957) Unit 1a and metamorphic rocks of the Nisling Terrane exposed north of the Yukon border, than the Boundary Ranges metamorphic suite.

The relationship between the Florence Range and Boundary Ranges metamorphic suites is unclear as they are separated by the Hale Mountain granodiorite and the Wann River gneiss. The Boundary Ranges suite may be a distal equivalent of the Florence Range suite; it may have some affinity with Stikinia; or it could be allochthonous to both. The timing of juxtaposition of these four units, their ages, protolith ages, and the relationship of structures within them to terrane accretion, will be addressed by continued mapping to the south and west, and by structural, geochronological and isotopic studies. A comparative study of detrital zircons from the metasedimentary rocks of the Florence Range and those from Nisling Terrane rocks in southern Yukon will investigate the relationship between rocks in these two areas.

ACKNOWLEDGMENTS

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NOTES

THE GEOLOGY OF THE ATLIN AREA (DIXIE LAKE AND TERESA ISLAND) (104N/6 and parts of 104N/5 and 12)

By Mary Anne Bloodgood and Kim A. Bellefontaine

KEYWORDS: Regional geology, Atlin Terrane, Stikine Terrane, Nahlin fault, Cache Creek Group, Laberge Group, structure, thrust faults, tear faults, McKee thrust, O'Donnel thrust, Silver Salmon thrust, mineralization.

INTRODUCTION

The Dixie Lake (104N/6) and Teresa Island (104N/5) map areas are located approximately 20 kilometres south of Atlin in northwestern British Columbia (Figure 1-20-1). Geological mapping at 1:25 000 scale was conducted between mid-June and mid-September and compiled at a scale of 1:50 000. This work marks the second year of a regional mapping project begun during the summer of 1988. Work completed on Atlin Mountain (part of 104N/12) joins with 104M/9 mapped by Mihalynuk and Mountjoy (1990, this volume).

The main contributions of 1989 mapping have been to examine some of the complex structures that exist in Cache Creek rocks and to speculate on the possible boundary relationships between the Atlin and Stikine terranes. An understanding of the structural history is critical to the evaluation of mineral occurrences and economic potential of Atlin and surrounding areas. Lithologies mapped during the 1989 season closely resemble those previously described by Bloodgood *et al.* (1989) with the exception of strata belonging to the Stikine Terrane which outcrop west of the Nahlin fault.

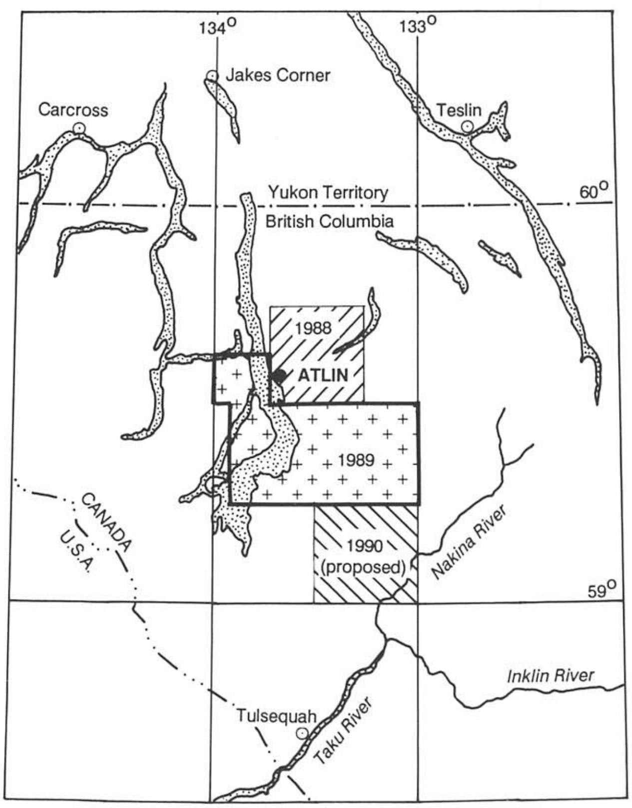


Figure 1-20-1. Location map showing 1988, 1989 and proposed 1990 field areas.

The region east of Atlin Lake is easily accessible by vehicle along a network of roads that follow most of the major creeks in the area. The western part of the map area is only accessible by boat or helicopter.

REGIONAL SETTING

The study area lies entirely within the Intermontane Belt and straddles the tectonic boundary between the Atlin and Stikine terranes. The northwest-trending boundary is defined by the Nahlin fault (Figure 1-20-2). The area west of this structure is underlain by Mesozoic and Cenozoic clastic rocks of the Inklin overlap assemblage (Wheeler *et al.*, 1988) which covers Jurassic Stuhini island arc volcanic rocks. These strata have been examined by Bultman (1979), Mihalynuk and Rouse (1988), Mihalynuk *et al.* (1989) and Mihalynuk and Mountjoy (1990, this volume) near Tagish Lake. Equivalent strata in the Yukon Territory have been studied by Doherty and Hart (1988) and Hart and Pelletier (1989a, b).

East of the Nahlin fault the area is underlain by Paleozoic clastic and volcanic rocks of the Cache Creek Group (Aitken, 1953, 1959; Monger, 1975). These rocks and associated Cretaceous plutons comprise the Atlin Terrane and are the focus of this study. The Cache Creek Group represents a major tectonostratigraphic unit within the Intermontane Belt. It extends almost continuously throughout the length of British Columbia (Figure 1-20-2) and is characterized by its structural and stratigraphic complexities (Monger, 1975, 1977a). The Atlin Terrane is the northernmost extension of the Cache Creek Group in the Canadian Cordillera. It is nearly everywhere fault bounded (Monger, 1975); the Thibert Creek fault marks the northeastern boundary of the terrane north of Dease Lake and extends to the northwest into the Teslin fault zone (Gabrielse, 1985). The extension of the Nahlin fault along the southern boundary of the terrane is characterized by a wide zone of thrust and reverse faults (Gabrielse and Wheeler, 1961; Gabrielse and Souther, 1962; Gabrielse *et al.* 1978, 1980). In the Dixie Lake map area, east of the Nahlin fault, structures characteristic of the southern terrane boundary are developed and occur as imbricate thrust sequences.

Relative timing of movement on the Nahlin fault is constrained by the Birch Mountain pluton which plugs the fault on Teresa Island. It is thought to be Tertiary in age based upon its similarity to dated Tertiary intrusions in northern British Columbia and southern Yukon (Aitken, 1959) and K-Ar ages of 56.6 ± 1.1 Ma and 46.3 ± 1.1 Ma based on biotite and hornblende, respectively (Bultman, 1979).

Monger (1975) established a broad stratigraphic succession based on his work throughout the Atlin Terrane. Mississippian to Pennsylvanian basalt of the Nakina Formation is the basal unit of the succession. It is overlain by and interfingers with chert, clastic sediments, minor carbonate and volcanic rocks of the Kedahda Formation. These are in turn gradationally overlain by a thick carbonate sequence of the upper Mississippian to upper Permian Horsefeed Formation. Locally, Mesozoic clastic rocks unconformably overlie the Cache Creek Group (Aitken, 1959; Monger, 1975, 1977b). Ultramafic rocks, including serpentinized harzburgite, dunite and gabbro, range in size from linear bodies many tens of kilometres in length to pods and slivers a few metres in extent (Figure 1-20-2). These rocks comprise some of the largest ultramafite bodies in British Columbia (Terry, 1977; Wheeler and McFeely, 1987) and were originally interpreted as ultramafic intrusions (Aitken, 1959). Later, Monger (1977) and Terry (1977) suggested that the large Nahlin ultramafite body may represent oceanic basement upon which the Nakina Formation was deposited. Recent work by Ash and Arksey (1990, this volume) supports this interpretation and indicates that thrust faulting has been the primary mechanism of emplacement for these slices of oceanic crust.

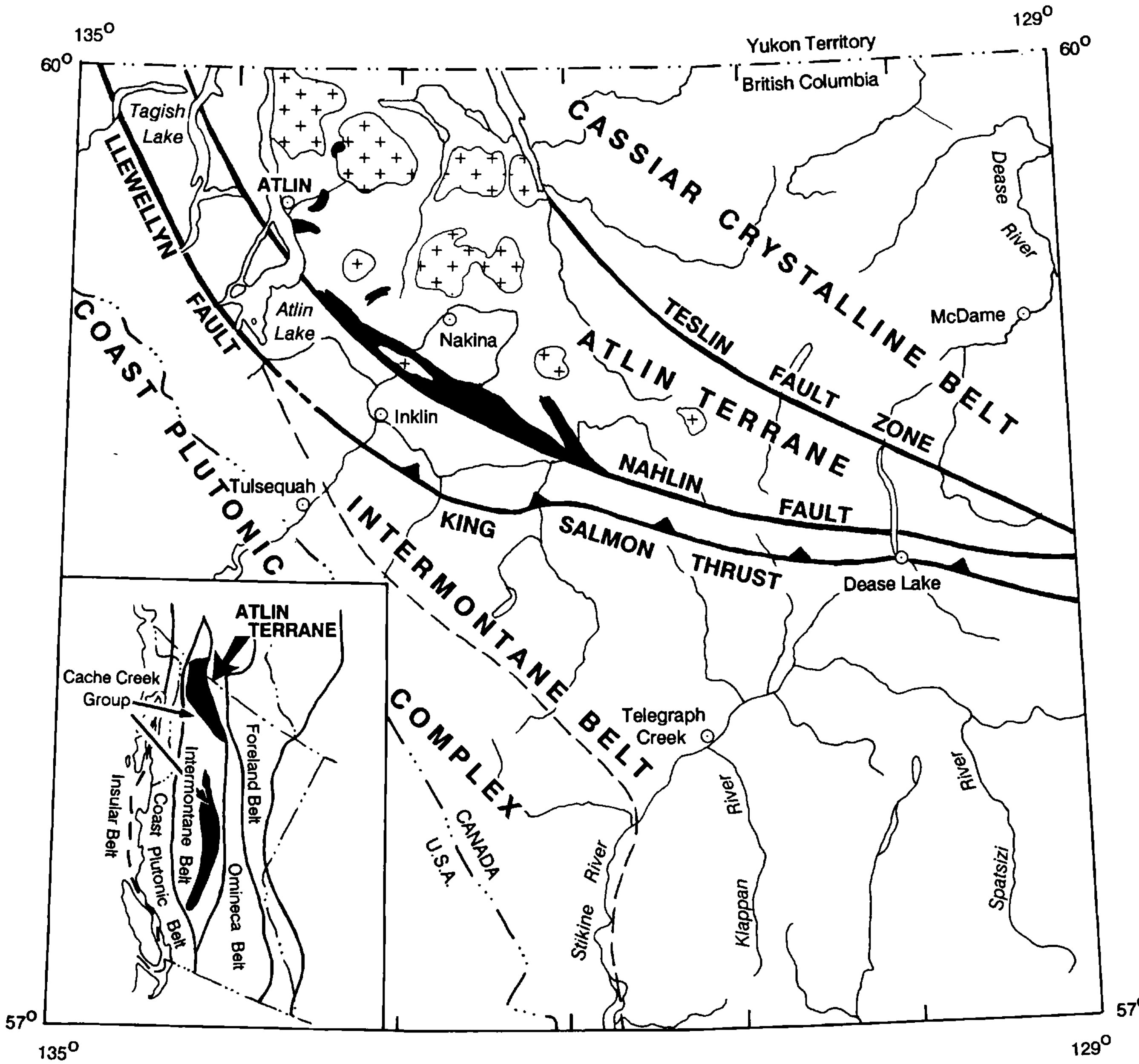


Figure 1-20-2. Regional geology of northwestern British Columbia (compiled from Aitken, 1959, Souther, 1971 and Wheeler *et al.*, 1989). Ultramafic rocks in the Atlin Terrane are shaded black and major plutons are represented by the crossed pattern. Inset map shows distribution of the Cache Creek Group in British Columbia.

LITHOLOGY

PALEOZOIC CACHE CREEK GROUP

CLASTIC SEDIMENTS

East of Wilson Creek (Figure 1-20-3), the map area is extensively underlain by a succession of clastic sediments consisting dominantly of fine-grained, locally siliceous mudstones, siltstones and sandstone/wackestones sometimes interbedded with minor impure chert. The sediments weather dark grey to rusty brown in colour and are typically dark grey to black on fresh surfaces. Bedding varies from fine internal laminations to distinctive centimetre-scale interbeds. Bedding-parallel cleavage varies in intensity from moderately to well developed slaty partings.

LIMESTONES

Massive limestone underlies an extensive area south of Sentinel Mountain where the exposures form large south-facing cliffs. Bedding is often difficult to discern in these large exposures but, from a more distant perspective, gross layering generally dips gently northward.

Grey, pale-weathering massive and featureless limestone characterizes much of the succession. The rocks are locally crinoidal and some corals, fusilinids, minor shell detritus and brachiopods have been recognized. Dark argillaceous limestone comprises one facies variation characterized by increasing silt and organic material accompanied by a distinctive fetid odour. Minor dolomitic limestones are also present.

Limestone within the volcanic and chert succession on Sentinel Mountain ranges from 1 to 10 metres in length and width. Some limestone bodies occur along contacts between other lithologies and are probably fault-bounded slices. Areas of extensive recrystallization and cleavage development along lithologic contacts suggest that these surfaces provided a locus for fault development. Zones of brecciation and recrystallization within the limestone bodies suggest that internal imbrication has also occurred. Limestone also occurs as blocks that are brecciated and completely enveloped by volcanics (Plate 1-20-1); these are interpreted as slumped masses incorporated during the deposition of the volcanic and chert succession.

CHERTS

Cherts underlie much of Sentinel Mountain where they are spatially associated with intermediate to mafic volcanics. They also outcrop north of Mount McMaster in the northeastern quadrant of the map area. The cherts are locally radiolarian-bearing and are characterized by a diverse range of colour; grey to black is the most common, but green, red, pink and cream varieties are also present. They show a wide variation from ribbon-bedded successions (Plate 1-20-2) to massive and unbedded chert.

North and west of the Sentinel Mountain summit repetitive interbeds of grey and red ribbon cherts vary from 1 to 10

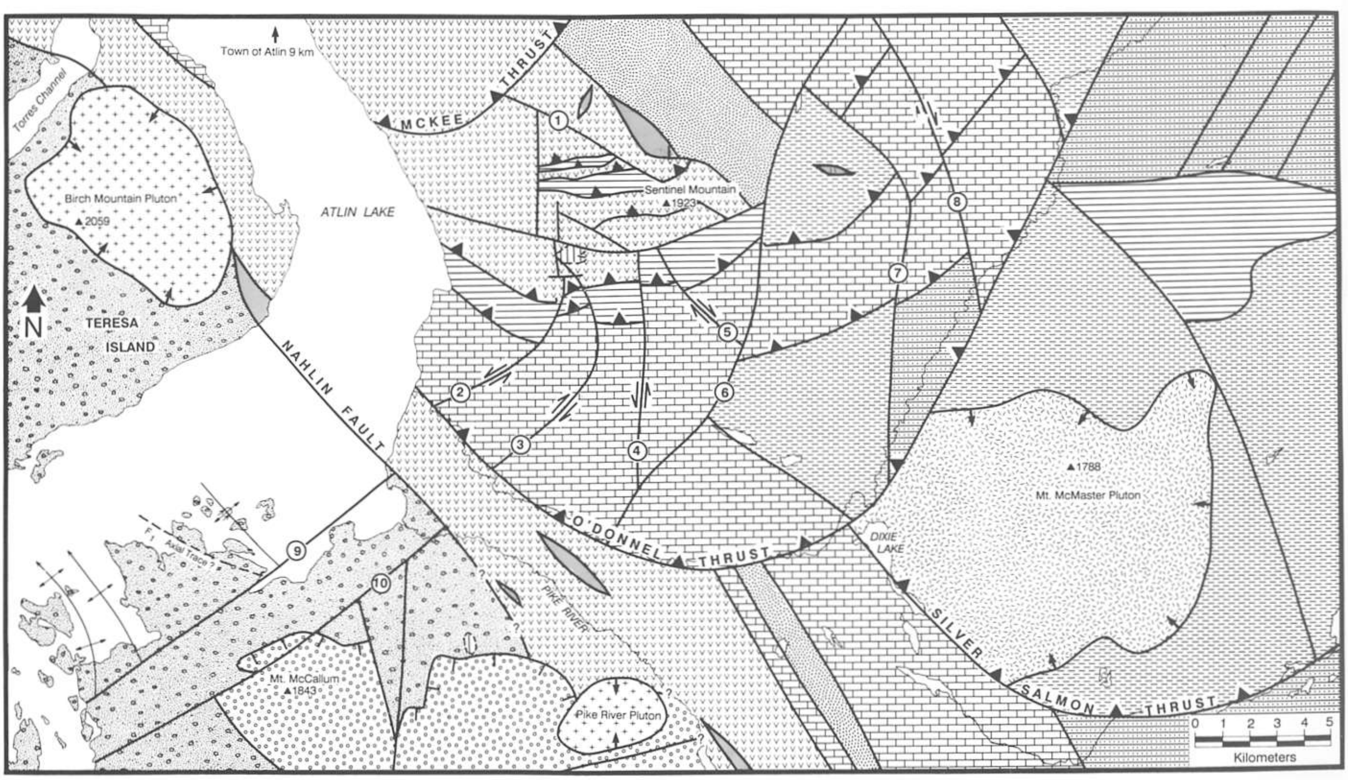
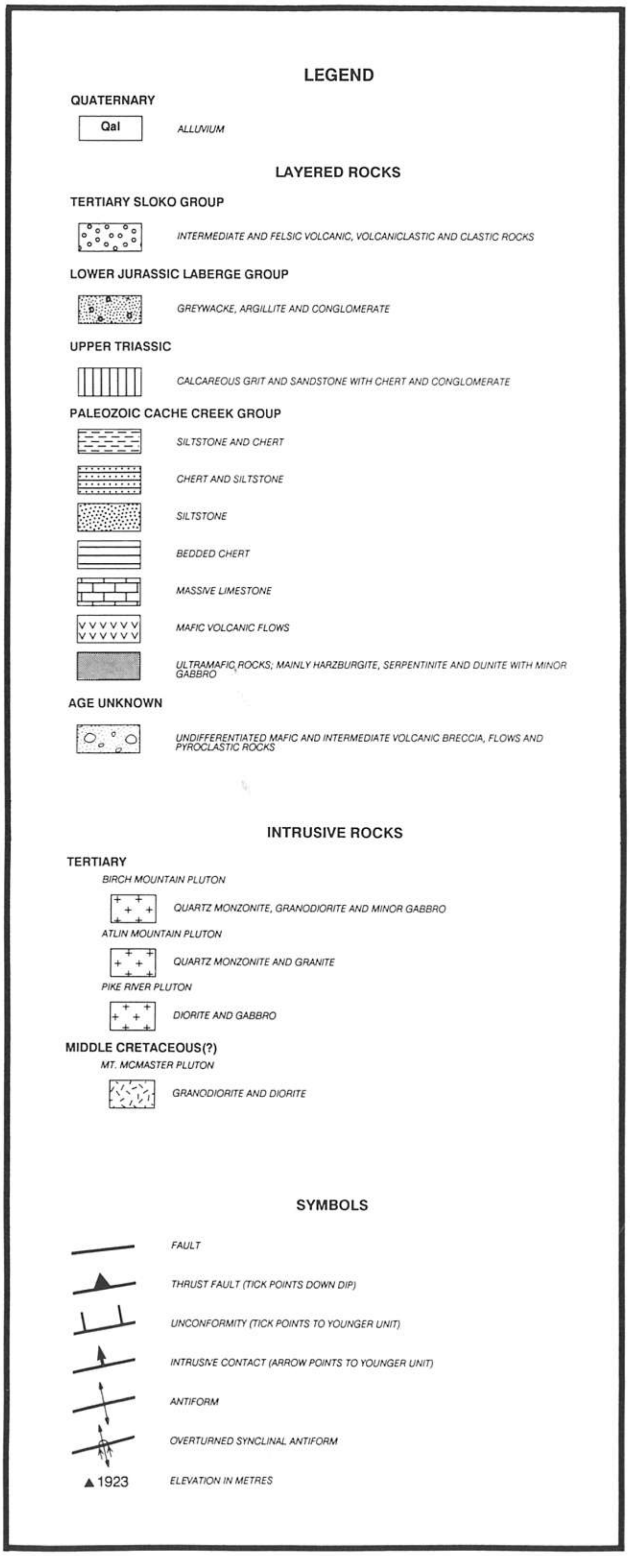


Figure 1-20-3. Simplified geology of Dixie Lake and Teresa Island map sheets (104N/5 and 6). Faults follow creeks and rivers and are numbered as follows: 1 – Eldorado Creek fault, 2 – Kennedy Creek fault, 3 – Jasper Creek fault, 4 – Burdette Creek fault, 5 – Little Creek fault, 6 – Wilson Creek fault, 7 – Canyon Creek fault, 8 – Baxter Creek fault, 9 – Anderson Bay fault, 10 – Mt. McCallum fault.



Legend to accompany Figure 1-20-3 and 1-20-4.

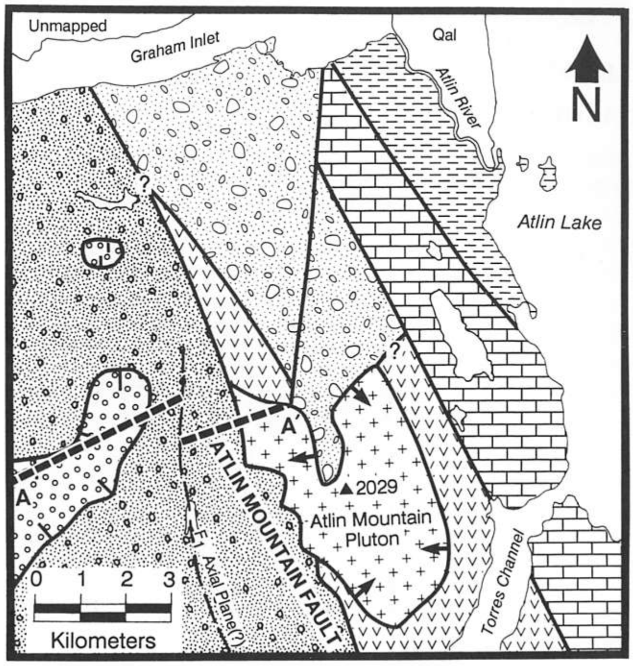


Figure 1-20-4. Simplified geology of Atlin Mountain (104N/12W).

centimetres in thickness. Fine colour variations between beds and colour laminations within individual beds are common. Both massive and laminated red-bedded cherts comprise a significant component of the succession immediately west of the summit. Along the western and southern flanks of the peak the cherts are poorly bedded or massive and are green, grey and black in colour. They are spatially associated with volcanic rocks.

VOLCANIC ROCKS

Volcanic rocks are well exposed on Sentinel Mountain (Figure 1-20-3) and underlie much of Atlin Mountain (Figure 1-20-4); they also outcrop along the northeastern shore of Teresa Island and north of the Pike River. Intermediate to mafic flows, tuffs and diabase comprise the volcanic assemblage.

The volcanic flows are medium to pale green and pale green weathering. Minor colour laminations, interpreted as flow banding, are sometimes present in addition to tuffaceous horizons. The volcanics are generally non-amygdaloidal with an aphanitic groundmass; rarely they are pyroxene porphyritic. Associated tuffs are pale green, locally laminated and contain small angular pumice fragments in a fine-grained matrix.

Pillowed basalts have been recognized along a north-trending ridge west of the Sentinel Mountain summit. At the north end of this ridge, pillows have a maximum dimension of 1 to 1.5 metres; elsewhere pillows range from 10 to 50 centimetres in diameter and grade into pillow breccias. Flattened pillow margins indicate that the sequence is right way up. They are locally interlayered with bedded cherts and in one locality the pillows are silicified immediately adjacent to red cherts, indicating a primary stratigraphic relationship.

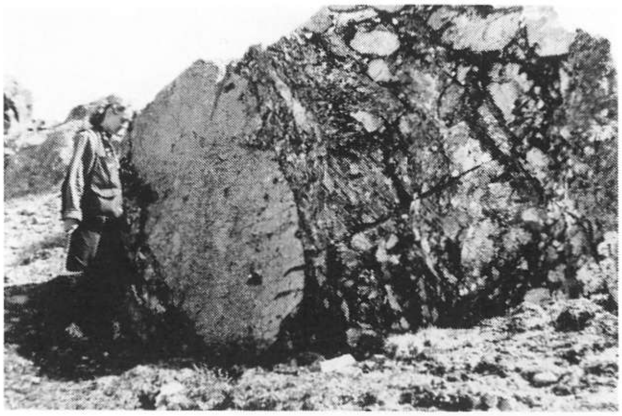


Plate 1-20-1. Limestone blocks enveloped by a submarine mafic volcanic flow. Well-preserved primary flow textures are deflected around the limestone clasts.

Plate 1-20-2. Outcrop photograph of ribbon-bedded cherts on Sentinel Mountain.

Massive diabase occurs locally on Sentinel Mountain, west and northwest of the summit. It is medium green on fresh surfaces but weathers a distinctive orange colour. Chilled margins along contacts with chert extend into the diabase for up to 10 centimetres; the adjacent chert is baked and recrystallized.

ULTRAMAFIC ROCKS

Ultramafic rocks comprise only a small proportion of the Cache Creek Group in the map area. They occur as small, isolated fault-bounded exposures of serpentinite, harzburgite with lesser dunite, and gabbro. On Sentinel Mountain ultramafic rocks outcrop north of Eldorado Creek and several small bodies occur southwest of the summit; one isolated exposure was found east of Wilson Creek. Sporadic exposures have been mapped along the northwest trend of the Nahlin fault along the Pike River, and on Teresa Island and Atlin Mountain, however many of these bodies are too small to appear on Figures 1-20-3 and 1-20-4.

North of Eldorado Creek a relatively large, poorly exposed ultramafic body consists of recessive serpentinite. To the west of the main body, carbonatized harzburgite with minor dunite occurs in a series of north-trending outcrops locally cut by quartz-carbonate stockwork. Several small gabbro bodies are associated with the serpentinite and mylonitized gabbro occurs at the southern limit of exposure.

Southwest of Sentinel Mountain, several ultramafic bodies are present within a structurally complex area. They are localized along an east-trending thrust fault offset by two later, orthogonal tear faults. The ultramafites are strongly serpentinized and exhibit recessive weathering.

Along the eastern shore of Teresa Island ultramafic rocks appear to mark the base of the Atlin Terrane although the contact with the Jurassic Laberge Group is not exposed. At this locality the ultramafics are strongly altered to listwanite; mariposite and pyrite are pervasive.

East of the Pike River strongly carbonatized ultramafics occur along a northwest-trending linear belt parallel to the Nahlin fault. Exposures are sporadic but have been traced for approximately 5 kilometres. The primary lithology is harzburgite with lesser dunite occurring as small pods with finely disseminated chromite. Quartz-carbonate veins up to 0.5 metre wide are developed parallel to the regional trend.

TRIASSIC CLASTIC SEDIMENTS

A distinctive suite of sedimentary rocks, recognized on Sentinel Mountain and in a small exposure southwest of the Pike River, are believed to be of Triassic age (Figure 1-20-3). They are lithologically similar to a succession exposed east of the confluence of the Nakina and Taku rivers (Figure 1-20-1) in the Hardluck Peaks area which has been assigned to the Triassic (Aitken, 1959). Well-laminated pale siltstones and sandstones characteristic of this unit are associated with gritty limestone, maroon chert and coarser clastic sediments ranging from greywackes to conglomerates.

Southwest of the Pike River these sediments are unconformably overlain by the Laberge Group. Brown to tan-weathering, laminated gritty siltstones and fine sandstones are interbedded with grey and maroon chert and gritty dolomitic limestone. Millimetre-scale colour laminations define bedding within the siltstones and the cherts are thinly interbedded with tan-weathering mudstone containing minor carbonaceous debris. Silty to sandy laminated limestone and dolomitic limestone weather grey to tan. No fossils or shell debris have been recognized.

West of the main peak of Sentinel Mountain laminated siltstones are interbedded, on a metre-scale, with grey to tan-weathering sandy limestones that contain conodonts of Norian age (GSC Sample #168203). Bedding in the limestone is defined by coarser grit beds, colour laminations and minor fragments of chert and mudstone.

To the north, a widely variable conglomerate is confined to a north-trending graben. The eastern boundary is a mélange zone containing pods and lenses of chert, volcanics and ultramafics; to the west, a sharp fault contact separates conglomerate from Cache Creek volcanic rocks. Facies variations within the unit are rapid and range from pebble to cobble conglomerate. Clasts are generally well-winnowed, and rounded to subrounded, however, some are angular, including ripped-up blocks identical to the host conglomerate.

Clasts of all Cache Creek lithologies except the ultramafics have been recognized. Fragments derived from the Triassic sedimentary succession are well represented, occurring primarily as angular clasts varying from 5 centimetres to 2 metres in size. Two distinctive types of limestone clasts are observed: a crinoidal limestone thought to be part of the Cache Creek Group, and a gritty laminated limestone believed to be derived from the Triassic succession. Samples of each have been collected for microfossil identification. Numerous cobbles and boulders of a felsic plutonic rock are also present but do not appear to be locally derived. The clasts are leucocratic, contain abundant quartz eyes and bear no resemblance to any of the plutons in the area.

JURASSIC LABERGE GROUP

Laberge Group rocks belonging to the Inklin overlap assemblage (Wheeler *et al.*, 1988) underlie much of the area along the southern and western shoreline of Atlin Lake, outcropping north of Mount McCallum, on Teresa Island (Figure 1-20-3) and Atlin Mountain (Figure 1-20-4). The contact of the Laberge Group with the Cache Creek Group is everywhere marked by a fault, either the Nahlin fault or the Atlin Mountain fault.

Massive grey to brown-weathering greywackes comprise a significant portion of the Laberge Group and are interbedded with grey to black siltstones, mudstones and shales; the shales rarely contain plant fossils. Individual beds vary in thickness from 0.5 to 3 metres and a bedding-parallel fissility is present in finer grained lithologies. Facing indicators include graded bedding, crossbedding and sole marks.

On Mount McCallum, brown to grey Laberge greywackes are massive, hard and weather orange to brown. They are bedded on a scale of 10 to 50 centimetres with abundant fine plagioclase laths within a sandy, poorly sorted matrix. Uncommon siltstone beds and lenses vary from 10 centimetres to 2 metres in thickness.

TERTIARY SLOKO GROUP

Volcanics and associated volcaniclastic rocks of the Sloko Group, are well exposed on Mount McCallum and Atlin Mountain and unconformably overlie folded Laberge Group (Figures 1-20-3, 1-20-4 and 1-20-5). The sequence may range from Late Cretaceous to Eocene in age (Bultman, 1979). Porphyritic rhyolite, andesite and minor basalt dominate the Sloko Group, but volcaniclastic sediments including tuffs, greywackes and volcanic conglomerates also occur. Numerous pale coloured, buff to rusty weathering felsic dikes crosscut the Sloko Group on both Mount McCallum and Atlin Mountain.

On Mount McCallum the Sloko volcanics are characteristically brown, mauve or dark grey and weather brown. Intermediate volcanic flows underlie much of the area and spectacular columnar basalts cap several ridges west of the main peak. A broad range of facies are present and facies variations are rapid. Aphanitic to feldspar-porphyritic intermediate flows containing quartz eyes and minor acicular mafic phenocrysts are associated with ash tuffs and volcanic-derived clastic sediments. The volcaniclastic sediments contain detrital quartz and feldspar and vary from brown, grey, and green to mauve in colour, and generally weather brown.

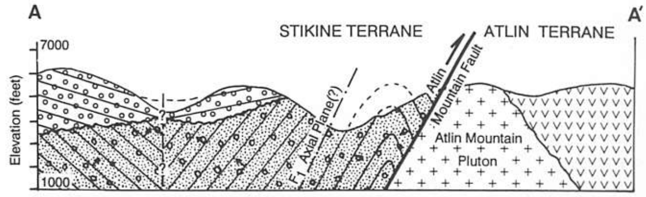


Figure 1-20-5. Cross-section of Atlin Mountain (*see* Figure 1-20-4 for location of cross-section; symbols as in legend). The Atlin Mountain fault places Laberge strata over Cache Creek volcanics and the Atlin Mountain pluton. A north-trending lineament west of the summit may correspond to an F_1 axial trace. Laberge strata west of this lineament are upward facing, whereas strata exposed east of it are overturned. A synclinal antiform occurs near the Atlin Mountain fault; it may be a drag fold associated with motion on the fault. To the west of the fault, the Laberge Group is unconformably overlain by Sloko volcanics and clastics. The occurrence of Sloko strata on mountain ridges and in valleys indicates the presence of significant topography in the Laberge rocks when the Sloko Group was deposited.

West of Atlin Mountain volcaniclastic conglomerates unconformably overlie Laberge sediments (Figure 1-20-4 and 1-20-5). Clasts vary from pebbles to boulders in a dominantly volcanic-derived matrix. Pebble conglomerate grades rapidly into cobble and boulder conglomerate. Abundant white quartz and dark grey chert pebbles varying from 1 to 20 millimetres in size are supported by a coarse wacke matrix with discontinuous silty horizons. Coarser conglomerates are locally clast supported and contain abundant well-rounded sandstone, greywacke and siliceous clasts up to 20 centimetres across.

Common crosscutting felsic dikes trend north and northeast. They are generally pale in colour, weather pale to rusty, and are locally completely bleached. Small white plagioclase laths are abundant and associated with small vitreous quartz eyes and minor mafic phenocrysts. The thickness of the dikes varies from 1 to 3 metres, with a maximum observed thickness of 5 metres.

TERTIARY OLIVINE BASALT

Olivine basalt outcrops on the lakeshore, north of Mount McCallum. It is fine grained, green to brown in colour and weathers dark grey or brown. Vertical columnar joints, up to 20 metres in height, are well developed and indicate the basalts are upright, flat lying and probably unconformaby overlie Jurassic Laberge strata. A K-Ar whole-rock age of 27 ± 4 Ma was obtained by Bultman and he cites another age of 16.2 ± 2 Ma on a whole-rock fraction of the same sample (Armstrong and Harakal, unpublished data).

UNDIFFERENTIATED VOLCANIC ROCKS/AGE UNKNOWN

A heterolithic succession of volcanic rocks of possible Mesozoic age (Aitken, 1959) underlies the area north of the

Atlin Mountain summit (Figure 1-20-4). These rocks are thought to be correlative with a similar succession exposed to the north of Graham Inlet on Table Mountain and Mount Minto. In part they comprise a roof pendant in the Atlin Mountain pluton and consist of fragmental rocks of intermediate composition. Grey-weathering, locally amygdaloidal breccias and tuffs grade into purple and grey trachytes.

INTRUSIVE ROCKS

MOUNT MCMASTER STOCK

The Mount McMaster stock underlies an extensive area east of the O'Donnel River and north of the Silver Salmon River (Figure 1-20-3). It consists of quartz diorite and granodiorite and intrudes sediments of the Cache Creek Group. Adjacent to the contact the sediments are metamorphosed to quartz muscovite schists and phyllites. Based upon lithologic similarities, the stock is correlated with the Fourth of July batholith north of Atlin from which Christopher and Pinsent (1979) obtained K-Ar ages of 73.3 ± 2.6 Ma and 110 ± 4 Ma, from biotite and hornblende respectively. The Mount McMaster stock weathers white to pale orange and is characterized by a medium to coarse-grained texture with local variations in the relative abundance of felsic and mafic minerals represented by quartz, plagioclase, biotite and hornblende. Potassium feldspar is a lesser constituent and several quartz potassium feldspar pegmatites were mapped east of the peak. Diorite xenoliths occur locally and are more abundant in the granodioritic phase of the stock.

ATLIN MOUNTAIN PLUTON

The Atlin Mountain pluton consists of fine-grained, white-weathering quartz monzonite. Plagioclase and potassium feldspar comprise the phenocryst assemblage, occurring as small chalky to orange-weathering laths supported by an aphanitic groundmass. The pluton intrudes Cache Creek volcanic rocks and is truncated along its western margin by the Atlin Mountain fault (Figure 1-20-4). Small apophyses of the intrusion extend into the fault zone and the Laberge strata, indicating that the pluton may be synkinematic. The pluton is believed to be Tertiary in age (Aitken, 1959) and related to the Birch Mountain and Pike River plutons. Samples collected for U-Pb isotopic dating may provide more specific constraints on the timing of the emplacement of these bodies.

BIRCH MOUNTAIN PLUTON

Underlying the northern part of Teresa Island, the Birch Mountain pluton is dominated by leucocratic quartz monzonite with lesser diorite and gabbro. The main phase is porphyritic, locally varying to phaneritic or aphanitic, and contains fine-grained granodiorite to diorite xenoliths. Diorite and gabbro constitute a minor phase near the southern margin of the stock, and appear to be similar to gabbro and diorite in the Pike River pluton.

PIKE RIVER PLUTON

A small dioritic to gabbroic stock outcrops south of the Pike River, near the southern margin of the map area and is thought to be correlative with the Atlin Mountain and Birch Mountain intrusions. The rocks are medium to coarse grained, with abundant dark plagioclase laths and a weakly developed foliation. This and several other small stocks of similar composition mapped south of the study area are thought to be related to the mafic volcanic members of the Sloko Group (Aitken, 1959).

STRUCTURE

Structures in the map area can be divided into three distinct groups; those east of the Nahlin fault in the Atlin Terrane, those west of the fault in the Stikine Terrane, and structures along the Nahlin fault itself.

ATLIN TERRANE

Faulting, rather than folding, is the dominant form of deformation in Cache Creek rocks of the Atlin Terrane. The most prominent structural features of the map area are numerous low-angle thrust faults and associated tear faults. The presence of thrusts in the Atlin area is expected as ocean-floor ultramafic rocks (ophiolites) outcrop throughout the region (Ash and Arksey, 1990, this volume; Monger, 1975; Terry, 1977). The presence of thrusts is also supported by previous work; Monger (1975) suggested that most, if not all of the lithological contacts on Sentinel Mountain show some degree of bedding-parallel shear, and Cole (1989) described the occurrence of horizontal faults near the summit of Sentinel Mountain. Homestake Mineral Development Company interpreted shallow northwest-dipping faults near Atlin in drill core at the Pictou showing (MINFILE 104N 044), the Yellowjacket property (MINFILE 104N 014) and the Heart of Gold property (MINFILE 104N 101) (personal communication, D. Marud, 1989). In addition, Ash and Arksey (1990, this volume) have mapped several thrust faults on Monarch Mountain.

Thrusts of all scales occur in the Dixie Lake map area. The three largest are the McKee, O'Donnel and Silver Salmon thrusts, which are named after the major drainages they follow (Figure 1-20-3). The curvature of their traces, structural measurements and strong air photo linears suggest that these large thrust faults dip gently northwest. The surface expression of each is a major mélange zone consisting of slivers of all Cache Creek lithologies. The O'Donnel thrust is exposed on the O'Donnel River, 3 kilometres west of Dixie Lake. It outcrops as a spectacular northwest-dipping fault zone 20 metres wide and cored by 1.5 metres of fault gouge (Plate 1-20-3). The Silver Salmon thrust was not seen in outcrop in the map area but is inferred from structural data and well-defined air photo linears. A series of northeast-trending slivers of intercalated ultramafic, sedimentary and volcanic rocks occurs at Chikoida Mountain (Aitken, 1959; 104N/3 – southeast of the map area). This map pattern suggests these strata lie in the footwall of the Silver Salmon fault and may be part of a complex imbricate thrust system.

Internally each thrust sheet consists of numerous imbricate slices. Several are exposed throughout the map area (Plate 1-20-4) but most are indicated by outcrops of small lenticular pods of sheared ultramafite and other Cache Creek lithologies, mylonitic fabrics, subhorizontal fractures, minor folds and slickenside lineations on bedding and fracture surfaces. Although contacts between cherts and volcanics on

Sentinel Mountain are primary, most lithological contacts show evidence of some degree of bedding-parallel slip. The traces of thrust faults are often marked by topographic depressions along mountain ridges; south of Sentinel Mountain several north-facing dip slopes are underlain by footwall strata of thrusts exposed in cols.

Cache Creek lithologies are also cut by northwest to northeast-trending, high-angle tear faults, which are an inherent part of the thrust deformation. These faults include the Kennedy Creek, Jasper Creek, Burdette Creek, Little Creek, Wilson Creek, Canyon Creek, Baxter Creek and several other unnamed faults. Marker units allow determination of relative motion on several of them, and displacement on most does not appear to be in excess of 1 or 2 kilometres. In addition to thrusts and tear faults, several high-angle east-trending faults are mapped. These structures truncate both thrusts and tear faults and are therefore later, however, their exact age and significance are not known.

Overall, Cache Creek rocks in the Dixie Lake map area display only one major phase of deformation. Deformation was previously thought to be Triassic to Late Jurassic in age (Monger, 1975), however mapping in the southeast part of the area has shown that the mid-Cretaceous(?) Mount McMaster stock is internally deformed and is confined to the Silver Salmon thrust sheet. This suggests that the intrusion may be riding on the fault and that thrusting may be as late as post mid-Cretaceous in this part of the Atlin Terrane. Structural data indicate that the McKee Creek, O'Donnel and Silver Salmon thrust sheets were once rooted somewhere to the northwest and have moved toward the southeast, placing older rocks on top of younger strata. This interpretation is supported by fossil ages obtained by Monger, 1975, 1977b and preliminary ages from 1989 mapping. The amount of movement on these major structures is not known with certainty, but displacement is thought to be large. Data are too limited to allow a predeformational reconstruction of Cache Creek rocks at this time, however, limestone and chert samples collected for conodont and radiolarian microfossils may provide sufficient age constraints to permit restoration of the collage of thrust sheets and fault blocks that are characteristic of the Dixie Lake area.

Stikine Terrane

Jurassic Laberge Group strata are folded into a series of upright anticlines west of the Nahlin fault. These folds plunge moderately to the northwest and have wavelengths of approximately 2 kilometres. Folds are mapped on the basis of regional limb orientations; one minor fold was observed in the field. This deformation is thought to be Late Jurassic in age and related to the collision of Superterrane I with cratonic North America (Tempelman-Kluit, 1979). Eocene deformation is indicated by northeast-trending faults such as the Anderson Bay and the Mount McCallum faults which crosscut the Tertiary Sloko stratigraphy and Laberge strata in the Teresa Island map area.

Nahlin Fault

Structural trends of both the Atlin and Stikine terranes are more complicated near the Nahlin fault. According to Souther (1971) the Nahlin fault is a high-angle, northwest-trending, east-dipping structure that places Cache Creek rocks over strata belonging to the Stikine Terrane. However, observations made in the Dixie Lake and Teresa Island map areas indicate that it is a wide, complex zone with both a history and a geometry that are still unclear. The trace of the Nahlin fault is probably more complex than Aitken (1959) identified and it may be offset on northeast-striking Eocene faults. In addition, exposure is very poor along the fault; any inferences of the attitude of this structure based on its apparent surface trace are tenuous.

Cache Creek rocks near the Nahlin fault strike northwest and dip steeply. This trend is consistent with the structural grain of the Cordilleran orogen but diverges significantly from the southeast-directed thrust sheets common in the eastern part of the map area. This raises the problem of how Cache Creek rocks can be transported for long distances to the southeast and be emplaced over Laberge strata to the southwest, without having major transcurrent motion on the fault between them, yet only one dextral transcurrent kinematic indicator was found in the area (Plate 1-20-5).

Plate 1-20-3. The O'Donnel thrust outcrops as a spectacular 20-metre fault zone that dips 25° northwest and contains 1.5 metres of fault gouge. View to the southwest.

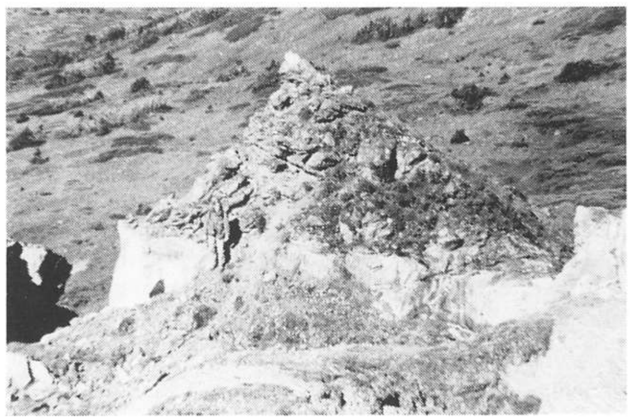


Plate 1-20-4. Horizontal thrust contact between sheared mafic volcanics (hangingwall) and massive limestone (footwall) on Sentinel Mountain. View to the west.

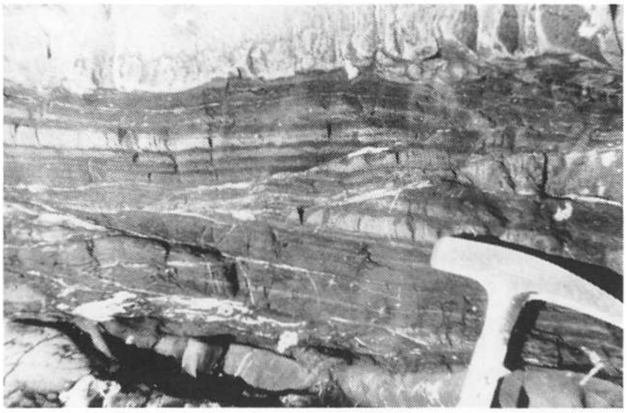


Plate 1-20-5. Kinematic indicator in steeply dipping Laberge wackes showing a minor component of dextral motion on the Nahlin fault. Outcrop is on the southern shore of Teresa Island.

Plate 1-20-6. The Atlin Mountain fault outcrops on several ridges west of the summit of Atlin Mountain (arrows point to fault; view to the southeast). It is a southwest-dipping thrust that places Laberge Group strata against Cache Creek volcanic rocks and the Atlin Mountain pluton.

Laberge rocks within 1 to 7 kilometres of the Nahlin fault display two phases of coaxial folding. Synclinal antiforms and overturned strata occur on Atlin Mountain (Figure 1-20-4), Teresa Island and on the lakeshore north of Mount McCallum (Figure 1-20-3). This structural orientation has not been previously described in the Laberge Group; it may be related to compression during or after convergence of the terranes. Because the geometry of the Nahlin fault is so poorly constrained in this area, it is difficult to assess whether these structures indicate that the Stikine Terrane has locally overridden the Atlin Terrane.

The Nahlin fault zone is crosscut by several intrusions including the Birch Mountain, Atlin Mountain and Pike River plutons. The Birch Mountain pluton plugs and distorts the trace of the Nahlin fault on Teresa Island. On Atlin Mountain, the pluton and the Nahlin fault are cut by the Atlin Mountain fault which outcrops on several ridges west of the summit (Plate 1-20-6). It is a northeast-directed reverse fault that dips approximately 50° southwest and places Laberge Group strata over Cache Creek volcanic rocks and the Atlin Mountain pluton (Figure 1-20-5). A synclinal antiform in the Laberge Group next to the fault may be a drag fold associated with motion on it.

The relationship between the Atlin Mountain fault and the Nahlin fault is not certain although they appear to be separate and distinct structures based on the width of the zone of deformation associated with each. The Nahlin fault is represented by a complex zone of imbrication approximately 5 kilometres wide in the Pike River, whereas the Atlin Mountain fault is confined to a zone only a few metres in width. The Atlin Mountain fault is interpreted as a later structure, possibly representing a back thrust that truncates the Nahlin fault zone. Samples collected from the plutons for U-Pb isotopic dating may help constrain the age of these structures.

MINERALIZATION

Known mineral occurrences in the Dixie Lake area are limited to placer gold deposits. Small placer operations are active on McKee Creek, Wilson Creek and the O'Donnel River. All of these deposits occur along major fault zones. Bloodgood *et al.* (1989) and Ash and Arksey (1990, this volume) have demonstrated that many lode deposits are associated with the occurrence of ultramafic rocks along major faults in the Atlin area and have suggested that these structures are the pathways for mineralizing fluids. This is supported in the Dixie Lake area by the development of pervasive pyrite mineralization and minor carbonatization within the O'Donnel thrust. However, the lack of known lode occurrences in the Dixie Lake area may reflect the relative absence of ultramafic rocks. The Atlin Provincial Park occupies a large segment of the Teresa Island map sheet and is closed to mineral exploration.

Altered ultramafic rocks, including listwanite and carbonatized harzburgite and dunite, occur intermittently along the entire trace of the Nahlin fault in the Dixie Lake, Teresa Island and Atlin Mountain areas. The Nahlin fault zone is thought to be a deep penetrating structure, and like the Llewellyn fault (Mihalynuk, *et al.* 1989) may serve as a conduit for extensive fluid flow. Thus the Nahlin fault zone may have significant economic potential especially where sizeable ultramafic bodies occur.

Potential for epithermal mineralization exists along east-northeast trending Eocene faults which crosscut the Sloko Group and the Nahlin fault. North of the Pike River, sporadic exposures of ultramafic rocks occur within a linear belt at least 11 kilometres in length. Extensive quartz-carbonate stockworks are developed parallel to the regional trend and are crosscut by chalcedonic veins trending consistently east-northeast. Similar relationships have been observed farther south along the Nahlin fault, and in the Atlin area an epithermal overprint has been observed at the Pictou showing. Mineralization on the Atlin Ruffner property is confined to a series of structures with a similar northeast trend. Areas of alteration and mineralization associated with the Nahlin fault zone which are crosscut and overprinted by these later northeast-trending epithermal features may provide specific targets for mineral exploration.

CONCLUSIONS

- Lithologies similar to those described by Monger (1975) outcrop in the Dixie Lake map area and primary lithologic contacts have been recognized within Cache Creek rocks.
- Southeast-directed thrust faults are important structural features east of the Nahlin fault. Faults correspond to major drainages, and movement is frequently accommodated along lithologic contacts, placing older over younger strata within imbricate sequences.
- Tear faults are an integral part of thrust deformation and have significantly influenced the map pattern within the Cache Creek rocks.
- The Nahlin fault is thought to extend to deep structural levels. Its surface expression is characterized by a wide zone of deformation and imbrication.
- A minimum age of movement on the Nahlin fault is constrained by the Birch Mountain pluton which plugs the fault on Teresa Island. The precise age of this intrusion has yet to be determined.
- Areas of highest mineral potential are structurally complex zones in which ultramafic rocks occur. The southern extension of the Nahlin fault and the Nahlin ultramafic body are good targets for gold exploration. In addition, the potential for epithermal gold mineralization has been recognized along late east-northeast-trending structures.

ACKNOWLEDGMENTS

The authors would like to express their thanks to Ingrid Gertz and Louise Maddison for their excellent assistance which contributed greatly to the success of the field season. Jay Jackson of the University of Arizona, Chris Ash, Ron Arksey of the Geological Survey Branch, and the staff of Homestake Mineral Development Company, especially Darcy Marud and Joanne Bozek, are gratefully acknowledged for their willingness to share information and discuss the geology of the Atlin area. Special thanks are due to Fabrice Cordey of the Laboratoire de Stratigraphie, Paris for an educational and enjoyable field visit. Enlightening discussions with Bill McMillan, Don MacIntyre, Dave Lefebure and Tom Schroeter are greatly appreciated, and sincere thanks are extended to Gordon Heynen for safe and reliable helicopter service.

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NOTES

THE BLUE DOME FAULT: THE EVOLUTION OF A TRANSFORM STRUCTURE INTO A THRUST FAULT IN THE SYLVESTER ALLOCHTHON, CASSIAR MOUNTAINS, BRITISH COLUMBIA* (104O/9, 16; 104P/12, 13)

By JoAnne Nelson

KEYWORDS: Structural geology, ophiolites, Blue Dome fault, Sylvester allochthon, transform faults.

INTRODUCTION

The Blue Dome fault plays a major role in the architecture of the Sylvester allochthon. It runs nearly the entire length of the allochthon (Figure 1-21-1), a northwest-trending zone of anastomosing faults, in general dipping more than 45° to the southwest. Prior to the 1989 field season it was known to extend as far south as the Dease River, and as far north as Blue Dome. Limited field observations suggest it does not pierce the basal thrust of the allochthon either to the south (H. Gabrielse, personal communication 1988) or to the north in the Tootsee Lake map area. This field relationship, which would establish it as a pre-emplacement fault, was investigated by detailed field mapping in the area northwest of Blue Dome in 1989.

NATURE OF THE BLUE DOME FAULT ZONE

The Blue Dome fault zone is generally steeply dipping, in contrast to the myriad Sylvester thrust faults, which are flat to gently dipping. In several localities it truncates such thrusts, for instance the Division II/Division I contact on Mount Pendleton and Blue Dome (Figure 1-21-1). It and its splays contain pods of scaley serpentinite which tend to be serpentinite mélanges containing blocks of gabbro and polymictic depositional breccia with clasts of ultramafic and lower crustal rocks (Plate 1-21-1a). Displacement indicators from the serpentinites (Nelson and Bradford, 1989; and field observations with V. Hansen, 1989) show predominantly dextral transcurrent motion.

The Blue Dome fault marks a Mississippian volcanic facies boundary (Nelson and Bradford, 1989). East of it, Division I, the lowest sedimentary division of the Sylvester allochthon, is very thick because of the presence of Unit IMsi, a sequence of Mississippian black argillite, chert, sandstone and calcarenite. This unit is structurally overlain by Pennsylvanian-Permian basalts (Unit IIPPvs) at the base of Division II. West of the Blue Dome fault, Division I is very thin; the basal unit of Division II, Unit IIMvs, is a sequence of Mississippian sediments identical to those in Unit IMsi, but interbedded with 10 to 90 per cent basalts.

Unpublished aeromagnetic data at 1:25 000 scale from the Sylvester allochthon, kindly provided by Brinco Ltd. (now Western Canadian Mining Corporation), reinforce the concept of a steep dip and the anastomosing nature of the Blue Dome fault zone. In contrast with gently dipping ultramafic sheets such as the Zus Mountain and Blue River bodies, the magnetic signature of the Blue Dome fault is a string of linear highs flanked to the east by magnetic lows. This string may bifurcate, as west of Hot Lake, where two linear magnetic highs coincide with the two sides of Unit IIMPvsu, a slivered supracrustal unit which Nelson and Bradford (1989) included within the Blue Dome fault zone. Similar splays are indicated in the area of limited exposure southeast of the Blue River in the Blue Dome map area. One of these splays, located 5 kilometres east of the Blue River, contains serpentinite and gabbro breccia. It is overlain by a small outlier of Triassic Table Mountain limestone, from which Norian conodonts have been recovered (M.J. Orchard, personal communication, 1987). This relationship suggests that motion on the Blue Dome fault was pre-Late Triassic.

NATURE OF THE BLUE DOME FAULT ZONE IN 1989 MAP AREA

Field mapping in 1989 covered the northwestern edge of the Blue Dome map area (104P/12), the northeastern corner of the Chromite Mountain map area (104O/09), the southeastern corner of the Tootsee Lake map area (104O/16) and the southwestern corner of the One Ace Mountain map area (104P/13). Like most map-boundary regions, this contained much essential information. The Blue Dome fault zone at this location is roughly 1 kilometre wide. It consists of scaly, tectonized serpentinite that encloses blocks and slivers of coarse-grained gabbro and polymictic breccias. The polymictic breccias are similar to those described elsewhere along the Blue Dome fault (Nelson *et al.*, 1988a, Nelson and Bradford, 1989). Clasts include coarse-grained gabbro, serpentinite, basalt, diabase, greenschist and chert. At several exposures breccias consisting of sand-sized clasts are interbedded with grey argillaceous cherts that contain sponge spicules. Conodont determinations on the cherts are in progress.

On the ridge 15 kilometres northwest of Blue Dome, the Blue Dome fault zone dips steeply west but flattens abruptly eastward, in continuous outcrop, to form a tongue-like structure as shown in cross-section A-A′ Figure 1-21-2. It forms several flat klippen on the crest of the ridge. These klippen appear on the 1:25 000 map (Nelson *et al.*, 1988b), but were not considered to be part of the Blue Dome fault zone. Continuing north, Rocky Top, the southernmost ridge in the Tootsee Lake map area, mapped in 1986 as microdiorite (Nelson and Bradford, 1987), and later assigned to Division III, is underlain by gabbro breccias and diabases of the Blue Dome fault zone. This ridge exposure extends 5 kilometres in a southwesterly direction, perpendicular to the strike of the Blue Dome fault zone. Given the structural

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

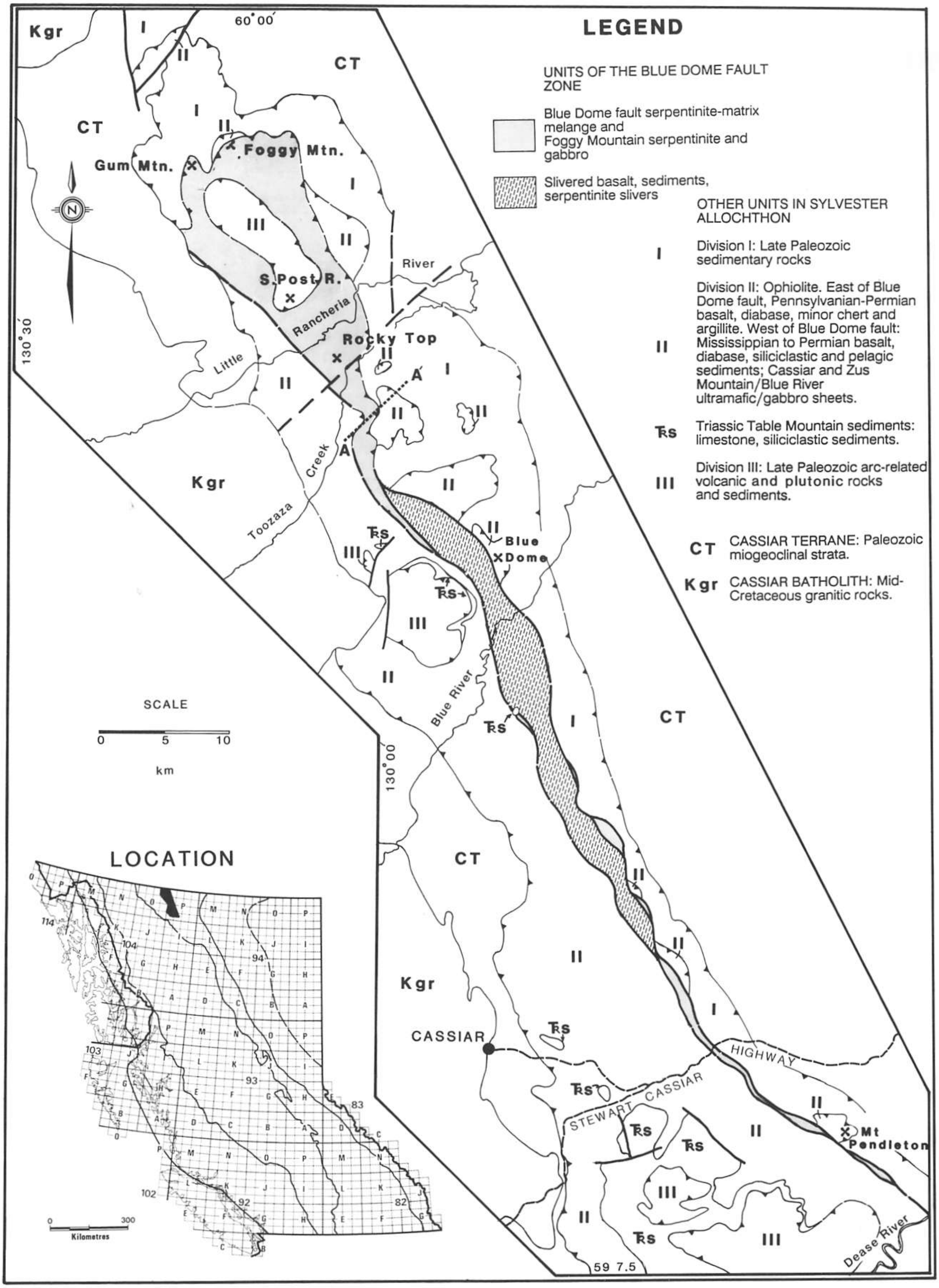


Figure 1-21-1. Regional map of the Blue Dome fault zone (south of Rocky Top) and the Foggy Mountain serpentinite mélange (Rocky Top and north). These two units are contiguous and correlative, differing only in that the Blue Dome fault zone is a crosscutting feature while the Foggy Mountain body is a gently dipping, sheet-like unit bounded by thrusts.

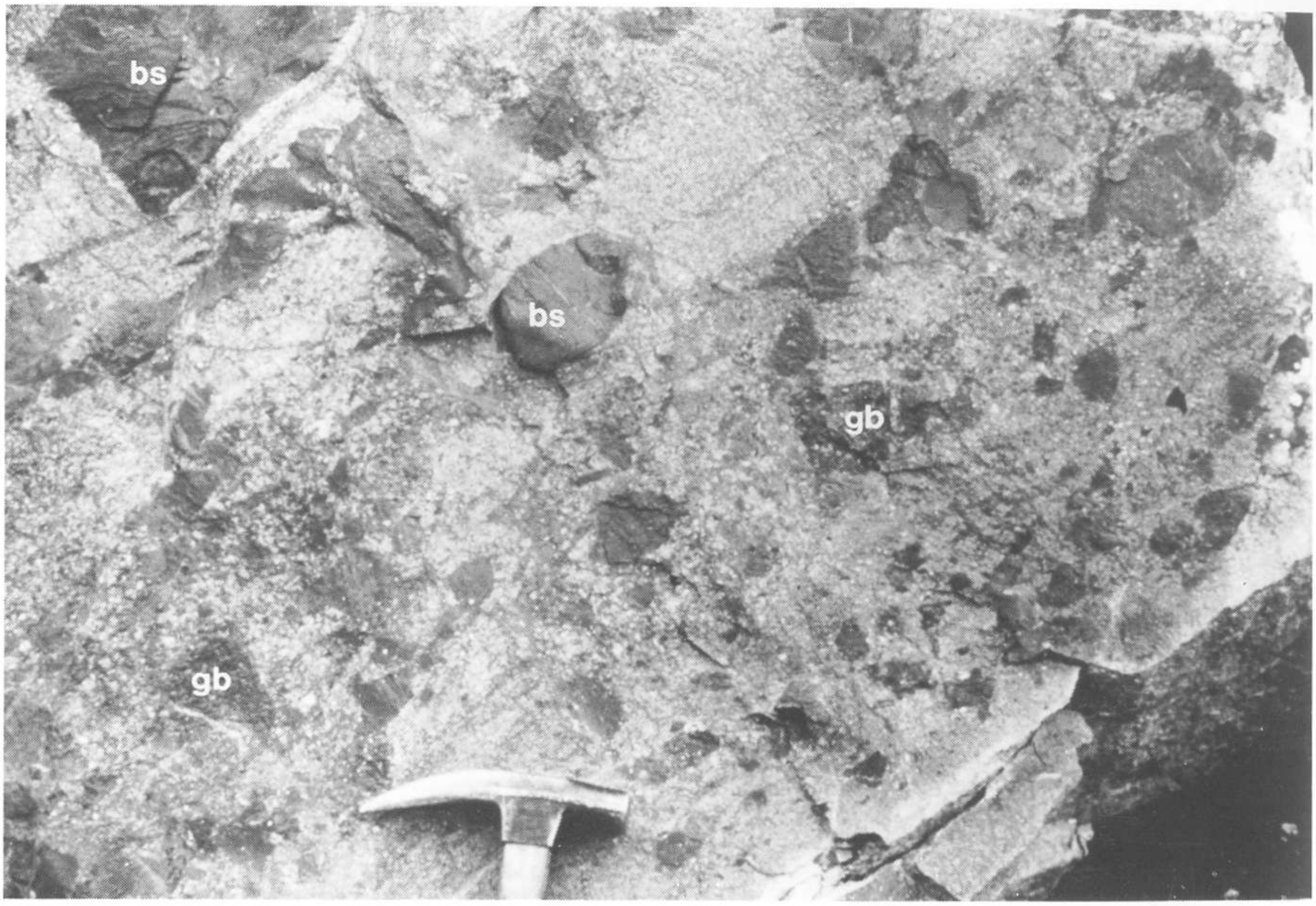


Plate 1-21-1a. Typical polymictic depositional breccia from a sliver within the Blue Dome fault zone, containing elements of oceanic lower crust and upper mantle (bs = basalt, gb = gabbro).

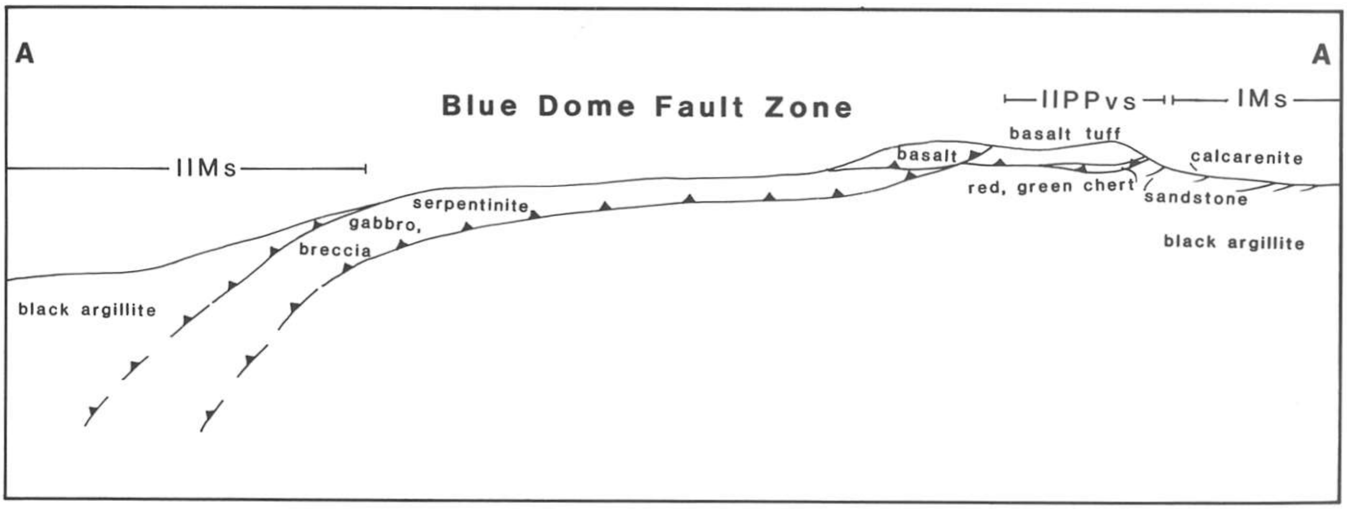


Figure 1-21-2. Cross-section of northern Blue Dome fault zone, in the area where it flattens in an easterly direction to merge into the Foggy Mountain body. IMs – Mississippian siliclastic sediments; IIMs = Mississippian siliciclastic sediments, interbedded with basalts to west; IIPPvs = Pennsylvanian – Permian basalt, diabase, chert, argillite.

flattening of the fault zone immediately south of it, it is reasonable to show the Blue Dome fault on Rocky Top is an essentially flat thrust sheet. This sheet would then project northward into the serpentinites of western South Post Ridge, which correspond to the Foggy Mountain gabbro and ultramafite (*see* Nelson and Bradford, 1987).

The Foggy Mountain body is a typical dismembered ophiolite within the Sylvester allochthon. It is a flat sheet that rests either directly on cherts and argillites of Division I or is separated from them by thin slivers of basalt. It is structurally overlain by the Division III trachyandesite-pyroclastic package on South Post Ridge. The Foggy Mountain body is a tectonized serpentinite that contains mountain-sized blocks of coarse-grained gabbro, amphibolite (mylonitic gabbro), a basaltic dike complex, and a variety of breccias: gabbro breccia, epidote-rich brecciated actinolite-quartz mylonite, and a block on Gum Mountain of depositional polymictic breccia in a limestone host (Plate 1-21-1b) which has yielded Permian conodonts (M.J. Orchard, personal communication, 1988). The correlation of the Foggy Mountain unit with the Blue Dome fault zone is an important step in unravelling the deformational history of the Sylvester allochthon.

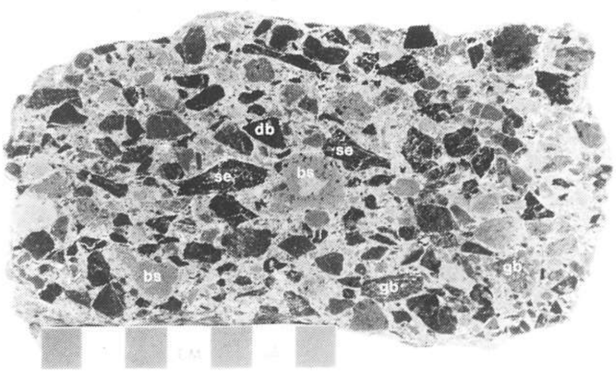


Plate 1-21-1b. The Gum Mountain breccia sliver. The limestone matrix has yielded Permian conodonts. Clasts include basalt, coarse-grained gabbro identical to gabbro that outcrops of Foggy Mountain, diabase and red and green chert.

DISCUSSION AND CONCLUSIONS

Substantial evidence can now be brought together to constrain the timing and nature of motion on the Blue Dome fault zone and its evolving role as a major crustal break within the Sylvester allochthon. Its earliest motion is recorded by the polymictic depositional breccias. The Permian age of the Gum Mountain breccia is now seen as the age of a breccia sliver that formed within the fault zone. This age is equivalent to basalts and sediments in the crustal packages that surround it. It is likely that the Blue Dome fault was initially a transform structure within the marginal basin in which the basalts and sediments of Divisions I and II were accumulating. Diapiric emplacement of serpentinite mélange into high levels in the fault zone occurred at this time. These diapirs, termed "protrusions" by Saleeby (1984) probably reached the sea floor. The breccias formed at the bases of fault scarps and serpentinite protrusions. The spiculitic cherts interbedded with the polymictic breccias, observed in 1989, are a further link between the breccias and typical Late Paleozoic Sylvester sedimentation.

The low-angle dextral displacement indicators in serpentinites – slickenfibres, slickensides and S-C structures, may represent this early motion or, given the ductile nature of serpentinite, any later phase. The fact that the Blue Dome fault truncates thrust faults indicates that at least part of its motion succeeded initial shortening within the allochthon; transcurrent strain may have been partitioned into it during a transpressional event (V. Hansen, personal communication, 1989). Concurrently, or later, the steep Blue Dome fault zone itself was carried and flattened to the east by easterly vergent thrusting. It now consists of two parts, a steep root and a flat thrust composed of serpentinite mélange. In general, it carries Mississippian to Permian volcanics and sediments of Division II (Units IIMvs and IIPPvs) in its hangingwall. On South Post Ridge in the Tootsee Lake map area, however, these hangingwall units are missing and it is directly overlain by Division III. The abrupt disappearance of Division II above it can be accounted for by assuming one thrusting event that brought Division II over the Blue Dome fault zone/ Foggy Mountain body; an interval of erosion; then thrust emplacement of Division III over everything. The time of the interval between these two separate shortening events has been inferred to be Late Triassic, as shown by relationships on Table Mountain in the Cassiar area (Nelson and Bradford, 1988). The onlapping Triassic limestone on the Blue River fault zone southeast of the Blue River provides further evidence for pre-Late Triassic motion. The major motion on the Blue Dome/Foggy Mountain thrust was, then, a Sonoman-aged event.

The root zone of the Blue Dome fault is truncated by the base of the Sylvester allochthon. This "out-of-sequence" relationship – that is, truncation of a thrust by a second thrust located further toward the hinterland – can again be explained by two separate shortening events, one pre-Late Triassic, the second the emplacement of the allochthon after Late Triassic but before mid-Cretaceous time. These events are summarized in the sketches in Figure 1-21-3.

The inferred history of the Blue Dome fault zone, consists of four phases:

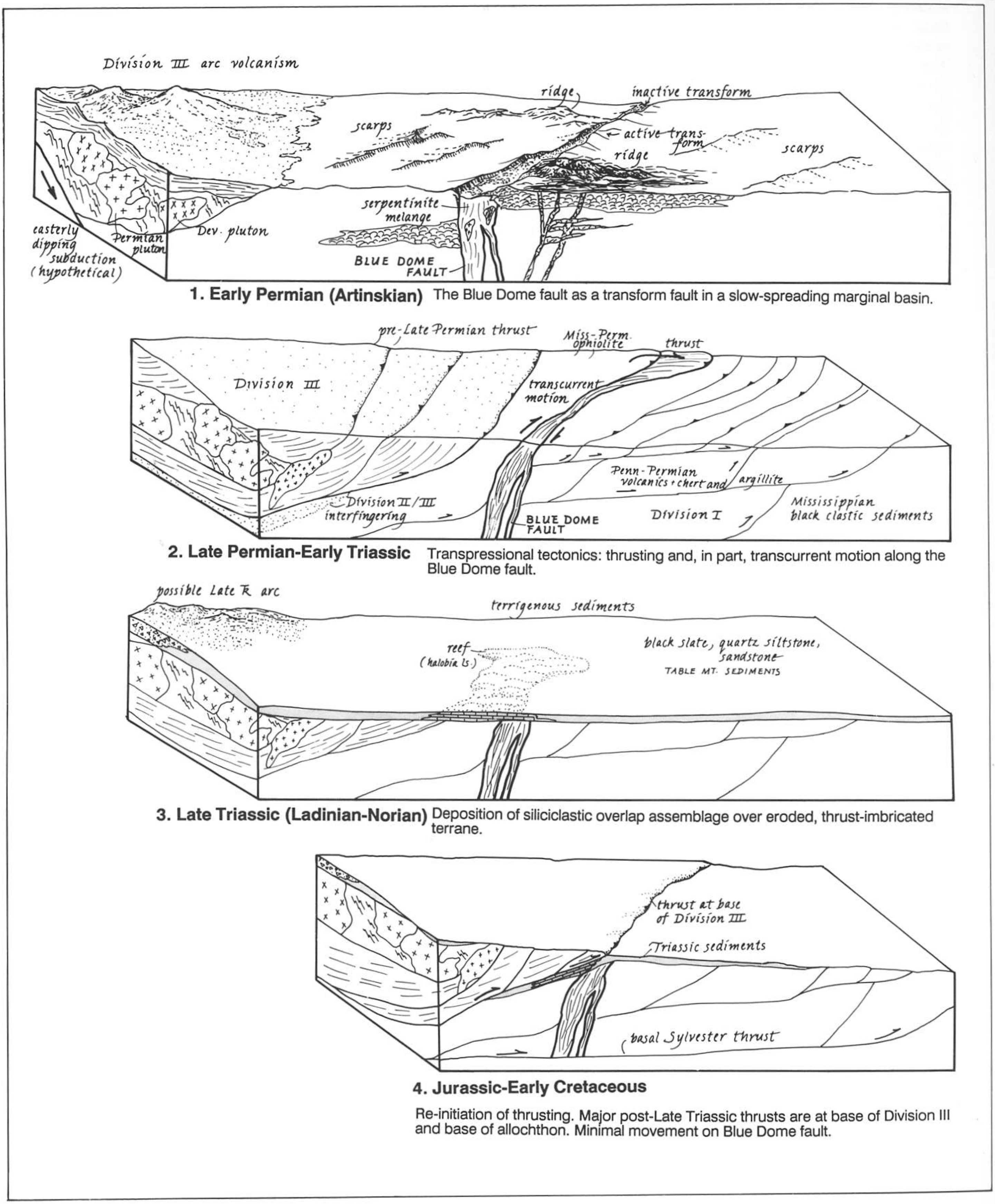
(1) Early Permian and earlier(?): oceanic transform fault within the marginal basin in which Divisions I and II of the Sylvester allochthon were formed.

(2) Late Permian to Early Triassic: transcurrent fault during Sonoman crustal shortening of the marginal basin; finally conversion to a flat easterly verging thrust, the Foggy Mountain serpentinite mélange.

(3) Late Triassic: erosional truncation of early structures; deposition of Table Mountain sediments on top of the thrust-imbricated Late Paleozoic units.

(4) Early Jurassic(?): Blue Dome fault carried east as passive structure during thrusting of Division III on Division II and of the Sylvester allochthon on the North American continental margin.

The steep root of the Blue Dome fault happens to be preserved within the Sylvester allochthon. Given that the allochthon is a very narrow klippe that contains a number of



1. Early Permian (Artinskian) The Blue Dome fault as a transform fault in a slow-spreading marginal basin.

2. Late Permian-Early Triassic Transpressional tectonics: thrusting and, in part, transcurrent motion along the Blue Dome fault.

3. Late Triassic (Ladinian-Norian) Deposition of siliciclastic overlap assemblage over eroded, thrust-imbricated terrane.

4. Jurassic-Early Cretaceous

Re-initiation of thrusting. Major post-Late Triassic thrusts are at base of Division III and base of allochthon. Minimal movement on Blue Dome fault.

Figure 1-21-3. Inferred history of the Blue Dome fault.

sheets of ultramafite and serpentinite mélange, it is likely that these also once had steep roots which have now been removed by erosion of more westerly parts of the allochthon. The existence of the Blue Dome fault, with its complex history, suggests that all of these mantle-derived sheets – the Zus Mountain/Blue River and Cassiar sheets (Nelson and Bradford, 1989) – may once have lain along transform faults separating different crustal segments. It rationalizes their presence within the allochthon, interleaved with sequences of upper crustal rocks. It explains why the allochthon does not contain complete lower crustal-mantle ophiolite sequences, but rather chips, blocks, pods and fragments such as are found along modern transform faults. Finally, as suggested by Karson and Dewey (1978) and Saleeby (1984), the Blue Dome fault provides a clear example of the evolution of a pre-accretion crustal discontinuity into a major thrust.

ACKNOWLEDGMENTS

The ideas expressed in this paper developed substantially during discussions with Vicki Hansen, who predicted the flattening of the Blue Dome fault.

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THE STRUCTURAL GEOLOGY OF THE MOUNT McDAME AREA, NORTH-CENTRAL BRITISH COLUMBIA (104/P)

By David S. O'Hanley

KEYWORDS: Structural geology, Sylvester allochthon, Cassiar platform, Mount McDame, Marble Creek fault, Cassiar asbestos deposit.

INTRODUCTION

Mount McDame is situated in north-central British Columbia, near the town of Cassiar, 15 kilometres northwest of the Stuart-Cassiar highway (Figure 1-22-1). Mapping in the area at 1:25 000 scale was completed in 1988 by Nelson and Bradford (1989). They identified an area of anomalous structure centred on Mount McDame and based on the juxtaposition of successively lower units of the Cassiar platform and the Sylvester allochthon. Mapping at 1:6000 scale was undertaken in 1989 to rationalize the anomalous structure and to relate the geology to that of the nearby Cassiar chrysotile asbestos mine as described by O'Hanley and Wicks (1987) and O'Hanley (1988).

GEOLOGIC SETTING

Mount McDame is composed of both platformal and eugeosynclinal rock sequences. Its lower slopes are underlain by Lower Cambrian to Devonian carbonates and shales

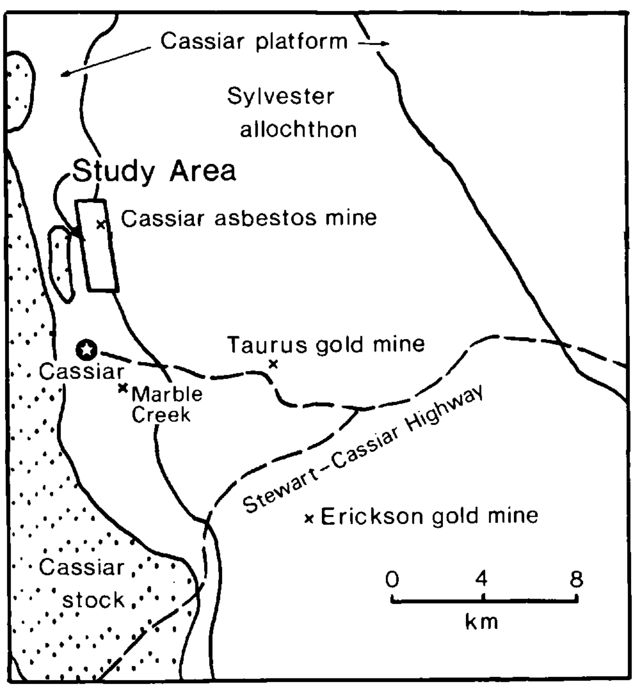


Figure 1-22-1. Location of the study area and the Cassiar asbestos mine on Mount McDame, in north-central British Columbia.

of the Cassiar platform, which is part of the ancestral North American continental margin. The upper part of the mountain consists of Pennsylvanian-Permian (?) argillite and chert, and Mississippian argillite, chert, limestone, greywacke, and greenstone sequences of oceanic and island arc affinities, and serpentinite, all part of the Sylvester allochthon emplaced in Jurassic time (Harms, 1986; Nelson and Bradford, 1989).

The base of the allochthon is the roof thrust in a duplex structure involving some of the platform units. The duplex has its roots in the Road River Group, and thrust sheets are composed of the Road River Group, the Tapioca sandstone and the McDame and Earn groups.

The Cassiar and McDame asbestos deposits occur on Mount McDame within the Sylvester allochthon, close to the basal contact with the platformal units. The Cassiar serpentinite lies between the chert-argillite and the Mississippian sequences.

The Earn Group is found immediately below the allochthon both to the north and south of the study area. Within the map area successively older units are in contact with the base of the Sylvester allochthon from south to north. Nelson and Bradford (1989) projected the Marble Creek fault, a late-stage, high-angle fault, into this contact near the Cassiar mine. No evidence of the Marble Creek fault was found during this study and it is not shown on Figure 1-22-2.

STRATIGRAPHY

The stratigraphy of the map area has been described by Gabrielse (1963) and by Nelson and Bradford (1989), and is only briefly summarized here. However, several marker units were used to document the presence or absence of stratigraphic continuity.

PLATFORM SEQUENCES

The Lower Cambrian Rosella Formation, exposed in the southwest corner of the map area, is recognised by its characteristic white-blue-grey banding in limestone and dolomite. The Cambro-Ordovician Kechika Group is absent. The contact between the Road River Group and the Rosella Formation seems conformable just south of the McDame adits but a quartz vein 20 metres wide marks this contact farther to the south (Figure 1-22-2). Dark grey limestone, a distinctive carbonate-altered volcaniclastic rock and the presence of graptolites identify the Road River Group. The "Silurian siltstones" mark the contact between the Ordovician Road River Group and the Devonian Tapioca sandstone.

The Tapioca sandstone contains a distinctive light grey quartzite, 3 metres wide, that is stratigraphically above

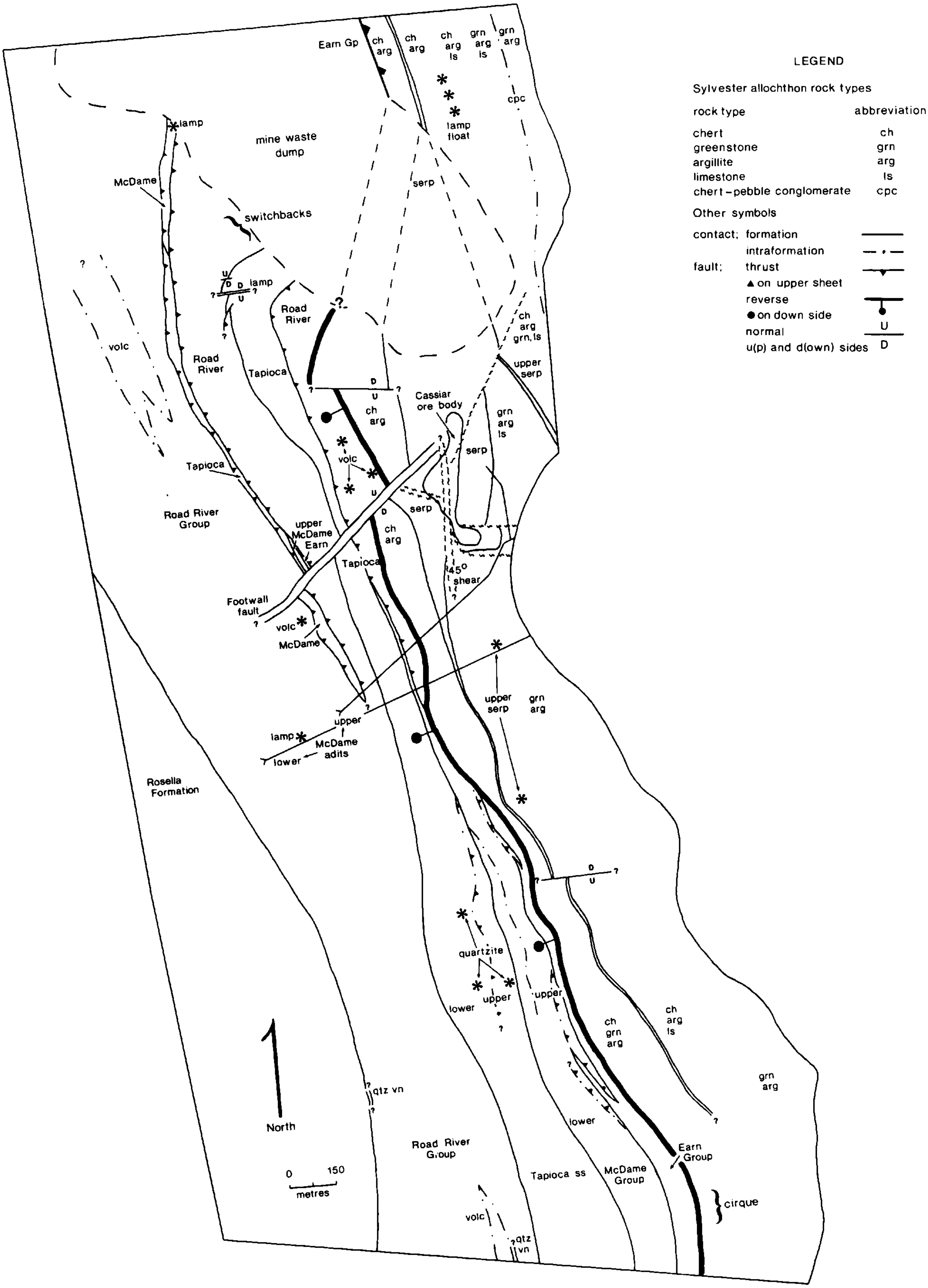


Figure 1-22-2. Interpretive geologic map of the western face of Mount McDame, based on outcrop maps compiled by Lyn O'Hanley (1988, unpublished) and Nelson and Bradford (1989).

interbedded Tapioca sandstone and quartzite. A structural repetition of the Tapioca sandstone south of the McDame adits is suggested by repetition of the quartzite and rotation of bedding, and a thrust fault has been inferred in this locality (Figure 1-22-2). This interpretation is supported by rotation of bedding in the Tapioca sandstone, from 354/72E to 025/47SE as the fault is approached from the west. Immediately north of this area the fetid, fossiliferous unit of the lower McDame is absent, and is not seen again until it crops out in a thrust panel on the switchback road west of the Cassiar orebody. In between these areas, nonfetid, cross-bedded dolomite assigned to the Tapioca sandstone is disconformably overlain by upper McDame limestones, which are in turn overlain by shales of the Earn Group (Plate 1-22-1). This evidence suggests that, at least locally, the lower McDame unit was removed by erosion before deposition of the upper platey limestones. Thus some of the McDame Group west of the mine was absent before deformation occurred; paleo-erosion of the McDame Group has been previously documented by Gabrielse (1963).

SYLVESTER STRATA

Nelson and Bradford (1989) identified two rock sequences in the area that belong to the allochthon. A Mississippian suite consists of chert, argillite, limestone, greenstone and a distinctive chert-pebble conglomerate. A second suite, thought to be of Pennsylvanian to Permian age, consists of chert and argillite. These two packages were thought to be separated by a fault containing the Cassiar serpentinite; with the Pennsylvanian-Permian rocks in the footwall, and the Mississippian rocks in the hangingwall of the serpentinite in the mine. Detailed mapping north of the mine located chert-argillite sequences, similar to those in the footwall of the mine, stratigraphically above outcrop consisting of talc and listwanite cobbles, representing altered serpentinite. South of the mine altered greenstone or tuff pods in argillite are found stratigraphically below serpentinite.

Traces of the upper serpentinite, defined in the hangingwall of the Cassiar orebody (Figure 1-22-2), are exposed to

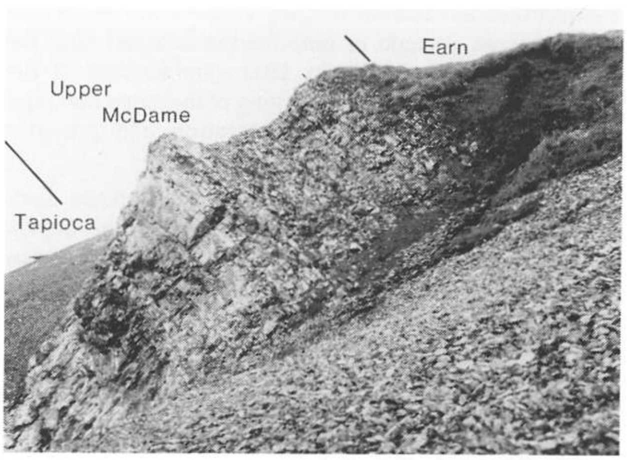


Plate 1-22-1. View to the north of outcrop containing karst features in Tapioca sandstone, which is overlain by upper McDame bedded limestones and Earn shales. McDame limestones are approximately 5 metres wide.

the south. Both occurrences are stratigraphically above outcrop consisting of serpentinite and talc schist that correlate with the footwall serpentinite in the mine. A normal fault just south of this outcrop offsets the serpentinite and suggests that the serpentinite to the south is part of the lower rather than the upper body.

The fact that the serpentinite does not separate chert-argillite sequences from those containing limestone and greenstone suggests that the serpentinite cuts obliquely across a Mississippian sequence which records a facies(?) change, from chert-argillite to chert-argillite-limestone to argillite-limestone-greenstone. Alternatively, the sequence may consist of rocks of both Pennsylvanian and Mississippian age.

LAMPROPHYRE DIKES

Biotite-rich lamprophyre dikes crop out in four localities and are present as float in three others (Figure 1-22-2). They cut all lithologies above the Rosella Formation. One dike, 4 metres wide, fills an east-striking normal fault that offsets the Road River – Tapioca contact. This dike, and one other found in the McDame limestones on the switchbacks west of the mine, is undeformed. In contrast, two dikes found in the Road River Group above the McDame adits have schistose textures at their margins.

STRUCTURAL GEOLOGY

Several types of faults are identified in the study area based on orientation and the presence or absence of fault fabrics. North-northeast-striking faults dip steeply to gently; they are recognised by changes and discontinuities in bedding orientation. Their present orientation suggests normal displacement but, if the affect of later tilting is removed, they become low-angle reverse faults. Several of the faults near the upper McDame – Earn contact define frontal ramps (Plate 1-22-2), and indicate west-over-east, east-directed transport. These faults are associated with duplex formation during emplacement of the allochthon (Harms, 1986). Faults in the Road River Group on the mine switchbacks, with similar orientations to those described above, contain fault fabrics (slickensides, fault cleavage and gouge) that indicate early normal displacement and subsequent reverse displacement. This interpretation is based on the assumption that fault gouge, which in some faults is similar to the foliated cataclasites of Chester *et al.* (1985) and Chester and Logan (1987), is younger than fault-zone cleavage.

East to northeast-striking, high-angle normal faults are recognised by offsets of rock contacts (Plate 1-22-3) and the presence of lamprophyre dikes; gouge is present in some faults. The two best examples are the lamprophyre-filled fault mentioned earlier, and the Footwall fault in the footwall of the Cassiar orebody (O'Hanley, 1988; Figure 1-22-2). These structures, which may or may not contain lamprophyre dikes, offset all other faults. In particular, the Footwall fault offsets the inferred Marble Creek fault.

These faults are probably more common than shown in Figure 1-22-2 for two reasons; they are difficult to identify due to the lack of fault fabrics and discontinuous outcrop; the presence of lamprophyre float suggests the existence of

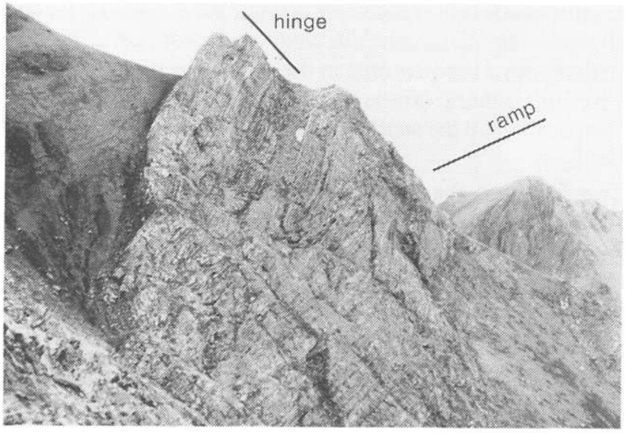


Plate 1-22-2. Frontal ramp in upper McDame limestones recognized by the difference in bedding orientation across the fault trace.

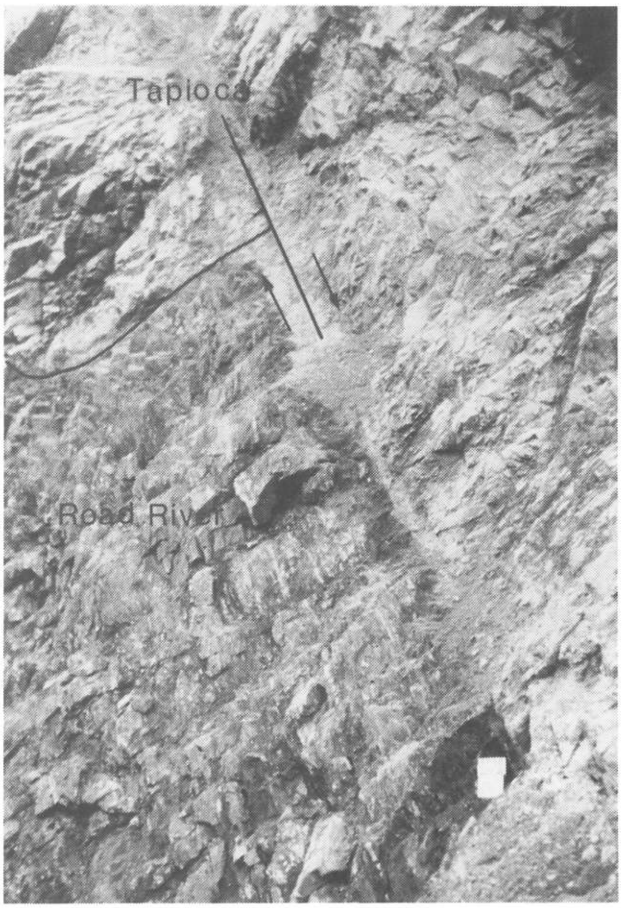


Plate 1-22-3. Normal, late-stage fault cutting the Road River–Tapioca contact. Scale bar is 1 metre.

normal faults, although the faults were not identified. The float occurs in drainage gulleys, as do known normal faults, so it is possible that every gulley represents a fault trace. A normal fault may explain the sharp change in width of the Cassiar serpentinite that occurs beneath the mine waste dump. The above discussion suggests that the continuity of some faults and rock contacts in Figure 1-22-2 is questionable.

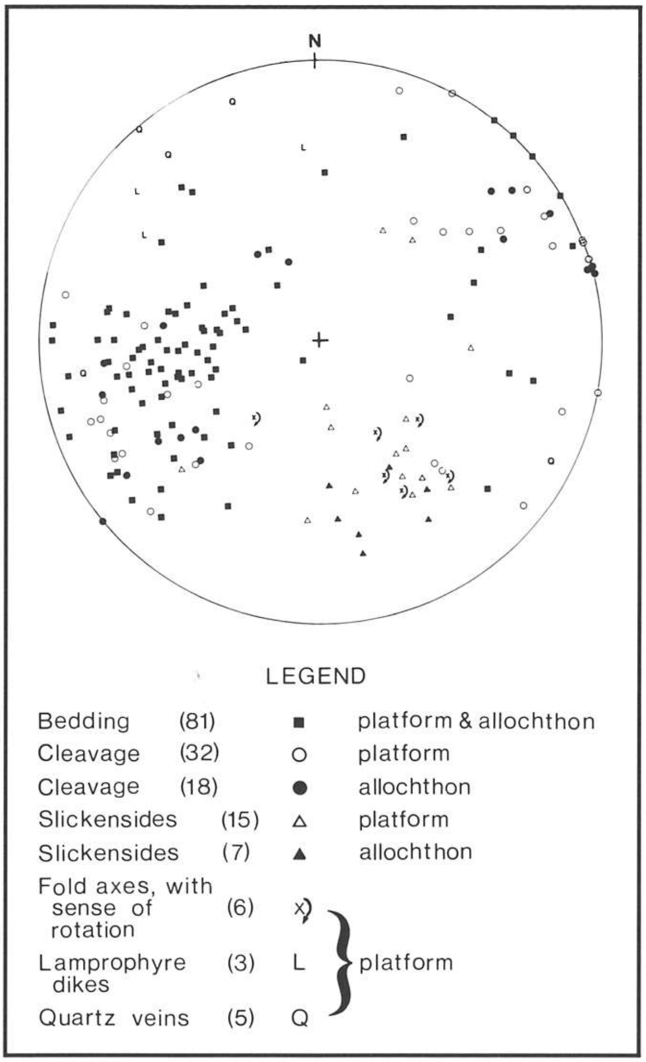
The south wall of the cirque at the southern limit of the map area consists of a structurally unrepeated platformal sequence, except that the Kechika Group is absent. North of the mine area, the Road River Group is structurally repeated three times, indicating that the duplex is well established (Nelson and Bradford, 1989). The presence of frontal ramps involving the Earn and upper McDame groups, where there is no repetition of the underlying strata, suggests the basic structural pattern is explained by increasing penetration of duplex-related deformation and faulting from north to south. Thus thrust faults are more common on the switchbacks west of the mine than farther to the south, and penetrate deeper into the platformal sequence.

STRUCTURAL DATA

The dominant bedding attitude in the study area has a northwesterly strike with a steep northeasterly dip (Figure 1-22-3). The Road River Group contains bedding-parallel cleavage, and both dextral and sinistral folds of calcite veins, which suggest that it is a flattening cleavage. The Sylvester chert-argillite sequence contains axial-planar cleavage that has the same orientation as the cleavage in the Road River Group. The more massive dolomites in the Tapioca and the McDame contain quartz veins orientated perpendicular to bedding and cleavage (Figure 1-22-3). The lamprophyre dikes and the normal faults have the same orientation as the quartz veins. All of these data are consistent with duplex-related structures.

Asymmetric folds of Road River cleavage and calcite veins, and asymmetric quartz-fibre pressure shadows around pyrite grains, are consistent with southwest-directed, east-over-west motion after the formation of the cleavage. This sense of motion is also consistent with the reverse displacement indicated by fault fabrics in the Road River Group. Slickensides, stretching directions, and fold axes both in the platform and allochthon rocks also suggest east-over-west, southwest-directed motion (Figure 1-22-3). All these data indicate that an episode of deformation affected both the platform and allochthon rocks after emplacement of the allochthon. It is attributed to the tilting of the strata into their present steeply east-dipping orientation and predates emplacement of the lamprophyre dikes.

The observations made during this study indicate three deformation events: an early event associated with duplex formation during emplacement of the allochthon, an intermediate event associated with tilting of the strata, and a late event associated with normal faulting and emplacement of lamprophyre dikes. As some of the lamprophyre dikes are physically deformed, deformation must have continued after dike emplacement but no other structural data that might reflect this late-stage deformation were documented. It is possible however, that the late-stage deformation is kinematically consistent with the earlier deformations, and that some structural features found in faults within the Road River Group formed during the late-stage deformation rather than the episode that tilted the strata.



The Marble Creek Fault

The Marble Creek fault was defined by Nelson and Bradford (1989) in the Marble Creek drainage basin south of the town of Cassiar. In this area the Kechika Group is not present between the Rosella Formation and the Road River Group; Nelson and Bradford invoked high-angle faulting to account for its absence. The juxtaposition of Road River against Sylvester cherts in the Cassiar mine, and the absence of the McDame and Earn groups in between Road River and Sylvester rocks in the lower McDame adit, could both be explained by a high-angle fault. Based on these observations Nelson and Bradford extended the Marble Creek fault northward obliquely across Mount McDame.

Surface mapping above the adits identified a distinctive chert-argillite sequence (red and black chert: R&B in Figure 1-22-4) that is also present in the upper adit, but not in the lower adit, which contains only ribbon chert. This observation is important for two reasons. First, correlation of the chert at the surface with that in the upper adit constrains the dip of the inferred Marble Creek fault and essentially restricts it to being a bedding-parallel structure, rather than a vertical fault. Second, this observation suggests the chert-argillite unit is missing both down dip and along strike to the south of the McDame adits.

Other observations argue against the existance of the Marble Creek fault. The Footwall fault offsets the inferred trace of the Marble Creek fault; this relationship indicates that the Marble Creek fault could not be a late-stage event. The absence of the red and black chert in the lower adit and at the south end of the map area, indicates that some Sylvester units are also missing, in addition to the platform units mentioned earlier. Sylvester units disappear from north to south, and platform units disappear from south to north; the missing units can be accounted for by invoking dextral-reverse displacement on the basal Sylvester fault that brings the allochthon to the southwest obliquely across the platform units. However, no data from fault fabrics within the basal fault were found that either confirmed or denied this hypothesis.

This model, together with the evidence that the duplex is incipiently developed at the south end of the map area, explains the observed distribution of rock types. A second line of reasoning supporting this interpretation is that the stratigraphic separation between the Rosella Formation, which is below the duplex, and the base of the allochthon, which is above it, is constant. If the Marble Creek fault operated as suggested by Nelson and Bradford, the platform sequence should be thinner to the north because part of it would have been faulted out. Furthermore, the chert-argillite unit should still be present near the cirque at the south end of the map area.

In summary, a fault is needed to explain the 1:25 000 mapping of Nelson and Bradford (1989), and the 1:6000 mapping completed during this study. It should be placed at the base of the allochthon and considered a nearly bedding-parallel fault. The absence of the Kechika Group from the Marble Creek area was not addressed by this study, but the presence of carbonate-altered volcanic rocks in the Road River Group in the map area, and nowhere else, and the apparent conformable contact between the Road River Group and the Rosella Formation in one locality, suggests that the Kechika is missing due to a facies change rather than as the result of faulting.

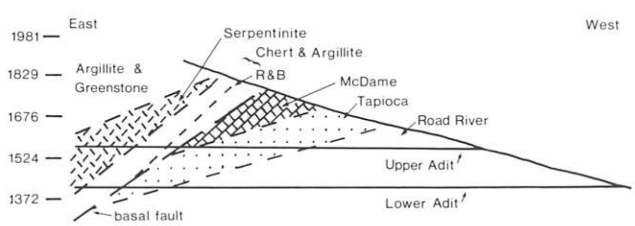


Figure 1-22-4. Schematic composite cross-section through Mount McDame parallel to the upper McDame adit. Section includes data from both the upper and lower adits, and surface mapping done during the current study. Horizontal scale equals the vertical scale. Correlation of red and black argillite (R&B) from surface to upper adit limits dip of inferred Marble Creek fault (Nelson and Bradford, 1989). Note that neither the red and black argillite nor the McDame Group is present in the lower adit.

CASSIAR AND McDAME ASBESTOS MINES

The ore zone in the Cassiar mine consists of specific serpentine minerals (chrysotile and antigorite) and textures (interlocking texture) developed during shear-zone controlled fluid flow and deformation (O'Hanley and Wicks, 1987; O'Hanley *et al.*, 1989). The fluids could be either of magmatic or metamorphic origin based on $\delta^{18}O$ fluid values calculated at 300°C from serpentine-magnetite mineral pairs (O'Hanley *et al.*, 1989). O'Hanley (1988) reported two episodes of faulting in the 45° shear, based on fault-zone fabrics; an earlier east-side-down displacement, and a later dextral-reverse displacement. The formation of the ore-zone textures and the asbestos veins is associated with the latter event.

The work reported here permits the earlier work of O'Hanley and Wicks (1987) and O'Hanley (1988) to be interpreted in the context of the geology of Mount McDame. The 45° shear, prevously interpreted as a normal fault, becomes a low-angle reverse fault when the affect of tilting is removed. Its orientation and sense of movement are consistent with emplacement structures, but it is not certain that it is in fact an emplacement-related structure. The ore zone formed during the later episode of deformation, characterised by dextral-reverse faulting that tilted the strata. This interpretation suggests that strike-slip faulting is not as important in the formation of chrysotile asbestos deposits as previously thought (O'Hanley 1988), although a change in stress regime still seems necessary to explain the observations from the Cassiar mine, in particular, the existence of two generations of asbestos veins with different orientations.

TIMING OF EVENTS

Duplex formation occurred during allochthon emplacement in Jurassic time (Harms, 1986). Emplacement must predate the Cassiar batholith, which is 89 to 110 Ma (Sketchley *et al.*, 1986). Dextral-reverse fault displacement is constrained to be older than the normal faulting that was accompanied by emplacement of lamprophyre dikes. The few lamprophyre dikes from other areas that have been dated are 110 ± 4 Ma (A. Panteleyev, cited in Sketchley *et al.*, 1986). Thus, asbestos formation, and tilting of the strata, predates the intrusion of the Cassiar batholith and the nearby Cassiar stock (69.3 to 76.5 Ma: Sketchley *et al.*, 1986). However, the physical deformation of the lamprophyre dikes could be due to emplacement of the Cassiar stock.

CONCLUSIONS

Structural data obtained during this study can be related to three episodes of deformation. The oldest structures are consistent with east-directed, west-over-east thrust faulting accompanying duplex formation during emplacement of the Sylvester allochthon. This deformation is accommodated by faults in the limestones with little fabric, and by faults in the Road River Group with cleavage. A younger deformation, characterized by dextral-reverse fault displacement on many of the same faults, is attributed to tilting of the strata into their now steeply east-dipping orientation. This deformation is older than the normal faulting and lamprophyre dike emplacement, which predate the Cassiar batholith. The formation of the Cassiar and McDame asbestos deposits occurred during the tilting event.

ACKNOWLEDGMENTS

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EVIDENCE FOR A CRYPTIC INTRUSION BENEATH THE ERICKSON-TAURUS GOLD-QUARTZ VEIN SYSTEM, NEAR CASSIAR, B.C.*
(104P/4, 5)

By JoAnne L. Nelson

KEYWORDS: Economic geology, Erickson-Taurus, gold-quartz veins, alteration assemblages, metamorphic facies, Sylvester allochthon, fluid inclusions.

INTRODUCTION

The Erickson and Taurus mines are located near Cassiar (Figure 1-23-1) on the Stewart-Cassiar Highway 120 kilometres south of the Alaska Highway junction. At the time of writing, these mines are closed pending evaluation of potential new reserves. Ore production at Erickson and Taurus has come from gold-quartz veins that are part of the larger Erickson-Taurus system (Nelson *et al.*, 1989). This system has been previously well described by Diakow and Panteleyev, 1981; Panteleyev and Diakow, 1982; Sketchley, 1986; Sketchley *et al.*, 1986, Dussell, 1986 and Boronowski, 1988. Veins occur in basalts of Division II of the Sylvester allochthon and along the base of the Triassic Table Mountain sediments which overlie them (Nelson *et al.*, 1989; Harms *et al.*, 1989). Productive veins are limited to a few hundred metres below the base of the Triassic sediments. The Erickson-Taurus system consists of a set of east-northeast-trending vein zones, in which individual steeply dipping veins strike 045° to 070°; and of flat veins such as the Vollaug vein at the base of the Table Mountain sediments (Panteleyev and Diakow, 1982). The vein system is slightly elongated in a northerly direction (Figure 1-23-1).

THE PROBLEM

The Erickson-Taurus veins show typical mesothermal characteristics. They are white bull quartz with orange-weathering carbonate alteration envelopes in basalt; some, notably the Vollaug vein, are finely ribboned with carbon. Sketchley (1986) studied alteration patterns at Erickson and identified alteration assemblages in basalts including ankerite, siderite, kaolinite, dolomite, pyrite, carbon, titanium oxide, arsenopyrite and sericite. Higher gold grades are associated with slivers of serpentinite along the base of the Table Mountain sediments. Most of these serpentinites are altered to talc-ankerite schists and quartz-ankerite-mariposite listwanites. Ore mineralogy in the veins includes free gold, pyrite, tetrahedrite, chalcopyrite, arsenopyrite, and sphalerite. Gold/silver ratios average slightly greater that 1. Fluid inclusions in the bull quartz are H_2O-CO_2-NaCl solutions, generally three phase at room temperature; total homogenization temperatures cluster in the range 250 to 300°C (Nelson and Bradford, unpublished data).

In terms of these characteristics, the Erickson-Taurus veins strongly resemble classic mesothermal deposits such as the California Mother Lode (Bohlke and Kistler, 1986; Weir and Kerrick, 1987), Bralorne (Leitch and Godwin, 1988), the Juneau belt veins (Goldfarb *et al.*, 1988), and Archean volcanic-hosted deposits such as the Sigma (Sibson *et al.*, 1988) and Giant Yellowknife mines (Allison and Kerrich, 1980). Their structural setting is a point of very significant

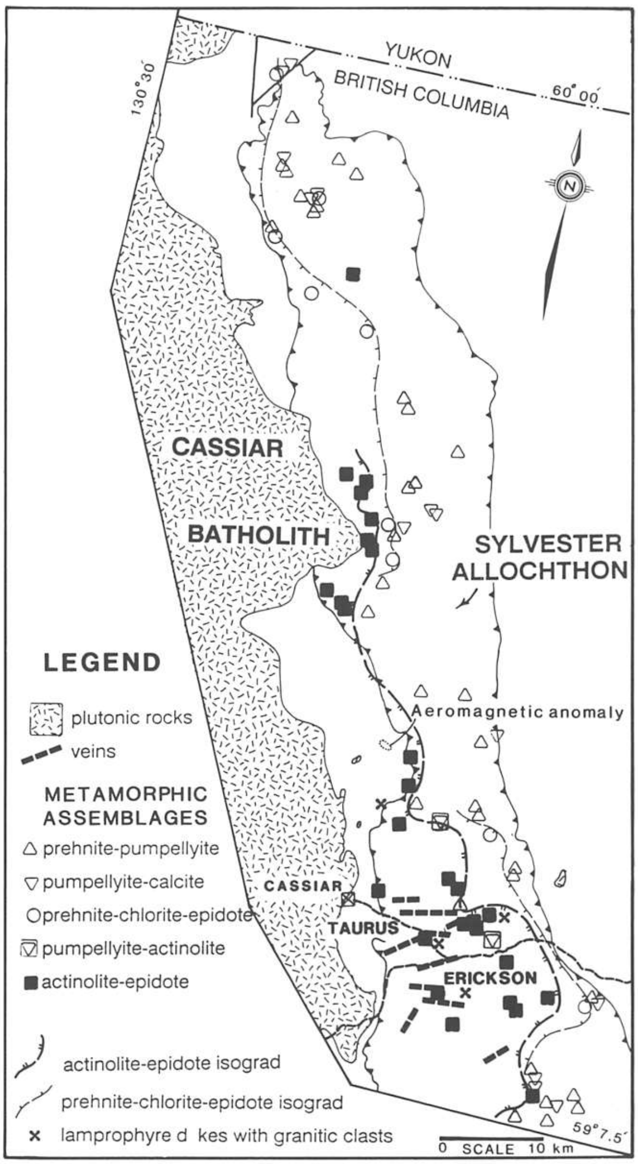

Figure 1-23-1. Regional map of the Erickson-Taurus system and its surroundings : veins, metamorphic assemblages, isograds, and granite-bearing lamprophyre dikes.

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

difference, however. Most mesothermal deposits are formed in or adjacent to steeply dipping major faults or even crustal-scale sutures such as the Melones fault zone in California, the Cadwallader fault zone at Bralorne, the Coast Range mega-lineament near Juneau, the Porcupine-Destor and Kirkland Lake–Cadillac breaks in the Superior Province, and the Con and Campbell shears near Yellowknife. Deposit models call on deep fluid circulation along these faults (Sibson *et al.*, 1988, McHaig, 1988). By contrast, major structures within and bounding the Sylvester allochthon are flat; the allochthon is a thin klippe perched on miogeoclinal strata. The east-northeast trending structures that host the Erickson veins are minor faults and fractures with minimal offsets. Neither they nor the thrust faults provide conduits for deep, large-scale fluid circulation. The suggestion of Nesbitt *et al.* (1985), that the veins "formed from deep circulation of meteoric water in major fault zones", is rendered unlikely by the lack of any major steeply dipping fault zones in the area.

Abbott (1984) has suggested that the dominantly north-northeasterly to northeasterly fractures that host epithermal silver-lead veins in the Cassiar Mountains are extensional features related to dextral motion on northwest-trending major faults. The orientation and distribution of the Erickson-Taurus structures have resisted such regional kinematic interpretation. Although the veins formed at about 130 Ma (Sketchley *et al.*, 1986) when the regional strain pattern was probably dextral-transcurrent on major faults such as the Tintina, Kechika and Cassiar faults (Gabrielse, 1985), they are oriented at high angles to the main faults and occur in the compressive rather than extensional regime of the strain ellipsoid. They are not en echelon in plan view but form a box-shaped array (Figure 1-23-1). They may be en echelon in cross-section and could have formed in response to near-horizontal motion on the fault at the base of the Table Mountain sediments (R. Britton, personal communication, 1988). Top-to-the-south movement on this fault during mineralization is shown by extensional duplex structures and slickensides within the Vollaug vein.

Thus important questions remain unresolved concerning the structural regime that gave rise to the Erickson-Taurus vein structures and the nature and driving mechanism of the fluid system that introduced the ore.

NEW EVIDENCE

A study of metamorphic assemblages in mafic rocks of the Sylvester allochthon was conducted as part of regional mapping of the Midway-Cassiar area. Based on this work, three syn to postemplacement metamorphic facies have been identified within the allochthon: prehnite-pumpellyite, transitional prehnite-chlorite-epidote and actinolite-epidote. The interpretation of assemblages and the identification of isograds is based on the experimental and theoretical work of Liou *et al.* (1985) and Cho and Liou (1987) (Figure 1-23-2). Figure 1-23-1 shows the regional distribution of the three facies. In general, the actinolite-epidote facies is restricted to a narrow band along the margin of the Cassiar batholith. It is assumed to have developed through contact metamorphic upgrading of regional prehnite-pumpellyite assemblages. North of Cassiar the actinolite-epidote facies widens somewhat around small intrusive bodies east of the Cassiar batholith – the Lamb Mountain and Contact stocks. Another buried intrusion may be indicated in this area by a negative magnetic anomaly centred 2 kilometres northeast of the Lamb Mountain stock; by abundant contact-metamorphic andalusite up to 3 kilometres from the margin of the Cassiar batholith; and by granitic clasts in a nearby lamprophyre dike (Figure 1-23-1).

East of Cassiar the actinolite-epidote isograd swings abruptly east to enclose the entire Erickson-Taurus system. The thermal peak in this area is postkinematic, as shown in thin sections by actinolite sprays growing across small shear zones and zones of chlorite fabric. Although of apparent contact metamorphic origin, this anomalously wide zone of relatively high-grade metamorphic assemblages is unlikely to be related to the Cassiar batholith. A study of the assemblages bordering Erickson veins shows a strong retrograde overprint, with actinolite replaced by pumpellyite. The Erickson veins were emplaced roughly 20 Ma before the mid-Cretaceous cooling age of the Cassiar batholith; given the

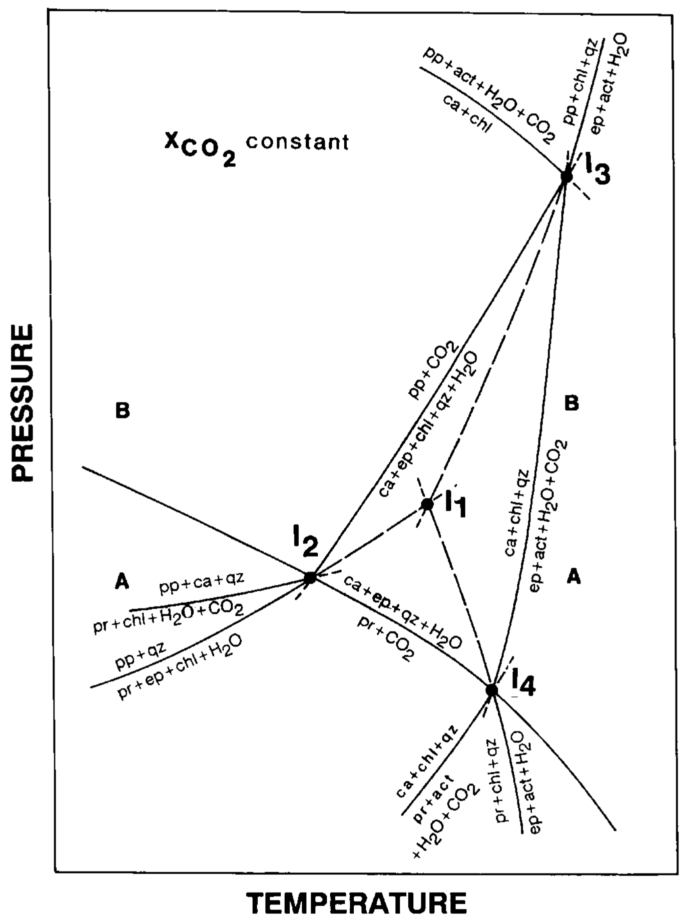

Figure 1-23-2. Schematic P-T phase relationships for the model basaltic system $CaO-MgO-Al_2O_3-Fe_2O_3-H_2O-CO_2$, from Cho and Liou (1987). $X_{CO_2}<0.1$. With increasing CO_2, the triangle I_2-I_3-I_4 expands, making the assemblage ca + cp + chl + qz stable over a wider range of pressures and temperatures. Increasing amounts of iron shift the grid to lower pressures and temperatures.

textural and age relationships, the actinolite near Erickson cannot have been generated by the batholith.

Lamprophyre dikes are common in the Erickson-Taurus camp. Where relationships are demonstrated, they are post-mineral, but tend to follow fractures surrounded by carbonate alteration and are themselves somewhat carbonate-altered. "Fresh" dikes contain prehnite-pumpellyite assemblages. Panteleyev (unpublished data, cited in Sketchley, 1986) obtained one K-Ar age of 110 ± 4 Ma on a lamprophyre dike, slightly younger than the sericites in the veins. Three lamprophyre dikes in the camp are known to contain granite xenoliths: one near the Taurus mine (Peter Read, unpublished report for Taurus Resources Limited), one on the east side of Table Mountain near the Erickson veins (Sketchley, 1986), and one north of Snowy Creek discovered in the course of this study (Figure 1-23-1). The dike north of Snowy Creek occupies a fracture surrounded by orange-weathering carbonate alteration (Plate 1-23-1a), although the lamprophyre itself is fresh. The inclusions are strongly clay-carbonate-altered medium-grained granite (Plate 1-23-1b), that originally consisted of plagioclase, orthoclase and quartz intergrown in a cumulate-like mosaic with minor biotite and other unidentified mafic minerals. In contrast, a lamprophyre dike in basalts of the allochthon north of the Erickson-Taurus system contains abundant clasts of subjacent miogeoclinal units, but no granite.

INTERPRETATIONS AND CONCLUSIONS

It is proposed that a large granitic body exists at depth below the Erickson-Taurus system. Contact metamorphism related to it generated the broad actinolite-epidote zone that coincides with the vein system. Constant-pressure cross-section A-A on Figure 1-23-2 shows the most likely prograde path based on the observed transitional assemblages, which passes below the invariant point I_1. Figure 1-23-3 is a T-X_{CO_2} diagram along this cross-section. On it are shown paths of prograde and also of retrograde metamorphism which accompanied veining. It is seen as a constant-temperature event, with X_{CO_2} decreasing progressively away from the veins. A typical sequence of assemblages passes outward from Sketchley's (1986) outer carbonate zone, which contains iron-magnesium carbonate, through chlorite-calcite-(epidote), into prehnite-pumpellyite-calcite-chlorite.

This sequence can be explained in terms of fluid/rock reactions. During the later stages of cooling, a hydrothermal convection cell centred on the granitic body. Fluids, consisting of mixed H_2O-CO_2 solutions in which gold was carried in sulphite compounds, ascended along a series of fracture zones in the cupola. They reacted with the mafic country rocks, producing carbonate halos that pass outward into carbonate-poor assemblages. The fluids were ponded at the base of the Table Mountain sediments, which formed a combination physical-chemical trap due to their carbon-rich

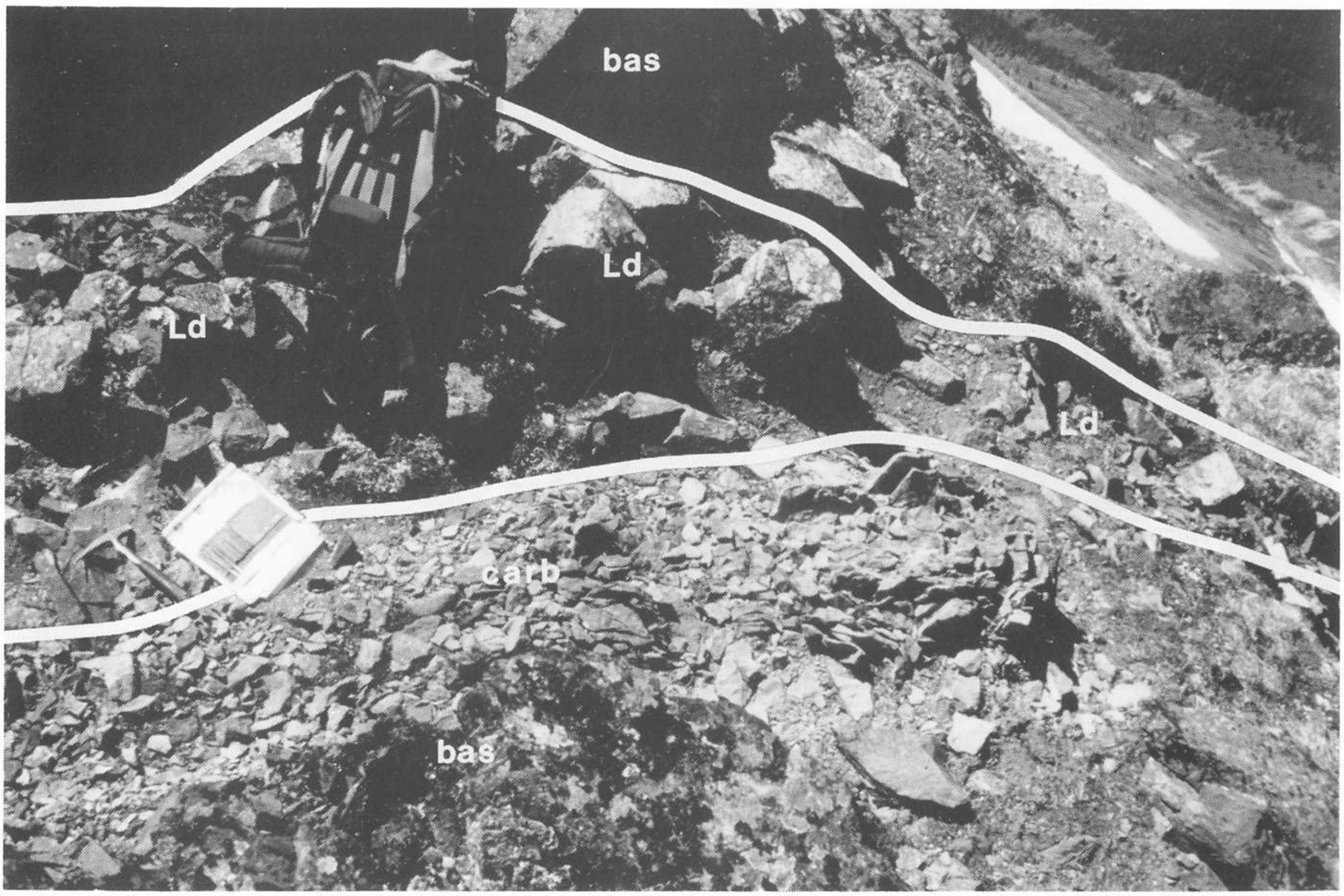

Plate 1-23-1a. Carbonate alteration zone (carb) north of Snowy Creek which hosts lamprophyre dike (Ld) containing granite clasts within basalt (bas).

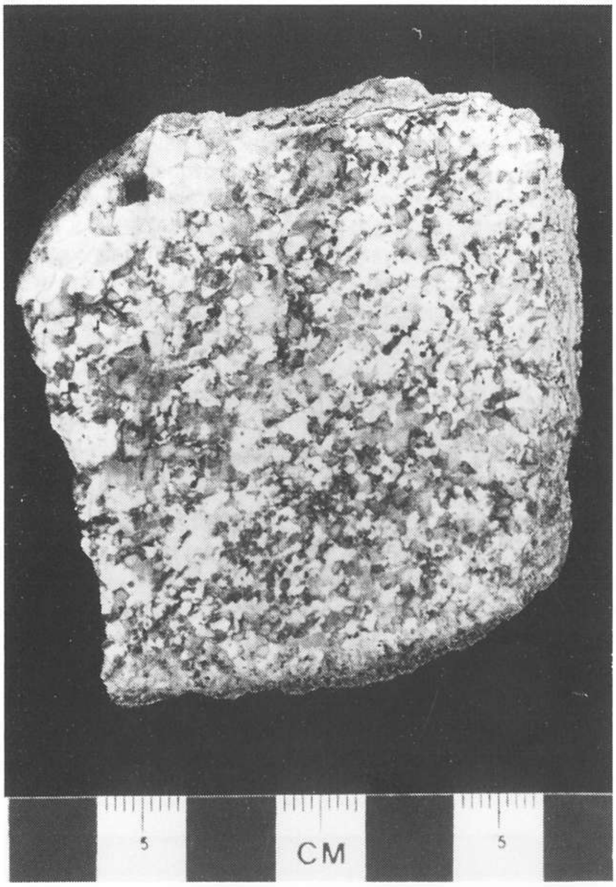

Plate 1-23-1b. Granite clast. Unlike rocks of the Cassiar batholith, this mafic-poor granite is medium grained and equigranular with a graphic texture.

composition and inability to sustain fractures. Gold precipitation probably occurred as a result of cooling, carbon-controlled reduction and absorption of sulphur in pyrite and arsenopyrite in basalt and particularly ultramafic wallrocks. Fluid inclusion data (Nelson and Bradford, unpublished) do not reveal the presence of coexisting immiscible H_2O-CO_2 fluids, so boiling was not a contributing mechanism.

Although cryptic intrusions have not ben previously linked to volcanic-hosted mesothermal gold-quartz veins in the northern Cordillera, they are an important element in models of another key type of deposit in the Cassiar and Pelly Mountains: mantos. Bradford (1988) summarizes evidence for a buried intrusion near the Midway silver-lead-zinc manto deposit, including quartz-sericite alteration, skarn development at depth, rhyolite dikes and a negative magnetic anomaly. The Butler Mountain deposit 20 kilometres north of Midway contains synmineralization rhyolite dikes that may represent the upper levels of a similar but younger pluton (Nelson and Bradford, 1987). Abbott (1977, 1986) attributes anomalous block uplifts in the Seagull and Ketza River districts and at the Mount Hundere manto zinc deposit to subjacent granites.

Remarkably, every significant precious metal deposit in the region, no matter what its immediate setting, seems to be linked to unexposed granite. By combining existing manto deposit models with observations of the Erickson-Taurus system, an extensive list of the surface expressions of these intrusions can be made. This list includes:

- Unexplained postkinematic metamorphic culminations.
- Anomalous horsts.
- Felsic dikes.
- Granitic inclusions in dikes.
- Negative magnetic anomalies.
- Alteration: carbonate/listwanite in basic and ultrabasic hosts, quartz-sericite in pelites.

These surface expressions are exploration parameters of general application. The local host determines what type of deposit will occur, whether gold-quartz veins in eugeosynclinal settings, or mantos in platformal carbonates. The strong influence of the local host on deposit character is particularly well demonstrated by the Erickson-Taurus system, where veins and alteration assemblages, indistinguishable from those in purely eugeosynclinal settings like the

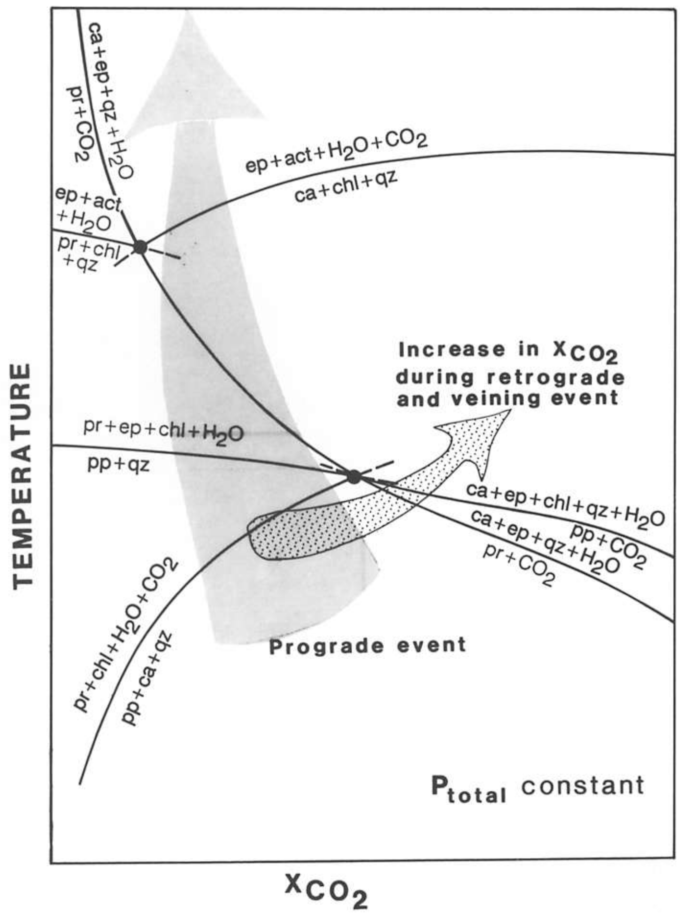

Figure 1-23-3. X_{CO_2} diagram representing cross-section A-A on Figure 1-23-2, below invariant point I_1. X_{CO_2} in this schematic diagram is less than 0.1. With higher X_{CO_2}, as in the alteration zones around the gold-quartz veins, additional carbonate phases such as dolomite and ankerite will develop. The retrograde arrow represents increase of X_{CO_2} towards, but not into, such zones.

California Mother Lode, have formed in a thin eugeosynclinal sheet, less than 600 metres above the underlying miogeocline.

This model for the Erickson-Taurus system accounts for both its similarities to and differences from other mesothermal gold-quartz deposits. The similarities – vein mineralogy and texture, and alteration assemblages – are due to local hostrock character, fluids consisting of H_2O, CO_2, NaCl, sulphur species and metals, and ambient pressure of 200-250 megapascals (Nelson and Bradford, unpublished data). Mixed H_2O-CO_2 fluids also occur in the manto deposits, as shown at Midway (Bradford, 1988). The major difference is the lack of a deep controlling structure for Erickson-Taurus. But, as shown above, an intrusion-driven, rather than a regional-scale hydrothermal cell adequately accounts for the presence and the distribution of the veins. The east-northeasterly fractures can be linked to extension in the cupola of a northerly-elongate granite. The top-to-the-south motion on the Vollaug vein, near the southern end of the system, is consistent with this interpretation.

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NOTES

Mineral Deposit
Studies

GOLD SKARNS: THEIR DISTRIBUTION, CHARACTERISTICS AND PROBLEMS IN CLASSIFICATION*

By G.E. Ray, A.D. Ettlinger and L.D. Meinert

KEYWORDS: Economic geology, gold skarn, precious metal enriched skarn, back arc basin, Proterozoic mobile belt, skarn classification, bismuth tellurides, porphyry copper districts.

INTRODUCTION

Historically, skarn deposits throughout the world have been an important source of iron, copper, molybdenum, lead, zinc, tin and tungsten. However, some skarns also contain economically recoverable amounts of gold, silver and rarely, platinum; the importance of this class of skarn deposit as either a primary or byproduct source of precious metals has only recently been widely recognized. All mineralized skarns contain some precious metals, ranging from parts per billion levels up to economic quantities. In the latter case the precious metals may be the primary commodity recovered but most gold and nearly all silver produced from skarns have been derived as byproducts of base or ferrous metal mining. It is not yet possible to present a precise definition of "gold skarn" or "precious metal enriched (PME) skarn" based either on mineral content, geological environment of formation or precious metal grade. However for the purpose of this paper, gold skarns are defined as those in which gold is the predominant, and in some cases only economic mineral present, while PME skarns include any ferrous or base metal skarns that contain gold, silver or platinum in sufficient quantities to be economically recoverable.

Skarn deposits worldwide have produced more than 1000 tonnes of gold (Meinert, 1988) and the skarns of British Columbia have contributed nearly 10 per cent of this total (Ettlinger and Ray, 1989). Between 1904 and 1961 the gold skarns in the Hedley mining camp (Figure 2-1-1) produced over 50 tonnes of gold, mostly from the Nickel Plate mine. Despite this 60-year history of Canadian gold production from skarns, only recently were gold skarns recognized as a distinct class of deposit (Orris *et al.*, 1987; Meinert, 1988). Gold skarns were not distinguished in either the classical paper by Einaudi *et al.*, (1981), which outlines and defines the six major classes of ferrous and base metal skarns (Fe, Cu, W, Pb-Zn, Mo and Sn), or in the geological synopsis of Canadian mineral deposits by Eckstrand (1984). However, recent discoveries of major precious metal enriched (PME) skarn deposits such as the Fortitude and McCoy in Nevada (Tingley and Smith, 1982; Wotruba *et al.*, 1986; Kuyper, 1987), and the Red Dome (Torrey *et al.*, 1986; Ewers and Sun, 1988) in Australia (Figure 2-1-2), and the reopening of the Nickel Plate mine in British Columbia as an open-pit operation (Simpson and Ray, 1986) indicates PME-skarns represent exploration targets with both high grade and large tonnage potential.

DISTRIBUTION PME-SKARNS

The worldwide distribution of some PME-skarn deposits is shown in Figure 2-1-2 and available data concerning their sizes and grades is outlined in Table 2-1-1. Although base metal skarns are developed in a wide variety of geological environments, tectonic regimes and hostrock lithologies (Zharikov, 1970; Einaudi *et al.*, 1981; Kwak, 1987), gold skarns throughout the world are more restricted, and are mostly concentrated in the Phanerozoic mobile belts. They are commonly found in the same geological environment as copper and iron skarns, and there is a worldwide spatial and temporal association between gold skarns and the copper porphyry provinces (Figure 2-1-2). The temporal relationship is suggested by the observation that both the gold skarns and porphyry copper deposits of North and South America are largely Mesozoic or Cenozoic in age, while the gold skarns of Australia and Russia, like their spatially associated porphyry provinces, are predominantly Paleozoic.

The close association between gold skarns and porphyry copper districts is seen in the Canadian Cordillera; many of the alkalic and calcalkalic porphyry systems lie close to the location of the Mesozoic initial 0.704-0.705 $^{87}Sr/^{86}Sr$ ratio lines, as defined by Armstrong (1988) (W. McMillan, personal communication, 1989) which mark the transition between Phanerozoic ensimatic crust to the west and Precambrian sialic crust to the east. Similarly, the gold skarns of the Hedley camp, the Dividend-Lakeview mine, and the Oka occurrence (Ettlinger and Ray, 1989) lie close to this transition. An examination of the Mesozoic and Cenozoic initial $^{87}Sr/^{86}Sr$ ratios may assist exploration for iron, base metal and gold skarns throughout the North American Cordillera.

The location and distribution of PME skarns in British Columbia is shown in Figure 2-1-1. Over 90 per cent of gold production from skarn in the province was derived from mineralization hosted in carbonate-bearing oceanic island arc, back arc or marginal basin assemblages presem in the Intermontane and Omineca belts, while only 9 per cent was produced from the westernmost Insular Belt. The latter belt contains low-potassium immature island arc rocks as well as calcalkaline continental margin arc rocks and tholeiitic oceanic flood basalts, none of which are as favorable for PME skarn developement as the mature, shoshonitic arc and marginal basin sequences present in the Intermontane and Omineca belts. Despite the presence of abundant carbonate sequences and some tungsten and tin skarns, no skarn-related gold has been recovered from the Foreland Belt which is underlain by cratonic basement.

Almost all of the gold skarn mineralization in the Canadian Cordillera is genetically related to high to intermediate level, I-type, late oceanic island arc plutonism. Much of the miner-

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

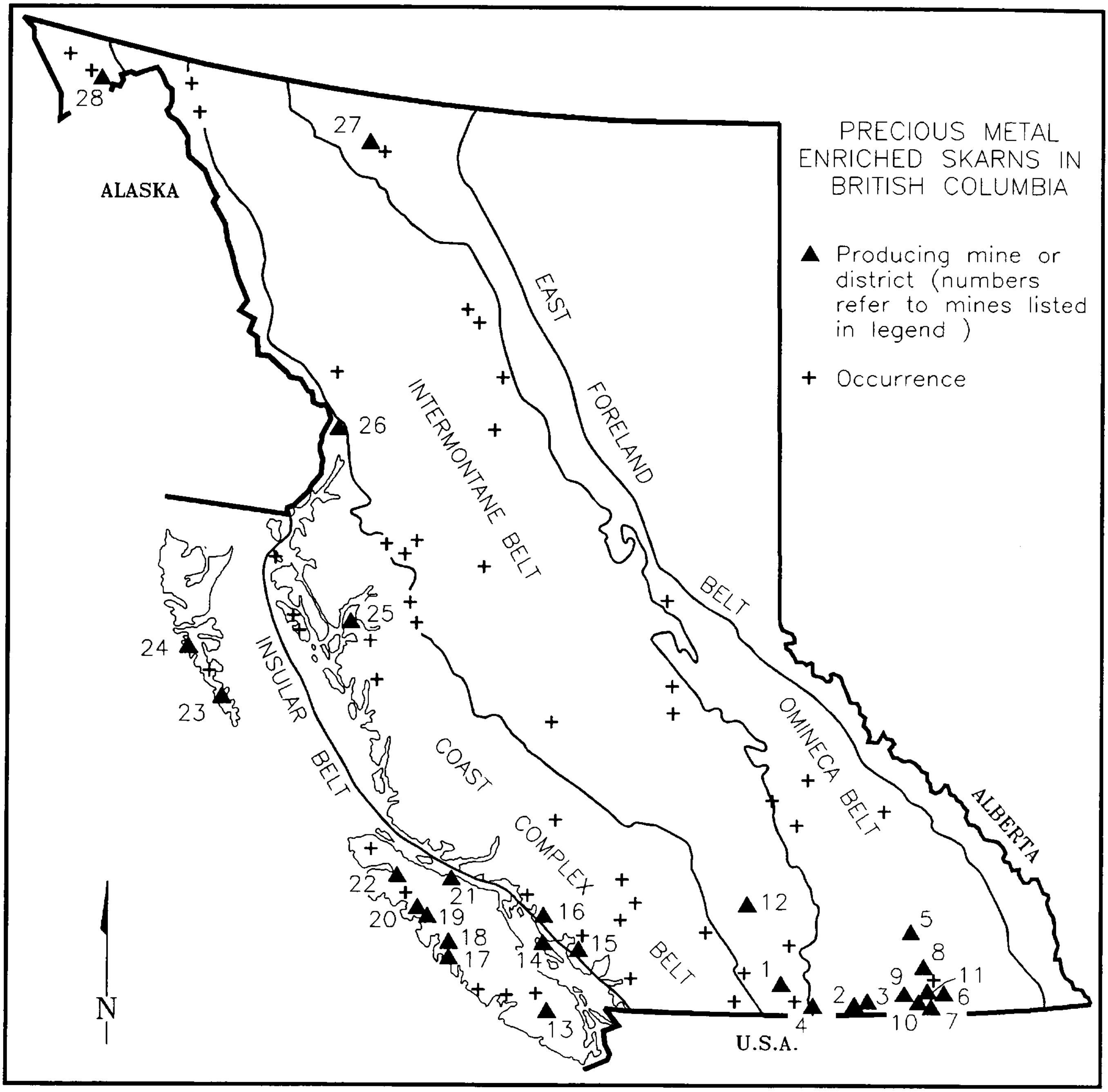

Figure 2-1-1. Distribution of PME skarns in British Columbia.

alization is also hosted in coeval arc superacrustals, although some precious metal enriched copper and iron skarn deposits such as those on Texada and Vancouver islands are exceptions; they are associated with Early to Mid-Jurassic continental margin intrusions of the Bonanza magmatic arc, and are hosted in Triassic or older, lithologically favorable oceanic rocks that were unrelated to the arc. Plutons associated with plate collision and accretion, or those exposed in the deeply eroded root zones of arcs, as possibly present in the Coast Belt of British Columbia (Figure 2-1-1) are less favorable for gold skarn mineralization; only 0.001 per cent of the total gold from skarn has come from the Coast Belt. Some gold skarns, such as those in the Hedley camp, are developed in calcareous clastic sediments that were deposited across a structural hinge zone marking the fracture-controlled edge of a back arc or marginal basin. Such rifted basin margins are particularly favorable for gold skarns because the controlling basement structures can preferentially channel the arc-related plutons into suitable carbonate-rich host rocks. Over 80 per cent of the PME-skarn occurrences in the British Columbia are associated with impure, often organic-rich and permeable limestone or marble-rich

Legend for Figure 2-1-1

Precious metal enriched (PME) skarn producing mines in British Columbia.

1 = Hedley camp (Nickel Plate, Hedley Mascot, French, Good Hope and Canty mines).

2 = Greenwood camp (Phoenix, Marshall, Motherlode, Greyhound, Morrison, Emma, Sunnyside and Oro Denoro mines).

3*= Loyal Canadian.

4 = Dividend-Lakeview.

5*= Tillicum Mountain camp (Heino-Money zone and Silver Queen).

6 = Salmo-Malarctic and Jackpot mines.

7*= Mormon Girl.

8 = Orinoco and Elk mines.

9*= Sir Douglas Haig.

10*= Rely 1.

11 = Second Relief.

12*= Lucky Mike.

13 = Blue Grouse.

14 = Texada Island camp (Little Billie, Paxton, Prescott, Yellow Kid, Cornell, Marble Bay and Copper Queen mines).

15 = Cambrian Chief.

16*= Roadside.

17 = Dewdney.

18 = Silverado.

19*= Geo-Star.

20*= Beano.

21 = Hab.

22 = Old Sport and Merry Widow mines.

23 = East Copper, Lily and Morning mines.

24 = Tasu.

25*= Gribble Island.

26*= Molly B.

27*= Contact.

28 = Maid of Erin.

*= Minor producers (<5 kg Au or 10 kg Ag).

sequences that also contain some argillite, tuff or minor volcanic flow componants. Regionally, some mature arc or back arc sequences favorable for gold skarns, such as the Triassic Nicola and Takla groups of British Columbia, include potassium-rich volcanic shoshonites and limestone-boulder breccias or olistostromes; the latter reflect the unstable sedimentary environment that prevailed in the basin-margin facies.

CHARACTERISTICS OF GOLD SKARNS

Gold and silver enrichment is mainly associated with calcic skarns; PME-magnesian skarns are very rare. The overall silicate mineral assemblages associated with gold skarns are similar to those in end-member iron and some base metal skarns, and thus cannot neccessarily be used to distinguish skarns with precious metal potential. However, gold skarns tend to be richer in pyroxene relative to garnet, and compared to copper skarns, their pyroxenes are commonly more iron-rich. Economic gold mineralization is more common in the exoskarn than the endoskarn, and the amount of exoskarn alteration associated with precious metal mineralization varies considerably from narrow zones less than 10 metres wide up to large envelopes many hundreds of metres thick; at the Nickel Plate deposit for example, the envelope totals between 0.75 to 1.5 cubic kilometres of alteration. The morphology of the envelopes varies from stratiform to subcircular to vein-like and sharply discordant, and the gold-sulphide mineralization is commonly found close to the outer, pyroxene-rich margins of the skarn.

The intrusive rocks associated with gold skarns range compositionally from granite to gabbro although quartz diorite and diorite are the most common. The intrusions, which are believed to have been emplaced at shallow to intermediate depths, vary from large stocks to narrow sills and dikes which may occur as swarm complexes. Some recent workers (Dawson *et al.*, 1990, this volume) suggest that the sill-dike complex at Hedley was intruded into wet, unconsolidated sediments which would imply that they and the related skarn fluids were emplaced at a high level under relatively low confining pressures. However, this conclusion is controversial since fluid inclusion data (Ettlinger *et al.*, in press) and textural features suggest that the Nickel Plate deposit was emplaced at high temperatures and intermediate depths.

The majority of the PME skarns in the North American Cordillera are genetically related to subalkalic, I-type intrusions with calcalkaline affinities. Many intrusions are also porphyritic with phenocrysts of hornblende and/or plagioclase; the latter may exhibit marked oscillatory and reverse composition zoning that may indicate the intrusions were derived from hybridized magmas or volatile-rich melts. The low initial $^{87}Sr/^{86}Sr$ ratios of these rocks suggest they were derived either from upper mantle material, or more likely represent recycled oceanic crust. Although no gold skarns associated with alkalic rocks have yet been identified in British Columbia, some high-level alkalic intrusions related to gold-bearing porphyry copper deposits are associated with skarn-like garnet-pyroxene-epidote-scapolite alteration assemblages (Preto, 1972; Hodgson *et al.*, 1976)

Many PME skarns exhibit variable mineralogical zoning patterns similar to those described in iron and base metal skarns. Recognition of these zones, often manifest as garnet-dominant proximal zones and pyroxene and/or wollastonite-rich distal zones, may assist in exploration. The Fortitude gold skarns (Myers and Meinert, 1989) and some of the smaller Hedley skarns include an outermost alteration halo of biotite and potassium feldspar that predates the garnet-pyroxene assemblages. Also at Nickel Plate, the gold mineralization was coeval with widespread scapolitization; this and the presence of chlorine-rich amphiboles in some other PME skarns (Ettlinger and Ray, 1989) suggests that hypersaline fluids may be important for the transportation and precipitation of gold in some skarns.

The degree of retrograde alteration (chlorite, epidote, tremolite-actinolite) overprinting the silicate assemblages varies enormously in PME skarns, and cannot be used to

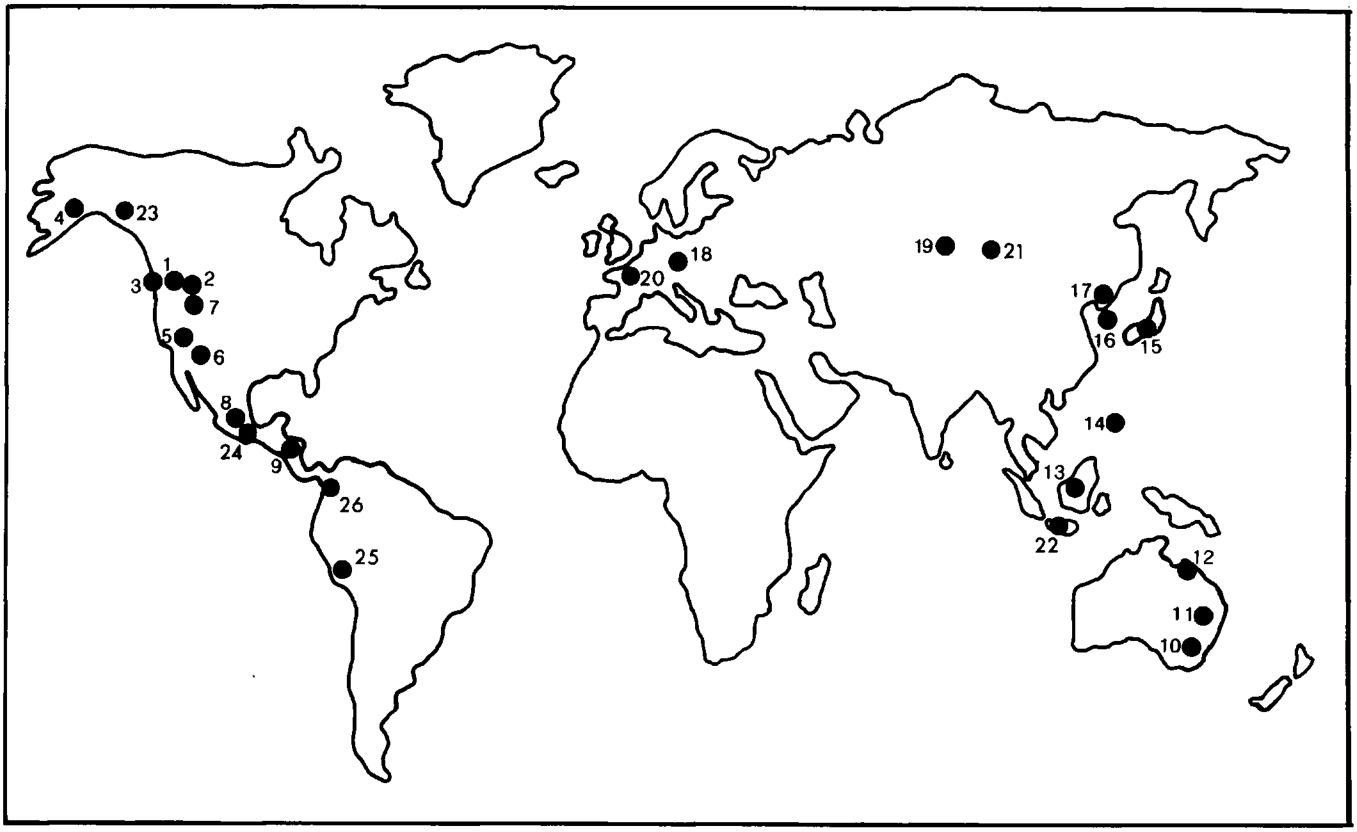

Figure 2-1-2. A: World distribution of some PME skarns. B: World-wide distribution of porphyry copper provinces. For legend *see* Table 2-1-1 (from McMillan and Panteleyev, 1988).

indicate gold potential. Most gold skarns, like iron skarns, are characterized by low-manganese (<1 weight per cent MnO_2) grandite garnets and low-manganese (<4 weight per cent MnO_2), iron-rich hedenbergitic pyroxenes. Data are sparse, but there appears to be no consistent core-to-rim compositional zoning in garnets in gold skarns; the Nickel Plate garnets exhibit variable zoning although grossular cores and andraditic rims are most common, while at the Fortitude gold skarns the reverse is seen (Myers and Meinert, 1989). The gold skarns at Tillicum Mountain in British Columbia are notable in containing manganese-rich garnets that range from 45 to 68 mole per cent pyralspite (Ray *et al.*, 1986; Ettlinger and Ray, 1989). These may represent gold-enriched mineralization with lead-zinc skarn affinities, or magmatically related skarn-like mineralization controlled within brittle-ductile shears, similar to those described by Mueller (1988) in Western Australia.

Gold mineralization in skarn is usually associated with opaque minerals that were mainly introduced after the prograde skarn assemblages. Pyrrhotite, arsenopyrite, chalcopyrite, pyrite, bornite, sphalerite, magnetite, and bismuth and/or tellurium minerals are the most common opaques in PME skarns. Less commonly, cobaltite, scheelite, molybdenite and galeua are also present. Gold commonly occurs as microscopic inclusions in the sulphides, and in some deposits it is difficult to visually distinguish ore from waste. There is also a highly variable trace element association; PME skarns may be enriched in As, Te, Bi, Ag, Cu, Co, Zn, Sb, Fe, W, Pb, Ni, Mo or Pt. Many gold skarn deposits exibit systematic geochemical variations throughout the envelope; at the Fortitude and Nickel Plate mines, for example, the copper grades in the skarn increase toward the endoskarn intrusions. The presence of lead, silver, nickel or, most commonly, bismuth tellurides, is a distinctive characteristic of gold skarns, as is the common presence of native bismuth and other bismuth minerals; the tellurides recorded include hedleyite, tetradymite, altaite and hessite. Other minerals reported in gold skarns include bismuthinite, wittichninite, breithauptite, lollingite and maldonite.

Sparse fluid-inclusion data on PME skarns such as Red Dome and Fortitude (Ewers and Sun, 1988; Myers and Meinert, 1989) suggest prevailing temperatures of 260°C to 450°C and fluids with salinities of 2 to 26 weight per cent NaCl equivalent. Higher temperatures for the Nickel Plate gold skarn are suggested (Ettlinger *et al.*, in press) with the garnet-pyroxene skarn crystallizing between 460°C and 480°C, but with local temperatures reaching 800°C; homogenization temperatures in scapolite associated with gold indicate the later mineralization occurred in the range of 320°C to 400°C.

Garnet and pyroxene compositions suggest PME skarns can develop in a variety of oxidizing states (Figure 2-1-3). The Texada Island and Red Dome PME skarns formed under intermediate to relatively oxidizing conditions, similar to those determined by Einaudi *et al.*, 1981 and Einaudi (1982) for iron skarns and copper porphyry related skarns. By

Erratum: To accompany Ray *et al.*, page 240, Geological Fieldwork 1989, Paper 1990-1

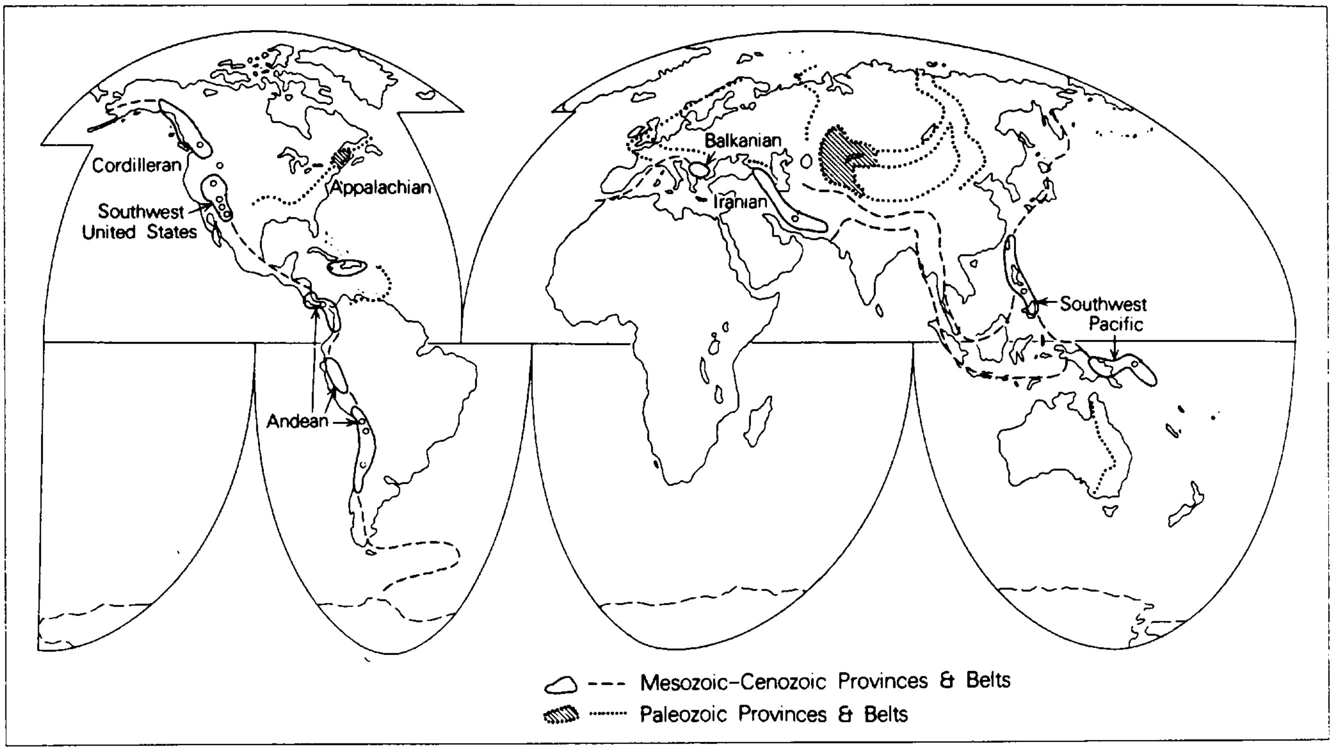

Figure 2-1-2B. World-wide distribution of porphyry copper provinces.

TABLE 2-1-1
SELECTED PME-SKARNS OF THE WORLD (*see* Figure 2-1-2)

Number in Figure 2-1-2	DEPOSIT	SIZE (tonnes)	Au (g/t)	Ag (g/t)	Cu (%)
1	Nickel Plate and Hedley-Mascot, B.C. (underground)	3 600 000	14.0	1.4	0.1
1	Nickel Plate Hedley, B.C. (open pit)	8 900 000*	4.5*	3.0**	0.1
2	Phoenix, B.C.	26 956 000	1.1	7.1	0.9
3	Texada Island, B.C. (Cu-Au skarns)	310 000	2.4	16.0	3.0
4	Zackly, Alaska	1 200 000	5.5	30.0	2.7
4	Nixon Fork, Alaska	NA	NA	NA	NA
5	Fortitude, Nevada	10 300 000	6.9	24.7	0.1
5	McCoy, Nevada	8 700 000	1.9	NA	0.1
6	Carr Fork, Utah	61 000 000	0.4	10.7	1.8
7	Cable, Montana	1 000 000	6.0	5.0	3.0
7	Golden Curry, Montana	930 000	8.5	4.2	0.33
7	Southern Cross, Montana	400 000	13.0	16.0	0.1
8	Golfo de Oro, Mexico	5 000 000	4.5	10.0	NA
8	Concepcion del Oro, Mexico	15 000 000	1.7	NA	2.0
9	La Luz, Nicaragua	16 000 000	4.1	1.2	0.44
10	Browns Creek, Australia	740 000	7.5	9.0	0.4
11	Mount Biggendon, Australia	500 000	15.0	NA	NA
12	Red Dome, Australia	13 800 000	2.0	4.6	0.46
13	Bau, Malaysia	2 400 000	7.2	0.1	NA
14	Siana, Philippines	5 400 000	5.1	10.0	NA
14	Thanksgiving, Philippines	1 700 000	6.4	40.6	0.4
15	Rokuromi, Japan	160 000	4.1	1.0	NA
16	Suian, North Korea	530 000	13.0	4.9	NA
17	Tul Mi Chung, North Korea	400 000	12.0	NA	NA
18	Reicher Trost, West Germany	<10 000	20.0	0.1	NA
19	Siniukhinskoe, USSR	NA	8.0	37.7	3.9
20	Salsigne*** France	1 500 000	13	33	0.15
21	Lebedskoe (Kaurchack) USSR	120 000	4.0	NA	NA
22	Pagaran Siayu Indonesia	110 000	5.6	2.5	0.2
23	Nabesna Alaska	80 000	23	NA	NA
24	Naica, Mexico	10 000 000	0.4	180	0.4
25	Huarca, Peru	500 000	2.0	10	3.0
25	Katanga Peru	2 000 000	6.1	46.5	3.0
26	Vieja Colombia	450 000	1.0	35	1.7
26	El Sapo Colombia	330 000	11.5	79.8	5.1

Data source for most deposits see Orris *et al.*, (1987); Ettlinger and Ray (1989); Meinert (1988). Data source for No. 19: Ettlinger and Meinert (*in review*); for No. 23: Wayland (1943); for Nos. 24-28: G.J. Orris, personal communication, 1989.

NA = values not available.

* Recently downgraded to 8 250 000 tonnes grading 3.02 g/t Au (Corona Corporation announcement, Dec. 1988).

** Estimated silver grade.

*** There is some uncertainty whether Salsigne represents a skarn deposit (G.J.Orris, personal communication (1989)).

contrast, all the gold skarns in the Hedley camp developed in a relatively reduced state (Figure 2-1-3); this conclusion is supported by the abundance of pyrrhotite and arsenopyrite, the presence of native bismuth, and the general scarcity of pyrite at Hedley. At the porphyry-related Fortitude deposit, the composition of the pyroxene-garnet assemblages (Myers and Meinert, 1989) suggests that the proximal copper-rich skarn developed under relatively oxidized conditions, while the distal gold mineralization formed under more reduced conditions (fields A and B respectively, Figure 2-1-3).

Based largely on spatial relationships, many workers have suggested that the intrusions were the primary source of both the skarn-forming fluids and metals in base and ferrous metal skarns. At the Nickel Plate mine, the gold is also believed to have originated from magmatic fluids (Dolmage and Brown, 1945; Ray *et al.*, 1988), and studies indicate the fluids were saline and high temperature (Ettlinger *et al.*, in press).

PROBLEMS WITH THE CLASSIFICATION OF GOLD AND PME SKARNS

Base and ferrous metal skarns have been classified using a number of criteria, including the predominant metal commodity present, as outlined by Einaudi *et al.*, (1981).

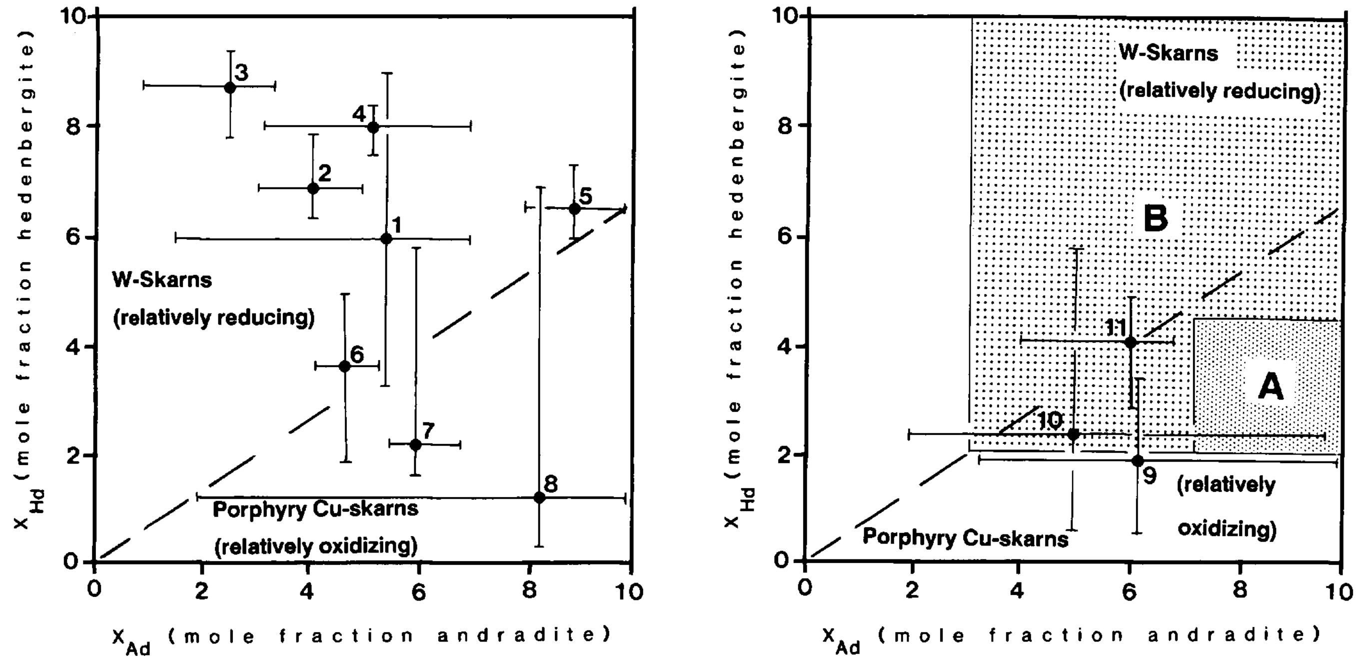

Figure 2-1-3. Plot of mole fraction Hd versus mole fraction Ad, illustrating the relative oxidation states of some gold, PME-copper and PME-iron skarns. Upper left field = relatively reducing W-skarns, lower right field = relatively oxidizing porphyry copper skarns, after Einaudi, (1982); (plot made on two diagrams due to density of data): 1 = Nickel Plate (Au), Hedley, BC; 2 = Canty (Au), Hedley, BC; 3 = Good Hope (Au), Hedley, BC; 4 = Peggy (Au), Hedley, BC; 5 = French (Au), Hedley, BC; 6 = Bob skarn (Au), Banks Island, BC; 7 = TP skarn (Au), Atlin, BC; 8 = Red Dome (Cu, Au), Australia; 9 and 10 = Little Billie (Cu, Au) and Florence (Au,Cu), Texada Island, BC; 11 = Texada Iron (Fe, Au), Texada Island, BC. Shaded area A = proximal Cu-rich skarn (West ore body), Copper Canyon, Nevada; shaded area B = more distal Au-rich skarn (Fortitude), Copper Canyon, Nevada. Note relatively reduced states for the Hedley and Fortitude gold skarns and oxidized to intermediate states for the Copper Canyon Cu-rich skarn, the TP gold skarn and the PME skarns at Red Dome and Texada Island. Data for 1 to 7 and 9 to 11 from Ettlinger and Ray (1989) and Ray and Dawson, (in preparation), for 8 from Ewers (personal communication, 1988), for shaded fields A and B from Myers and Meinert (1989). Bar lines show range of X Hd and X Ad values; centre points = average values except for Red Dome which shows mean values.

However, gold skarns are difficult to classify in relation to the six main skarn classes as the base and ferrous metals are commonly present in percentage quantities while gold is an extremely high-value commodity that generally only reaches concentrations of a few parts per million. Thus, small changes in either gold concentration or metal prices can progressively alter the classification of a deposit from an end-member ferrous or base metal skarn to an end-member gold skarn. A continuum probably exists between end-member gold skarns such as the Nickel Plate and Fortitude deposits, and end-member copper and iron skarns with little or no precious metal enrichment. Compared to some copper skarns, iron skarns tend to have less overall gold enrichment; if present, gold is usually concentrated in relatively small, isolated areas of the deposit where the magnetite ore is sulphide and copper rich. Although gold is sufficiently enriched in many copper skarns, such as the Phoenix deposit in British Columbia (Figure 2-1-1), to provide a major economic support to the mining operation, it is generally much lower grade in iron skarns and only recovered as a byproduct.

In the North American Cordillera gold enrichment at economically recoverable grades is rare in molybdenum and tin skarns and uncommon in lead, zinc and tungsten skarns. However, since gold-enriched tungsten skarns are reported in the Yukon (Brown, 1985), Japan (Shimazaki, 1980) and the U.S.S.R. (Khasanov, 1982) these skarns can provide exploration targets for precious metal deposits.

Orris *et al.* (1987) examined the gold grades and tonnages of some producing skarn deposits and designed a useful classification scheme. By their definition "gold skarns" average 1 ppm gold or more and must have been exploited primarily for gold, while "byproduct gold skarns" include any skarn mined, primarily for base or ferrous metals, where significant amounts of byproduct gold were recovered. However, this classification is limited since it is restricted to producing deposits and cannot be used to classify mineralized skarn occurrences.

Another method of classifying gold, copper and iron skarns is by comparing the Cu/Ag versus Cu/Au ratios of mineralized skarns. This method has the advantage that metal ratios can be determined from either assay or production data, although grab sample assays or older production data are often unreliable. The Cu/Au and Cu/Ag ratios of 40 deposits from around the world, described in the literature by various authors as "gold", "copper" or "iron" skarns, are listed in Tables 2-1-2, 2-1-3 and 2-1-4. The ratios of these deposits are plotted in Figure 2-1-4 and the main fields for gold, copper and iron skarns are empirically determined from the clustering of points. Deposits described as "gold skarns" tend to have Cu/Ag and Cu/Au ratios less than 1000 while most of those described as "copper skarns" have Cu/Au ratios ranging mainly between 2000 and 25 000, and Cu/Ag ratios ranging from 500 to 2500. In contrast the "iron skarns" generally have the highest Cu/Au and Cu/Ag ratios, ranging from 20 000 to 160 000 and 2500 to 5000, respectively (Figure 2-1-4).

Four deposits described in the literature and listed as gold skarns in Table 2-1-2 do not fall within the gold skarn field outlined in Figure 2-1-4; these are the Cable (No. 5), Red Dome (No. 8), Surprise (No. 9) and La Luz (No. 16). It is possible that the Red Dome deposit with its average grades of 2 grams gold per tonne and 0.46 per cent copper (Torrey *et al.*, 1986) represents a gold-enriched copper-skarn, rather than a true end-member gold skarn.

SUMMARY

Precious metal enriched skarns and gold skarns are preserved and exposed worldwide in Phanerozoic mobile belts, where they are preferentially hosted in carbonate-rich assemblages deposited in back arc and marginal basin environments, or in mature oceanic arcs that may contain potassium-rich volcanic sequences. Rifted basin margins are favorable sites for gold skarns because the controlling structures can preferentially channel the arc plutons into suitable basin-edge carbonate sediments; the unstable sedimentary environments in the basin margin facies may result in the deposition of slump carbonate deposits, limestone-boulder breccias or olistostromes. By contrast, immature island arcs, arcs formed along sialic continental margins and the deeply eroded root zones of magmatic arcs are less favorable for gold skarns.

Gold skarns in the Canadian Cordillera are mostly related to intermediate to high-level, I-type island arc plutonism of calcalkaline, subalkaline affinities and dioritic composition. In many cases the intrusions are coeval with the arc volcanism, but some PME skarns are formed when younger plutons intrude considerably older but lithologically favorable hostrocks.

Gold skarns show an overall spatial and temporal association with the world's copper porphyry districts, and some deposits, such as the Fortitude in Nevada, are intimately related to porphyries. Others, such as the Nickel Plate deposit are probably not directly associated with porphyry systems although often they too are developed in regions of copper porphyry mineralization. The worldwide spatial relationship between gold skarns and the porphyry districts outlines potential areas for gold skarn exploration. Besides

TABLE 2-1-2
Cu/Au AND Cu/Ag RATIOS OF DEPOSITS DESCRIBED AS "GOLD SKARNS" (*see* FIGURE 2-1-4)

Number in Figure 2-1-4	Name	Cu/Au	Cu/Ag	Reference or Data Source
1	Hedley Mascot, B.C.	125	510	Ettlinger and Ray (1989).
2	Nickel Plate, B.C. (underground)	23	235	National Mineral Inventory production data; Ray *et al.* (1988).
3	Nickel Plate, B.C. (open pit)	222	333	Mascot Gold Mines Ltd. report – Nov. 1987.
4	Browns Creek, Australia	533	444	Meinert (1988).
5	Cable, Montana	5000	6000	Earll (1972).
6	Fortitude, Nevada	200	56	Blake *et al.* (1984). Wotruba *et al.* (1986).
7	Minnie-Tomboy, Nevada	357	111	Blake *et al.* (1984). Theodore *et al.* (1986). Meinert (1988).
8	Red Dome, Australia	2300	1000	Torrey *et al.* (1986).
9	Surprise, Nevada	2500	372	Meinert (1988).
10	Northeast Extension Nevada	379	73	Wotruba *et al.* (1986). Orris *et al.* (1987).
11	Pagaran Siayu, Indonesia	357	800	Orris *et al.* (1987).
12	Thanksgiving, Philippines	162	35	Bryner (1969).
13	Southern Cross, Montana	68	55	Earll (1972). Orris *et al.* (1987).
14	Salsigne, France	115	45	Elevatorski (1981)
15	Golden Curry, Montana	388	785	Orris *et al.* (1987).
16	La Luz, Nicaragua	1073	3666	Orris *et al.* (1987).

TABLE 2-1-3
Cu/Au AND Cu/Ag RATIOS OF DEPOSITS DESCRIBED AS "COPPER SKARNS" (*see* FIGURE 2-1-4)

Number in Figure 2-1-4	Name	Cu/Au	Cu/Ag	Reference or Data Source
17	Clifton District, Utah	33333	213	Elevatorski (1982). Meinert (1988).
18	Frankie, Utah	23529	1270	Elevatorski (1982). Meinert (1988).
19	Geo-Star, B.C.	24000	2118	Ettlinger and Ray (1989).
20	Moncocco, Utah	33333	211	Elevatorski (1982). Meinert (1988).
21	Phoenix, B.C.	7611	1198	Church (1986).
22	Phoenix, B.C.	12000	800	Production data quoted by Meinert (1988).
23	Rosita, Nicaragua	17778	2667	Meinert (1988).
24	Victoria, Nevada	79000	1692	Atkinson *et al.* (1982). Meinert (1988).
25	Whitehorse Copper District	19821	1568	Tenney (1981); Meinert (1986).
26	Yaguki, Japan	2666	51	Einaudi *et al.* (1981).
27	Copper Queen, Texada Island, B.C.	3851	509	Peatfield (1987).
28	Zackly, Alaska	4909	900	Production data quoted by Meinert (1988).
29	Cornell, Texada Island, B.C.	2906	623	Ettlinger and Ray (1989).
30	Little Billie, Texada Island, B.C.	2256	683	Ettlinger and Ray (1989).
31	Marble Bay, Texada Island, B.C.	4397	537	Ettlinger and Ray (1989).

TABLE 2-1-4
Cu/Au AND Cu/Ag RATIOS OF DEPOSITS DESCRIBED AS "IRON SKARNS" (*see* FIGURE 2-1-4)

Number in Figure 2-1-4	Name	Cu/Au	Cu/Ag	Reference or Data Source
32	Iron King	160000	20000	Warner *et al.* (1961). Meinert (1988).
33	It, Alaska	20000	2667	Warner et al. (1961). Myers (1984, 1985a, 1985b).
34	Larap	1000	200	Einaudi *et al.* (1981).
35	Magnetite Cliff.	13333	3200	Elevatorski (1981).
36	Mamie, Alaska	30166	4525	Warner *et al.* (1961). Myers (1984, 1985a, 1985b).
37	Mount Andrew, Alaska	38625	2809	Warner *et al.* (1961). Myers (1984, 1985a, 1985b).
38	Poor Man	3000	1500	Warner *et al.* (1961). Myers (1984, 1985a, 1985b).
39	Prince of Wales District, Alaska.	29625	3656	Production data quoted by Meinert (1988).
40	Texada Iron mines B.C.	30112	1130	Meinert (1984); Peatfield (1987).

the relatively well known North American Cordillera, the porphyry copper belts of South America, the Himalayas, Iran and southeast Asia, particularly where carbonate successions are preserved, can be expected to contain economic gold skarn deposits; in Thailand for example, numerous gold skarn prospects are currently being explored (Pisutha-Arnond *et al.*, 1984).

Gold skarns occur in the same geological environment as iron skarns and oceanic island arc hosted copper skarns; like iron skarns they are commonly related to primitive, ocean-crust-derived dioritic intrusions with low initial $^{87}Sr/^{86}Sr$ ratios, tend to be enriched in arsenic and cobalt, and are sporadically associated with late scapolite alteration. Similarly, they are characterized by manganese-poor grandite garnets (<1 weight per cent MnO_2) and manganese-poor, iron-rich hedenbergitic pyroxenes (<4 weight per cent MnO_2). However, they are atypical of base metal skarns in commonly containing arsenic, bismuth and telluride minerals. Precious metal enrichment can occur in skarns formed in either reduced or oxidized conditions, although preliminary evidence suggests end-member gold skarns favour more reduced states. Many PME skarns associated with iron skarns or copper porphyries tend to develop in more oxidizing environments while all the Hedley gold skarns formed under reduced conditions. The presence of scapolite and chlorine-rich amphiboles in some deposits suggests chlorine-

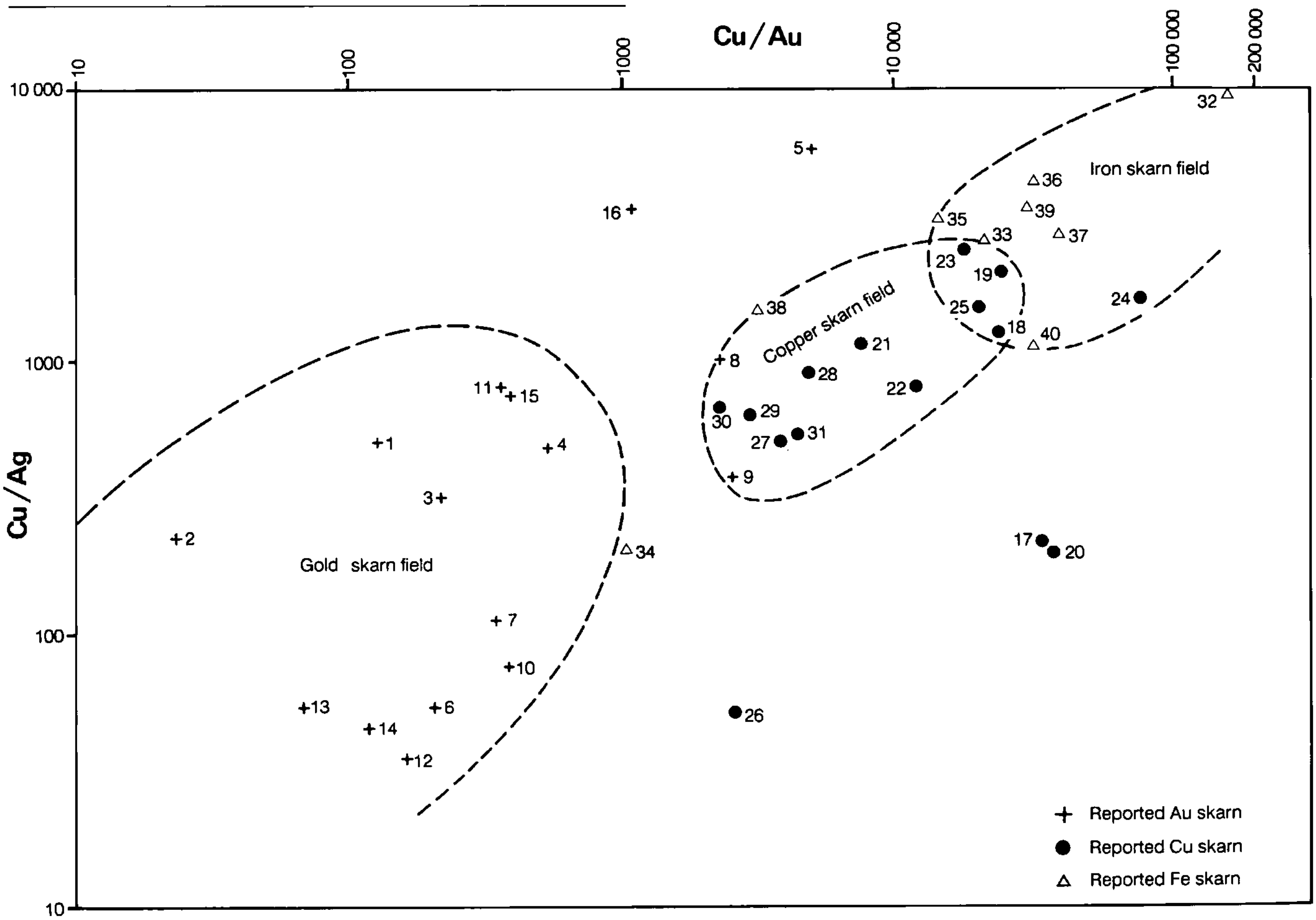

Figure 2-1-4. Plot showing Cu/Ag versus Cu/Au ratios of the 40 deposits listed in Tables 2-1-2, 2-1-3 and 2-1-4 and described by various authors as "gold", "copper" or "iron" skarns. Note: fields have been empirically drawn around the main clustering of points to outline the three skarn classes.

rich fluids were important for the transportation and precipitation of gold in skarns, and the gold is believed to have been derived from, and largely carried in magmatic fluids.

For a variety of reasons, gold skarns cannot be adequately classified using the criteria employed to classify and define base and ferrous metal skarns, although using Cu/Ag versus Cu/Au ratios can broadly differentiate gold, copper and iron skarns. Further study may indicate that gold skarns, like copper porphyries, can be divided into different subclasses dependant on such features as the predominant base or ferrous metal content (if any), the association with alkalic or calcalkaiic plutonism, and the environment and depth of formation. To most exploration geologists, however, PME-skarns can be regarded as skarns that contain gold, silver or rarely, platinum, in sufficient quantities to be economically recoverable as either primary or byproduct commodities. A continuum probably exists between end-member gold skarns and end-member copper and iron skarns containing little or no precious metal enrichment.

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PRELIMINARY REPORT OF RESEARCH IN THE SHEEP CREEK CAMP, SALMO, BRITISH COLUMBIA (82F/3, 6)

By Robert Hardy
University of Alberta

KEYWORDS: Economic geology, Sheep Creek, quartz veins, mesothermal gold, stable isotopes.

INTRODUCTION

The Sheep Creek gold camp is situated at latitude 117°09′W, longitude 49°08′N within the Nelson Range of the Selkirk Mountains in southeastern British Columbia (Figure 2-2-1). The camp consists of ten significant mining properties and is located within the Nelson (West Half) map area, where more than 300 mining properties have been documented (Little, 1960, 1963, 1964) of which many have produced considerable amounts of base and precious metals in concentrates.

Interest in the lode gold deposits in the area was aroused as early as 1885 when the Hall brothers staked claims near the headwaters of Ymir Creek. The Yellowstone and Queen veins of the Sheep Creek camp, located about 18 kilometres south of Ymir Creek, were staked in 1896 and since that time the camp has produced approximately 28.9 million grams of gold, 15.5 million grams of silver and modest amounts of lead and zinc (Schroeter *et al.*, 1986; Mathews, 1953). Mining activity was intense from the turn of the century until the outbreak of World War I, began again in 1928 and reached its peak in 1937. There was little activity in the camp from about 1942 until recently, when interest was renewed by Gunsteel Resources Incorporated.

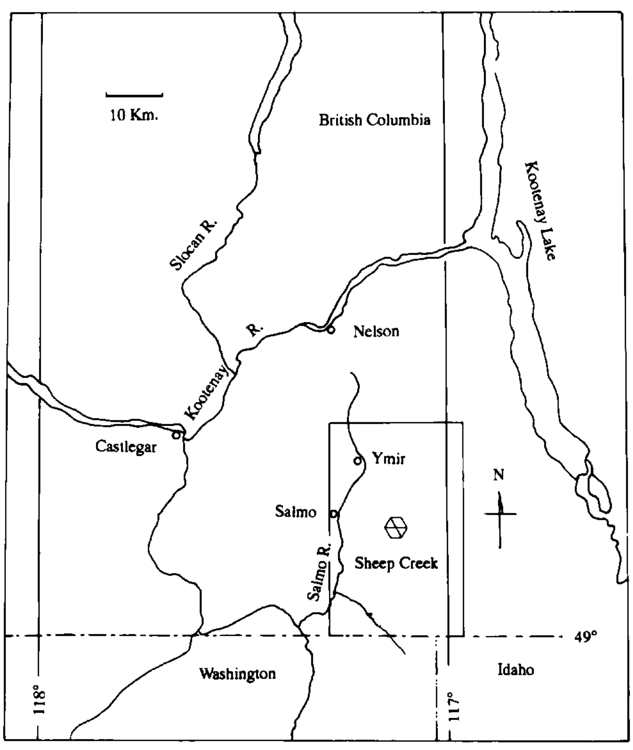

Figure 2-2-1. Location map for Sheep Creek Camp.

In 1986 the camp was ranked seventh of British Columbia's gold camps in terms of total historic production (Schroeter, 1986); this significant rank, in conjunction with interest by Gunsteel Resources and the camp's geological and geochemical similarities with other lode gold deposits in the Canadian Cordillera, prompted this research project.

The principal objective of this study is to investigate the geochemical nature of the gold mineralization. To accomplish this a sampling program was designed and executed to investigate the lateral and vertical zoning in fluid composition, hostrock and vein alteration, and temperature of mineralization.

Significant work at the Sheep Creek camp has been done by private and public interests over a period now approaching a century and surrounding districts have also been well documented. The principal studies are: Mathews, 1953; Fyles and Hewlett, 1959, and Little, 1960. These papers have proved indispensible in understanding the geology of the region.

GEOLOGY OF THE SHEEP CREEK CAMP

The Sheep Creek camp straddles the eastern boundary of the Kootenay arc. This arcuate belt of deformed and metamorphosed rocks has had a complex history of Mesozoic and Paleocene compressional and transpressional deformation overprinted by Eocene extensional tectonics. The rocks are believed to represent both a distal continental margin and part of a late Paleozoic to early Mesozoic back-arc basin (Klepacki, 1986; Parrish *et al.*, 1988).

The Quartzite Range Formation is comprised of massive white quartzites intercalated with minor argillites and argillaceous quartzites. It overlies the Three Sisters Formation of the Proterozoic Windermere Supergroup and consists principally of grey grit and lesser amounts of white grit and blue-white quartzite. Most of the gold production has been from the Quartzite Range Formation; no gold-bearing veins are known in the Three Sisters Formation.

Overlying the Quartzite Range Formation is the Reno Formation. These argillites, grading to argillaceous quartzites and dark bluish or greenish quartzites, are complexly folded and show extreme variations of thickness. The Reno Formation is hostrock for extensive mineralization in the Reno vein and two mineralized veins in the upper part of the Gold Belt mine.

The Laib Group in the Sheep Creek area is a moderately thick (300 metres) succession of six argillite and limestone members overlying the noncalcareous rocks of the Reno Formation. Other sediments in the region are the Nelway and Active formations which are black to dark grey to grey limestones, calcareous argillites, slates and phyllites, and minor quartzites.

This apparently conformable miogeoclinal succession is folded into a sequence of north-northeast trending, overturned anticlines and synclines which are considered significant structural controls in the camp. The Windermere Supergroup is exposed to the east of the camp and Mesozoic eugeosynclinal rocks of the Rossland Formation outcrop to the west. The geology of the district is further complicated by a host of intrusive bodies including: granites, quartz monzonites, quartz porphyry sills and dikes and lamprophyre dikes (Mathews, 1953; Fyles *et al.*, 1959).

Four well-defined sets of faults are recognized in the camp but only the northeasterly trending, dextral strike-slip faults have significant associated mineralization. Quartz veins of variable thickness (less than 1 centimetre to more than 2 metres) and extensive vertical and lateral continuity are the source of all production to date. Gold occurs as small (<30 microns) isolated particles within the veins and is most often associated with sphalerite, pyrite or galena. As a generalization, ore-grade gold mineralization is confined to veins with quartzite forming either or both footwall and hangingwall. Significant pyrite is present in the wallrocks and in some areas, particularily the Reno mine, extends for many metres away from any known veins. Galena and sphalerite are usually associated with carbonates or calcareous argillites.

Metamorphism in the region is for the most part low-grade greenschist facies. Sericite is the most common accessory mineral in the quartzites; local chlorite alteration is associated with veining in more argillaceous rocks. Silicification is widespread throughout the region. Mathews (1953) describes argillic alteration, in the form of kaolinite-rich zones and pseudomorphs of feldspars.

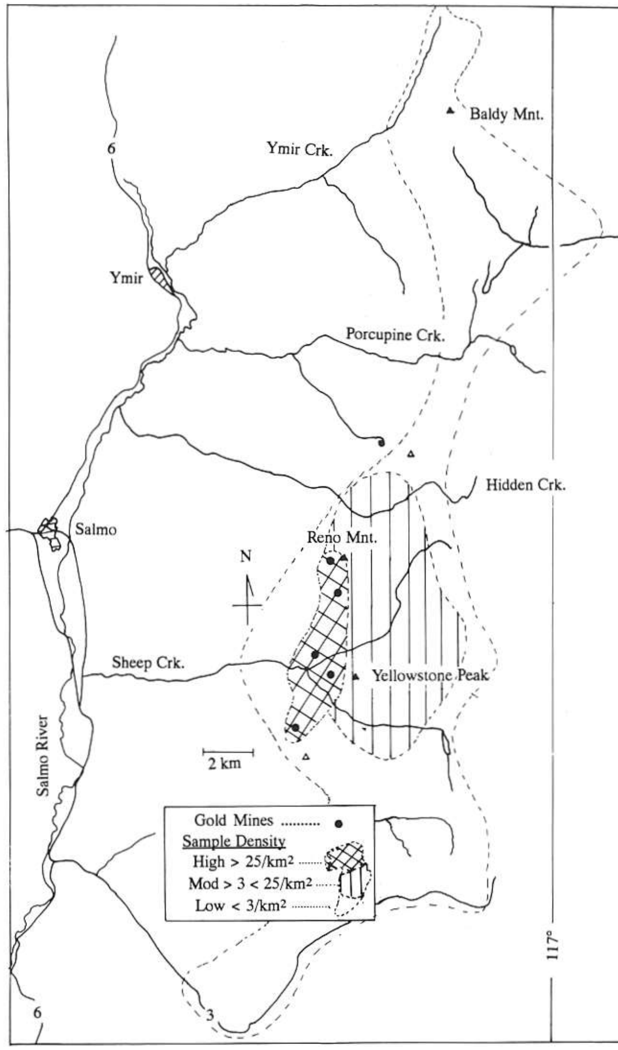

Figure 2-2-2. Sample Density map for Sheep Creek Camp.

SAMPLING

In keeping with the principal objective of the study, an attempt was made to recover as many representative examples of vein material as possible and to make complete observations of the vein characteristics and relationships with the country rock. The observed geological characteristics will provide a framework for interpretation of fluid inclusion data and stable isotope signatures.

A suite of 353 samples was collected (Figure 2-2-2) comprised of: 78 mineralized (defined as "anomalous concentrations of sulphides") and 102 apparently barren quartz veins; 31 mineralized and 79 apparently barren quartzite samples; 17 mineralized and 24 apparently barren argillites and phyllites; 4 mineralized and 10 apparently barren carbonate samples and 18 miscellaneous rocks.

OBSERVATIONS

The active workings of Gunsteel Resources provided an opportunity to examine and sample highly mineralized (pyrite, oxidized sulphides and gold) veins and adjacent hostrocks underground. Vein characteristics range from massive, milky white quartz with disseminated pyrite through semiregularly fractured blue-grey quartz veins and irregularly fractured sulphide-rich (disseminated and stringer mineralization) veins to highly comminuted gouge material. The majority of these veins strike east-northeast, crosscutting the northeast-trending beds.

Surface outcrops of quartz veins are generally unmineralized (minor to no observable sulphides) and if they have continuity, are oriented parallel to bedding strike or are crosscutting the beds in a northwesterly orientation. Veins in massive bedded quartzite in the eastern parts of the camp are generally massive, milky white quartz. Veins in argillaceous hostrocks are often very irregular in thickness and orientation and may bifurcate locally to incorporate wallrock lenses in the veins. Quartz veins throughout the region vary in size from extremely fine, hair-like fracture fillings as observed in an outcrop of en echelon fractures on the west side of Yellowstone Peak, to huge (3 metres wide) massive outcrops on the east slope of Reno Mountain. Both of these examples are hosted by quartzites.

RESEARCH PLAN

The sample suite is presently being prepared for fluid inclusion studies, which will be performed using a USGS heating/freezing stage apparatus. Data from the fluid inclusion study will contribute to the understanding of the compositions and temperature of the ore-forming fluids. The suite is also being prepared for oxygen, hydrogen and carbon stable isotope analyses. Relevant samples are being prepared as thin sections, polished mounts and polished thin sections. Assay data supplied by Gunsteel Resources Incorporated will be supplemented where necessary by additional analyses.

It is anticipated that data from fluid inclusion and isotope work, as well as petrographic observation, will aid in relating the Sheep Creek camp to other Phanerozoic mesothermal lode gold deposits in ancient, tectonically active continental margins (Nesbitt, in press).

Preliminary $\delta^{18}O$ values for a group of five samples average 14.3‰, which is similar to other mesothermal vein deposits in the Canadian Cordillera (Nesbitt, in press) but it is inappropriate to draw further conclusions until a more complete representation of the sample suite has been analysed and data from fluid inclusions have been ascertained.

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NOTES

PRELIMINARY REPORT ON THE SILVANA MINE AND OTHER Ag-Pb-Zn VEIN DEPOSITS, NORTHERN KOKANEE RANGE, BRITISH COLUMBIA (82F, 82K)

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KEYWORDS: Economic geology, vein, silver, lead, zinc, Sandon, Ainsworth, Slocan, Nelson batholith, Silvana mine, Kokanee Range.

INTRODUCTION

More than 100 years ago, the Bluebell deposit was located by R.E. Sproule on the east shore of Kootenay Lake. This drew interest to the area and led, in 1883, to the staking of silver-lead prospects around Ainsworth. The discovery of the Payne vein in 1891, by Eli Carpenter and John L. Seaton, was followed by a staking rush and the development of the Sandon and Slocan areas into one of the more prominent mining districts in Canada at the turn of the century.

All but a few of the deposits selected for study in this project are located in the northern Kokanee Range; the remainder are in the southern Goat Range. The study area is north of Nelson, and is bounded on the west by Slocan Lake and on the east by Kootenay Lake (Figure 2-3-1). There are 370 silver-lead-zinc vein and replacement deposits within this area, the majority of which are clustered in three former mining camps: Slocan (Sandon area), Slocan City and Ainsworth (Sinclair, 1979). Although most of the deposits yielded low tonnages of high-grade ore, 39 of them produced over 10 000 tonnes of ore. At present, only the Silvana Division of Dickenson Mines Limited is operating. The deposits have a wide distribution, occurring both within the Nelson batholith and surrounding Phanerozoic sedimentary and volcanic rocks up to 12 kilometres from surface exposures of the batholith.

Regional mapping and mineral deposit investigations in the study area began in the late 1880s (Dawson, 1890) and have been described in numerous reports by the British Columbia Geological Survey Branch and the Geological Survey of Canada (Schofield, 1920; Cairnes, 1934, 1935; Maconachie, 1940; Hedley, 1945, 1952; Little 1960; Fyles, 1967; Höy, 1980). These reports comprise regional geological maps and extensive mineral deposit descriptions which remain the basic source of information on Kokanee Range geology and mineral deposits. More recently, Brown and Logan (1988) mapped the geology of Kokanee Glacier Park area and evaluated its mineral potential.

The spatial distribution of the deposits led previous geologists to almost unanimously link mineralization genetically with emplacement of the Nelson batholith (c. 165 Ma). However, geological evidence in the Ainsworth area indicates mineralization is younger than mid-Cretaceous (95 Ma) metamorphism and deformation (Fyles, 1967; Archibald *et al.*, 1984). A fluid inclusion Rb-Sr isochron at Bluebell suggests that the mineralizing event is Miocene (c. 19 Ma; Changkakoti *et al.*, 1988). Although granite-related silver-lead-zinc vein systems commonly display some zoning outward from a nearby intrusion, previous attempts to define such zoning in the study area resulted in conflicting zoning patterns (Sinclair, 1967; LeCouteur, 1973; Lynch, 1988).

The large number and areal distribution of deposits, all of which are described and coded in the MINFILE database, provide a good opportunity for a regional research project. The intent of the project is to investigate zoning patterns of metals and metal ratios, minerals, fluid inclusion temperatures and salinities, and stable and radiogenic isotopes, between the deposits and relative to the Nelson batholith, and to determine age(s) of mineralization.

METHODOLOGY

Because the study area covers parts of six 1:50 000 map sheets (82F/10, 11, 14, 15; 82K/2, 3), there is no single geological map covering the area at this scale (1:50 000). This has impeded the investigation and comparison of mineral deposits on a regional basis. Consequently, as part of this project, a new 1:50 000 regional geology map of the study area has been compiled from previous work and will be available as an Open File (Beaudoin, in prep.).

During six months of field work in the summers of 1988 and 1989, 55 deposits (Figure 2-3-1, Table 2-3-1) were selected for study. These included the 39 deposits with more than 10 000 tonnes production each; the remainder were chosen either because they were a focus of exploration activity during the project or were required to produce a more even sampling distribution for zoning studies. Samples were selected to represent the deposit paragenesis observed in outcropping veins or reported by previous investigators. In many instances, outcropping mineralization has been mined out and samples from the dump are the only material available. Alteration is rare in sedimentary but common in granitic hostrocks. Where present, alteration zones were sampled for petrographic and chemical analyses. Five weeks of detailed underground mapping and sampling were carried out in the Silvana mine (No. 50, Figure 2-3-1) over the two field seasons.

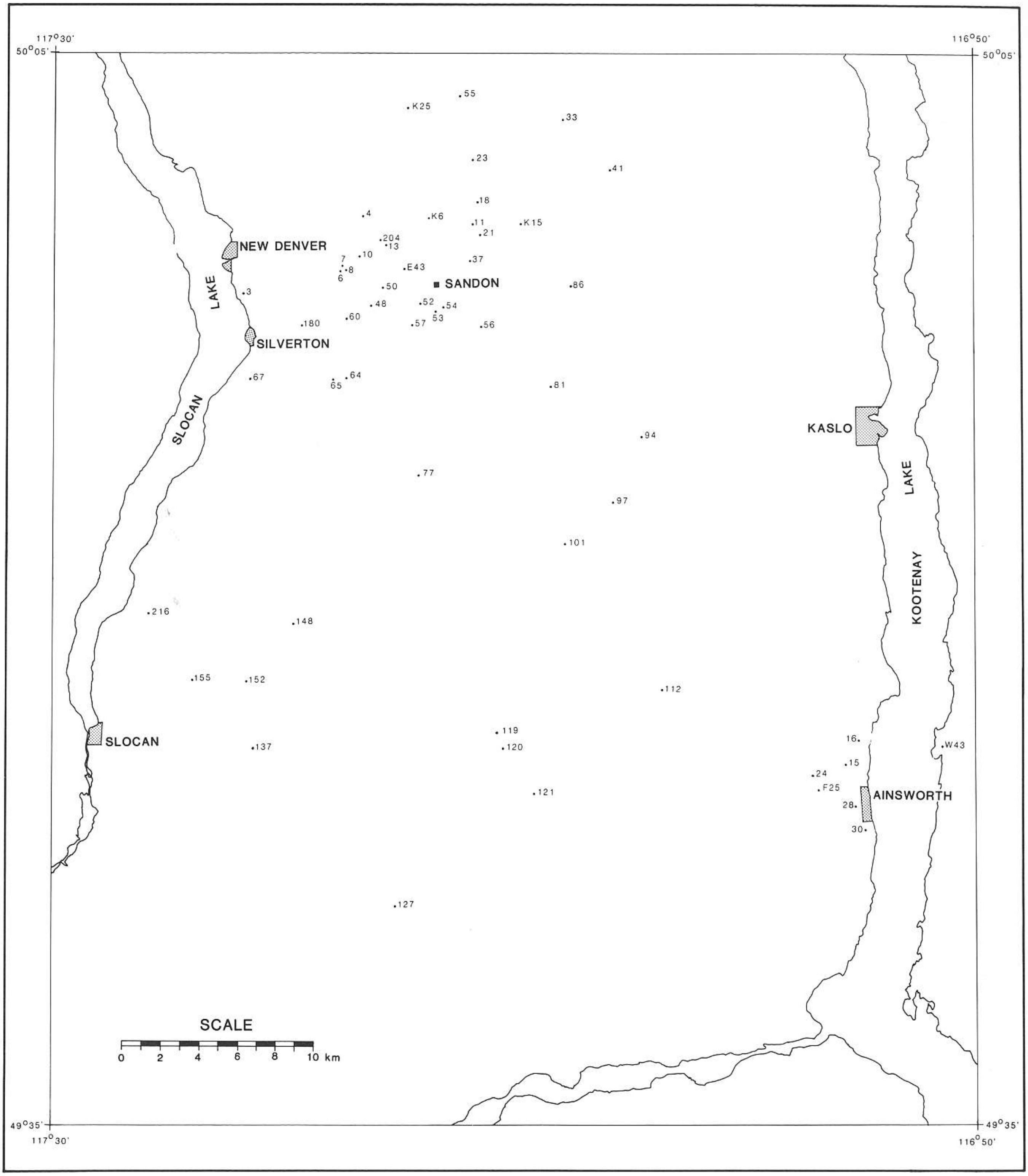

Figure 2-3-1. Location of deposits selected for the study. Numbers refer to deposits indexed in Table 2-3-1.

TABLE 2-3-1
DEPOSITS SELECTED FOR THIS STUDY
(for locations, *see* Figure 2-3-1)

No.	Name	MINFILE No.
3	BOSUN	082FNW003
4	MONITOR	082KSW004
K6	PAYNE	082KSW006
F6	SILVER BELL	082FNW006
7	IDAHO	082FNW007
8	ALAMO	082FNW008
10	QUEEN BESS	082FNW010
11	ANTOINE	082KSW011
13	HINCKLEY	082FNW013
15	HIGHLAND	082FNE015
K15	NORTHERN BELLE	082KSW015
16	FLORENCE	082FNE016
18	RAMBLER	082KSW018
21	SURPRISE	082FSW021
23	LUCKY JIM	082KSW023
24	SILVER HOARD	082FNE024
K25	McALLISTER	082KSW025
F25	NUMBER ONE	082FNE025
28	SPOKANE	082FNE028
30	HIGHLANDER	082FNE030
33	WHITEWATER	082KSW033
37	NOBLE FIVE	082FNW037
41	CALEDONIA	082KSW041
W43	WONDERFUL	082FNW043
E43	BLUEBELL	082FNE043
48	CARNATION	082FNW048
50	SILVANA	082FNW050
52	RUTH-HOPE	082FNW052
53	SILVERSMITH	082FNW053
54	RICHMOND-EUREKA	082FNW054
55	PANAMA	082KSW055
56	NOONDAY	082FNW056
57	IVANHOE	082FNW057
60	MAMMOTH	082FNW060
64	VAN ROI	082FNW064
65	HEWITT	082FNW065
67	GALENA FARM	082FNW067
77	COMSTOCK	082FNW077
81	MOUNTAIN CON	082FNW081
86	UTICA	082FNW086
94	CORK-PROVINCE	082FNW094
97	WINTROP	082FNE097
101	INDEX	082FNE101
112	SCRANTON	082FNW112
119	SLOCAN CHIEF	082FNW119
120	SMUGGLER	082FNW120
121	MOLLY GIBSON	082FNW121
127	ALPINE	082FNW127
137	METEOR	082FNW137
148	ENTERPRISE	082FNW148
152	ARLINGTON	082FNW152
155	OTTAWA	082FNW155
180	STANDARD	082FNW180
204	VICTOR	082FNW204
216	MORNING STAR	082FNW216

A major problem to be addressed in this metallogenic study is the age(s) of the mineralizing event. This is being investigated by Ar-Ar dating of micas from mineralized veins and sericitic alteration in adjacent wallrocks. Further constraints on the age(s) of mineralization are being sought by K-Ar dating of lamprophyric and gabbroic dikes, some of which are younger, and others older, than the veins.

GEOLOGY OF THE SILVANA MINE

The Silvana mine is located near Sandon, about 10 kilometres east of New Denver. Cumulative production as of June 1988 was 376 750 tonnes of ore from which 194 million grams of silver (514.7 g/t), 21.7 million kilograms of lead (5.8% Pb), and 19.3 million kilograms of zinc (5.1% Zn) were recovered.

The mine is situated in about the middle of the Main Lode fault zone which has been traced from the Standard deposit on the west (No. 180, Figure 2-3-1) to the Richmond-Eureka deposit on the east (No. 54, Figure 2-3-1). The Main Lode is a zone of faulting and brecciation up to 50 metres wide. In the currently producing eastern part of Silvana orebody, it is a rather narrow and well-defined fault zone with little, if any, penetrative fabric at its margins. The western part of the orebody, now mostly inaccessible, consists, in contrast, of a wide zone of sheared graphitic rocks. The Main Lode strikes east with a shallow dip to the south, averaging 45°; the average dip of the Silvana orebody, within the Main Lode, is about 35°. The Main Lode fault zone displays a normal and left-handed sense of movement but the amount of displacement is poorly constrained because of a lack of markers (Hedley, 1952). Although a normal and left-handed sense of shear was also determined in many locations in the eastern part of the Silvana orebody, using shear bands, the displacement could not be estimated.

DESCRIPTION OF MINERALIZATION

The orebody consists of siderite, galena and sphalerite lenses which rapidly pinch and swell in all directions. The lenses are within, and parallel with, the Main Lode fault zone. The footwall of the fault zone commonly contains subvertical siderite-sphalerite tension veins with minor galena. Hangingwall rocks rarely contain mineralized tension veins.

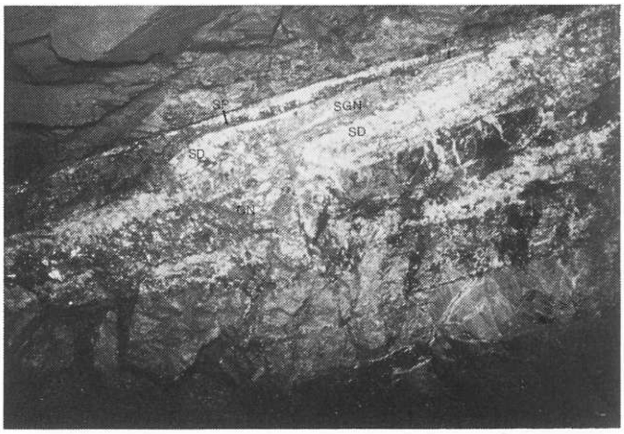

Plate 2-3-1. Siderite (SD) vein (about 30 cm thick) with a band of sphalerite (SP) at the upper margin. Contained within the siderite are scattered grains of sphalerite and galena. The lower part of the sphalerite band is cut by a fault (F) subparallel to the vein wall. An s-shaped tension opening in the siderite vein is filled by coarse-grained galena (GN). The galena is sheared (SGN) close to the fault plane. (GSC photo 205017).

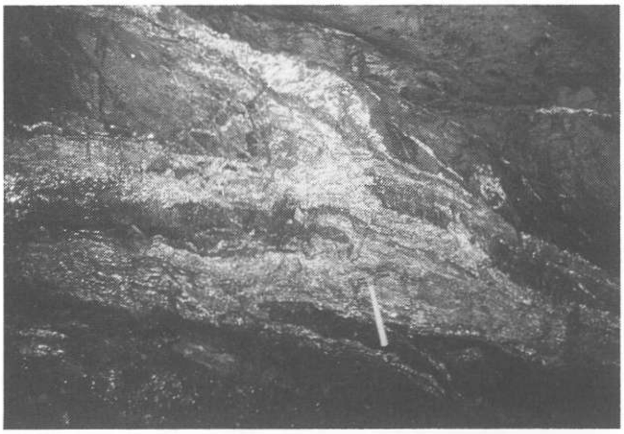

Plate 2-3-2. Vein of strongly foliated massive galena (GSC photo 205016).

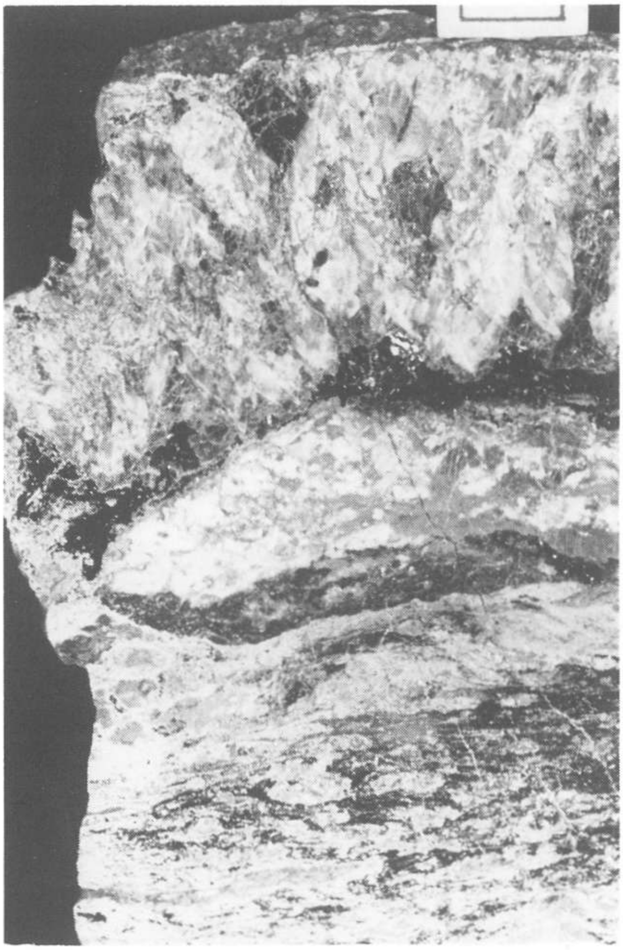

Plate 2-3-3. Polished hand-sized specimen of a vein, the top third of which is comprised of coarse-grained euhedral siderite rhombs (light grey) growing from the vein wall, intergrown with, and molded by, sphalerite (medium grey). The siderite-sphalerite zone is molded on the bottom by a narrow band of foliated galena (black). The rest of the vein consists of a chaotic aggregate of strongly deformed galena, sphalerite and siderite (scale bar is 1 cm) (GSC photo 205014).

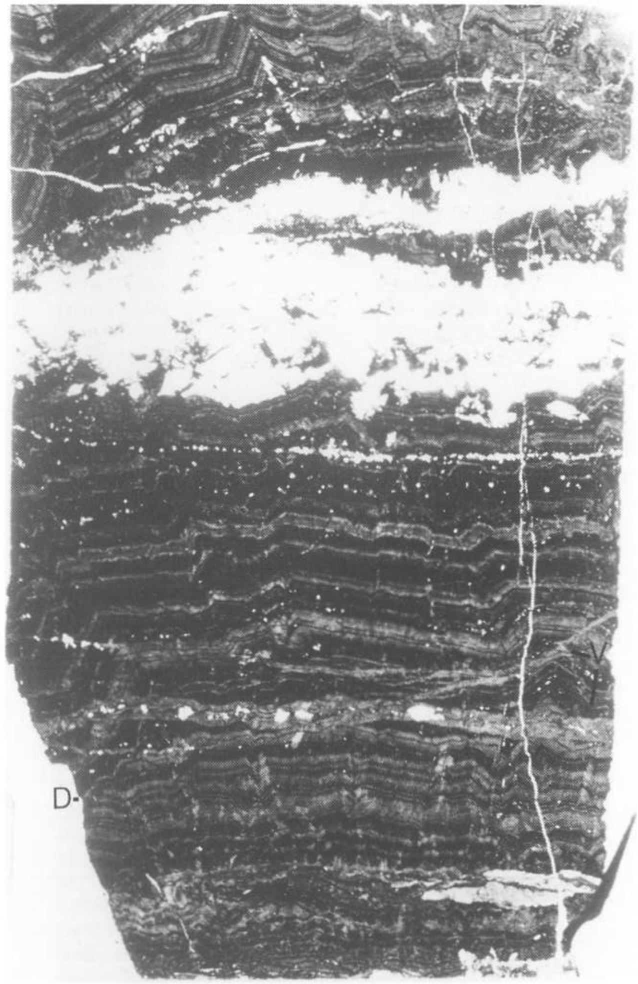

Plate 2-3-4. Photograph of a thin section of finely banded, light to dark brown sphalerite cut by a vein of carbonate (white). The lower part of the plate shows a zone of fine bands of clear sphalerite cut by a dissolution surface (D). The dissolution surface is molded by a zone of thicker and darker bands of sphalerite. The banded sphalerite is cut by veins of clear sphalerite (V) (GSC photo 205015).

The ore lenses are usually less than 2 metres thick with down-dip and strike lengths up to tens of metres. They are separated by thin (<10 cm) intervals of weakly mineralized to barren fault zones of variable length. The core of a lens may consist of massive siderite with scattered grains of sphalerite and galena (Plate 2-3-1) or massive galena (Plate 2-3-2), whereas the margins of the ore lenses are commonly composed of alternating or intergrown centimetre-wide bands of siderite and sphalerite (Plate 2-3-3). Paragenetic sequences are difficult to decipher because late movement in the fault zone has resulted in deformation of the ore lenses. Particular examples, such as illustrated in Plate 2-3-1, suggest that an initial stage of scattered galena in siderite was followed by opening of the lens and precipitation of coarse-grained galena. Late deformation resulted in sheared galena near the plane of movement, but not in areas sheltered by massive siderite and sphalerite. Foliation in galena wraps around siderite-sphalerite fragments ripped from the lens and which now "float" in foliated galena. In hand specimen, foliated galena may mold undeformed euhedral siderite prisms intergrown with, and overgrown by, sphalerite, or contain chaotic aggregates of dislocated grains of sphalerite and siderite (Plate 2-3-3). The foliation in galena is consistent with oblique normal and left-handed movement, similar to the direction of movement on the Main Lode fault zone.

Sphalerite is finely banded (Plate 2-3-4) whether it forms bands or intergrowths with siderite. Colour banding in sphalerite, oriented parallel to the vein walls, is common not only in Silvana but has been noted in other deposits in the study area, a feature not previously reported in the literature. Preliminary investigation of sphalerite banding has revealed the existence of precipitation cycles, dissolution surfaces and fracturing and veining by neosphalerite. Attempts will be made to define a local and, if possible, district-wide "sphalerite stratigraphy".

ACKNOWLEDGMENTS

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NOTES

GEOLOGY AND MINERAL DEPOSITS OF NORTHERN TEXADA ISLAND (92F/9, 10, and 15)

By I.C.L. Webster and G.E. Ray

KEYWORDS: Economic geology, Texada Formation, Marble Bay Formation, Jurassic plutonism, Bonanza arc, iron skarn, copper-gold skarn, moss-mat geochemistry.

INTRODUCTION

Texada Island, which is situated approximately 100 kilometres northwest of Vancouver between the Georgia and Malaspina Straits, is 50 kilometres long and averages 6 kilometres in width (Figure 2-4-1). The 1:20 000 mapping of northern Texada Island forms part of a 4-year project by the British Columbia Geological Survey Branch to study the province's important gold, iron and base metal skarn camps. The project was initiated by the work of Ray *et al.* (1987, 1988) and Ettlinger and Ray (1988, 1989) and this study will continue to examine the distribution and metallogeny of skarns in relation to the tectonic belts and geological terranes of the Canadian Cordillera. It will also look at the controls of mineralization and attempt to establish genetic models for skarn formation.

The northern part of Texada Island was selected as an area for detailed mapping because it contains a varied suite of polymetallic skarns that between 1896 and 1976 produced 10 million tonnes of magnetite iron ore, 35 898 tonnes of copper, 39.6 tonnes of silver and 3.3 tonnes of gold. There are also numerous precious and base metal vein deposits and occurrences and an almost inexaustible supply of pure limestone suitable for the manufacture of lime and cement.

PREVIOUS WORK

The discovery of magnetite in the northern part of the island in 1873 marked the beginning of a long history of mining activity. An examination of the shoreline geology was undertaken by Dawson (1885), a comprehensive report and geological map was made by McConnell (1914), and a study of the iron skarns was made by Swanson (1925). More recent work on various aspects of the island's skarn mineralization includes that by Bacon (1952), Muller and Carson (1968), Sangster (1969) and Ettlinger and Ray (1989). Mathews and McCammon (1957) examined the limestones. The 1:20 000 mapping study of northern Texada Island (Webster and Ray, 1990) which forms the basis of this paper, was completed during the summer of 1989.

GEOLOGY

INTRODUCTION

Texada Island lies at the eastern edge of both the Wrangellia Terrane and the Insular physiographic belt. It is mostly underlain by volcanic rocks of the Middle to Late Triassic Texada Formation which, in the northern part of the island, is conformably overlain by massive limestone of the Marble Bay Formation (Figure 2-4-1). These are respectively correlated with the Karmutsen and Quatsino formations of the Vancouver Island Group. Paleozoic Sicker and Buttle Lake Group rocks underlie the Texada Formation and are exposed in a narrow band on the south end of the island. Various stocks and minor intrusions, ranging in composition from gabbro through diorite to quartz monzonite, intrude the volcanics and limestones; many of these are I-type, calcalkaline intrusions related to the Middle Jurassic Bonanza magmatic arc (Ettlinger and Ray, 1989), and some are associated with iron and copper-gold skarn mineralization. Poorly exposed Cretaceous sediments of the Nanaimo Group crop out around Gillies Bay; these represent the eastern margin of the Comox basin.

TEXADA FORMATION

Middle to Late Triassic volcanic rocks of the Texada Formation largely comprise pillowed and massive basaltic flows with thick units of plllowed breccias; the flows are often amygdaloidal and spherulitic. Most of the island is underlain by this formation; the most extensive section in the study area is exposed on Surprise Mountain where it reaches 300 metres in thickness.

Although the base of the formation is not seen on northern Texada Island, a stratigraphic succession is recognized. This consists of a lower sequence of predominantly massive basaltic flows with subordinate amounts of pillow lava and pillow breccia. The flows are characterized by amygdules, feldspar phenocrysts and spherulitic "snowflake" textures. Higher in the succession, pillowed flows and pillow breccias predominate. The top of the sequence, immediately underlying the Marble Bay Formation, often comprises thick units of fine hyaloclastite breccia and coarse pillow breccia. The pillow breccias are monomictic with angular to subrounded volcanic clasts up to 30 centimetres across; some breccias display alternating layers of finer and coarser grained material. Particularly well-preserved pillowed flows and pillow breccia outcrop along the west coast between the south end of Crescent Bay and Favada Point (Figure 2-4-1). In this area highly elongate pillows are seen and the layering in the breccias indicates the rocks dip moderately northeast.

The volcanics in the Surprise Mountain area include a 15-metre-thick subhorizontal unit of columnar-jointed basalt, that overlies rhythmically layered amygdaloidal, feldspar-porphyritic and spherulitic flows. Columns vary between 1 and 1.5 metres in diameter and up to 15 metres in height; it is uncertain whether this unit represents an intrusive sill or a slow-cooling flow, although the latter is favoured. The discreet zones of pillowed and massive flows and the rare columnar jointed basalts suggest areas of submarine ponding developed during extrusion of the lavas.

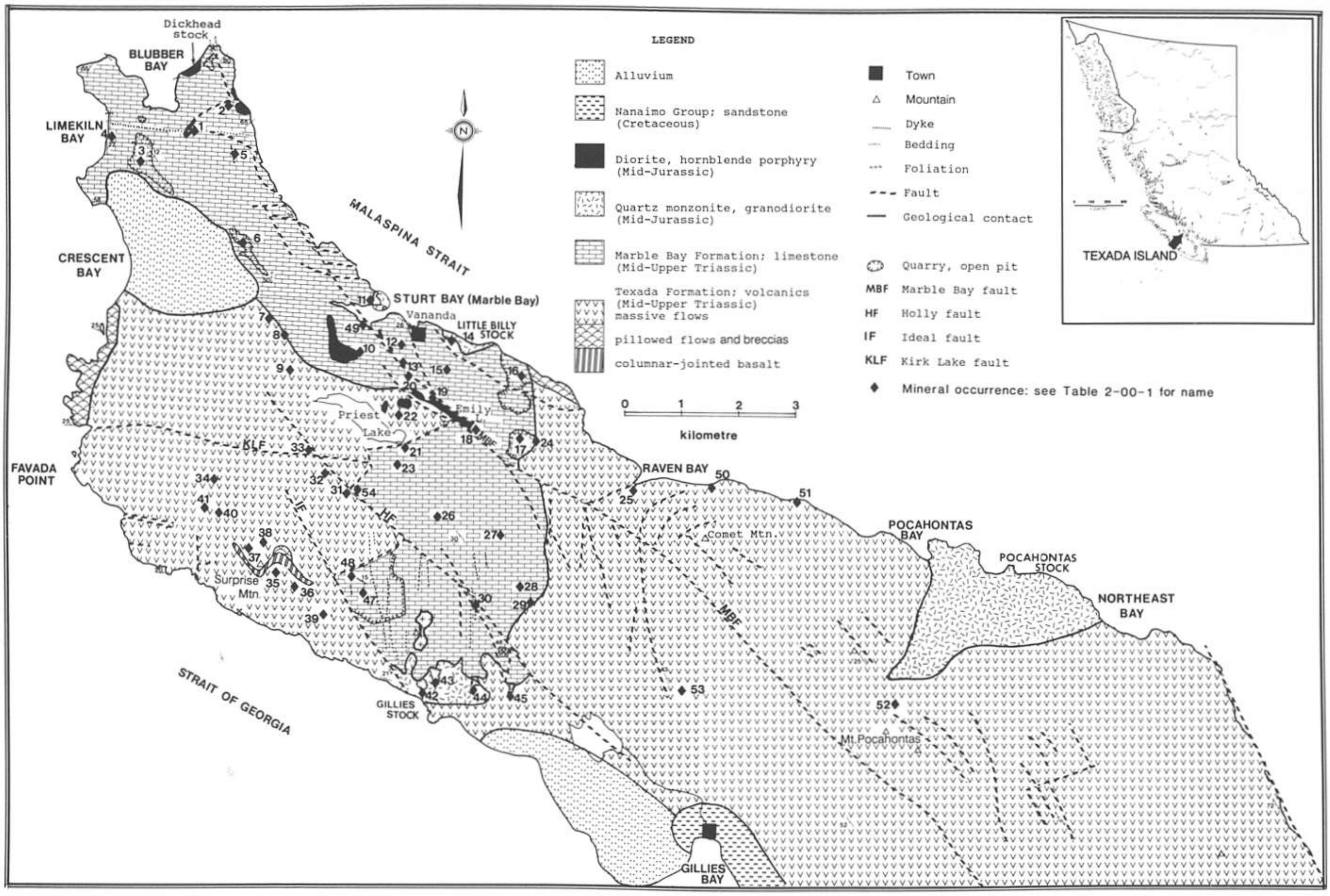

Figure 2-4-1. The geology and location of mineral occurrences on northern Texada Island.

Near the top of the Texada Formation, close to its contact with the overlying Marble Bay Formation, the flows and pillow breccias contain beds of fossiliferous grey limestone up to 3 metres thick, together with with rare, thin beds of dark shale. Crinoid and bivalve fragments are abundant and the tops of the limestone beds often contain irregular and nodular cherty concretions up to 5 centimetres in diameter. These thin beds of fossiliferous limestone are best exposed and least altered on the upper slopes of Surprise Mountain; elsewhere they tend to be recrystallized to marble or overprinted by skarn alteration.

Marble Bay Formation

The Upper Triassic Marble Bay Formation is a limestone sequence 60 to 520 metres thick that occupies a belt 3 kilometres wide extending northwest from Gillies Bay to Vananda and Blubber Bay (Figure 2-4-1). It conformably overlies the Texada Formation and mainly comprises pure, massive to poorly bedded, grey, featureless calcareous and dolomitic limestone. Exposed contacts between the limestone and underlying volcanic rocks are usually marked by steep faults. Recrystallization and bleaching to white marble, together with intense jointing and local shearing make the recognition of primary sedimentary features difficult. The least-altered limestone and best-preserved bedding is exposed along the shoreline of Limekiln Bay and on the east coast north of Vananda, where impure silty interbeds are present.

Mathews and McCammon (1957) divided the limestones of the Marble Bay Formation into three members, based on differences in chemical composition. Calcium-rich limestones, which occupy the lowest part of the succession, are overlain in turn by two members that become increasingly magnesium-rich. These dolomitic units, which contain up to 15 per cent MgO, at first thicken eastward from the Blubber Bay No. 6 quarry and then progessively thin towards the coast; dolomitization appears to be discordant to the limestone stratigraphy. The lower lime-rich member, which at the Blubber Bay No. 6 quarry exceeds 99 per cent $CaCO_3$, is ideal for lime and cement production.

Intrusive Rocks

Numerous sills and dikes, as well as several larger stocks, intrude the Texada and Marble Bay formations. Some mafic sills may represent feeders for the Triassic volcanic rocks, but many are younger as they cut the Marble Bay limestone. The stocks are I-type calcalkaline rocks ranging in composition from quartz monzonite to gabbro, and are locally associated with skarn mineralization (Ettlinger and Ray, 1989). The more mafic stocks are generally smaller than the felsic bodies and tend to be concentrated along the northwest-trending Marble Bay fault (Figure 2-4-1); this fault zone extends from south of Emily Lake, northwards to Sturt Bay and may extend yet farther north to the Canada trench and Paris mine area. The amphibole-rich dioritic stocks and sills in this belt vary from fine to very coarse grained, and are characterized by mafic xenoliths and coarse hornblende megacrysts; numerous skarn occurrences and several past-producing mines are associated with these intrusions. The dioritic Cornell stock, at the south end of the belt, contains up to 35 per cent hornblende and pyroxene phenocrysts set in a fine-grained plagioclase-rich matrix. Similar mafic intrusions, some of which are locally apatite-rich, outcrop near the Florence and Security copper-gold skarns on the north side of Emily Lake. The mafic intrusions outcropping along the northeast coast below the Loyal mine dump contain rounded to angular mafic xenoliths of coarse hornblendite and gabbro up to 30 centimetres across, as well as smaller, rarer xenoliths of massive magnetite. The Dickhead stock (Figure 2-4-1), along the north shore east of Blubber Bay, exhibits compositional layering with abundant xenoliths of magnetite and hornblende-rich material and lesser amounts of paler siliceous material.

The felsic stocks are exposed along the northeast and southwest coasts and include the Gillies, Little Billy and Pocahontas stocks. The Gillies stock, which has yielded a zircon U-Pb radiometric age of 178 Ma (Ettlinger and Ray, 1989), is genetically associated with several magnetite-rich skarn deposits. It mainly comprises a grey, medium-grained equigranular quartz monzonite that contains amphibole, biotite and occasional pyroxene phenocrysts. A late potassium feldspar rich phase is also present. The stock and the surrounding limestones are cut by sets of northerly and easterly trending feldspar-porphyritic dikes that reach 10 metres in thickness and postdate skarn mineralization. The Little Billy stock, on the north coast, is spatially associated with copper-gold skarn mineralization at the Little Billy mine. It comprises a light grey, fine to medium-grained equigranular tonalite (Ettlinger and Ray, 1989) containing up to 10 per cent biotite and hornblende. The Pocahontas stock, farther southeast, is a light grey, equigranular, quartz-rich intrusion that contains approximately 10 per cent biotite and 5 per cent hornblende. Unlike the Gillies and Little Billy stocks, it is not associated with known skarn mineralization.

Other small granodioritic bodies outcrop near the Marble Bay mine, at the head of Marble Bay and at the entrance to Sturt Bay. Like the more mafic intrusions associated with copper-gold skarn mineralization, these felsic bodies sporadically contain subrounded xenoliths of hornblendite up to 5 centimetres in diameter. A distinct, easterly trending quartz porphyry dike that transects the island from Limekiln Bay, through the Paris mine, to the east coast, may be Cretaceous in age. This dike reaches 7 metres in width and appears to postdate the major northwest-trending faults. The margins of the dike contain spherical to oval-shaped cavities up to 5 centimetres in diameter. These probably represent gas bubbles that were concentrated along the edges of the dike during its intrusion. All of the stocks sampled to date give Middle Jurassic radiometric ages (Ettlinger and Ray, 1989).

STRUCTURAL GEOLOGY

Texada Island lies close to the eastern margin of the Insular Belt, and may possibly represent a horst along the eastern edge of the Comox basin (T. Hamilton, personal communication, 1989). Only one major episode of folding (F_1) has been recognized; this resulted in the limestones and, to a lesser extent, the underlying volcanics, being deformed into a series of broad, northwest-trending open folds that plunge northwards. Three subparallel, northwesterly striking lineaments are also recognized, reflecting zones affected by both ductile and brittle movement; from south to north these lineaments coincide with the Ideal, Holly and Marble Bay faults (Figure 2-4-1); These faults cut a set of northeasterly striking faults, one of which forms the northwest boundary of the limestone in the Ideal quarry. Locally the younger, northwest-trending structures are represented by a single fault but elsewhere they contain numerous splays of brittle fractures, as seen along the Marble Bay fault zone near Sturt Bay and Blubber Bay. Slickenside measurements and offset of the limestone-volcanic contact along the Marble Bay and Holly faults indicate up to 600 metres of subhorizontal sinistral movement. A noticeable increase in the intensity of F_1 folding occurs within and adjacent to the northwest-trending structural lineaments, particularly toward the Marble Bay fault. In the Limekiln Bay area the limestones are gently folded, but toward the Marble Bay fault the folding becomes tighter, and the development of bleached marble is more common. Localized ductile deformation along the northwest-trending lineaments includes the development of a marked penetrative shear fabric and ductile flow folds in the marbles, as well as the formation of stretched boudin structures in some dikes that intrude the marble.

The Marble Bay fault, and to a lesser extent the Ideal fault, have also apparently controlled the emplacement of some of the Jurassic intrusions and their associated skarn mineralization; this suggests that the plutonism occurred along lines of structural weakness and that the northwesterly trending lineaments are at least Middle Jurassic in age. The Cornell, Copper Queen, Florence-Security, Little Billy, Charles Dickens and Marble Bay copper-gold deposits are all located on or close to the Marble Bay fault, while the Gillies stock and its associated iron-skarn deposits lie close to the Ideal fault. Locally at the iron mines, the volcanic-limestone contact is highly deformed (Figure 2-4-2, 3A, 3B) and these structures have partly controlled the distribution of the magnetite ore. The abundant xenoliths in the mafic plutonic suite associated with the Marble Bay fault, and the morphology of the intrusions associated with the iron-skarn deposits close to the Ideal fault (Figures 2-4-3A and 3B), suggest that diapiric and explosive intrusive activity occurred. It is uncertain, however, whether the intense deformation of the volcanic-limestone contact at the iron mines resulted from localized F_1 regional folding, thrust faulting or forcible diapiric emplacement of the intrusions. This uncertainty is illustrated at the Paxton mine where both the limestones and the relatively competent volcanics are deformed into a series of tight fold-like structures (Figures 2-4-3A and 3B).

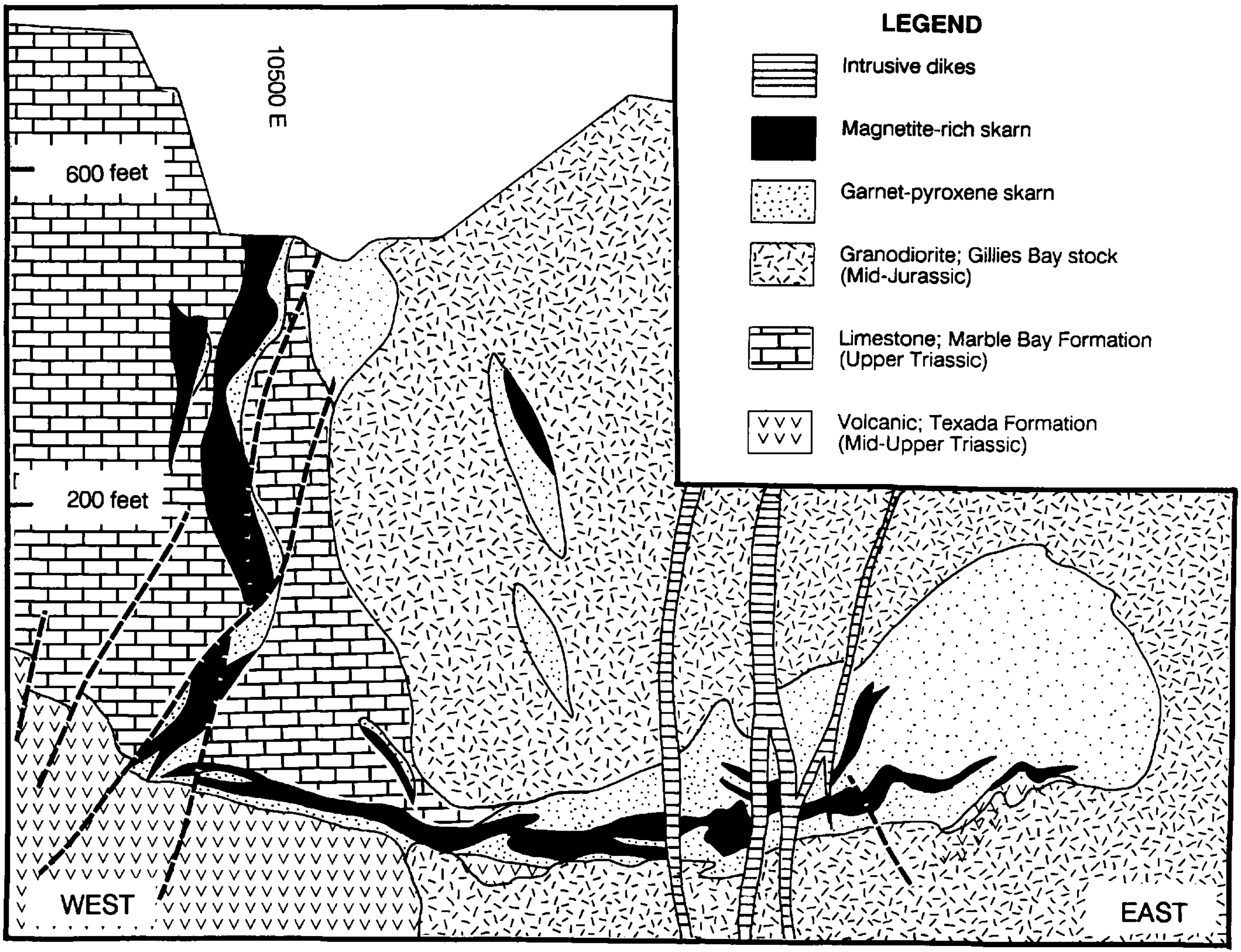

Figure 2-4-2. East-west geological cross-section through the Prescott pit at 11135 north, Texada Iron mine. Data courtesy of Ideal Cement Company.

MINERALIZATION

The mineral deposits and occurrences on northern Texada Island are listed in Table 2-4-1. Mineralization mainly comprises skarns and veins, both of which carry base and precious metals. Ten million tonnes of iron ore and substantial amounts of copper, gold and silver have been produced from the skarns but the quartz and carbonate veins, which received considerable attention during the early mining history of the island, have so far been relatively unproductive.

SKARNS

Skarn mineralization can be broadly divided into iron-rich, which is associated with the felsic Gillies stock, and copper-gold-rich, which is mostly related to a suite of mafic intrusions. All skarns on the island are associated with high-level calcalkaline plutonism and were formed under oxidizing to intermediate conditions (Ray *et al.*, 1990, this volume). Small occurrences of skarn alteration are common throughout the Marble Bay Formation; in some cases the hornblende porphyry dike suite intruding the limestone exhibits varying degrees of endoskarn alteration but exoskarn halos are generally less than 1 metre thick and, in many cases, are totally lacking. In numerous localities the limestones have been intensely bleached as in the Imperial and White Rock quarries. Bleaching often follows subvertical joints and gently dipping bedding planes, and was probably caused by the movement of fluids and gases produced by skarn development at depth.

IRON SKARNS

Iron-skarn mineralization is seen at the Prescott, Yellow Kid, Paxton and Lake mines where it is generally developed close to or along the margin of the Gillies stock (Figure 2-4-1). Mineralization is concentrated along either the Marble Bay–Texada Formation contact, the margins of the Gillies stock or within limestone and volcanic rocks some distance from the stock where the skarn-forming fluids were controlled by subvertical brittle fractures (Figure 2-4-2). Magnetite

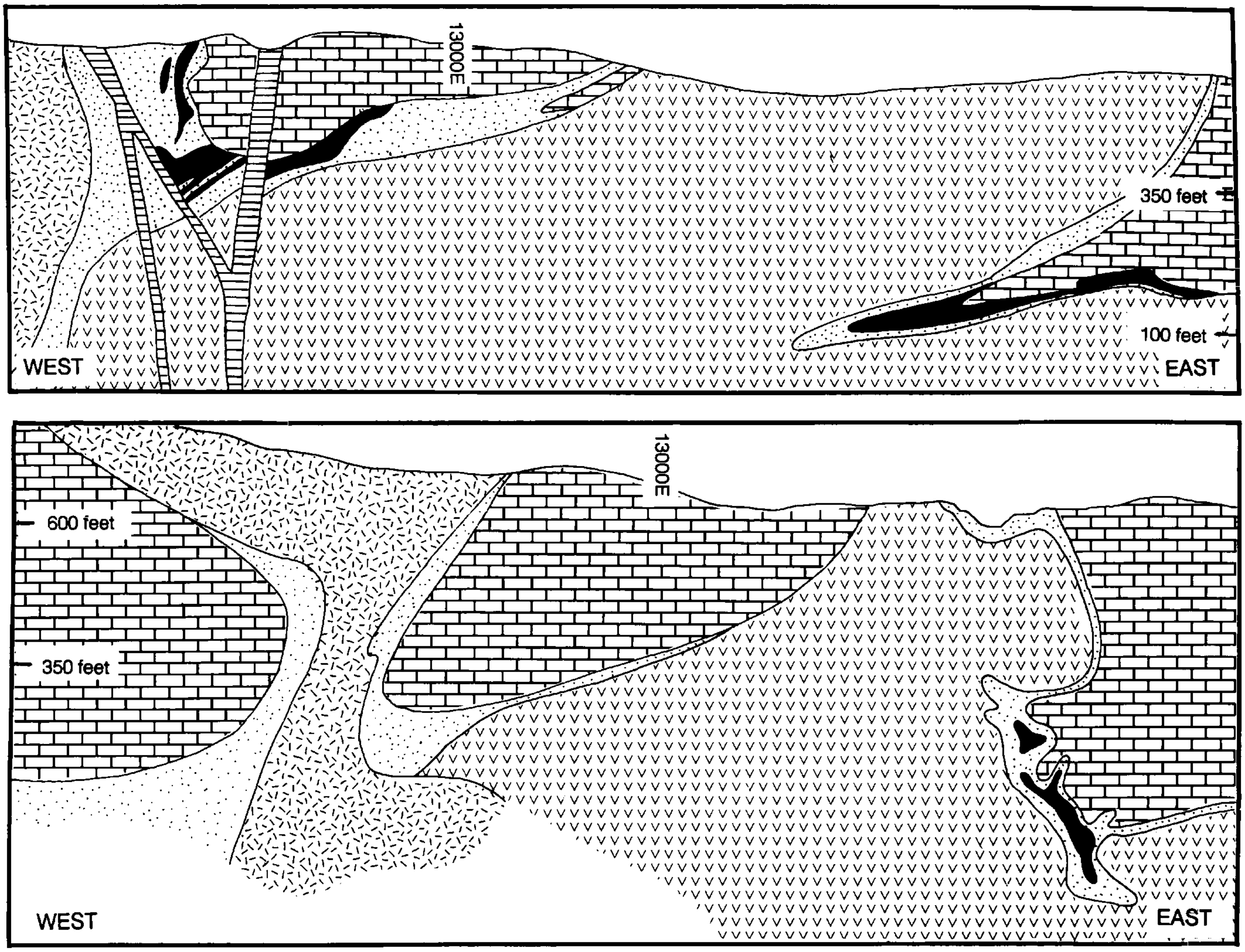

Figure 2-4-3A and 3B. East-west geological cross-sections at 11300 north and 11900 north, Texada Iron mine. Legend as in Figure 2-4-2.

orebodies adjacent to the stock are generally associated with abundant garnet-pyroxene-amphibole skarn, while the more distal, structurally controlled, subvertical deposits have less extensive skarn envelopes. The massive magnetite occurs with reddish brown garnet, pyroxene, epidote, actinolite and sporadic chalcopyrite, pyrite and pyrrhotite. No arsenic or bismuth enrichment is reported (Ettlinger and Ray, 1989) and gold values are generally less than 0.5 gram per tonne. The skarn alteration and mineralization overprints all phases of the Gillies stock and, to a lesser degree, the limestone and volcanic rocks, although it is often difficult to distinguish between exoskarn and endoskarn. Contacts between the skarn and unaltered rocks are generally sharp. Mineralogical zoning is recognized and, where fully developed, comprises barren skarn close to the intrusion, grading outwards to magnetite-rich skarn and then into marble. Locally, chalcopyrite and pyrite occur close to the outer margins of the skarn envelope, adjacent to limestone or marble. Magnetite veinlets commonly cut garnet-pyroxene skarn (Sangster 1969) and in the Yellow Kid mine, veinlets of pyrite and chalcopyrite crosscut the magnetite. Thus, early garnet-pyroxene assemblages were followed, in turn, by the introduction of magnetite and late sulphide mineralization.

Minor iron skarn also occurs farther north, near the Imperial Limestone quarry and at the Raven Bay showings (Figure 2-4-1; Table 2-4-1). These showings, which lie close to the contact between the Marble Bay and Texada formations, comprise magnetite and chalcopyrite with high (0.17 per cent) cobalt ; some erythrite-bearing float is seen in this area.

COPPER-GOLD SKARNS

Copper-gold skarns are more widely distributed on the island than the iron skarns, and are also more variable in their mineralogy and chemistry. Copper and gold were won from a number of deposits including the Marble Bay, Little Billy, Copper Queen and Cornell mines southeast of Vananda (Figure 2-4-1). Mineralization often forms irregular pipe-like bodies that plunge moderately, subparallel to the contacts

TABLE 2-4-1
LIST OF MINERAL DEPOSITS AND OCCURRENCES

Map Number	Occurrence and Commodity	92 F MINFILE Number	Type
1	Paris, Fe, Cu, Zn	266	skarn
2	Loyal, Cu, Ag, Pb, Zn	265	skarn
3	Blubber Bay No. 6, lst	397	quarry
4	Limekiln Bay, lst	407	quarry
5	Canada, Fe, Cu	267	skarn
6	Hiesholt, lst	*	quarry
7	Bolivar, Au	364	u
8	Marjory, Ag, Au	109	vein
9	Saga	*	vein
10	Volunteer, Fe	268	skarn
11	Oke, Cu, Zn, Ag	374	vein
12	Marble Bay, Au, Ag, Cu	270	skarn
13	Charles Dickens, Cu, Zn	295	u
14	Little Billy, Cu, Au, Ag	105	skarn
15	Copper Queen, Cu, Ag, Au	271	skarn
16	LaFarge-Beale, lst	396	quarry
17	Imperial, lst	394	quarry
18	Cornell, Cu, Au, Ag	112	skarn
19	Florence, Cu	210	skarn
20	Security, Fe, Cu	269	skarn
21	Wolfe	*	u
22	Yew, Cu, Au, Fe	*	skarn
23	Belle, Au	*	u
24	Midas, Fe, Co	*	u
25	Raven, Cu, Fe, lst	111	u
26	Lucky Jack	*	u
27	Vauxhall, Au, Ag, Cd	*	u
28	Sentinel, Cu, Pb, Zn	113	vein
29	Aladdin	*	vein
30	Sandy, Zn, Pb, Ag, Au, Cu	373	u
31	Holly	321	vein
32	Gem, Au	*	vein
33	Victoria, Au	264	vein
34	Iron Horse	*	vein
35	Surprise, Au, Ag, Cu, Zn	262	vein
36	Copper King, Au, Ag, Cu	263	vein
37	Silver Tip, Au, Ag, Cu, Zn	261	vein
38	Nancy Bell	*	vein
39	Retriever, Cu, Ag, Au	357	vein
40	Tyee	*	vein
41	Lion, Cu	*	vein
42	Prescott, Fe, Cu	106	skarn
43	Yellow Kid, Fe, Cu, Au, Ag	258	skarn
44	Paxton, Fe, Cu	107	skarn
45	Lake, Fe, Cu	259	skarn
46	White Rock, lst	*	quarry
47	Ideal, lst	395	quarry
48	Manto, Au, Zn, Pb, Cu	*	u
49	Marble Bay, lst	095	quarry
50	Malaspina, Fe	273	u
51	Lucky Lead	*	u
52	Black Prince, Au, Cu, Ag	108	skarn
53	Cap Sheaf, Cu, Fe	274	skarn
54	Maude Adams	*	u

* = no MINFILE number.

u = unknown type of mineralization.

between limestone and intrusive rocks. Compared to the iron-skarn deposits, which developed close to the base of the Marble Bay Formation, the copper-gold skarns southeast of Vananda occur throughout the thick limestone succession. The composition of the intrusion genetically related to the Little Billy skarn is uncertain. The main Little Billy stock is tonalitic (Ettlinger and Ray, 1989); this is atypical of most gold-copper skarns on the island which are generally associated with more mafic dioritic plutonism. Recent drilling by Freeport-McMoRan Gold Company indicates some of the skarn is spatially associated with a suite of amphibole-rich mafic dikes that may postdate the tonalitic Little Billie stock. The Little Billy skarn comprises coarse, light tan grossularite and light green and dark brown andradite garnet that varies compositionally from Ad_{35} to Ad_{70} mole per cent (Ettlinger and Ray, 1989), as well as wollastonite, clinopyroxene, tremolite, quartz and feldspar. The main ore minerals are chalcopyrite and bornite with variable but minor amounts of molybdenite, pyrite, magnetite and sphalerite. Compared to the other copper-gold skarns, the Little Billy mineralization is locally pyrite-poor. Bornite sometimes occurs as coarse euhedral crystals intergrown with garnet, and the higher gold values are commonly found with the higher copper concentrations (C.N. Forster, personal communication, 1989). Ettlinger and Ray, (1989) report the gold occurs as minute 20 to 50 micron blebs that are attached to grains of bornite. Chalcopyrite and bornite are interstitial to bladed wollastonite. Other minerals at the Little Billy include galena, scheelite and native silver as well as the tellurides hessite, petzite and wehrlite (A. Panteleyev, personal communication, 1989).

The Marble Bay orebody is reported to have carried chalcopyrite, bornite and native silver within extensive, steeply dipping, skarn-altered fracture zones that cut brecciated limestone (McConnell, 1914). The sulphides, gold and silver tend to be concentrated along one margin of these zones at the contact between skarn and marble or skarn and unaltered limestone; the other margin is commonly occuppied by barren garnet-pyroxene-epidote-tremolite-calcite skarn.

The ore at the Cornell and Copper Queen mines is dominated by chalcopyrite and bornite. However, the copper-gold skarn mineralization in the Florence and Security area, north of Emily Lake, is locally more magnetite rich. This mineralization, like the iron-skarns, lies close to the faulted and unconformabie contact between the Marble Bay limestones and the underlying Texada Formation, although it is related to small mafic, amphibole-phenocrystic sills and stocks that are locally apatite rich and epidotized. Mineralization, which may be massive, is commonly developed in the basal gritty limestones; it consists primarily of magnetite with some chalcopyrite, pyrite and chalcocite in a garnet-pyroxene gangue.

In the Blubber Bay area, copper-gold skarn containing pyrite, bornite, chalcopyrite, pyrrhotite, sphalerite, galena and variable amounts of magnetite occurs at the Paris and Loyal mines and at the Canada trench. These properties are associated with a suite of elongate hornblende-rich dioritic intrusions that commonly contain mafic xenoliths and occupy major fractures. In the Loyal area, northeasterly trending skarn-altered mafic dikes, more than 250 metres long, carry minor sulphides and a little gold, but the exoskarn halos associated with these intrusions seldom exceed 1 metre in thickness.

With the exception of the Paris mine, where crystalline native arsenic has recently been identified by x-ray diffraction (M. Chaudhry, personal communication, 1989) in marbles adjacent to the outer margins of the skarn, there are no reports of arsenic enrichment or arsenopyrite present in the skarns on Texada Island. Bismuth analyses are generally low except at the Paris skarn, and tellurides have only been identified at the Little Billy mine.

VEIN MINERALIZATION

Numerous quartz and carbonate veins, carrying a varied suite of base and precious metals, were discovered and worked in the early 1900s, particularly in the Surprise Mountain area (Figure 2-4-1). Many of these are described by McConnell (1914) and listed in Table 2-4-1. Most veins are located in or adjacent to north or northwest-trending faults or shear zones that cut the Texada Formation.

At the Lion's trench (Table 2-4-1), drusy quartz veins containing angular fragments of wallrock cut amygdaloidal basalts. These veins carry pyrite, chalcopyrite and anomolous gold values (P. Sargeant, personal communication, 1989). The Silver Tip prospect (Table 2-4-1) is a northwest-trending quartz-carbonate vein, 75 centimetres wide, occupying a shear zone that cuts weakly chloritized feldspar-porphyritic volcanic rocks. Mineralization includes pyrite, chalcopyrite, galena, sphalerite and gold.

At the Nancy Bell showing, on Surprise Mountain, a northwest-trending shear zone 3 to 4 metres wide cuts silicified volcanic rocks and a thin interbed of limestone. A grab sample of pyrite-sphalerite-chalcopyrite-galena-mineralization returned analyses of 32.3 grams per tonne gold, 96 grams per tonne silver, 1.65 per cent copper, 0.2 per cent lead, 6.37 per cent zinc, 47 ppm cobalt, 55 ppm molybdenum, 219 ppm arsenic and 102 ppm bismuth.

MOSS-MAT STREAM GEOCHEMISTRY

Forty-four moss-mat samples were collected from streams throughout the island (Figure 2-4-4) for geochemical analysis of contained silt. The collection and preparation of the samples was conducted using the procedures outlined by Matysek and Day (1988). A 31-element ICP and 6-element hydride ICP analysis was performed and the preliminary results for Au, Ag, Cu, Pb, Zn, Ni, Co, As and Hg are listed in Table 2-4-2. For comparative purposes, the mean, 90th and 95th percentile values for these elements in the Karmutsen Formation, determined from the Regional Geochemical Survey of the Alert Bay/Cape Scott area of northern Vancouver Island (RGS Open File 23) are also listed.

Due to time constraints, only one or two samples were collected from most streams, however two streams were sampled in more detail to provide some orientation data and to detect any changes in element concentration along their courses. The preliminary results show sporadic gold anomalies in streams draining the Priest Lake – Vananda area where the Marble Bay, Little Billy and Florence-Security copper-gold skarn mineralization occurs. Some weak to strong gold and copper anomalies also occur in streams draining the southern half of the island. The highest gold value (Sample 20-1) was obtained from a creek close to the southern tip of the island (Figure 2-4-4); the stream bed at the sample site contains abundant float of quartz-vein material. The complete analytical results together with the conclusions on this moss-mat sampling program will be presented at a later date.

SUMMARY AND CONCLUSIONS

Texada Island, which lies adjacent to the eastern margin of Wrangellia, may contain clues regarding the style and history of tectonism along the contact between the Insular Belt and the Coast plutonic complex. As the island possibly contains the most easterly known equivalents of the Sicker, Buttle Lake and Nanaimo groups and Karmutsen and Quatsino formations, further study of these rocks may reveal facies changes compared to rocks on Vancouver Island and provide additional data concerning both massive sulphide mineralization in the Sicker and Buttle Lake groups and skarn mineralization in the Triassic succession.

TABLE 2-4-2. ANALYTICAL RESULTS OF MOSS-MAT GEOCHEMICAL SAMPLES (for locations *see* Figure 2-4-4).

ID No.	Au	Hg	Ag	Cu	Pb	Zn	Ni	Co	As
	— ppb —					ppm			
2-1	4	60	0.1	51	9	71	12	9	8
2-2	136	40	.1	34	10	53	11	7	4
2-3	11	50	.1	37	11	67	13	8	3
2-4	1	40	.1	36	11	56	11	7	4
2-5	3	50	.1	32	5	83	15	9	6
2-6	1	20	.1	20	5	37	11	6	2
2-7	2	70	.1	60	16	159	29	15	19
2-8	2	110	.1	39	10	131	17	11	18
3-1	33	30	.1	23	10	72	6	4	14
3-2	1	20	.1	8	5	57	3	3	13
3-3	1	40	.1	10	10	68	3	3	16
3-4	2	60	.1	19	17	86	6	4	25
3-5	1	20	.1	13	7	38	3	2	6
3-6	80	30	.1	59	9	42	5	4	11
3-7	1	50	.1	11	6	46	4	3	18
3-8	9	150	.1	33	21	217	6	5	108
4-1	4	80	.1	42	8	41	11	7	3
4-2	1	30	.1	26	4	28	7	5	3
5-1	4	70	.1	47	15	43	12	11	2
6-1	1	20	.1	45	7	43	20	10	3
7-1	1	50	.1	29	10	51	12	7	2
7-2	1	40	.1	33	2	40	9	7	2
8-1	1	50	.1	37	11	38	11	6	2
9-1	2	30	.1	37	6	32	11	6	3
10-1	1	100	.2	80	11	40	18	7	7
10-2	71	30	.1	28	2	41	14	8	3
11-1	55	40	.1	88	9	87	51	25	8
12-1	4	100	.1	131	31	101	36	17	6
13-1	3	80	.1	143	15	68	42	18	2
14-1	4	60	.1	177	10	71	45	18	7
15-1	21	90	.1	153	14	98	46	23	5
16-1	27	40	.1	102	5	55	33	16	6
17-1	6	50	.2	118	7	57	33	17	2
18-1	4	130	.3	79	43	152	33	67	10
19-1	5	80	.2	54	3	54	25	11	7
20-1	320	140	.1	113	28	135	60	30	8
21-1	13	60	.1	126	29	166	64	33	9
22-1	23	80	.3	129	41	85	40	22	6
25-1	7	60	.1	23	21	41	9	8	2
26-1	126	50	.1	42	26	50	20	10	3
27-1	38	180	.7	94	88	188	20	19	17
27-2	8	100	.4	93	44	188	38	21	22
28-1	8	80	.3	59	70	74	13	10	13
29-1	1	170	.8	91	157	79	17	17	16

	Au ppb	Cu ppm
90th percentile	34	149
95th percentile	80	166
Mean	18.2	90.0
(MEMPR BC RGS 23 1988)		

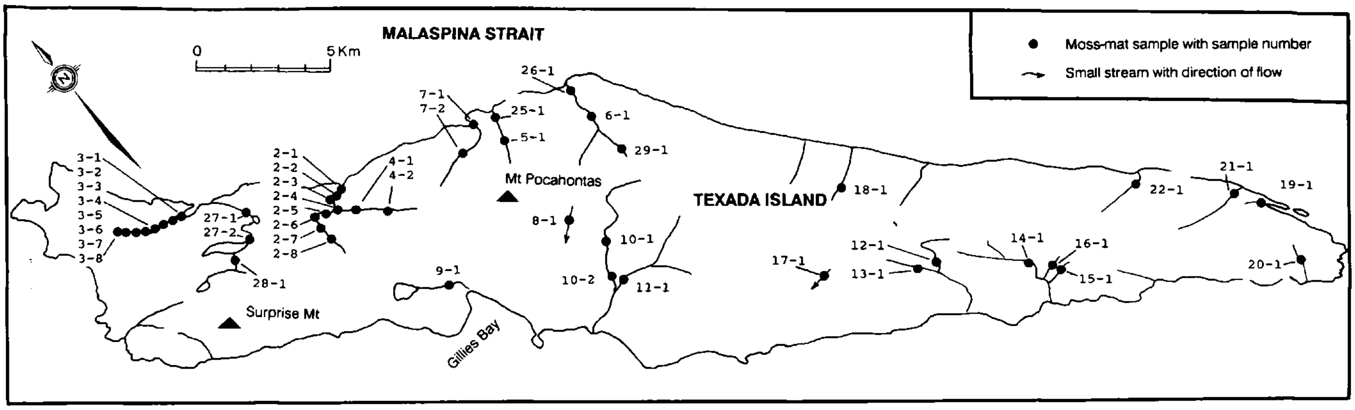

Figure 2-4-4. Location of moss-mat geochemical samples (*see* Table 2-4-2 for analytical results).

The northern part of Texada Island was affected by an episode of deformation that generally produced broad open, northwesterly striking folds in the volcanics and limestones. However, the island geology has been influenced by at least three northwesterly trending, subparallel structural zones that have undergone both ductile and brittle deformation. Evidence for ductile movement includes the development of stretched boudins in some dikes, and the presence of flow folds and a penetrative shear fabric in some marbles. These narrow belts coincide with brittle transcurrent fault zones with sinistral offset atypical of the mostly dextral faults present elsewhere in the province. Many copper-gold and iron-skarn occurrences, and their related intrusions, are associated with these belts. Thus, on a district scale, the intensity of deformation, emplacement of the intrusions, and skarn formation were controlled by mid-Jurassic zones of structural weakness. However, it is possible that the intense deformation in these zones was not entirely due to folding but partly resulted from the forcible diapiric intrusion of the stocks into the limestones.

The iron and copper-gold skarns are stratigraphically and structurally controlled, and are genetically related to a varied suite of I-type intrusions that formed part of the mid-Jurassic Bonanza magmatic arc. Intrusions of similar age farther west, at Nanaimo Lakes and Zeballos on Vancouver Island, are also associated with skarns (Ettlinger and Ray, 1989). All of the iron and some of the copper-gold skarn mineralization on Texada Island lies close to the base of the Marble Bay Formation. However, copper-gold skarns, including the Marble Bay and Little Billy deposits, occur throughout the limestone succession.

The iron and copper-gold skarns are believed to be coeval and related, and both formed within a similar high-level, oxidized to intermediate environment (Ray *et al.*, 1990, this volume). The gold-bearing skarns are unusual in only containing sporadic arsenic, bismuth or tellurium enrichment. The massive, impermeable nature of most Marble Bay limestone on the island, and its pure composition, are unfavorable for skarn formation, and this has inhibited the development of wide exoskarn halos. The presence of extensive bleaching in the limestones may indicate skarn mineralization at depth and this feature can possibly be used as an exploration guide. Moss-mat sediment sampling indicates sporadic precious and base metal anomalies that warrant follow-up investigation in several streams on the island.

ACKNOWLEDGMENTS

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NOTES

INTERPRETATION OF GALENA LEAD ISOTOPES FROM TEXADA ISLAND* (92F)

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and
Gerry E. Ray and Ian C.L. Webster

KEYWORDS: Galena lead isotope, deposit age, deposit origin, plutonogenic, Wrangellia, Coast plutonic complex, Texada Island.

INTRODUCTION

This paper demonstrates how a galena lead isotope model can be used to interpret the origin of mineralization on Texada Island, southwestern British Columbia. The showings analyzed do not contain micas or other materials suitable for traditional geochronometry, such as K-Ar or Rb-Sr dating. The speculation that some precious metal mineralization on northern Texada Island might be Tertiary in age motivated collection of some of the samples.

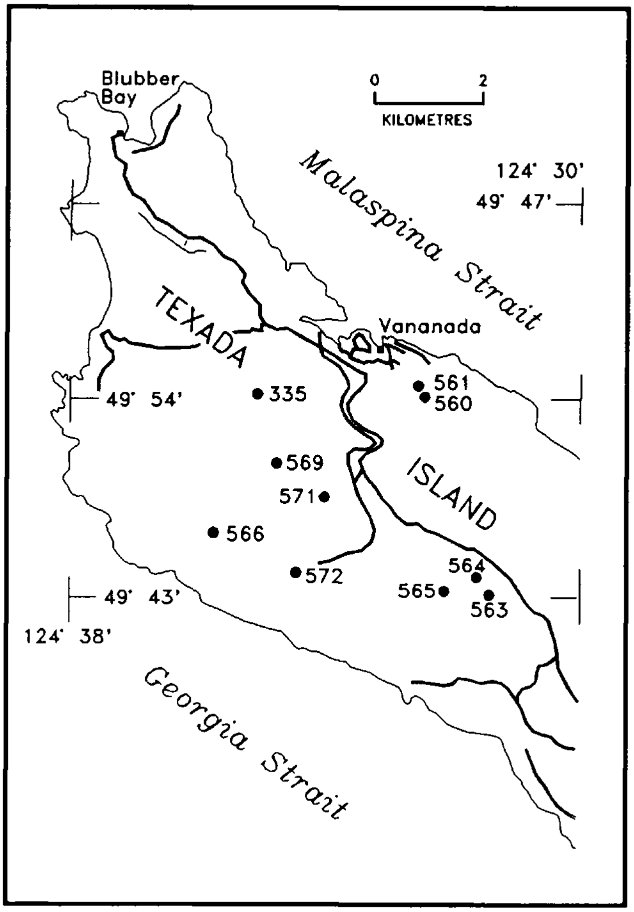

Figure 2-5-1. Northern Texada Island, showing the locations of showings with galena that were analyzed (codes for showings are prefixed by 30 in the tables that follow). Geology of the showings is described in Table 2-5-1. Data, listed in Table 2-5-2, are plotted in Figure 2-5-2.

GENERAL GEOLOGY

Texada Island (Figure 2-5-1) is within Wrangellia. Calc-alkaline plutonism of the Coast plutonic complex is superimposed on this terrane.

The northern part of the island is made up mainly of basaltic amygdaloidal flows and pillow lavas of the Texada Formation (Karmutsen Formation) overlain by massive grey limestone and white marble of the Marble Bay Formation (Quatsino Formation). These Upper Triassic formations have been intruded by diorite to quartz monzonite stocks and dikes of Jurassic age [Ettlinger and Ray, 1989; the Gillies stock, about 2 kilometres south of sample site 30565, Figure 2-5-1, has had zircons dated at Middle Jurassic (178-180 Ma years)]. Tertiary intrusions have not been identified.

Geological details of the showings are in Table 2-5-1. Most of the showings examined here are associated with sulphide-rich veins that commonly are gold and silver rich. These sulphide-rich deposits generally contain traces of galena, which facilitated this study. Exploration is currently in progress on some of these precious metal occurrences.

THE PLUTONOGENIC MODEL

Mixing-line isochrons as defined by Andrew *et al.* (1984), have been used by Reddy (1989), Leitch (1989) and Leitch *et al.* (in press) to describe linear arrays of galena lead data obtained from showings generated by plutons related to the Coast plutonic complex. Reddy (1989) called such showings "plutonogenic" to clearly draw attention to the close temporal and spatial relationships between granitic plutons and the deposits.

Deposits defined as plutonogenic by Reddy are in the Harrison Lake–Whistler–Squamish–Sechelt area of southwestern British Columbia (92G and the south-central part of 92J). These deposits, generally veins in Jurassic to Cretaceous granitic plutons or veins in adjacent country rock, yielded galena lead isotopes that plot as linear arrays on $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ plots. A line that approximates this trend is

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

TABLE 2-5-1
GEOLOGY OF TEXADA ISLAND SHOWINGS
ANALYSED IN TABLE 2-5-2

30335 NUTCRACKER (known as the GEM mine since at least the early 1920s) comprises quartz veins and veinlets within a shear zone that is about a metre wide. The mineralization is hosted by brownish feldspar-porphyritic and amygdaloidal volcanics of the Texada Formation. The veins and veinlets reportedly carried good gold values near the surface.

30560 SHOWING is a steeply dipping massive sulphide vein in Marble Bay limestone. Mineralogy of the vein is mainly pyrite, sphalerite, galena and chalcopyrite.

30561 SHOWING has the same geological character as 30560 SHOWING — it is possibly an extension of it.

30563 ALLADIN is a showing in a 400 by 200 metre zone of hydrothermal breccia in Marble Bay limestone. Galena and pyrite occur as veinlets and disseminations.

30564 SENTINEL is a banded and brecciated polymetallic quartz vein in Marble Bay limestone. Mineralogy of the vein is mainly pyrite, sphalerite, galena and tetrahedrite.

30565 SANDY is a steeply dipping massive sulphide vein, up to a metre wide, in Marble Bay limestone. Mineralogy includes: pyrite, arsenopyrite, sphalerite, galena and tetrahedrite.

30566 SILVER TIP is a northwesterly striking subvertical quartz-carbonate vein in a shear zone that is up to a metre wide. It is hosted by feldspar-porphyritic volcanics of the Texada Formation. Mineralization includes sphalerite, pyrite, chalcopyrite, galena and gold.

30569 VICTORIA is a quartz and carbonate vein system occurring at the intersection of the Holly and Kirk Lake faults. It is hosted by amygdaloidal basalts of the Texada Formation. Mineralization includes pyrite, sphalerite, galena and gold.

30571 HOLLY FAULT mineralization occurs on the north side of the Holly fault in brecciated limestone of the Marble Bay Formation. Mineralization from the old workings carries pyrite, galena and malachite.

30572 RETRIEVER is a wide, west-striking, quartz-bearing shear zone in amygdaloidal basalt of the Texada Formation. Mineralization associated with the quartz includes: pyrite, disseminated chalcopyrite, galena and gold.

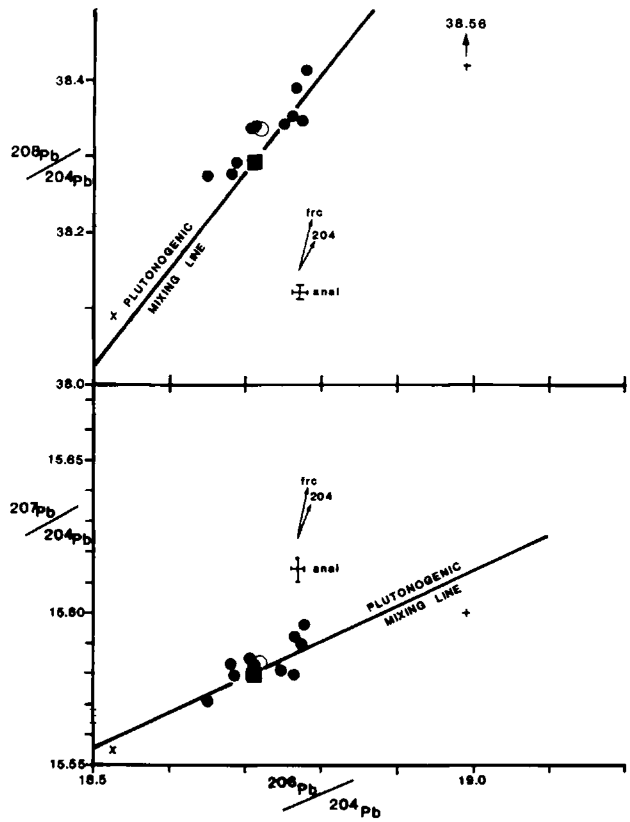

Figure 2-5-2. Lead-lead plots of galena lead isotopes from mineral deposits on northern Texada Island. The slightly elongate trend to the data plots along the "plutonogenic" trend of Reddy (1989). This implies a close temporal and genetic relationship to the plutons of the Coast plutonic complex. Data, from Table 2-5-2, are represented as follows: dot = analyses from showings on Texada Island, circle = average of analyses from Texada Island, solid square = average of plutonogenic data from Reddy (1989), cross = average of Vancouver Island Jurassic mineralization, plus = average of Tertiary Catface intrusion related mineralization from Vancouver Island.

drawn on Figure 2-5-2 and is labelled "plutonogenic mixing line". Leitch (1989), and Leitch *et al.* (in press), evaluated a similar line that was obtained from analysis of galena from showings in the Bridge River (92J) and Black Dome Mountain gold camps (92O). They concluded that the line represented a mixture of mantle and upper crustal reservoirs (compare to: Doe and Zartman, 1979, and Godwin *et al.*, 1988). They also showed that mineralization represented by this line was generated by Cretaceous to Tertiary plutons. These plutons formed near to and young away from the eastern margin of the Coast plutonic complex. Galena lead from the younger deposits generally plots in the more radiogenic portions of the line.

The lead-lead isotopic fingerprint of the galena collected from northern Texada Island is shown in Figure 2-5-2. The average for galena from Texada Island matches almost exactly the average from Reddy's plutonogenic data. The data from Texada Island are also aligned along the plutonogenic mixing line. It is significant that the area studied by Reddy is adjacent to, but east of Texada Island.

DISCUSSION

The tight cluster of galena lead isotope data for the deposits analyzed on Texada Island indicates that all the deposits are closely related. They were generated from a similar source – all at about the same time.

The plutonogenic model, and data from the Harrison Lake–Whistler–Squamish–Sechelt areas, closely describe the data obtained from Texada Island. Because the means are indistinguishable statistically, the deposits are probably the same age as those studied by Reddy (1989). The Texada mineralization therefore is Jurassic to Cretaceous in age; because several granitic bodies on Texada Island have been dated as Jurassic, this date is preferred.

The galena lead isotope model for Vancouver Island by Andrew and Godwin (1989) does not apply directly to data presented here for Texada Island. However, this model confirms that the Texada Island galena lead: (1) is not related to the Paleozoic Sicker Group, because it has a $^{206}Pb/^{204}Pb$ ratio that is greater than 18.64, and (2) has $^{206}Pb/^{204}Pb$ ratios that

TABLE 2-5-2
GALENA LEAD ISOTOPE ANALYSES[1] FROM SHOWINGS, TEXADA ISLAND

Lab Number[2]	Deposit Name	Lat. N	Long. W	^{206}Pb/^{204}Pb(er%)	^{3207}Pb/^{204}Pb(er%)	^{208}Pb/^{204}Pb(er%)	^{207}Pb/^{206}Pb(er%)	^{208}Pb/^{206}Pb(er%)
30335-001	NUTCRACKER	49.72	124.55	18.704 (0.017)	15.573 (0.019)	38.294 (0.031)	0.832610 (0.006)	2.047403 (0.023)
30335-001	NUTCRACKER	49.72	124.55	18.722 (0.019)	15.596 (0.021)	38.376 (0.025)	0.833066 (0.005)	2.049843 (0.010)
30335-001	NUTCRACKER	49.72	124.55	[18.712 (0.018)]	[15.584 (0.020)]	[38.339 (0.028)]	[0.832859 (0.006)]	[2.049121 (0.017)]
30560-001	SHOWING	49.75	124.53	18.777 (0.041)	15.606 (0.041)	38.430 (0.047)	0.831114 (0.009)	2.046621 (0.021)
30560-001AD	SHOWING	49.75	124.53	18.763 (0.011)	15.587 (0.014)	38.371 (0.020)	0.830709 (0.006)	2.045029 (0.011)
30560-AVG2	SHOWING	49.75	124.53	[18.766 (0.026)]	[15.592 (0.026)]	[38.388 (0.034)]	[0.830868 (0.008)]	[2.045585 (0.016)]
30561-001	SHOWING	49.75	124.54	18.776 (0.017)	15.596 (0.019)	38.415 (0.024)	0.830637 (0.006)	2.045942 (0.013)
30563-001	ALADDIN	49.72	124.52	18.791 (0.057)	15.605 (0.039)	38.381 (0.071)	0.830447 (0.042)	2.042543 (0.041)
30563-001AD	ALADDIN	49.72	124.52	18.766 (0.018)	15.583 (0.020)	38.330 (0.023)	0.830425 (0.006)	2.042556 (0.009)
30563-AVG2	ALADDIN	49.72	124.52	[18.772 (0.038)]	[15.590 (0.030)]	[38.343 (0.047)]	[0.830428 (0.024)]	[2.042554 (0.025)]
30564-001	SENTINEL	49.72	124.53	18.748 (0.019)	15.580 (0.021)	38.351 (0.024)	0.831034 (0.005)	2.045651 (0.009)
30564-001AD	SENTINEL	49.72	124.53	18.749 (0.015)	15.582 (0.015)	38.341 (0.027)	0.831039 (0.011)	2.044897 (0.018)
30564-AVG2	SENTINEL	49.72	124.53	[18.749 (0.017)]	[15.581 (0.018)]	[38.346 (0.026)]	[0.831036 (0.008)]	[2.045403 (0.014)]
30565-001	SANDY	49.72	124.53	18.763 (0.023)	15.579 (0.024)	38.365 (0.030)	0.830299 (0.009)	2.044785 (0.014)
30565-001AD	SANDY	49.72	124.53	18.761 (0.014)	15.581 (0.017)	38.344 (0.021)	0.830537 (0.005)	2.043874 (0.010)
30565-AVG2	SANDY	49.72	122.53	[18.762 (0.019)]	[15.580 (0.021)]	[38.353 (0.026)]	0.830452 (0.007)]	[2.044254 (0.012)]
30566-001	SILVER TIP	49.73	124.59	18.650 (0.017)	15.571 (0.019)	38.276 (0.030)	0.834919 (0.007)	2.052344 (0.022)
30566-001AR	SILVER TIP	49.73	124.59	18.650 (0.018)	15.571 (0.020)	38.273 (0.023)	0.834936 (0.005)	2.052215 (0.009)
30566-AVG2	SILVER TIP	49.73	124.59	[18.650 (0.018)]	[15.571 (0.020)]	[38.274 (0.027)]	[0.834929 (0.006)]	[2.052253 (0.016)]
30569-001	VICTORIA	49.76	124.56	18.680 (0.013)	15.583 (0.016)	38.277 (0.021)	0.834248 (0.006)	2.049160 (0.010)
30571-001	HOLLY FAULT	49.73	124.56	18.676 (0.015)	15.564 (0.016)	38.261 (0.023)	0.833359 (0.010)	2.048645 (0.012)
30571-001AR	HOLLY FAULT	49.73	124.56	18.689 (0.019)	15.572 (0.019)	38.291 (0.029)	0.833231 (0.011)	2.048893 (0.018)
30571-001BD	HOLLY FAULT	49.73	124.56	18.690 (0.017)	15.597 (0.019)	38.307 (0.022)	0.833509 (0.006)	2.049574 (0.009)
30571-AVG3	HOLLY FAULT	49.73	124.56	[18.684 (0.017)]	[15.577 (0.018)]	[38.286 (0.025)]	[0.833397 (0.009)]	[2.049105 (0.013)]
30572-001	RETRIEVER	49.71	124.56	18.707 (0.018)	15.585 (0.020)	38.339 (0.024)	0.833119 (0.007)	2.049458 (0.009)
TEXADA: OVERALL AVERAGE[3] (N=10)		49.72	124.55	[18.720 (0.228)]	[15.583 (0.045)]	[38.334 (0.102)]	[0.832399 (0.194)]	[2.047553 (0.138)]
PLUTONOGENIC[4]: OVERALL AVERAGE[3] (N=10)		—.—	—.—	[18.71]	[15.58]	[38.29]	[0.8327]	[2.0465]
VANCOUVER ISLAND JURASSIC[4] OVERALL AVERAGE[3] (N=2)		—.—	—.—	[18.53]	[15.53]	[38.06]	[0.8381]	[2.0539]
VANCOUVER ISLAND TERTIARY[4] OVERALL AVERAGE[3] (N=10)		—.—	—.—	[18.99]	[15.60]	[38.56]	[0.8215]	[2.0305]

[1] All analyses have been normalized to the National Bureau of Standard sample NBS981 with accepted values (absolute error) of: ^{206}Pb/^{204}Pb = 16.004 (0.006); ^{207}Pb/^{204}Pb = 15.390 (0.007); ^{208}Pb/^{204}Pb = 35.651 (0.017), ^{207}Pb/^{206}Pb = 0.961635 (0.000567); ^{208}Pb/^{206}Pb = 2.227631 (0.001351).

[2] Suffixes on Laboratory Number: (1) A and B are additional analyses, (2) R is a repeated analysis from the specimen, (3) D is a duplicate analysis of the solution of a sample.

[3] Averages are weighted by analytical error; N = number of analyses in average. Errors quoted are expressed in per cent; er% = analytical error, except for "overall averages" where the error is one standard deviation of ratios used in average.

[4] Plutonogenic data are from Reddy (1989); Vancouver Island Jurassic data are from Godwin *et al.* (1988) for 30432 – Utluh Creek and 30699 – Island Copper; Vancouver Island Tertiary data are the average of the gold-quartz veins related to the Zeballos Tertiary intrusions (Godwin and Andrew, 1988).

are markedly lower than those from Tertiary galena deposits on Vancouver Island, which have ^{206}Pb/^{204}Pb ratios that are greater than 18.8. The galena lead isotope ratios for the Jurassic deposits on Vancouver Island (Table 2-5-2) plot close to the plutonogenic mixing line in Figure 2-5-2, but at a point that is less radiogenic. Tertiary galena lead isotopes from Vancouver Island plot along the more radiogenic extension of the plutonogenic mixing line at a point that is apparently younger. This position is analogous to the position of Tertiary lead on the Bralorne–Pioneer to Black Dome mixing line (Leitch, 1989; Leitch *et al.*, in press).

The spread of data along the mixing line is not related simply to the age of the deposits analyzed, although Tertiary deposits cluster near the more radiogenic end of the line.

Variations in the values for Jurassic and Cretaceous lead isotopes apparently reflect the relative abundance of upper crustal component involved in generation of the plutons that generated the mineralization. Consequently, different metallogenic characteristics might be reflected by the variations in lead isotopes. This is under investigation.

CONCLUSIONS

The galena lead isotope compositions from vein showings on Texada Island show that they all originated at the same time, from a common source. The deposits are plutonogenic: they are related genetically to intrusions of the Coast plutonic complex. The age of the mineralization is probably Jura-Cretaceous Tertiary ages are unlikely.

ACKNOWLEDGMENTS

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GEOLOGY OF THE GOOD HOPE – FRENCH MINE AREA, SOUTH-CENTRAL BRITISH COLUMBIA* (92H/8)

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Corona Corporation

KEYWORDS: Economic geology, Hedley, gold skarn, French mine, Good Hope mine, ore controls, mineralogy.

INTRODUCTION

The Good Hope and French mines, in the Hedley mining camp, are gold skarn deposits hosted by French Mine formation limestones of the Upper Triassic Nicola Group. They are located approximately 5 kilometres east of the town of Hedley in south-central British Columbia, about 5 kilometres south of the Nickel Plate open-pit mine. Access to the property is by gravel road off the Nickel Plate mine road. The property is owned by Golden North Resource Corporation and is being explored by Corona Corporation.

This report incorporates regional mapping by Ray and Dawson (1987, 1988), and detailed mapping by Hammack (1988) and Dawson in this study. This work forms the basis of an M.Sc. thesis by Dawson at The University of British Columbia.

REGIONAL GEOLOGY

The Hedley gold camp lies within Quesnellia in the Intermontane Belt of the Canadian Cordillera. The first regional mapping of the area by Bostock (1930, 1940a, b) has recently been updated by Ray and Dawson (1987, 1988).

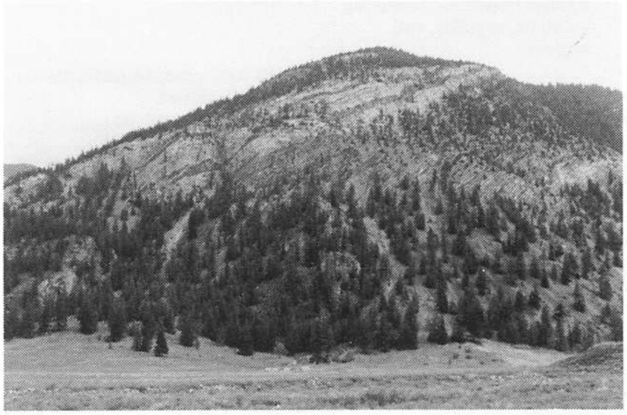

Plate 2-6-1. Possible sediment-sill complex in the Hedley area consisting of Hedley diorite sills intruding Late Triassic Hedley formation limestone and siltstone. Photo, half a kilometre east of Hedley township, was taken looking north from Highway 3.

The Table of Formations (Table 2-6-1) outlines the evolution of nomenclature in the camp.

A sill swarm exposed on the cliffs east of the township of Hedley (Plate 2-6-1) is one of the most visually striking features of the area. Sills of hornblende-porphyritic diorite are part of the Hedley intrusive suite believed to be Late Triassic or Early Jurassic (199 Ma based on U-Pb dates of zircons from the Banbury stock, 3.5 kilometres west of Hedley). In this location the sills, which vary from 1 to 25 metres in thickness, make up almost 50 per cent of the stratigraphic column where they intrude laminated limestone and siltstone of the Hedley formation. They occur both adjacent to the Toronto quartz diorite to gabbro stock and as far away as 2 kilometres from it. Auriferous skarn mineralization is spatially and genetically associated with the stock and the adjacent diorite sills (Billingsley and Hume, 1941; Dolmage and Brown, 1945; Ray *et al.*, 1986, 1987, 1988). A similar sill swarm is developed within the French Mine formation at the French mine and a single sill is associated with mineralization at the Good Hope mine.

LOCAL GEOLOGY

The Good Hope–French mine area is underlain by sedimentary and volcanic rocks of the Late Triassic Nicola Group and the Middle to Late Paleozoic and Triassic Apex Mountain complex (Figure 2-6-1). The Apex Mountain complex, a deformed ophiolite package, consists of greenstone, chert, argillite, siltstone and minor limestone (Milford, 1984).

Structure within the Good Hope–French mine area is relatively simple with units generally striking to the north-northeast and dipping gently west (Figure 2-6-1). Major faults include the Cahill Creek fracture zone and the Good Hope fault that were important in controlling intrusion of the Cahill Creek pluton. Major folds have not been identified, however, Hammack (1988) mapped numerous small northwest and northeast-trending small-scale flexures.

The Nicola Group has been informally subdivided into three stratigraphically distinct formations within the Good Hope–French mine area (Ray *et al.*, 1987, 1988): a lower volcanic package called the Peachland Creek formation, a middle carbonate package called the French Mine formation, and an upper volcanic package called the Whistle Creek formation. The contact between the Nicola Group and the

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

TABLE 2-6-1. TABLE OF FORMATIONS, HEDLEY AREA, SOUTH-CENTRAL BRITISH COLUMBIA.

Age	Camsell, C. (1910)	Bostock, H.S. (1930)	Bostock, H.S. (1940)	Ray and Dawson (1988 and in press)
Tertiary	aplite, rhyolite & andesite dikes, granodiorite		basalt flows, pyroclastics	Unit 13: basaltic flows
			conglomerate sandstone	Unit 12: conglomerate, sandstone
		——— Unconformity ———		
Early Cretaceous				**Spences Bridge Group** Unit 11: andesite – rhyolite pyroclastics, minor sedimentary rocks
			——— Intrusive contact ———	
				Unit 10: Verde Creek stock; granite to microgranite
			——— Intrusive contact ———	
Middle Jurassic				Unit 9: Ashnola Hill formation andesite – dacite pyroclastic rocks
			——— Intrusive contact ———	
				Unit 8: Cahill Creek pluton (168 Ma); granodiorite – quartz monzodiorite
				Unit 7: quartz-feldspar rhyolite porphyry dike (171 Ma)
Early Jurassic		granite	granite	Unit 6: Bromley batholith (198 Ma)
		granodiorite	granodiorite	granodiorite
Late Triassic(?) – Early Jurassic		diorite, gabbro	diorite, gabbro	Unit 5: Hedley intrusive suite; (199 Ma); phyric and aphyric quartz diorite – gabbro
		——— Intrusive contact ———		
Late Triassic		**Nicola Group** Unnamed section; volcanic & sedimentary rocks	**Nicola Group** Wolfe Creek formation; andesite – basalt tuff, minor sedimentary rocks	**Nicola Group** Unit 4: Whistle Creek formation; andesite ash & lapilli tuff, minor siltstone
				Unit 3d: Copperfield breccia, limestone breccia
		Aberdeen formation; quartzite limestone, argillite	Henry formation; argillite tuff, impure limestone	Unit 3c: Stemwinder Mountain formation (western facies); argillite, limestone
		Red Mountain formation; tuffs & breccias	Hedley formation; limestone, quartzites, argillite conglomerate, breccia, tuff	Unit 3b: Hedley formation (central facies); siltstone, limestone
		Nickel Plate formation; limestone & quartzite	Sunnyside formation; limestone	
		Redtops formation; limestone, quartzite, argillite, tuff, breccia, limestone	Redtop formation; limestone quartzite, argillite, tuff, breccia	
				Unit 3a: French Mine formation (eastern facies); limestone, limestone conglomerate
				Unit 2: Peachland Creek formation; basalt tuffs and flows, argillite, chert pebble conglomerate, limestone olistostrome
	——— Contact occupied by the Cahill Creek pluton ———			
Middle to Late Paleozoic and Triassic	**Cache Creek Group** Aberdeen formation; limestone, quartzite, argillite, tuff, volcanic breccia		Independence formation; chert, argillite, basalt – andesite flows, breccia	Unit 1: Apex Mountain complex; argillite, greenstone, limestone, chert
	Red Mountain formation; tuff, volcanic breccia, quartzite, argillite		Bradshaw formation; argillite, tuff, quartzite, breccia, andesite, limestone	
	Nickel Plate formation; limestone, quartzite, argillite, tuff			
	Redtop formation; limestone, quartzite, argillite, tuff, breccia			

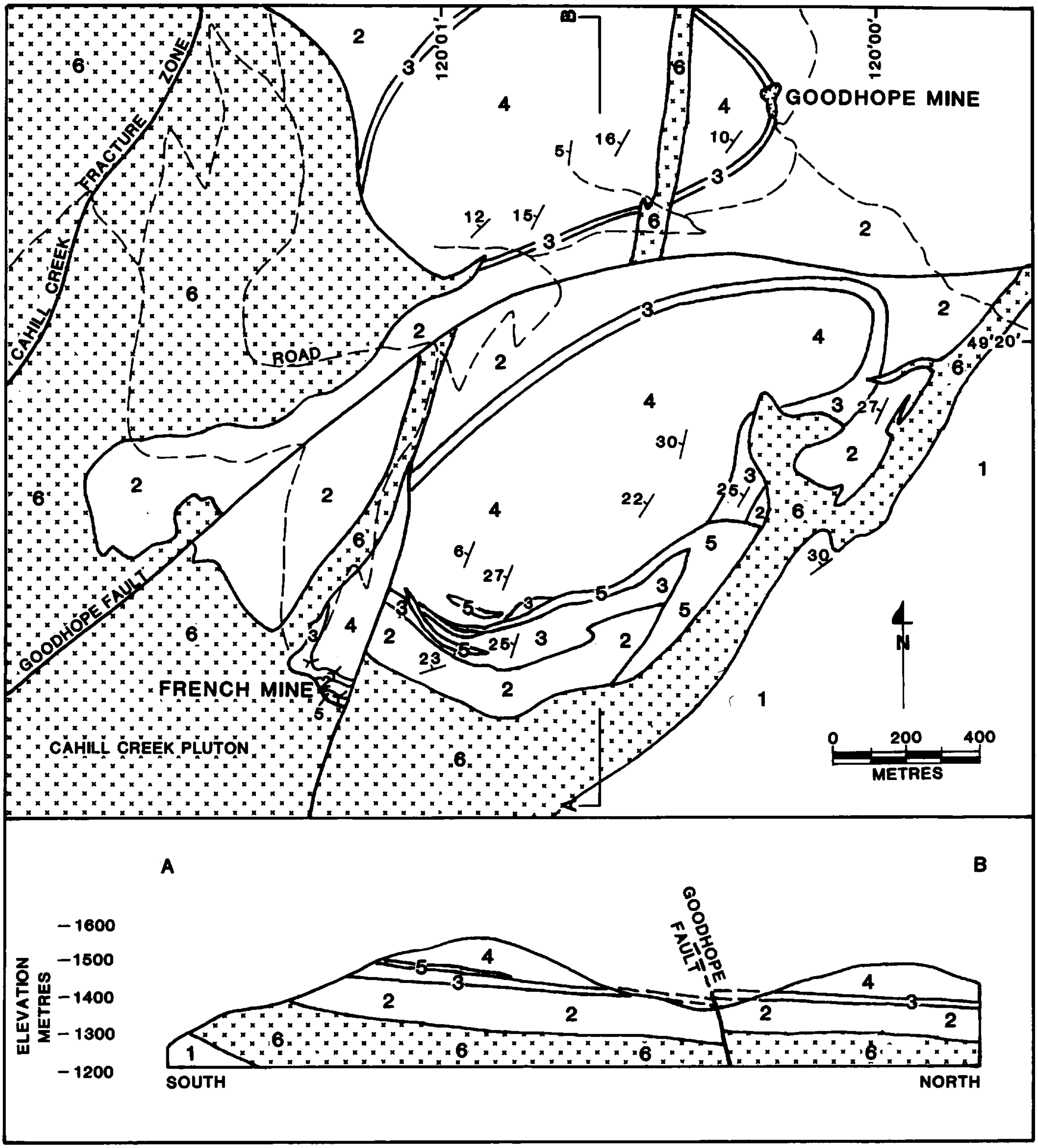

Figure 2-6-1. Geology of the Good Hope to French mine area, south-central British Columbia. Units on the figure, from oldest to youngest, are: 1 = Apex Mountain complex, 2 = Peachland Creek formation, 3 = French Mine formation, 4 = Whistle Creek formation, 5 = Hedley intrusive suite, 6 = Cahill Creek pluton, 7 = Quartz-feldspar rhyolite dike.

Apex Mountain complex is occupied by the Cahill Creek pluton. Consequently it is unknown whether the original contact was an unconformity or a suture.

The Peachland Creek formation comprises the oldest Nicola Group rocks identified in the study area. It is correlated with, and named after, a volcaniclastic sequence in the Pennask Mountain area approximately 30 kilometres west of Peachland (Dawson and Ray, 1988).

Massive to poorly bedded, andesitic to basaltic tuffs and volcanic flows with minor argillite and limestone comprise most of the sequence. The tuffs often contain sparse chert and recrystallized quartz grains. Rare, thin chert-pebble conglomerate beds may represent turbidite deposits derived from the Paleozoic Apex Mountain complex farther east. Algal-rich marble blocks, up to 5 metres in diameter and occurring throughout the sequence, are interpreted as olistostromes that were derived from carbonate reefs to the east. Bedding underlying the olistostrome is locally disrupted. Spherical argillaceous carbonate mud balls or oncolites, up to 2 centimetres in diameter, are found locally, indicating a shallow depositional environment. The base of the Peachland Creek formation is not exposed in the map area but the unit is at least 400 metres thick.

The French Mine formation stratigraphically overlies the Peachland Creek formation and consists of massive to poorly bedded limestone interlayered with limestone pebble to boulder conglomerate and minor limestone breccia. It has a maximum thickness of 100 metres and tapers westward towards the Cahill Creek fracture zone (Figure 2-6-1).

The limestone pebbles and cobbles in the conglomerates make up 95 per cent of the clasts; they are 5 to 50 centimetres in diameter, subangular to subrounded, and both clast and matrix supported. Rare clasts of tuff, argillite and aphyric mafite(?) occur within this unit. The limestone breccia clasts are angular, generally less than 5 centimetres in diameter, and are clast supported. The matrix of the limestone conglomerate and breccia is altered to massive garnet or garnet-diopside reaction skarn, which reflects the variable composition of the matrix and its high porosity. The limestone is invariably recrystallized to marble. This unit probably represents a shallow-water shelf environment of fore-reef or lagoonal facies. It hosts the gold skarn mineralization at the Good Hope and French mines.

Whistle Creek formation stratigraphically overlies the French Mine formation and consists primarily of laminated to massive tuffaceous siltstone and andesite tuff. The lower part of this unit is markedly epiclastic and often exhibits graded beds, flame textures and load casts. These features indicate that the unit is right-way-up and that paleocurrent directions are predominantly from the east. The section grades upwards into more thickly bedded to massive ash, lapilli and tuff breccia that is arc-related and includes both alkalic and subalkalic rocks of andesitic to basaltic composition (Ray and Dawson, in preparation). Biotite + pyroxene + potassium feldspar hornfels is common in the lower sedimentary section of the unit, probably because it is close to Cahill Creek intrusive rocks; higher permeability of the bedded units and the chemical gradients between individual laminations or beds often enhance alteration. The maximum thickness of this unit in the district is 1200 metres (Ray and Dawson, in preparation), but near the Good Hope and French mines it is about 200 metres thick.

Hedley intrusive rocks in the study area form both phyric and aphyric sills, dikes and stocks throughout the Nicola Group rocks, but are absent from the Apex Mountain complex. A U-Pb isotopic age of 199 Ma (Ray and Dawson, in preparation) from the Banbury stock indicates that they are Early Jurassic in age, however contact relationships within the study area suggest they may be as old as Late Triassic.

The phyric Hedley intrusive rocks are commonly calc-alkaline and dioritic in composition (Ray *et al.*, 1988). In hand sample they consist of medium-grained inequigranular feldspar-hornblende diorite and coarse-grained hornblende-porphyritic diorite. In thin section, the hornblende phenocrysts and matrix commonly contain very fine grained felted biotite and oscillatory zoned plagioclase.

Aphyric Hedley intrusive rocks are massive, dark brown to black, biotitic, aphanitic and sulphide rich. They generally occur as small sill and dike-like bodies or as margins to the larger phyric Hedley intrusions. They are interpreted (by the first two authors) as a quenched phase of the Hedley intrusive suite. Peperite-like textures (Plate 2-6-2; *cf.* Busby-Spera and White, 1987; Kokelaar, 1982) developed along some contacts suggest intrusion into wet sediment. The authors do not concur with this interpretation of the origin for these aphyric, biotite-rich rocks. Further detailed work is planned to investigate their origin and to differentiate them from non-bedded mafic tuffs and argillites in the area.

The Cahill Creek pluton consists of medium-grained biotite-hornblende granodiorite to monzodiorite of calc-alkaline composition (Ray *et al.*, 1988). It is the next youngest intrusive suite in the Good Hope–French mine area, and forms a large body with minor apophyses controlled in part by the Cahill Creek fracture zone. Uranium-lead isotopic dates from zircons give a mid-Jurassic age of 168 Ma (Ray and Dawson, in preparation). Minor late aplitic dikes occur both in the pluton and adjacent to it.

Quartz-feldspar rhyolite, the youngest intrusive rock identified in the study area, forms a dike less than 3 metres wide

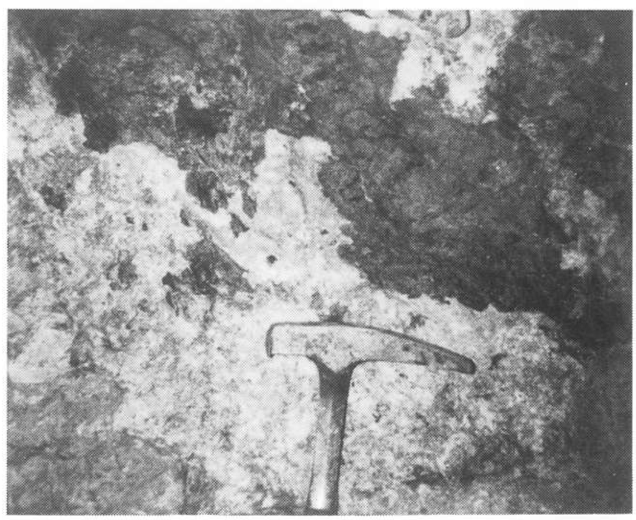

Plate 2-6-2. Possible globular peperite developed along the contact of aphyric mafite(?) Hedley sill and massive limestone of the French mine formation.

that cuts mineralization at the French mine. A similar intrusive unit, 11 kilometres southwest of Hedley, returned a U-Pb isotopic zircon age of Middle Jurassic (171 Ma, Ray and Dawson, in preparation). This dike suite may be a feeder system to previously unrecognized mid-Jurassic volcaniclastic rocks on Ashnola Hill 10 kilometres southwest of the project area, and on Lookout Mountain 7 kilometres to the north. These rocks were originally mapped as Nicola Group by Bostock (1940a) and as Early Cretaceous Spences Bridge Group by Ray and Dawson (1987, 1988).

ALTERATION AND MINERALIZATION

The Good Hope mine (MINFILE 92H 060) has produced 178 kilograms of gold, 120 kilograms of silver and 602 kilograms of copper from 11 410 tonnes of ore mined during the period 1946 to 1948 and in 1982. Production was from gold-enriched skarn developed along the contact between the French Mine formation and a Hedley diorite sill (Figure 2-6-2). In general the bedding in the area is gently dipping, but a broad synclinal structure is exposed within the trench area.

The diorite sill is approximately 2 metres thick and is composed of feldspar and hornblende crystals, less than 3 millimetres in diameter, set in a fine-grained matrix. The hornblende crystals and matrix contain fine-grained felted biotite with minor diopside occurring along fractures. Skarn

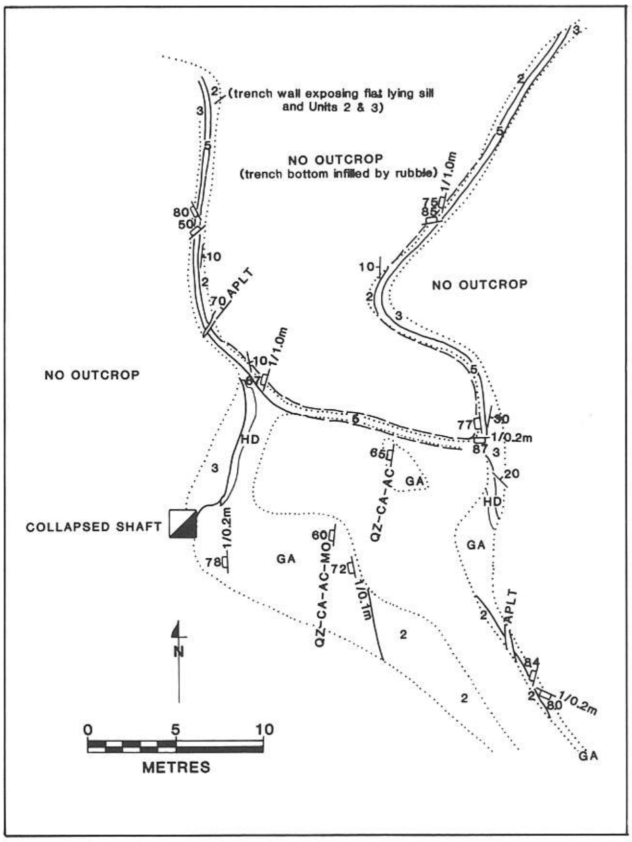

Figure 2-6-2. Sketch map of the Good Hope trench (*see* Figure 2-6-1 for legend), additional abbreviations are: APLT = aplite, GA = garnet, HD = hedenbergite, QZ = quartz, CA = calcite, AC = actinolite, MO = molybdenite, SC = scapolite).

is best developed in the hangingwall of the sill. A distinct mineralogical zonation is recognized from the sill contact upwards into the overlying marble: this consists of a massive garnetite zone adjacent to the sill and up to 2 metres thick; followed by a discontinuous zone, less than 0.3 metre thick, of large tabular hedenbergite crystals. The garnet crystals are reddish brown to black, subhedral to euhedral, less than 1 centimetre in diameter, anisotropic and exhibit sector twinning. Microprobe analysis of selected garnet grains show they are Ad_{10} to Ad_{30} mole per cent and enriched in manganese (11%) compared to other skarns in the Hedley area (Ettlinger and Ray, 1989). Tabular euhedral hedenbergite crystals, up to 10 centimetres long, are oriented perpendicular to the marble contact. Microprobe analysis show they range from Hd_{90} to Hd_{100} and are also enriched in manganese (10%).

Minor retrograde skarn consisting of calcite, epidote and sulphides occurs interstitial to the hedenbergite crystals. Sulphides consist of finely disseminated and massive pyrrhotite, arsenopyrite, pyrite, marcasite and chalcopyrite, with minor native bismuth and hedleyite. Grab samples from the hedenbergite-sulphide skarn assayed up to 94 ppm gold (Ettlinger and Ray, 1989). Local zones of jasperoid are developed along the upper contact of the sill with the marble.

A second period of mineralization crosscuts the auriferous skarn mineralization and consists of north-striking quartz + actinolite + epidote + calcite ± molybdenite ± scheelite veins bordering the aplitic dikes of the Cahill Creek pluton.

The French mine (MINFILE 92H 059) produced 1615 kilograms of gold and 124 kilograms of silver from 79 000 tonnes of ore during the periods 1950 to 1955, 1957 to 1961, and in 1983. Mineralization is confined to a broad anticlinal structure within a down-faulted block of the French Mine formation (Figure 2-6-1). Within the area of the mine workings, the French Mine formation consists dominantly of massive limestone with some limestone conglomerate and breccia layers present in the western end of the workings. The anticlinal structure strikes west to northwest and has been worked along two main stopes over a horizontal distance of 225 metres (Figure 2-6-3). Mineralization is terminated against the high-angle French fault on the west and the west-dipping Cariboo thrust fault on the east. Other northeast and northwest-striking high-angle faults have been identified underground, however displacements are generally less than 3 metres. The stopes are about 3 metres wide and are believed to be separated by biotite-rich aphyric mafite sills of Hedley diorite (the authors do not concur – the separating unit may be hornfelsed tuffs and argillites).

A distinct skarn mineralogical zonation is developed outwards from aphyric mafite sills and dikes. Zones consist of successive envelopes of: scapolite + potassium feldspar + quartz, followed by garnet + diopside, followed by massive marble. The scapolite + potassium feldspar + quartz envelope is up to 50 centimeters thick. The garnet + diopside envelopes are up to 1 metre thick and are composed of massive, fine-grained reddish brown isotropic garnet with minor diopside. Microprobe analysis of garnets within the ore zone shows that they are enriched in iron and range in composition from Ad_{80} to Ad_{100} mole per cent; garnets from the outer margin of the skarn envelope range from Ad_{13} to

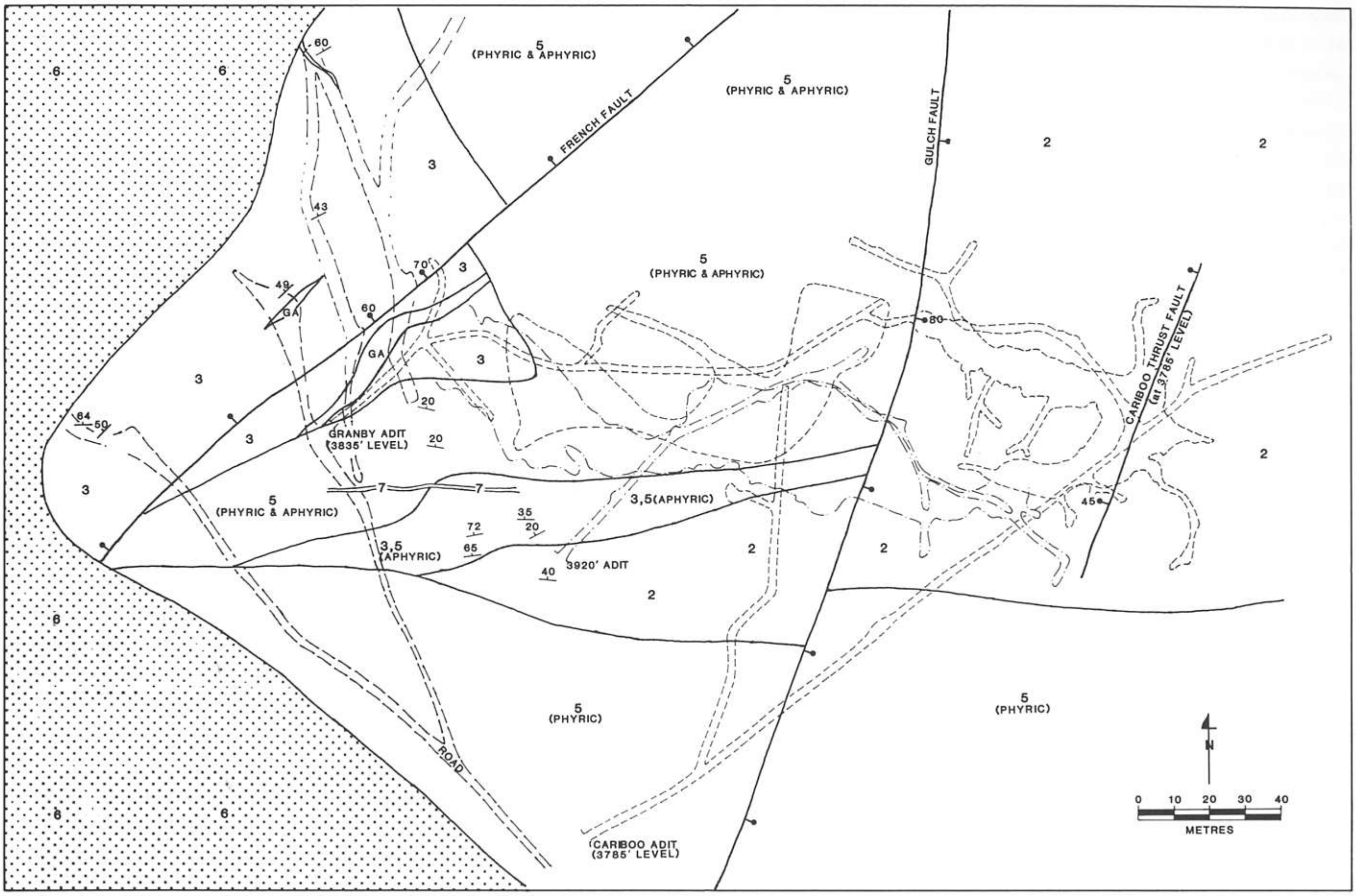

Figure 2-6-3. Sketch map of the French mine area (*see* Figure 2-6-1 for legend and Figure 2-6-2 for abbreviations; Note: short dash = lower stope and haulage level, and long dash and dot = upper stope and haulage level).

Ad_{25}. Pyroxene crystals range from Hd_{63} to Hd_{67} and have a low (less than 1%) manganese content. Associated skarn minerals include minor epidote, wollastonite and sulphides.

Sulphides average less than 5 per cent by volume throughout most of the deposit, except for the western part that was relatively rich in copper and low in gold. The major sulphides identified are pyrrhotite, chalcopyrite, bornite, covellite, pyrite and arsenopyrite. Minor cobaltite, erythrite, tellurides and native gold have been identified. In the lower stopes visible gold is associated with coarse telluride grains. Recent underground chip sampling by Corona Corporation has outlined zones of high grade gold mineralization over a strike length of 65 metres with several samples returning values over 35 grams per tonne gold over widths of 1 metre (Godfrey, 1989). Down-dip extentions of the ore horizons and the diplaced horizons underneath the Cariboo thrust are currently being tested by drilling.

Sporadic coarse scheelite and molybdenite are also reported. A 35-metre chip sample along an underground face averaged 0.68 per cent WO_3 (Ray *et al.*, 1988). The relationship of this mineralization to the major gold-bearing skarns remains uncertain, but it may be related to the underlying Cahill Creek pluton.

SUMMARY

The Good Hope–French mine area is underlain by the Upper Triassic Nicola Group consisting of the lower volcanic Peachland Creek formation, the middle carbonate-dominant French Mine formation and the upper volcaniclastic Whistle Creek formation. Calcsilicate reaction skarn, widely developed throughout the French Mine formation, may have been formed by the Hedley intrusive suite, the younger Cahill Creek granodiorite, or both. However, auriferous skarn mineralization at the Good Hope and French mines is genetically and spatially related to the Hedley intrusive suite. A second period of mineralization consisting of quartz + actinolite + calcite + molybdenite + scheelite veins crosscuts earlier auriferous skarn mineralization and may be associated with the aplitic phase of the Cahill Creek pluton.

At the Good Hope mine, auriferous skarn is best developed along the upper contact of a feldspar-hornblende-phyric Hedley diorite sill. Successive envelopes of garnet, diopside and hedenbergite skarn are developed outwards into the overlying marble of the French Mine formation. Sulphides and associated gold mineralization are concentrated in the coarse-grained hedenbergite envelope; this suggests iron-rich hydrothermal fluids were important in transporting gold. Jasperoid developed along the sill-marble contact and along pre-intrusion faults might be a late feature of the skarning process (*i.e.* fluids were not hot enough to produce calc-silicate mineralogy).

At the French mine, scapolite, garnet-diopside and marble envelopes are developed adjacent to numerous small Hedley aphyric mafite sills and dikes which have intruded limestone of the French Mine formation. Mineralogical zoning suggests hydrothermal fluids were confined to areas between individual sills and dikes resulting in multiple "box-like" zones of skarn alteration. Minor calcite + quartz + chlorite + sulphides, and associated gold mineralization, are found predominantly within the garnet-diopside skarn.

The recognition of possible aphyric mafite intrusions as a quenched mineralized phase of the Hedley intrusive suite, formed by intrusion into wet sediment, has important genetic and economic significance in gold skarn models. Some implications are: contemporaneous sedimentation and intrusive volcanism; shallow depth of intrusion and associated skarn formation; availability of large quantities of seawater that might facilitate chlorine complexing and transportation of metals; and depositional environment within an extentional regime, perhaps related to rifting in a back-arc basin.

Distinguishing barren calcsilicate reaction skarn from economic auriferous skarn mineralization is difficult. However, the presence of iron-rich prograde mineral assemblages such as andraditic garnet and hedenbergite pyroxene and retrograde minerals such as epidote, calcite, quartz, amphibole, chlorite and sulphides may indicate that the skarn is not isochemical and therefore has auriferous potential.

The amount of alteration and mineralization developed appears to be proportional to the number of sills present. The presence of only one sill at the Good Hope mine may explain its small size as compared to the Nickel Plate and French mines where sill swarms are more extensively developed.

ACKNOWLEDGMENTS

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NOTES

STOCKWORK MOLYBDENITE IN THE MISSION RIDGE PLUTON: A NEW EXPLORATION TARGET IN THE BRIDGE RIVER MINING CAMP* (92J/16)

By Robert G. Gaba

KEYWORDS: Economic geology, Bridge River, Cub claims, Mission Ridge pluton, Rexmount porphyry, mylonite, molybdenum, copper, gold, stockwork, quartz veins.

INTRODUCTION

The Mission Ridge pluton is a northwest-trending body of granodiorite 30 kilometres long that outcrops along Mission Ridge and the southern Shulaps Range approximately 200 kilometres north of Vancouver. The area is within the region mapped during the summer of 1989, the final year of the MDA-funded Taseko–Bridge River project (Schiarizza *et al.*, 1990, this volume). During the course of this work the author and D.A. Archibald of Queen's University discovered considerable exposures of silicic granodiorite containing disseminated molybdenite. The area was mapped in some detail and several dozen rock samples were collected for base and precious metal analysis. This report summarizes the results of the field study and addresses the nature and distribution of metals within the Mission Ridge granodiorite, specifically on the Cub 200 and adjacent mineral claims.

The area of interest (and much of the adjacent property) is covered by mineral claims owned or under option to MacNeill International Industries Inc. (formerly MacNeill Industrial Inc.) of Vancouver. The Spokane prospect (held 50 per cent by MacNeill International Industries Inc., 25 per cent by Enexco International Ltd. and 25 per cent by Julia Resources Corporation), which is approximately 3 kilometres to the west, had been the focus of MacNeill's attention and was being trenched and diamond drilled under the direction of Wright Engineering Ltd. at the time of our discovery. Our findings prompted MacNeill to examine the area of molybdenum mineralization. Surface samples collected by the ministry and by Wright Engineering Ltd. yielded significant gold in addition to molybdenum and copper; at the time of writing this report (late October) a $250000 diamond-drilling program was initiated to test the nature and distribution of the metals on the Cub 200 mineral claim (George Cross News Letter, Issue No. 201, 1989).

GEOLOGY OF THE SOUTHEAST PART OF THE SHULAPS RANGE

GENERAL STATEMENT

The regional geology of the area is described by Schiarizza *et al.* (1990, this volume). The southeast parr of the Shulaps Range is dominated by schists and phyllites of the Bridge River complex intruded by syn to post-tectonic granitic to felsic porphyry intrusions. These are structurally overlain by the Shulaps ophiolite complex, and near the Spokane prospect are imbricated with a belt of Shulaps-related serpentinite mélange and penetratively deformed metasedimentary rocks of the Cadwallader Group (Figure 2-7-1).

Granodiorite of the Mission Ridge pluton occupies the central part of the southern Shulaps Range Quartz feldspar porphyry, known as the Rexmount porphyry, is in contact with the northwest part of the Mission Ridge pluton and forms an irregular extension that continues northwest past Rex Peak to the Shulaps ultramafic complex.

MISSION RIDGE PLUTON

The Mission Ridge pluton consists of generally coarse-grained biotite granodiorite to quartz diorite. These rocks are usually massive within the interior of the pluton, although there is rare compositional layering, whereas marginal phases commonly have a slight to pervasive foliation that parallels contacts with country rocks. Pervasively foliated rocks are protomylonites defined by ribboned quartz paralleled by clots of biotite and chlorite and augened plagioclase grains and granodiorite clasts. Late aplite dikes are common and crosscut fabric trends.

Biotite from granodiorite of the Mission Ridge pluton has yielded a K-Ar age of 44 Ma (Woodsworth, 1977). Zircon and monazite have yielded U-Pb ages of 47.5 Ma for the Mission Ridge granodiorite and 46.5 Ma for deformed dikes of similar composition (M. Coleman, personal communication, 1989). The northern end of the pluton crosscuts north-dipping thrust faults that may be related to emplacement of the Shulaps ophiolite complex. Foliation and lineation along the margin of the pluton and within deformed dikes of similar composition are, however, related to the final stages of ductile deformation within the enclosing Bridge River schists, which may be related to Eocene dextral strike-slip faulting (Schiarizza *et al.*, 1990, this volume).

REXMOUNT PORPHYRY

The Rexmount porphyry consists of megascopic phenocrysts in a light grey to greenish aphanitic quartz-feldspar groundmass. Plagioclase, commonly a few millimetres in size, is the most abundant phenocryst and is generally partly altered to sericite and clay minerals. Smaller bipyramidal quartz and chlorite pseudomorphs after biotite and hornblende are present in variable proportions. Flow struc-

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

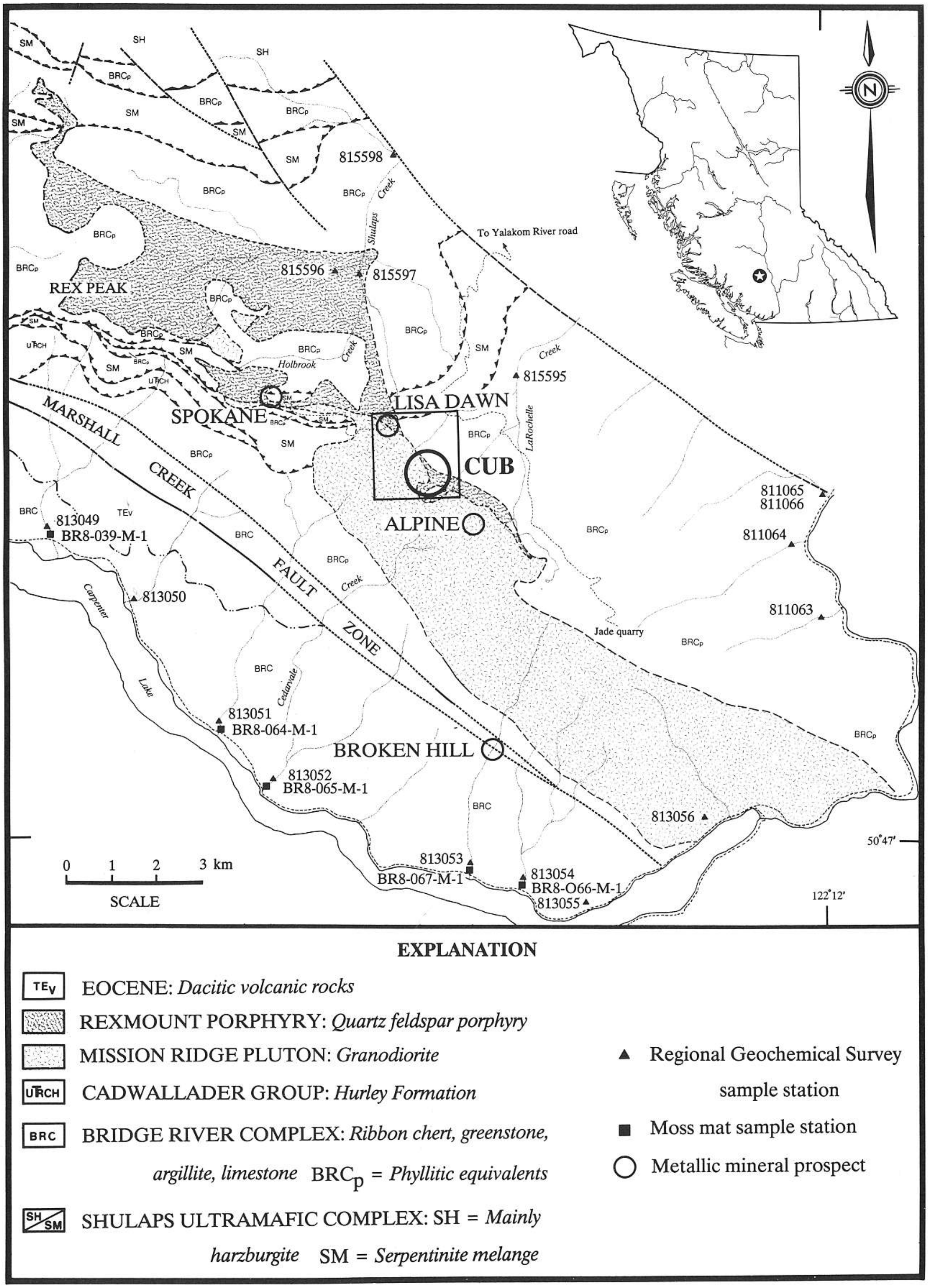

Figure 2-7-1. Geological setting and metallic mineral prospects of the Mission Ridge pluton. Area within square is location of Figure 2-7-2.

ture, defined by the alignment of phenocrysts, and columnar jointing are rare features in marginal phases.

The lower contact of the northwest part of the Rexmount porphyry with underlying phyllites of the Bridge River complex dips gently to the northeast and is exposed along the western slopes of Rex Peak and the prominent unnamed peak to the northwest. The underlying rocks, also cut by porphyry dikes, record no contact metamorphic effects related to the Rexmount porphyry. The northwesternmost lobe of the porphyry partly truncates serpentinite mélange along the southern margin of the Shulaps ultramafic complex.

The Rexmount porphyry is in contact with the Mission Ridge pluton at the headwaters of Holbrook and LaRochelle creeks (Figure 2-7-1). The contact is essentially a complex zone of granodiorite interfingered with and crosscut by abundant Rexmount porphyry dikes; some dikes contain small xenoliths of foliated granodiorite. Breccia, composed almost exclusively of angular fragments of Rexmount porphyry, occurs as irregular to elongate zones at some porphyry-granodiorite contacts.

Attempts to determine the age of the Rexmount porphyry by K-Ar methods have not been successful. Field relationships indicate porphyry emplacement was post-fabric development in granodiorite.

ECONOMIC GEOLOGY

Metallic mineral prospects in the southern Shulaps Range are within or adjacent to the Mission Ridge pluton (Figure 2-7-1) and spatially associated with dikes of Rexmount porphyry. The Cub and Lisa Dawn molybdenite prospects are combinations of veins, stockworks and disseminations within granodiorite adjacent to porphyry dikes. Other previously known prospects include the Spokane, Broken Hill and Alpine (Table 2-7-1).

The Spokane prospect comprises gold, silver and copper-bearing quartz veins within granodiorite and adjacent country rocks. The distribution of gold closely follows that of copper, and is commonly accompanied by anomalous bismuth and tungsten (Table 2-7-1). Host granodiorite is generally foliated and is crosscut by relatively fresh Rexmount porphyry dikes. Some quartz veins are also crosscut by porphyry; this indicates that vein formation predated the porphyry.

The Broken Hill prospect consists of a zone of polymetallic veinlets and disseminations in silicic sedimentary rocks. The metal concentrations are peripheral to dikes of granodiorite and porphyry, suggesting a possible genetic relationship.

GEOCHEMICAL SURVEYS

The Alpine molybdenite vein-prospect was first noted by Pollock (1983) in a report on a geological and geochemical survey carried out in the Holbrook-LaRochelle creeks area for Utah Mines Ltd. The results of the exploration program were generally discouraging and no areas of anomalous base or precious metals were found.

Stream sediment from the southern Shulaps Range was collected and analyzed as part of the Canada/British Columbia Regional Geochemical Survey of the Pemberton (92J) map area. No notably anomalous metal concentrations were obtained, except for the sample from Cedarvale Creek which yielded 10 ppm molybdenum (sample 813052, Table 2-7-2; Figure 2-7-1). This was verified when moss-mat sediments collected from the same location yielded 11 ppm molybdenum (sample BR8-065-M-1, Table 2-7-3; Church, 1989). Cedarvale Creek drains approximately 20 square kilometres of steep terrain along the southwest side of the Mission Ridge pluton: its northernmost headwaters are less than a kilometre

TABLE 2-7-1
METALLIC MINERAL PROSPECTS WITHIN OR ADJACENT TO THE MISSION RIDGE PLUTON
(SEE FIGURE 2-7-1 FOR LOCATIONS)

Name	Description	Approximate Dimensions Length (m)	Width (m)	Comments/References
Cub	mo as fine disseminations in brecciated and silicic gd and as flakes within and adjacent to qtz veins (± py) in variably silicic gd; vein qtz (as talus blocks up to 1.5 m thick) contains mo as stylolites and irregular concentrations, and cpy as blebs and veinlets	650	120	Dimensions indicate area where mo was observed on surface; anomalous Cu and Au (with Bi and Pb)
Lisa Dawn	mo as stylolites within qtz vein at contact between gd and qfp; mo as disseminated flakes in adjacent gd	30	1.5	Dimensions of exposed vein; extent of disseminated mo not known
Alpine	mo within qtz vein in extremely fractured, limonitic gd; gd contains cpy, mal and az	50	2.5	Pollock, 1983; weakly anomalous Au and Ag
Spokane	cpy, mal, az, po and py in massive to partly ribboned, vuggy qtz (? c/c) vein; wallrocks are gd and qfp: gd adjacent to vein is foliated, whereas qfp appears to be fresh	700	2	Appreciable Au, Ag, Cu, Bi and W content MINFILE 092JNE034
Broken Hill	py, gln, sp, cpy and mal occur as disseminations and narrow lenses and veinlets within silicic Bridge River complex sedimentary rocks adjacent to dikes of gd and qfp	500	18	Appreciable Ag, Au, Cu, Pb and Zn content MINFILE 092JNE087

Abbreviations: az = azurite, clc = calcite, cpy = chalcopyrite, gln = galena, mal = malachite, mo = molybdenite, po = pyrrhotite, py = pyrite, qtz = quartz, sp = sphalerite; gd = granodiorite, qfp = quartz feldspar porphyry.

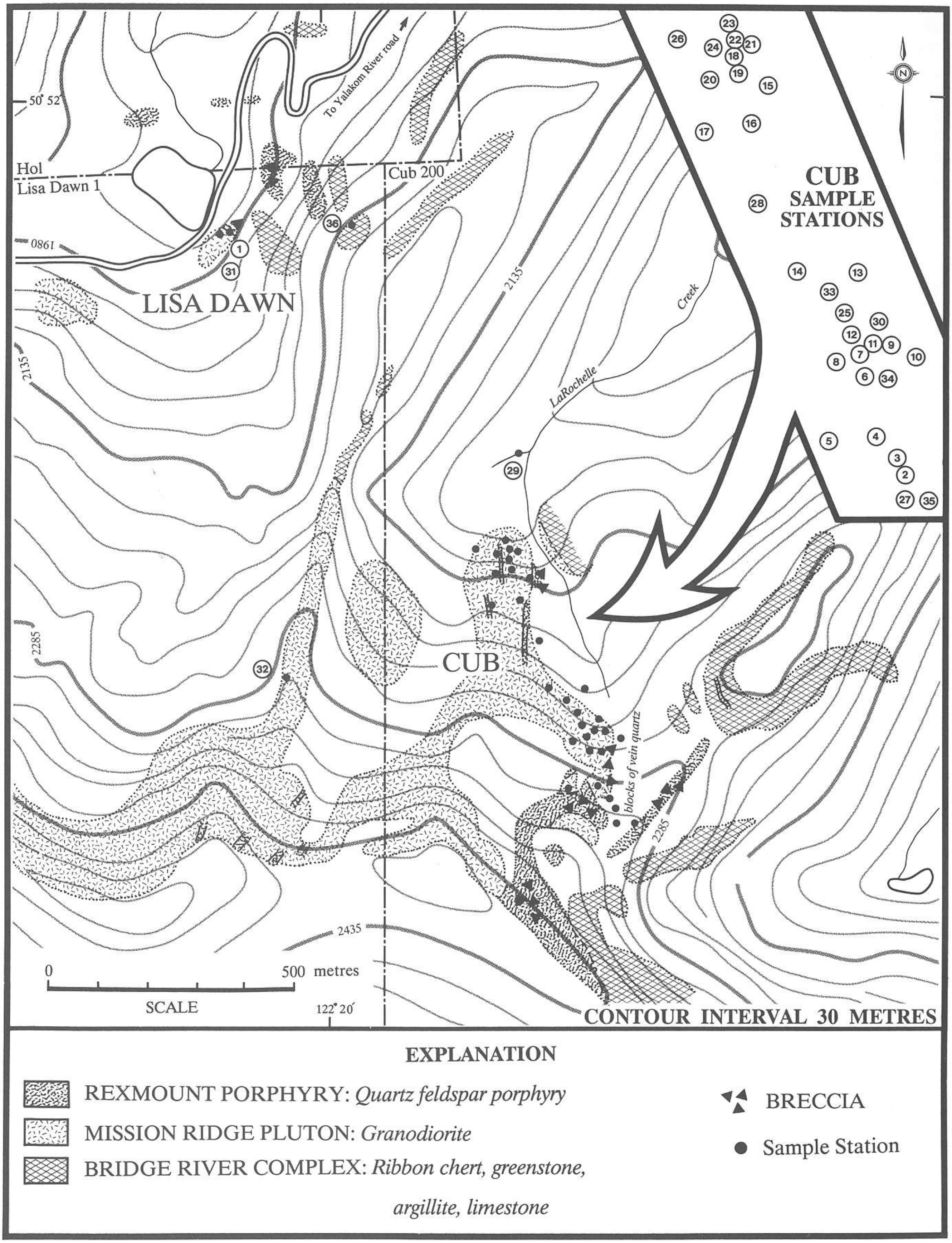

Figure 2-7-2. Geology in the area of the Cub and Lisa Dawn molybdenum prospects. Location is shown on Figure 2-7-2.

TABLE 2-7-2
STREAM SEDIMENT (RGS) GEOCHEMICAL ANALYSES
(SEE FIGURE 2-7-1 FOR SAMPLE LOCATIONS)

Sample Number	Zn	Cu	Pb	Ni	Co	Ag	Mo	W	As	Sb	Hg (ppb)
811063	95	56	—	150	17	0.4	3	1	19.5	0.2	30
811064	115	105	1	160	32	0.4	2	1	18.0	0.6	20
811065	120	90	3	90	27	0.2	3	1	14.5	0.6	20
811066	120	92	2	90	26	0.2	4	1	16.0	0.8	20
813049	86	54	3	73	17	0.2	2	1	15.0	1.8	30
813050	86	36	1	39	15	0.1	2	1	7.5	0.8	140
813051	115	82	5	101	20	0.2	4	1	11.5	1.2	260
813052	165	150	12	75	21	0.1	10	1	16.0	2.2	610
813053	92	82	6	70	32	0.1	2	1	10.0	1.2	120
813054	120	86	1	72	43	0.1	3	1	5.0	0.4	100
813055	50	38	1	17	9	0.1	1	1	7.0	0.4	100
813056	75	39	3	50	13	0.1	1	1	14.5	0.4	40
815595	105	50	7	330	30	0.3	3	1	19.5	1.8	50
815596	92	45	6	650	49	0.4	2	1	9.5	1.0	160
815597	97	56	1	340	30	0.4	3	2	35.5	1.0	50
815598	90	53	1	340	28	0.4	2	1	13.5	1.2	30

Data source: Regional Geochemical Survey, BC RGS-9, 1981 (GSC Open File 867).

All results recorded in ppm, except Hg in ppb.

All analyses performed by Chemex Labs Ltd., North Vancouver.

Analytical techniques: W by colorimetric determination, Hg by flameless cold vapour atomic absorption spectroscopy; all other elements by atomic absorption spectroscopy.

south of the Cub, Lisa Dawn and Alpine prospects and may be sampling similar molybdenite concentrations.

THE CUB MOLYBDENITE PROSPECT

The Cub prospect is wholly within the Cub 200 mineral claim at the headwaters of the southwesternmost tributary of LaRochelle Creek (Figures 2-7-1 and 2-7-2). At the ridge crest, the north-facing exposures of the Mission Ridge pluton, Rexmount porphyry and Bridge River complex are steep and craggy. Outcrop that extends down into the cirque is mostly granodiorite with smooth but steep surfaces over much of its exposure; a steep cliff along its eastern margin and water from runoff and melting icefields at higher elevations make access to much of the exposure hazardous. The lowermost outcrops are at treeline, and are now accessible by a road recently extended into the valley by MacNeill International Industries Inc. (Figure 2-7-1).

GEOLOGICAL SETTING

In the area of the Cub molybdenum prospect, Mission Ridge granodiorite is in contact with metamorphic rocks of the Bridge River complex, and is intruded by Rexmount porphyry (Figures 2-7-1, 2-7-2). All molybdenite seen in surface exposures is within granodiorite that is foliated to mylonitic and variably silicic. The granodiorite contains abundant quartz as stockwork veinlets to veins a few centimetres thick; the quartz veins are unevenly distributed and appear to be of several generations. Most outcrop surfaces are stained with manganese oxide.

Rexmount porphyry occurs as a large mass at higher elevations and as several dikes or apophyses, some of which clearly intrude the granodiorite. The porphyry is mostly feldspar and quartz-phyric, with minor hornblende altered to chlorite. It outcrops on the east side of the main zone of molybdenite mineralization at high elevations, but its contact with the granodiorite is largely obscured by talus of porphyry and Bridge River phyllite. The contact is marked by a ravine and a series of ribboned vein-quartz blocks up to 2 metres in size. The source of these blocks is not exposed but is thought to be close to the porphyry-granodiorite contact beneath talus.

An irregular and discontinuous zone of breccia fringes part of the granodiorite in contact with Rexmount porphyry. The breccia is largely composed of angular fragments of feldspar-phyric Rexmount porphyry, with lesser silicic granodiorite, quartz, and rare fragments of Bridge River complex rocks. It is generally adjacent to extremely silicic, brecciated and vuggy granodiorite (in part mylonitic). Vugs in granodiorite are lined with drusy quartz and chalcedony, and rarely by quartz pseudomorphous after calcite.

METAL DISTRIBUTION

The exposures of granodiorite in which molybdenite was observed occupy an area of at least 650 by 120 metres and span an elevation difference of at least 250 metres. The locations from which samples were obtained for analysis reflect the inaccessibility of the area between the 2150 to 2225-metre elevations (Figure 2-7-2).

At elevations between 2225 and 2315 metres, molybdenite, pyrite and possibly other sulphides are finely disseminated in extremely silicic, brecciated and partly vuggy grano-

TABLE 2-7-3
MOSS-MAT GEOCHEMICAL ANALYSES
(SEE FIGURE 2-7-1 FOR SAMPLE LOCATIONS)

Sample Number	Zn	Cu	Pb	Ni	Co	Ag	Mo	W	As	Sb	Cr	Au (ppb)	Pt (ppb)
BR8-039-M-1	132	83	9	106	21	0.3	2	1	34	2	104	592	1
BR8-064-M-1	161	101	15	97	20	0.2	5	1	16	2	72	5	1
BR8-065-M-1	169	130	15	65	20	0.3	11	1	17	2	29	6	1
BR8-066-M-1	119	82	7	69	29	0.1	1	1	6	2	83	1	1
BR8-067-M-1	105	98	3	72	29	0.2	1	1	11	2	91	1	1

All samples collected and submitted for analysis by B.N. Church, 1988.

All results recorded in ppm, except Au and Pt in ppb.

All analyses performed by Chemex Labs Ltd., North Vancouver. Analytical techniques: Hg by flameless cold vapour atomic absorption spectroscopy, Au and Pt by fire assay and mass spectroscopy; all other elements by inductively coupled plasma – atomic emission spectroscopy.

diorite. In some areas silicification and granulation are so intense that the character of the granodiorite is obliterated and the rocks are banded blue-grey silicic microbreccia or mylonite. The molybdenum content of these rocks is generally less than 400 ppm (Table 2-7-4), with less than 120 ppm copper and negligible gold.

Within less silicic foliated protomylonitic granodiorite, discordant vein and stockwork quartz contains up to a few per cent visible molybdenite and pyrite, but analyses generally indicate only up to 140 ppm molybdenum with negligible copper and gold. One outstanding exception to this is sample 33 (Table 1-7-4) which contains 0.50 per cent copper, 2120 ppb gold and anomalous bismuth and lead.

Large blocks of vein quartz (along the ravine at the Rexmount porphyry-granodiorite contact) are found up to an elevation of 2225 metres. The vein quartz is ribboned with molybdenite stylolites and veinlets and blebs of chalcopyrite. Samples of these blocks returned analyses as high as 0.55 per cent molybdenum, 1.36 per cent copper, 1.66 grams per tonne gold, in excess of 50 grams per tonne silver and, in some cases, anomalous bismuth, lead and zinc. In general, the distribution of gold closely follows that of copper, but not molybdenum (Table 1-7-4).

At elevations between 2070 and 2150 metres, molybdenite occurs as disseminated flakes and clots a few millimetres in size within and adjacent to irregular pyritic quartz stockwork

TABLE 2-7-4
TRACE METAL CONTENTS OF SAMPLES COLLECTED FROM THE CUB MOLYBDENUM PROSPECT AND ADJACENT AREAS
(SEE FIGURE 2-7-2 FOR SAMPLE LOCATIONS)

Sample Number	Field Station	Au	Ag	As	Bi	Cu	Fe	Mo	Pb	Zn	Sample Description
1	89BGA-34-6	120	2.5	14	—	17	0.73	2270	40	12	Quartz vein (Lisa Dawn prospect); mo, fmo
2	89BGA-35-10a	75	1.0	120	—	261	0.73	537	35	58	Quartz vein (talus); mo, fmo, cpy, mal, az
3	89BGA-35-10b	870	10.0	130	—	3610	1.50	1260	90	174	Quartz vein (talus); mo, fmo, cpy, mal, az
4	89BGA-35-10c	15	0.5	11	—	117	0.63	69	10	14	Bleached, silicic gd; py, hem
5	89BGA-36-2	<5	<0.5	5	—	37	0.49	17	15	12	Stylolitic quartz in gd; py
6	89BGA-36-4-1	<5	<0.5	4	—	11	0.41	12	10	4	Quartz stockwork in silicic gd; py, mo
7	89BGA-36-4-2	<5	1.0	3	—	23	0.45	137	165	10	Quartz stockwork in silicic gd; mo, py
8	89BGA-36-4-3	<5	<0.5	3	—	14	0.40	8	15	4	Silicic gd; mo, py
9	89BGA-36-5	<5	<0.5	3	—	14	0.53	394	40	14	Silicic gd; mo, py
10	89BGA-36-6	45	2.5	5	—	1070	0.58	5490	5	34	Quartz vein (talus); mo, fmo, cpy, mal, az
11	89BGA-37-1b	15	<0.5	7	—	13	0.71	54	10	12	Silicic gd – qfp contact; py
12	89BGA-37-2	<5	<0.5	3	—	21	0.66	107	5	8	Silicic gd, vuggy; mo
13	89BGA-37-3	<5	<0.5	4	—	27	0.75	50	5	4	Quartz stockwork in silicic gd; mo, py
14	89BGA-37-4	<5	<0.5	4	—	14	0.57	26	15	22	Quartz veins in gd
15	89BGA-37-7	<5	0.5	53	—	25	0.78	15	5	6	Quartz stockwork in silicic gd; py, mo
16	89BGA-37-8	60	3.0	39	—	514	2.32	866	155	48	Quartz veins in gd; mo, py
17	89BGA-37-9	<5	<0.5	15	—	25	0.71	26	20	8	Quartz stockwork in silicic gd; py, mo
18	89BGA-37-10	65	2.0	65	—	174	2.93	1310	5	12	Quartz stockwork in silicic gd; mo, py
19	89BGA-37-11	10	0.5	10	—	78	0.99	873	5	6	Quartz veins in silicic gd; mo
20	89BGA-37-12	15	1.0	7	—	68	0.74	1250	35	16	Quartz veins in silicic gd; mo, py
21	89BGA-38-1	10	1.0	14	—	64	0.88	558	5	6	Quartz veins in silicic gd; mo, py
22	89BGA-38-2	5	1.5	7	—	46	0.60	1620	5	6	Quartz veins in silicic gd; mo, py
23	89BGA-38-3	<5	1.0	6	—	42	0.72	462	5	2	Quartz veins in silicic gd; mo, py
24	89BGA-38-4	10	1.5	20	—	60	0.95	441	10	6	Quartz veins in silicic gd; mo
25	89BGA-38-5	<5	1.0	5	—	27	0.45	34	<5	8	Brecciated silicic gd, chalcedony; py
26	89BGA-38-6	55	1.5	38	—	42	1.19	1310	10	4	Quartz veins in silicic gd; mo, py (talus)
27	DAR-89-004	50	1.5	31	<3	821	0.60	0.46%	32	16	Quartz vein (talus); mo, cpy, py
28	DAR-89-006	20	1.9	<3	<3	26	0.44	0.14%	30	3	Quartz veins in silicic gd (talus); mo, py
29	DAR-89-008	<5	0.2	23	<3	32	2.07	31	26	57	Silt sample from LaRochelle Creek
30	DAR-89-015	1660	>50.0	8	144	0.96%	2.17	3	1236	51	Quartz vein (talus); cpy, py
31	DAR-89-020	150	0.1	47	3	366	1.12	876	20	5	Slightly silicic gd; mo, py
32	DAR-89-028	150	2.3	69	<3	517	1.93	162	26	27	Quartz vein in gd; py
33	DL-89-6	2120	0.1	<3	72	0.50%	1.14	3	769	22	Quartz stockwork in silicic gd; mo, cpy
34	08651	340	45.5	486	65	1.36%	2.78	158	635	483	Quartz vein (talus); cpy, mal, az
35	83-RKT-756	200	0.6	3	—	870	—	18	1	56	Chert-argillite (Bridge River Complex); py
36	T300	30	—	14	—	>1%	—	—	—	—	Argillite-greenstone (BRC); py, cpy, mal

Samples 1 to 26 collected by Geological Survey Branch staff; approximately 1 kg samples taken in-situ unless otherwise noted.

Samples 27 to 34 collected and submitted for analysis by Min-Ex Resource Consultants.

Samples 35 and 36 collected and submitted for analysis by Utah Mines Ltd. (Pollock, 1983).

All results recorded in ppm unless otherwise noted; Au in ppb.

All analyses, except for 27 to 34, performed by Chemex Labs Ltd., North Vancouver. Analytical techniques: Au by fire assay and atomic absorption spectroscopy finish; as by atomic absorption spectroscopy; all other elements by inductively coupled plasma – atomic emission spectroscopy.

Analyses 27 to 34 performed by Vangeochem Lab Ltd., Vancouver. Analytical techniques: Au by fire assay and atomic absorption spectroscopy finish, and all other elements by inductively coupled plasma spectroscopy.

Abbreviations: apy = arsenopyrite, az = azurite, cpy = chalcopyrite, fmo = ferromolybdite, hem = hematite, mal = malachite, mo = molybdenite, py = pyrite; gd = granodiorite, qfp = quartz feldspar porphyry.

and discontinuous quartz veins. Rare chalcopyrite accompanies pyrite concentrations in quartz veins. The host granodiorite is slightly silicic and has a well-defined protomylonitic fabric paralleled by conspicuous clots of chlorite and secondary(?) biotite. The molybdenum content of these rocks is generally between 0.05 and 0.16 per cent (Table 2-7-4) but ranges up to 0.32 per cent (J. Perry, personal communication, 1989). Anomalous copper is rare and correlates with slightly anomalous gold and silver.

SYNTHESIS

The Cub prospect is hosted by protomylonitic to mylonitic Mission Ridge granodiorite adjacent to Rexmount porphyry. Host granodiorite is extremely silicic at higher elevations and contains only geochemically anomalous metal contents; these rocks display textures characteristic of the upper part of a hydrothermal system. At lower elevations, quartz is present mainly as stockwork veins with associated molybdenum mineralization. Overall, there is a rough zonation from a high-level silicic cap with generally low but sporadic metal content to a deeper level of stockwork quartz-molybdenite-pyrite mineralization.

Vein quartz occurring as blocks in the granodiorite-porphyry contact zone contains the largest molybdenum, gold and copper contents. The zone of breccias at the granodiorite-porphyry contact records a complex history of repeated fracturing. The breccias appear to be essentially tectonic, but irregular pipe-like forms suggest hydrothermal events may have played a role in their development. The position of ribboned quartz veins at the contact between granodiorite and porphyry (as at the Cub and Lisa Dawn prospects) suggests that their origin may be related to episodic tectonism and hydrothermal activity along the contact zone.

REGIONAL METALLOGENIC IMPLICATIONS

Metal prospects within the Mission Ridge pluton are conspicuously aligned along the contact with the Rexmount porphyry (Figure 2-7-1). However, the participation of porphyry in the mineralizing events is not understood. Porphyry at the Cub prospect is unmineralized, and its fresh appearanee suggests postmineralization emplacement. Similarly, the porphyry at the Spokane prospect seems late.

The similar metal assemblages in the Spokane veins, the ribboned quartz vein at the Lisa Dawn prospect, the vein quartz blocks on the Cub prospect and the Alpine vein provide an important link across the area; the Spokane veins, however, are copper-rich with little(?) or no molybdenum. If the veins at the Cub prospect are an intregal part of the Cub stockwork molybdenite system, and the vein at the Lisa Dawn prospect is related to disseminated molybdenite in adjacent granodiorite, then the area between the Spokane and Alpine prospects is prime exploration ground for additional stockwork molybdenite.

ACKNOWLEDGMENTS

Doug Archibald of Queen's University assisted in the collection of data and shared in the thrill of discovering the Cub prospect (alias "the Stubby"). Rob Macdonald of Memorial University of Newfoundland whole-heartedly helped with the task of geochemical sampling. Field trips and discussions with P. Schiarizza, D.G. MacIntyre, W.J. McMillan, R.E. Meyers and W.R. Smyth of this ministry, and D. Lucas and J. Perry of Min-Ex Resouree Consultants were informative and helped focus attention on specific geological problems. The generous hospitality of the MacNeill International Industries Inc. field camp and F. Hilton in Vancouver is much appreciated. Constructive discussions and geochemical data from J. Perry greatly added to the content of this paper.

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NOTES

GEOLOGY OF THE SILVER QUEEN MINE AREA, OWEN LAKE, CENTRAL BRITISH COLUMBIA (93L)

By Craig H.B. Leitch, Christopher T. Hood, Xiao-lin Cheng and Alastair J. Sinclair The University of British Columbia

KEYWORDS: Economic geology, Silver Queen mine, epithermal, polymetallic vein, Tip Top Hill volcanics Kasalka Group, continental volcanism.

INTRODUCTION

The Silver Queen (Nadina, Bradina) mine of Pacific Houston Resources Inc. is near Owen Lake, 35 kilometres southeast of Houston, and 100 kilometres southeast of Smithers, in the Bulkley Valley region of central British Columbia (Figure 2-8-1). The geology of the 20 square kilometre area surrounding the deposit has been mapped, and results suggest that the stratified rocks hosting this epithermal gold-silver-zinc-lead-copper vein deposit (the Late Cretaceous Tip Top Hill volcanics; Church, 1971) may be correlative with rocks hosting the Equity Silver deposit, and are lithologically similar to the Kasalka Group of late-Early to early-Late Cretaceous age. The geological mapping is part of a more extensive project dealing with the geology and origin of polymetallic vein deposits in the Owen Lake area.

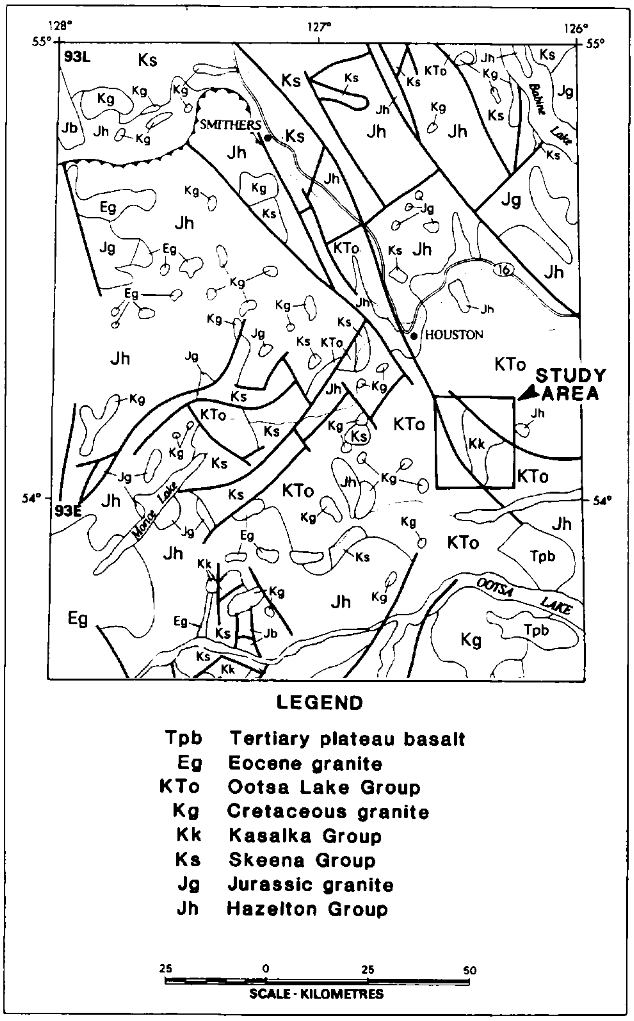

Figure 2-8-1. General geology of west-central British Columbia, showing the regional setting of the study area. Taken from MacIntyre (1985).

REGIONAL GEOLOGIC SETTING

West-central British Columbia lies within the Stikine Terrane, which includes submarine calcalkaline to alkaline immature volcanic islandarc rocks of the Late Triassic Takla Group, subaerial to submarine calcalkaline volcanic, volcaniclastic and sedimentary rocks of the Early to Middle Jurassic Hazelton Group, Late Jurassic and Cretaceous successor basin sedimentary rocks of the Bowser Lake, Skeena and Sustut groups, and Cretaceous to Tertiary calcalkaline continental volcanic arc rocks of the Kasalka, Ootsa Lake and Endako groups (MacIntyre and Desjardins, 1988). The younger volcanic rocks occur sporadically throughout the area, mainly in downthrown fault blocks and grabens. Plutonic rocks of Jurassic, Cretaceous and Tertiary age form distinct intrusive belts (Carter, 1981), with which porphyry copper, stockwork molybdenum and mesothermal and epithermal base and precious metal veins are associated.

The Kasalka Group (Armstrong, 1988) is considered to be a late-Early (Armstrong, 1988) or early-Late Cretaceous (MacIntyre, 1985) continental volcanic succession that is predominantly porphyritic andesite and associated volcaniclastic rocks. It is well exposed in the Kasalka Range type section near Tahtsa Lake. In the type area, it includes a basal polymictic conglomerate that is strikingly red in colour and lies in angular unconformity on older rocks. The unit is generally between 5 and 10 metres thick (locally 50 metres in channel-fill deposits), and includes interfingering lenses of sandstone. The conglomerate is overlain by a felsic fragmental unit over 100 metres thick, consisting of grey to cream-coloured, variably welded siliceons pyroclastic rocks (lithic lapilli tuff, crystal and ash-flow tuff, minor breccia) with interbedded porphyritic flows. These fragmental rocks are in turn overlain by a major unit of columnar jointed, massive, greenish grey flows or sills of hornblende-feldspar-porphyritic andesite to dacite, at least 100 metres thick. The andesite flows are conformably overlain by a chaotic assemblage of volcanic debris flows (lahars), at least 200 metres thick, in which most clasts are identical to the

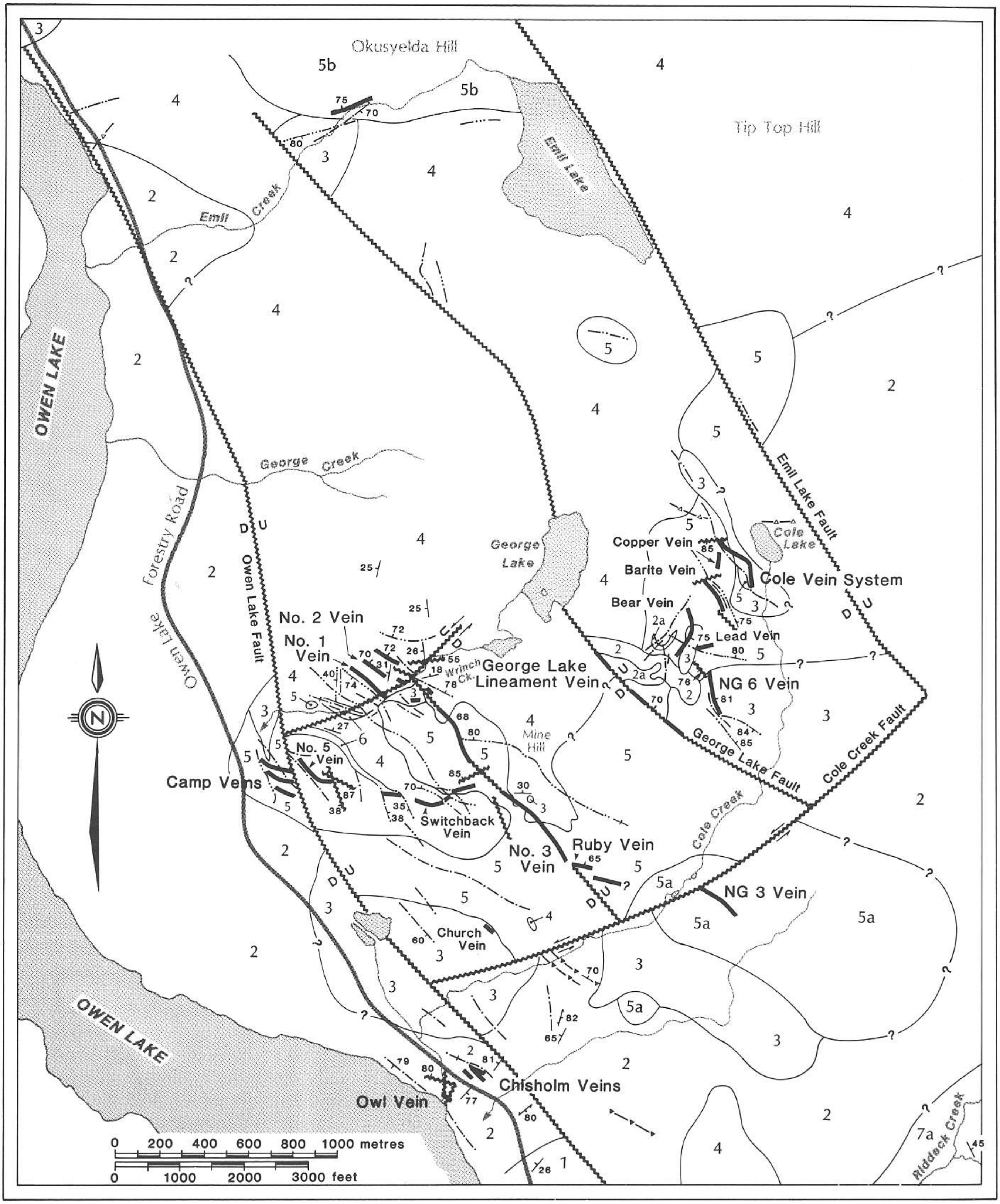

Figure 2-8-2. Detailed property geology of the Silver Queen property. Owen Lake area, west-central British Columbia. Units are defined in Table 2-8-1.

underlying flows and sills. Rhyolite flows and tuffs and columnar jointed basalt flows, together more than 100 metres thick, cap the succession (the basalts may be significantly younger: MacIntyre, 1985).

A mid to Late Cretaceous age is assigned to the Kasalka Group volcanic rocks because they unconformably overlie sedimentary rocks containing latest Early Cretaceous (Albian) fauna (Duffel, 1959). Dacitic lapilli tuffs near the base of the group give an isotopic age of 108 to 107 ± 5 Ma by K-Ar on whole rock, and intrusions dated at 87 ± 4 to 83.8 ± 2.8 Ma cut the stratified units (MacIntyre, 1985).

Volcanic rocks of similar age and lithology are not widely known in west-central British Columbia, but possible correlatives are rocks found in the Mount Cronin area northeast of Smithers (MacIntyre and Desjardins, 1988). The correlative rocks near Mount Cronin were formerly mapped as Brian Boru formation by Tipper and Richards (1976), and correlated to Brian Boru rocks as defined by Sutherland-Brown (1960) in the Rocher Déboulé Range northwest of Smithers. In the Mount Cronin area, MacIntyre and Desjardins separate the Kasalka Group into lower and upper divisions. As in the Kasalka Range, the succession begins with a heterolithic, maroon basal conglómerate with interbedded sandstone, siltstone and mudstone. This is followed by thin-bedded tuffs and epiclastics, mafic flows, and pyroelastic rocks that include lapilli tuff and breccia, bedded lahar, and siliceous ash-flow tuff and breccia. The upper division comprises a thick section of poorly bedded volcanic breccia with angular clasts, grading upward to hornblende-feldspar crystal tuff and interbedded to overlying hornblende-feldspar-porphyritic andesite, most of which are flows but some intrusive stocks and sills may also be present.

In spite of very similar lithology, the Tip Top Hill volcanics of the Buck Creek basin in the Parrot Lake and Owen Lake area, and volcanic rocks hosting The Equity Silver deposit, cannot be correlated with the Kasalka Group on the basis of currently available isotopic dates.

GEOLOGY OF THE BUCK CREEK BASIN

The Buck Creek basin has been characterized as a resurgent caldera, with the important Equity Silver mine located within a window eroded into the central uplifted area (Church, 1985). The Silver Queen mine lies on the caldera rim or perimeter of this basin, which is roughly delineated by a series of rhyolite outliers and semicircular alignment of Upper Cretaceous and Eocene volcanic centres scattered between Francois Lake, Houston and Burns Lake (*see* Figure 59 of Church, 1985). A prominent lineament 30 kilometres long and trending east-northeasterly from the Silver Queen mine towards the central uplift hosting the Equity mine, appears to be a radial fracture coinciding with the eruptive axis of the Tip Top Hill (Kasalka Group) volcanics and a line of syenomonzonite stocks and feeder dikes to an assemblage of "moat volcanics" that include the Goosly Lake formation (Church, 1985). Block faulting is common in the basin, locally juxtaposing the various ages of volcanic rocks found within it.

In broad outline, a Mesozoic volcanic assemblage is overlain by a Tertiary volcanic succession. The oldest rocks exposed within the basin are at the Equity Silver and Silver Queen mines. The sequence at the Equity mine has been characterized by Church (1984) as Jurassic Hazelton Group rocks of the Telkwa formation overlain with angular unconformity by Lower Cretaceous Skeena Group sedimentary rocks. However, Wetherell *et al.* (1979) and Cyr *et al.* (1984) correlate the sequence hosting the Equity orebodies with the Upper Cretaceous Kasalka Group, and Wojdak and Sinclair (1984) list as possible correlatives the Lower Cretaceous Skeena Group, the Kasalka Group and the Brian Boru formation. The geology of the Equity mine area is obviously as yet imperfectly known.

Large areas of Upper Cretaceous rocks are exposed westwards from the Equity mine to the Owen Lake area, where they host the Silver Queen deposit (Church, 1984). These rocks, which have been dated at 77.1 ± 2.7 to 75.3 ± 2.0 Ma by K-Ar on whole rock (Church, 1973) are described by Church (1984) to consist of a lower, acid volcanic unit overlain by the Tip Top Hill formation andesites to dacites. This subdivision is based on "rhyolitic volcanic rocks below the Tip Top Hill formation in the Owen Lake area in extensive drill holes in the vicinity of the Silver Queen mine" (Church, 1973), which he considers to be "lateral equivalents of quartz porphyry intrusions exposed nearby on Okusyelda Hill" (Figure 2-8-2). Current mapping indicates that the lower volcanic unit exposed in the drill holes may in part be a strongly altered equivalent of the Tip Top Hill volcanics. The quartz porphyry of Okusyelda Hill could correlate with dacitic quartz porphyry sills, dikes and laccoliths common within the type Kasalka Group section in the Tahtsa Lake area. Late quartz feldspar porphyry dikes are also found at the Equity mine (Cyr *et al.*, 1984; Church, 1985), although these are dated at 50 Ma and thus belong to the younger Ootsa Lake Group.

The Upper Cretaceous rocks are overlain by the Eocene Ootsa Lake Group, which includes the Goosly Lake and Buck Creek formations of Church (1984). The Goosly Lake andesitic to trachyandesitic volcanic rocks are dated at 48.8 ± 1.8 Ma by K-Ar on whole rock, and this is supported by dates of 49.6 ± 3.0 to 50.2 ± 1.5 Ma for related syenomonzonite to gabbro stocks with distinctive bladed plagioclase crystals (Church, 1973) at Goosly and Parrot lakes. The Buck Creek andesitic to dacitic volcanic rocks, which directly overlie the Goosly Lake formation, are dated at 48.1 ± 1.6 Ma by K-Ar on whole rock. These ages correlate with whole rock K-Ar ages of 55.6 ± 2.5 Ma for dacite immediately north of Ootsa Lake (Woodsworth, 1982) and 49.1 ± 1.7 Ma on biotite for Ootsa Lake Group rocks in the Whitesail Lake area immediately south of Tahtsa Lake (Diakow and Koyanagi, 1988).

Basalts of the upper part of the Buck Creek formation (Swans Lake member: Church, 1984) may correlate with the Endako Group of Eocene-Oligocene age. These rocks give whole rock K-Ar ages of 41.7 ± 1.5 to 31.3 ± 1.2 Ma on samples from the Whitesail Lake map area (Diakow and Koyanagi, 1988).

The youngest rocks in the Buck Creek basin are cappings of Miocene columnar olivine basalt, called the Poplar Buttes formation by Church (1984) and dated at 21.4 ± 1.1 Ma by K-Ar on whole rock (Church, 1973).

GEOLOGY OF THE STUDY AREA

The preliminary geology of the study area immediately surrounding the Silver Queen mine, as determined by fieldwork and petrological studies completed in 1989, is shown in Figure 2-8-2 (units are defined in Table 2-8-1). Relationships between the map units are shown diagrammatically in Figure 2-8-3. The succession is similar to that observed in the Kasalka Range and on Mount Cronin.

The rocks of the study area have been subdivided into five major units plus three dike types; Table 2-8-1 lists the map units defined to date. A basal reddish purple polymictic conglomerate (Unit 1) is overlain by fragmental rocks ranging from thick crystal tuff (Unit 2) to coarse lapilli tuff and breccia (Unit 3), and this is succeeded upwards by a thick feldspar-porphyritic andesite flow unit (Unit 4), intruded by microdiorite sills and other small intrusions (Unit 5). The stratified rocks form a gently northwest-dipping succcession, with the oldest rocks exposed near Riddeck Creek to the south and the youngest exposed in Emil Creek to the north (Figure 2-8-2). All the units are cut by dikes that can be divided into three groups: amygdaloidal dikes (Unit 6), bladed feldspar porphyry dikes (Unit 7), and diabase dikes (Unit 8). The succession is unconformably overlain by basaltic to possibly trachyandesitic volcanics that crop out in Riddeck Creek and farther south. These volcanics may be correlative with the Goosly Lake formation (Church, 1973).

Mineralization on the property is mainly restricted to quartz-carbonate-barite-specularite veins, 1 to 2 metres thick, that contain disseminated to locally massive pyrite, sphalerite, galena, chalcopyrite, tennantite and argentian tetrahedrite. Locally, in chalcopyrite-rich samples, there is a diverse suite of Cu-Pb-Bi-Ag sulphosalts such as aikinite, matildite (in myrmekitic intergrowth with galena), pearcite-arsenpolybasite, and possibly schirmerite (berryite, guettardite and meneghinite have also been reported but not yet confirmed). Native gold with unusually low fineness of 510 to 620 (actually electrum) is present in minor amounts. The veins cut the amygdaloidal, fine-grained plagioclase-rich dikes (Unit 6), and are cut by the series of dikes with bladed plagioclase crystals (Unit 7). Both these dike types are possibly correlative with the Ootsa Lake Group Goosly Lake volcanics of Eocene (approximately 50 Ma) age. The bladed feldspar porphyry dikes cut the amygdaloidal dikes, and both

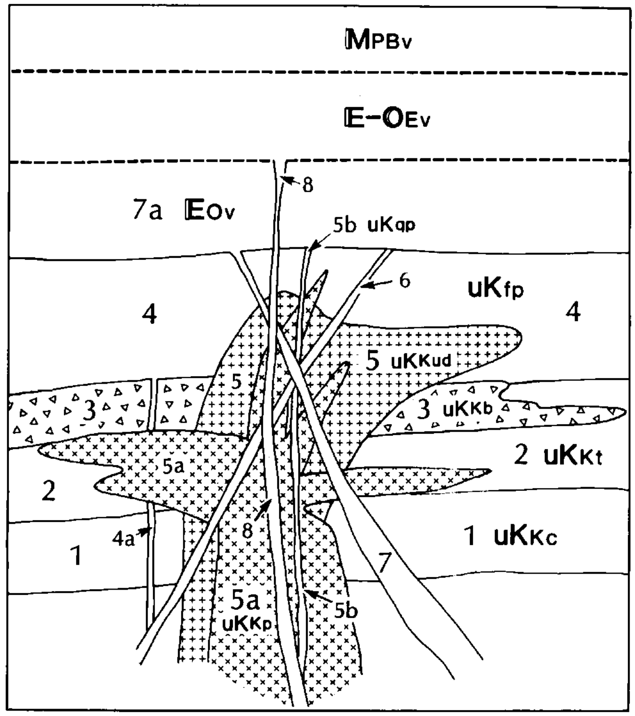

Figure 2-8-3. Schematic diagram of stratigraphic and intrusive relationships, Owen Lake area, west-central British Columbia. Units are defined in Table 2-8-1.

TABLE 2-8-1
TABLE OF FORMATIONS, OWEN LAKE AREA

Period	Epoch	Age (Ma)	Formation	Symbol	Unit	Lithology
TERTIARY	Miocene	21	Poplar Buttes	M_{PBv}		Olivine basalt
	Eocene-Oligocene	40-30	Endako Group	EO_{Ev}	8	Basalt, diabase dikes
	Eocene	56-	Ootsa Lake	E_{Ov}	7a	Trachyandesite, basalt
		47	Group		7	Bladed feldspar porphyry dikes
			— MINERALIZED VEINS —			
					6	Amygdular dikes
CRETACEOUS	(Late)		"Okusyelda"	uKqp	5b	Quartz-eye rhyolite dikes, stock
			Tip Top Hill volcanics	uKKp	5a	Intrusive porphyry sills, stocks
		75		uKKud	5	"Mine Hill" microdiorite
		77			4a	Feldspar biotite porphyry dikes
				uKKfp	4	"Tip Top Hill" feldspar porphyry (voluminous porphyritic andesite)
				uKKb	3	Medium to coarse tuff-breccia
				uKKt	2	Crystal tuff, local lapilli tuff
					2a	Fine ash tuff
				uKKc	1	Polymictic basal conglomerate, sandstone and shale interbeds

are cut by the diabase dikes that may correlate with Endako Group volcanism of Eocene-Oligocene (approximately 40 to 30 Ma) age.

TIP TOP HILL VOLCANICS

Units 1 to 5, as defined in the map area, fall within the Tip Top Hill formation (Church, 1984), but correspond closely with the units defined in Kasalka Group rocks elsewhere. The units are described in detail below, to facilitate comparison with other, possibly correlative rocks.

BASAL POLYMICTIC CONGLOMERATE (UNIT 1)

The basal member of the succession is a reddish to purple, heterolithic, poorly sorted pebble conglomerate that contains rounded to subangular small white quartz and grey-brown to less commonly maroon tuff and porphyry clasts. Local interbeds of purplish sandstone with graded bedding are found within the unit, as are rare black shaly partings. The matrix is composed of fine sand, cemented by quartz, sericite and iron oxides. The best exposure is found in a roadcut at the southern tip of Owen Lake, where the unit is about 10 metres thick and dips 25° to the northwest. The base is not exposed and the unit is in presumed fault contact with the younger volcanic rocks of the Ootsa Lake Group (Goosly Lake formation; Unit 7) exposed at higher elevations farther south along the road. In drill holes farther north, near the centre of the property, the upper contact of the conglomerate with overlying porphyry is sharp and appears conformable, but the porphyry may be an intrusion rather than a flow.

CRYSTAL-LITHIC TUFF (UNIT 2)

In outcrop, the next major unit is a sequence of mainly fragmental rocks that are mostly fine crystal tuffs with thin interbeds of laminated tuff, ash tuff, lapilli tuff, and less abundant breccia. The unit may be as much as 100 metres thick. The most widespread rock type is a massive, grey to white, strongly quartz-sericite-pyrite altered, fine crystal tuff that grades imperceptibly into a porphyry of similar appearance and composition; the latter may be partly flow, intrusive sill, or even a welded tuff. Only the presence of broken phenocrysts and rare interbeds of laminated or coarsely fragmental material suggest that the bulk of this unit is tuffaceous. In thin section, the rock is seen to be made up of 1 to 2-millimetre broken, altered plagioclase relics and 0.5-millimetre anhedral quartz grains (that may be partly to entirely secondary) in a fine matrix of secondary sericite, carbonate, pyrite and quartz. Drill-core exposures show that the basal contact of Unit 2 with the underlying conglomerate is commonly occupied by the porphyry rather than the tuff. The best exposures of Unit 2 are in the area of Cole Creek and the Chisholm vein (Figure 2-8-2), where thin (10 centimetre) interbedded laminated tuff bands occur, many with variable dips to near-vertical, although coarser lapilli tuff lenses, up to 1 metre thick, display gentle northerly dips. In drill core, sections of laminated tuffs with faint but discernible layering on a centimetre scale, may be up to 10 metres thick; angles with the core axis suggest a gentle dip for the banding.

Outcrops on the northeast side of the George Lake fault (Figure 2-8-2) have rare interbeds of a very fine, uniform "ash tuff" that are up to several metres thick (Unit 2a). Typically they are dark grey to medium grey-green and have a siliceous appearance. Locally they contain angular fragments of either mixed origins (heterolithic clasts) or of larger blocks that are only barely distinguishable from the matrix (monolithic clasts).

The composition of Unit 2 is not known from chemical analyses. Although it looks felsic (*cf.* Church, 1973) the highly altered nature may give a misleading impression of its original character; it may have been originally andesitic as are the overlying units.

COARSE FRAGMENTAL UNIT (UNIT 3)

A distinctive coarse fragmental unit overlies or in some places is interlayered with the upper part of Unit 2. It is composed of blocks and bombs(?) (*cf.* MacIntyre, 1985) of feldspar-porphyritic rock similar in appearance to both the underlying porphyry and the overlying porphyritic andesite. The clasts are mostly angular to subangular and about 2 to 5 centimetres in diameter, but some are much larger (up to 0.5 metre); the matrix makes up a widely variable percentage of the rock, from almost zero to 90 per cent, so that in places the rock has the appearance of an intrusive breccia with little or no rotation of fragments. In other places the fragments are clearly unrelated and "accidental" or unrelated clasts of chert or fine tuff are common, although still volumetrically minor; this has the appearance of a lahar.

In outcrop near the Cole veins (Figure 2-8-2), this breccia or lahar(?) unit forms discontinuous lenses generally less than 10 metres thick, with a suggestion of gentle northerly dips. The lenses appear to be conformable with the underlying or enclosing tuffs. In drill core, two distinctly different modes of occurrence are noted for this unit: in one, it appears to be conformably overlain by Unit 4 porphyritic andesites (the total thickness of the breccia unit is up to 30 metres); in the other, it appears to have subvertical contacts, implying it is an intrusive breccia. Good examples of the latter distribution are found in the Cole Lake area, the Camp vein system and around the southern end of Number 3 vein. There is thus a rough correlation between the subvertical breccia bodies and mineralized areas, just as there is between the microdiorite and mineralized areas (*see* below).

In thin section, the clasts of the breccia are seen to be composed of strongly altered feldspar porphyry, fine tuff and quartz or quartzofeldspathic rocks, enclosed in a fine tuffaceous matrix. Alteration in the mine area is usually carbonate-sericite-quartz-pyrite, and is intense enough to largely obscure the original texture.

FELDSPAR PORPHYRY (UNIT 4)

The fragmental rocks appear to be conformably overlain by a thick, massive unit of porphyritic andesite that outcrops over much of Mine Hill and is best developed north of Wrinch Creek (Figure 2-8-2). This unit is equivalent to the Tip Top Hill volcanics of Church (1970), although in most places on the property the porphyry is coarser and contains sparser phenocrysts than the exposures on Tip Top Hill. In exposures in Wrinch Creek canyon, a distinct flow lamination is developed by trachytic alignment of phenocrysts, best seen on weathered surfaces. This suggests that these porphyries are mostly flows, with gentle northerly to northwesterly dips. However, some of the coarsest material probably forms intrusive sills and stocks [*cf.* the type sections of MacIntyre and Desjardins (1988) and MacIntyre (1985)] and in many places the porphyry grades into intrusive microdiorite (Unit 5).

Parts of this unit, particularly in Emil Creek, west of Emil Lake, and on Tip Top Hill itself (Figure 2-8-2), may actually be crystal tuffs. In these exposures, the feldspar phenocrysts are smaller, much more crowded and in places broken, and rare lithic fragments are visible.

This unit has been dated at 77.1 ± 2.7 Ma by K-Ar on whole rock (Church, 1973). Rhyolite from Tsalit Mountain on the west side of Owen Creek valley, 10 kilometres northwest of the Silver Queen mine, gives a very similar isotopic date of 77.8 ± 3.0 Ma, also by K-Ar on whole rock (Church, 1973). Church correlated this rhyolite with the "Okusyelda" quartz porphyry (Unit 5b of this study, thought to be slightly younger than Unit 5 microdiorite) found in Emil Creek and on Okusyelda Hill (Figure 2-8-2).

In thin section, the feldspar porphyry is seen to contain abundant 2 to 3-millimetre euhedral crystals of andesine. Oscillatory zoning is present, but with little overall change in composition within a given specimen, from An_{45} to An_{35}. Mafic minerals include roughly equal amounts (about 5% each) of 1 to 2-millimetre clinopyroxene and hornblende, though both are strongly altered to carbonate, hydrobiotite and apatite. Euhedral 1 to 2-millimetre biotite phenocrysts are generally less altered. The groundmass is an aphanitic mesh of intergrown feldspar with minor opaque grains; primary magnetite is abundant in the fresh specimens.

The average composition of the feldspar porphyry is between andesite and dacite, as indicated by arc-fusion determinations and chemical analyses (Church, 1973). Apart from a lower potash content, the chemistry of the feldspar porphyry is remarkably similar to that of the microdiorite (Unit 5).

BIOTITE FELDSPAR PORPHYRY DIKES (UNIT 4A)

Rare, thin (1 metre or less) dikes with similar composition and appearance to the flows of Unit 4 probably represent feeders to overlying flows. They are distinguished by prominent scattered books of black biotite up to 3 millimetres across, as well as abundant 1 to 2-millimetre plagioclase phenocrysts. These dikes have only been recognized near the north end of Cole Lake and on the highway at the north end of Owen Lake (Figure 2-8-2), but they may be more extensive (they are difficult to recognize because of their similarity to Unit 4).

MICRODIORITE (UNIT 5)

Microdiorite forms subvolcanic sills, dikes, and possibly small irregular stocks on the Silver Queen mine property. These intrusions are centrally located in the two main mineralized areas, the No. 3 Vein and Cole vein areas (Figure 2-8-2). Contacts with the feldspar porphyry are indistinct or gradational over about 1 metre, but dikes are seen cutting older units. The gradational contacts probably caused earlier workers such as Marsden (1985) to propose two divisions of microdiorite, one with quartz and biotite and one without. With further work, it can now be seen that the biotite-bearing phase belongs to the feldspar porphyry (Unit 4).

Typically the microdiorite is a medium to fine-grained, dark greenish grey equigranular to porphyritic rock characterized by small (1 millimetre, but locally glomeratic to 4 millimetres) plagioclase phenocrysts and 0.5-millimetre mafic relics in a phaneritic pink feldspathic groundmass. Primary magnetite is found in the less-altered specimens. It is distinguished in outcrop by its relatively fine-grained, even-weathering texture, lacking flow structure compared to the feldspar porphyry. Because of the gradational relationship to the feldspar porphyry, mineralogical distinction is not reliable. In thin section, the plagioclase is the same as in the feldspar porphyry (oscillatory zoned andesine, An_{45-30}), and euhedral clinopyroxene phenocrysts, partly altered to carbonate, are the most abundant mafic. Apparent hornblende relics are completely altered to chlorite. No biotite is seen, but rare scattered quartz phenocrysts, displaying late-stage overgrowths of quartz, are observable ranging up to 1 millimetre in size (these are not visible in hand specimen). The groundmass is composed of fine (0.1 millimetre) quartz, plagioclase and potassium feldspar.

Chemically, the microdiorite is the same as the feldspar porphyry (Church, 1970, 1971). This relationship is the same as that observed by MacIntyre (1985) in the Kasalka Range near Tahtsa Lake. The chemistry compares closely to that of an average augite andesite (Daly, 1933, cited in Church, 1970) or quartz-bearing latite andesites from Chile (Seigers *et al.*, 1969, cited in MacIntyre, 1985). Because of the relatively high K_2O content, both the microdiorite and the feldspar porphyry classify as latite-andesites or dacites by the scheme of Streckeisen (1967; *cf.* MacIntyre, 1985).

The microdiorite has been dated isotopically at 75.3 ± 2.0 Ma by K-Ar on whole rock (Church, 1973). The two main outcrop areas of the microdiorite correlate with the two main areas of mineralization, but this relationship may be only coincidental.

PORPHYRY (UNIT 5A)

Large bodies of a coarsely feldspar-porphyritic rock, up to 1000 metres across, crop out in the vicinity of Cole Creek and are also found in drill core from the south end of the Number 3 vein system, where the porphyry body usually occurs between Units 1 and 3. The rock is composed of roughly 50 per cent variably saussuritized or sericitized plagioclase phenocrysts of up to 5 millimetres in diameter and 10 to 20 per cent smaller altered mafic relics in a fine feldspathic groundmass. The porphyry is distinguished from the feldspar porphyry, Unit 4, by its coarser texture and by the absence of flow textures. It probably represents subvolcanic or high-level intrusive bodies that were emplaced below or postdate the extrusive feldspar porphyry, but are related to the same magmatic event that produced it. Such subvolcanic intrusive bodies, with identical mineralogy to the extrusive porphyritic andesites, have also been noted in the Kasalka Group near Tahtsa Lake (MacIntyre, 1985).

QUARTZ FELDSPAR PORPHYRY (UNIT 5B)

Quartz feldspar porphyry that appears to be part of a subvolcanic intrusive stock crops out along Emil Creek and on Okusyelda Hill to the north of the creek. This unit was formerly called "Okusyelda" dacite (rhyolite) by Church (1970). Although its contact relationships are uncertain, it appears to intrude Unit 4 (Tip Top Hill volcanics). Church (1984) correlates the quartz porphyry intrusions on Okusyelda Hill with acid volcanic rocks in the Tchesinkut Lake and Bulkley Lake areas, and possibly with the Tsalit Mountain rhyolite of 77.8 Ma (*see* under Unit 4). However, in the Kasalka Range, MacIntyre (1985) found sills and dikes of quartz-porphyritic dacite and rhyolitic quartz-eye porphyry, commonly associated with mineralization, that cut stocks dated at approximately 76 Ma (Carter, 1981). Henee, the quartz porphyry is considered to be younger than the microdiorite/feldspar porphyry in the Owen Lake area. It is cut by thick calcite veins and quartz-sericite-pyrite alteration on the extension of the George Lake vein (Figure 2-8-2) and so is probably pre-mineral.

Thin sections show the quartz porphyry consists of 10 to 15 per cent 2-millimetre quartz phenocrysts and slightly smaller euhedral andesine plagioclase crystals, plus smaller relic mafic grains, in a microgranular groundmass of roughly equal amounts of quartz, plagioclase and potash feldspar. Quartz, and to a lesser extent plagioclase, also occur as angular fragments or shards.

AMYGDALOIDAL DIKES (UNIT 6)

Units 1 to 5 are cut by a series of variably amygdaloidal dikes that are concentrated in the two main areas of mineralization (No. 3 vein and Cole vein areas). They generally trend northwesterly, parallel to the mineralized veins, but north, east and northeast-trending examples are known. Dips are either subvertical to steep, or else gentle (as low as 20°). These dikes are irregular and anastamosing in some parts of the property, for example between the Camp and Switchback vein systems. Strongly altered examples are commonly found adjacent to and parallel to veins; elsewhere veins cut through these dikes. These dikes have been referred to previously as "pulaskite" at both the Silver Queen and Equity deposits, but this is a highly inappropriate term, implying an alkali-rich mineralogy including soda orthoclase, alkali pyroxene or amphibole and feldspathoids, and it is avoided in this study.

In underground exposures the dikes range from dark grey-green where fresh, to pale green or creamy buff where strongly altered; they are purplish in weathered surface outcrops. They are typically fine grained and are characterized by amygdules filled by calcite or, less commonly, iron oxides, particularly at their chilled margins (dikes less than 2 metres wide may lack the amygdules). Flow orientations are generally parallel to the walls, and provide an indication of attitude in surface outcrops, but in the larger dikes (up to 10 metres thick) the flow orientations are random.

In thin section, the most striking feature of these dikes is the abundance of fine trachytic-textured feldspar microlites that average about 0.25 millimetre long. Alteration to carbonate and sericite is extensive, but the texture is generally preserved and links these dikes to the trachytic-textured dikes of Unit 7, which have similarities to Goosly Lake volcanics (*see* below).

BLADED FELDSPAR PORPHYRY DIKES (UNIT 7)

Trachytic-textured porphyry dikes, 1 to 5 metres wide and characterized by coarse (up to 1 centimetre long) bladed plagioclase phenocrysts, cut and slightly offset the amygdaloidal dikes. The complete lack of alteration in the bladed feldspar porphyry dikes, and the fact that they distinctly crosscut mineralized veins (*e.g.*, the Bear Vein, Cole Lake area: Figure 2-8-2), indicates that they postdate mineralization. Their spatial distribution is similar to that of the amygdaloidal dikes, with concentrations in the two main mineralized areas; orientations are also similar, with subvertical dips.

The similarity of these post-mineral bladed feldspar porphyries to the Goosly and Parrot Lake syenomonzonite stocks, and bladed feldspar andesite dikes at Equity dated at 50.7 ± 1.8 Ma by K-Ar on whole rock, suggest that they are probably of the same age. The pre-mineral amygdaloidal dikes, although considerably finer grained, also have similar characteristics (trachytic-textured feldspar), but their age is not yet established.

In thin section, the bladed feldspar porphyry dikes are seen to be composed of large (4 to 10 millimetres) plagioclase phenocrysts and rare to locally abundant clinopyroxene crystals up to 5 millimetre across, set in a dark purplish groundmass of feathery, interlocking plagioclase microlites with interstitial quartz, alkali feldspar, opaques and skeletal rutile(?). The plagioclase forms strongly zoned, oscillatory crystals that range from cores of andesine (An_{50}) to rims of oligoclase (An_{15}). The pyroxene has a strong green colour and is probably iron-rich.

If the dikes of Unit 7 are feeders for the Goosly Lake volcanics or related to the Goosly and Parrot Lake syenomonzonite as postulated, then they probably have similar trachyandesite compositions (*see* analyses 3, 4 and 6 of Church, 1971).

DIABASE DIKES (UNIT 8)

Black fine-grained dikes of probable basaltic composition cut all other units on the property. They are much more limited in distribution than the older dikes, with subvertical dips and northwest or east-west strikes. However, they still seem to be concentrated in areas of veining, and are subparallel to the veins: for example, where a vein strikes east, as in Emil Creek (Figure 2-8-2), a diabase dike has the same orientation.

It is likely that these dikes were feeders to a younger volcanic group such as the Endako Group of Eocene-Oligocene age (40 to 30 Ma), but the possibility cannot be ruled out that they are related to the Buck Creek volcanic unit (48 Ma). There is little possibility that they are related to the Miocene Poplar Buttes volcanic rocks (21 Ma), as they lack olivine. Thin sections show they are composed of diabasic-textured plagioclase in clinopyroxene, with accessory opaques.

STRUCTURE

The structure of the Silver Queen mine area is dominated by a gently north to northwest-dipping homocline. There is no folding apparent at the scale mapped; the sequence presumably has been tilted 20° to 30° from the horizontal by block faulting.

Two prominent sets of faults displace this homoclinal sequence, cutting it into a series of fault panels: a northwest-trending set and a northeast-trending set. The former predates or is contemporaneous with mineralization, whereas the latter is mainly post-mineral. Most of the mineralized veins and the dikes follow the northwest-trending faults, whereas veins are cut off and displaced by the northeast-trending set. The northwest-trending faults dip 60° to 80° to the northeast, and the northeast-trending set appears to be subvertical.

The sense of motion on the northwest-trending faults is such that each successive panel to the east is upthrown, leading to successively deeper levels of exposure to the east. Thus, in the panel between the George Lake and the Emil Lake faults (Figure 2-8-2), there is considerably more of the lower fragmental rocks (Units 2 and 3) exposed than in the next panel to the west, between the Owen Lake and the George Lake faults. There does not seem to be much displacement across the No. 3 vein fault; slickensides seen underground on this structure suggest a reverse sense of movement.

The sense of motion on the the northeast-trending faults appears to be south side down, with a small component of sinistral shear. Offsets of No. 1 and 2 veins across fault along Wrinch Creek (Figure 2-8-2) suggest a few metres of left-lateral displacement, but the displacement of an amygdaloidal dike near the portals of the 2880 level suggests the south side must have dropped as well. The boundaries of this fault zone, and its dip, are not well constrained; in outcrops in Wrinch Creek, it appears as a vaguely defined zone up to 10 metres wide, with segments that have possible shallow to moderate dips to the north. The Cole Creek fault is not well exposed at surface; a splay from it may cause the change in orientation of the No. 3 vein to the Ruby vein (Figure 2-8-2). A considerable left-lateral offset of perhaps as much as 200 metres is suggested by drill-hole intersections of the NG3 vein, which may be a faulted extension of the No. 3 vein south of the Cole Creek fault. Underground, this fault is exposed at the southernmost extent of drifting as a gouge zone 1 to 2 metres thick (Plate 2-8-1). Other examples of minor northeast-trending faults are seen underground.

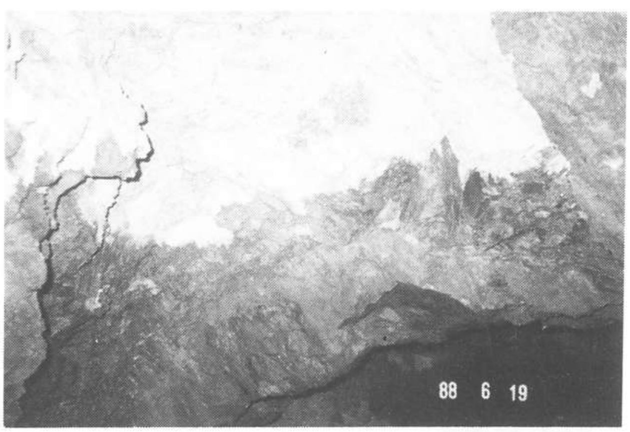

Plate 2-8-1. Late northeast-trending fault cutting the vein (here trending along the adit). Two-metre gouge zone (top of photo), slickensided face (bottom).

DISCUSSION

The sequence of rocks exposed in the Silver Queen mine area, mapped as Tip Top Hill formation (Church, 1984) is petrographically and stratigraphically similar to the Kasalka Group as defined in the Tahtsa Lake area by MacIntyre (1985) and the Mount Cronin area by MacIntyre and Desjardins (1988). The section in all three areas, comprises a sequence from a basal, reddish purple heterolithic conglomerate, upwards through a sequence of fragmental volcanic rocks, to a widespread, partly intrusive porphyritic andesite, all intruded by a distinctive microdiorite. It is similar to type sections of the Kasalka Group.

Potassium-argon isotopic dating suggests that the rocks in the Silver Queen mine area are of Late Cretaceous age; the porphyritic andesite volcanics are about 77 Ma, and are intruded by microdiorite of 75 Ma age (Church, 1973). This is younger than the Kasalka Group rocks in the type section near Tahtsa Lake, which give dates of 108 to 107 Ma near the base, and are cut by intrusions dated at 87 to 84 Ma (MacIntyre, 1985). These dates actually straddle the Early to Late Cretaceous boundary (Harland *et al.*, 1989). Thus, in spite of the similarities in lithology between the Silver Queen mine area and the Kasalka Group rocks elsewhere, a correlation is not supported by the available isotopic dating. Possibly the magmatic front associated with this Late Cretaceous volcanic activity took longer to arrive further inland [65 kilometres in 30 Ma gives a rate of advance of 0.22 centimetres per year, comparable to the rate of 0.25 centimetres annually suggested by Godwin (1975); *cf.* Armstrong (1988) and Leitch (1989)].

Mineralization in epithermal veins at the Silver Queen mine occurred between the time of deposition of the Late Cretaceous Tip Top Hill volcanics and intrusion of Early Tertiary dikes. The latter may correlate to the Goosly Lake trachyandesite volcanics (49 Ma) of the Ootsa Lake Group and syenomonzonite stocks (50 Ma) found at Equity Silver mine and Parrot Lakes (Church, 1973). The veins are also cut by diabase dikes that may correlate with the Buck Creek volcanics, dated at 48 Ma (Church, 1973), or to Endako Group volcanics dated at 40 to 30 Ma (Diakow and Koyanagi, 1988). Additional radiometric dating is in progress, including U-Pb on zircon from the quartz-eye rhyolite and K-Ar (whole rock and biotite separates) from other major units.

ACKNOWLEDGMENTS

Work on the Owen Lake project was supported by Pacific Houston Resources Inc. and by a cooperative research grant from the Natural Science and Engineering Research Council. Dr. D.G. MacIntyre of the British Columbia Ministry of Energy, Mines, and Petroleum Resources provided insight into regional correlation problems. Pacific Houston Resources expedited our field study; we particularly thank Mr. W. Cummings and Mr. J. Hutter for their assistance during fieldwork, and Mr. A. Petancic for his continuing interest and support.

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NOTES

GEOLOGY AND DESCRIPTIVE PETROLOGY OF THE MOUNT BISSON ALKALINE COMPLEX, MUNROE CREEK, BRITISH COLUMBIA* (93N/9E, 93O/12W, 5W)

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KEYWORDS: Economic geology, petrology, Mount Bisson, alkaline complex, Wolverine metamorphic suite, rare earth elements, syenite, pegmatite, alkali metasomatism.

INTRODUCTION

The Mount Bisson alkaline complex is located in the Wolverine Range, 64 kilometres northwest of the town of Mackenzie, British Columbia. The area studied during the 1988 and 1989 field seasons is covered by the southeastern corner of the Manson Creek map sheet (93N/9) and the west margins of map sheets 93O/12 and 93O/5 (Figure 2-9-1). Other alkaline complexes in the region (Figure 2-9-1A) include: the Lonnie (Halleran, 1980; Pell, 1987), Veril (Pell, 1987) and Aley carbonatite (Mäder, 1986, 1987) as well as the Prince and George carbonatites (Mäder and Greenwood, 1988). The Mount Bisson complex is unique in that there are no known associated carbonatites; it comprises mainly silica-saturated lithologies and consequently the host minerals to the rare earth element mineralization are dominantly silicates (allanite, cerorthite).

The objectives of this paper are: to summarize the field relationships of the alkalic rocks exposed in the Wolverine Range, to describe their petrology, and to describe the continuing research on their mineralogy and mineral chemistry.

PREVIOUS WORK

McConnell (1896) first described the lithologies of the Wolverine complex and proposed a geological boundary north of Manson Creek village between the Cache Creek Group and older rocks. This boundary remains virtually unchanged (Ferri and Melville, 1988). Later work includes: Dolmage's (1927) mapping of the Finlay River district north of Manson Creek, and Armstrong's (1949) work which provided the first petrologic data and age correlations for the Wolverine suite. Muller (1961) and Tipper *et al.* (1974) mapped the southernmost part of the Wolverine metamorphic suite and published age dates. More recently, geological mapping of the Manson Creek map area by Ferri and Melville (1988) has contributed to the understanding of the Mount Bisson lithologies by strengthening the regional geological framework.

Mineral exploration near Mount Bisson began with the discovery of graphite in carbonate units within the Wolverine rocks (Halleran, 1985). Rare earth mineralization was discovered in 1986 and 1987 (mineral occurrences on the Ursa, Will and Laura claims). Chevron Minerals Ltd. conducted a limited exploration program for rare earth elements in the alkalic rocks in 1988 (Halleran, 1988).

REGIONAL GEOLOGY

The Mount Bisson alkaline complex occurs within a part of the Omineca crystalline belt termed the Wolverine metamorphic suite. The rocks comprising the metamorphic suite are inferred to be Proterozoic in age, although K-Ar age determinations for the metamorphic rocks range from 69 to 43 Ma. (Tipper *et al.*, 1974). Ferri and Melville (1988) divided the Omineca crystalline belt into, (1) the Wolverine suite: comprising intensely metamorphosed and deformed high-grade calcsilicate, amphibolite and granitic gneisses, which are intruded by later felsic intrusions and, (2) relatively unmetamorphosed quartzite, argillaceous quartzite and schists to the west (Ingenika Group).

The Mount Bisson complex comprises a group of diverse rock types with common mineralogical characteristics, including:

- Modal primary quartz is rarely present in the alkalic rocks.
- Modal sodic-bearing ferromagnesian minerals (*e.g.* aegirine-augite) are abundant in all alkaline rock types.
- Modal sphene and/or rutile are common to all alkaline rocks.
- Rare earth elements are abundant in several alkalic lithologies and are a major component of the allanite pegmatites.

The rocks of the alkaline complex include small intrusions (*e.g.* syenite, monzonite), mappable alkalic dikes (*e.g.* syenite), pegmatite dikes, and metamorphic rocks of the Wolverine suite characterized by a strong alkalic overprint. At Mount Bisson, these alkalic rocks are exposed at five localities over a strike length of 10 kilometres (Figure 2-9-1), they coincide with a regional aeromagnetic anomaly, and contain rare earth element minerals. Unfoliated, fine-grained quartz monzonite to quartz syenite intrusions occur throughout the region. Mapping is incomplete but there are at least four large intrusions (1 by 3 kilometres in area) and numerous smaller satelite bodies. The relationship between these intrusions and the alkalic rocks is unclear.

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

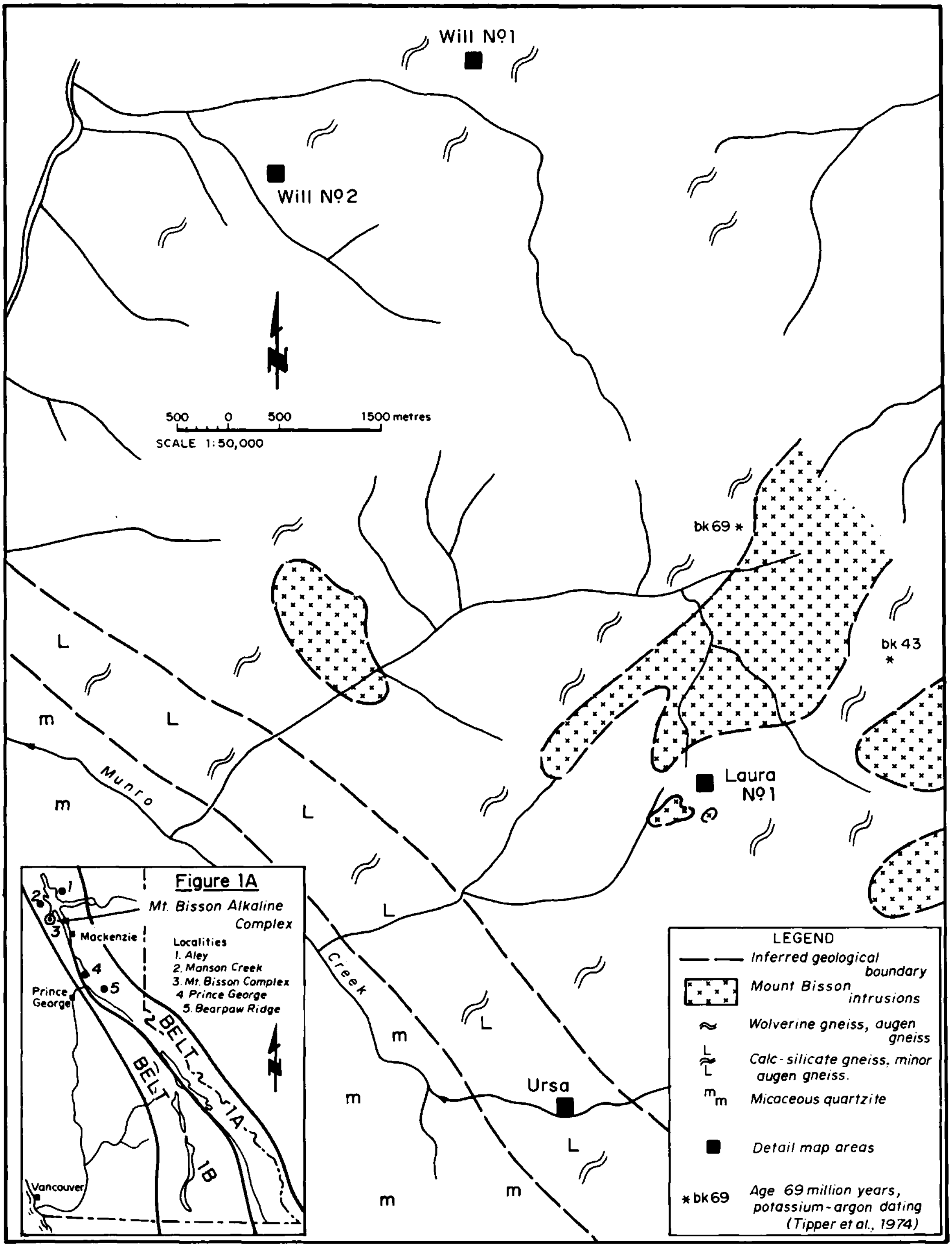

Figure 2-9-1. Geological map of the Mount Bisson area. Four occurrences of alkaline rocks include the Laura, Will (No. 1 and 2) and Ursa properties. Inset figure (1A) illustrates the location of the Mount Bisson alkaline rocks with respect to other alkaline rocks in the area. Belt 1A and Belt 1B are defined by their tectonic history and alkaline rock types (Pell, 1987). Belt 1A has subcircular to elliptical alkaline intrusions with extensive metasomatic alteration halos and which are hosted in Middle Cambrian to Middle Devonian rocks. Belt 1B alkaline intrusions are foliated, sill-like bodies, strongly deformed, metamorphosed to amphibolite facies, and hosted in Late Precambrian to Early Cambrian metasedimentary rocks.

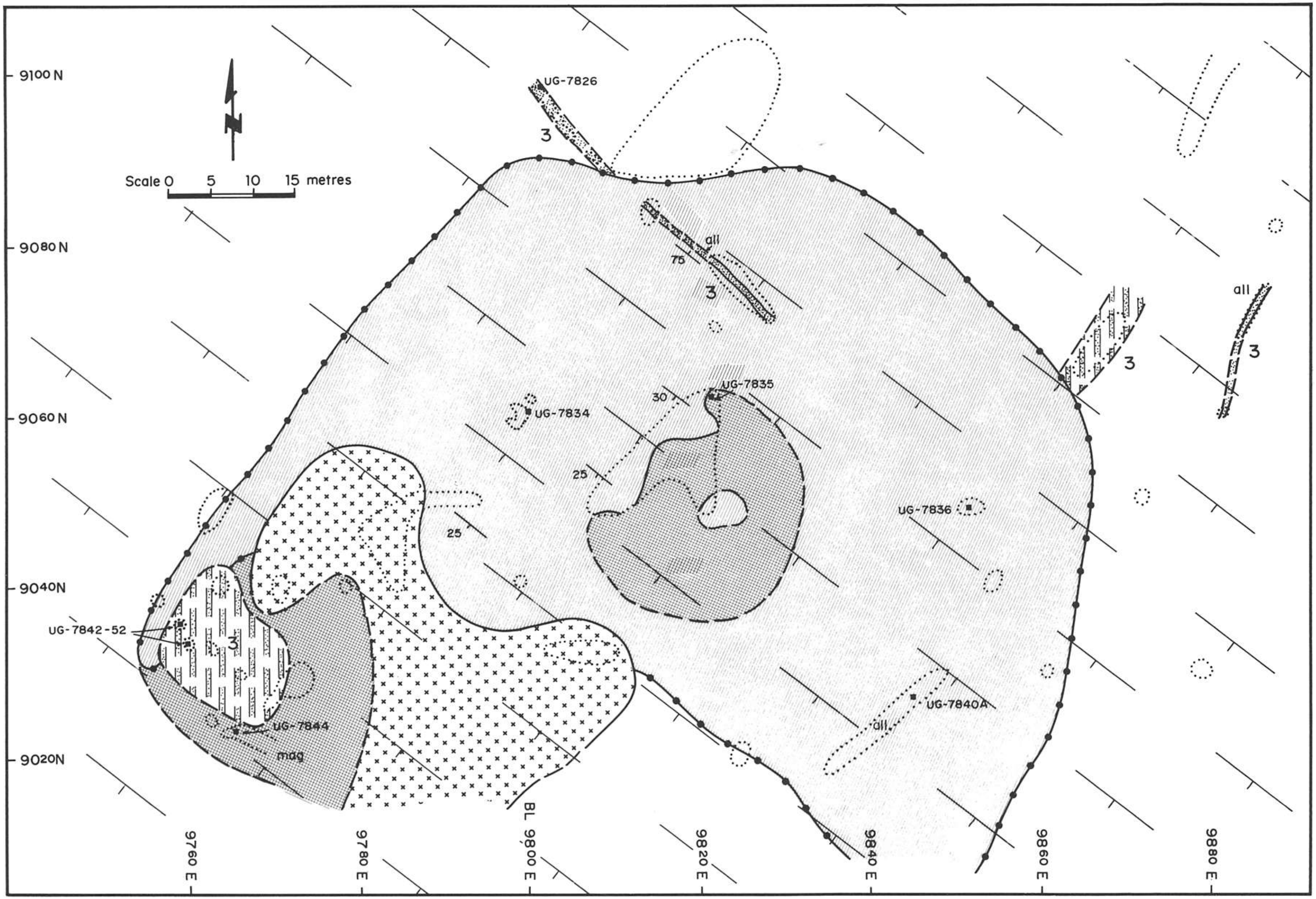

Figure 2-9-2. Detailed geological map for the Laura property. Lithologic and symbol legends are given in Figure 2-9-2A opposite.

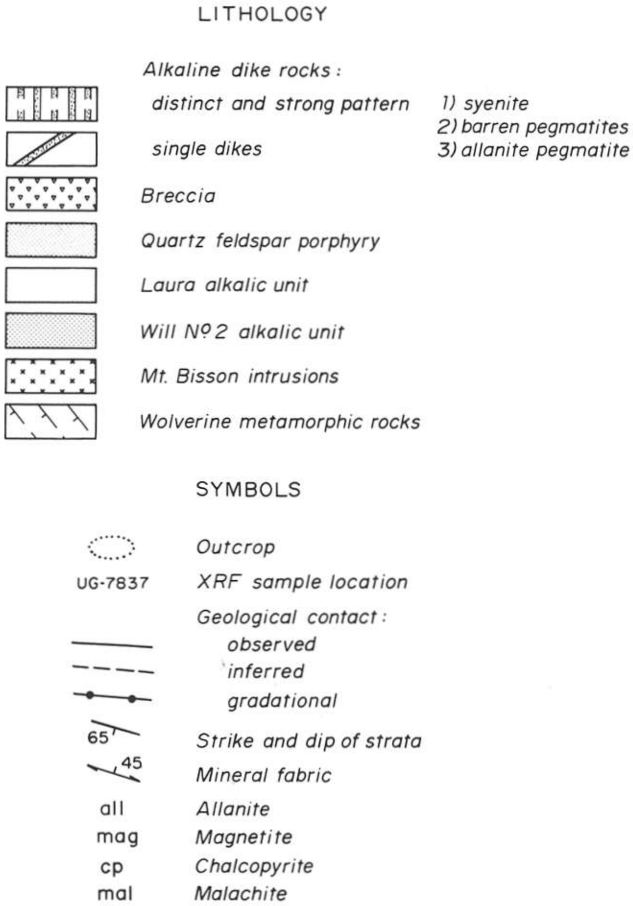

Figure 2-9-2a. Lithologic and symbol legends for Figures 2-9-2, 3 and 4.

LOCAL GEOLOGY

Four of the five localities have been mapped in detail (Halleran, 1988) and are summarized below. The fifth locality (Laura No. 2) has geology similar to that described for Laura No. 1.

LAURA NO. 1

Figure 2-9-2 is a detailed geologic map of the Laura No. 1 locality. The Laura No. 1 map area covers a series of intrusions which cut the metamorphic rocks of the Wolverine suite. The Wolverine metamorphic rocks include coarse-grained amphibolite, biotite schist and strongly foliated quartzofeldspathic gneiss. Locally, the gneisses are characterized by a metasomatic alkalic signature, here mapped as the Laura alkalic unit.

There are a variety of intrusive rock types which crosscut the structure of the Wolverine metamorphic rocks. Allanite pegmatites, 1 to 4 metres wide, are common and are enriched in rare earth elements. They have a minimum strike length of over 30 metres. Late quartz veinlets, 50 millimetres wide, cut the pegmatite. Mount Bisson rocks intrude the Wolverine suite and cut the metasomatic alkalic overprinting. No field relationships between the Mount Bisson intrusions and the pegmatite were observed.

The Laura alkalic unit (Figure 2-9-2) is a distinctive lithology comprising Wolverine metamorphic rocks, which have an alkaline character expressed by the presence of aegirine-augite and/or sphene, allanite and alkali feldspar. The map unit is massive, fine to medium grained and retains a strong fabric related to the original metamorphic fabric of the Wolverine suite. Specifically, the alkalic overprinted rocks are commonly banded on a millimetre to centimetre scale. Dark bands comprise aegirine-augite, hornblende, sphene and allanite, whereas more felsic bands are dominated by alkalic feldspar. This unit has a circular map pattern with a minimum diameter of 60 metres. The contact between the alkaline overprinting and the host Wolverine gneisses is gradational and some metamorphic lithologies (*e.g.* amphibolites) are more intensely metasomatized than others (*e.g.*, quartzofeldspathic gneisses). Furthermore, the replacement process commonly preserves the older regional structure. These observations suggest that the alkalic character is derived through preferential replacement of amphibolite gneisses of the Wolverine suite. The metasomatism may be related to a large, yet undefined, deep seated intrusion.

WILL NO. 1

The Will No. 1 map area is illustrated in Figure 2-9-3. There is little exposure and consequently the field relationships between lithologies are uncertain. The geology includes two separate aegirine-augite syenite dikes, which crosscut Wolverine gneisses and biotite schists, and several outcrops of Mount Bisson intrusion. The remaining lithology is a breccia with intrusive clasts supported by a green fine-

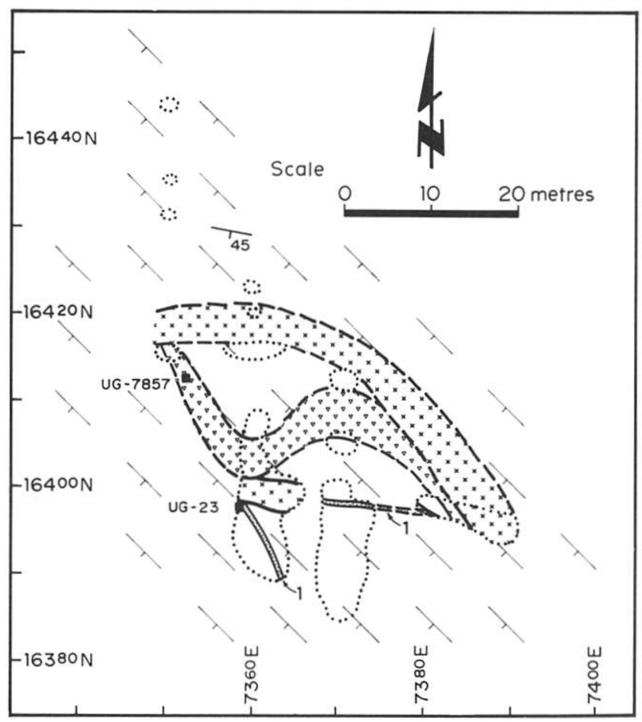

Figure 2-9-3. Detailed geological map for the Will No. 1 locality. Lithologic and symbol legends are given in Figure 2-9-2a.

grained matrix. In thin section the matrix comprises relic potassium feldspar (25 volume per cent), plagioclase (10 volume per cent), altered light yellow-green to blue-green pleochroic amphibole, trace sphene and apatite.

WILL NO. 2

The geology of the the Will No. 2 map area is shown in Figure 2-9-4. The main lithologies include: fine-grained Mount Bisson intrusions, a sequence of metasomatized Wolverine schists and gneisses (Will No. 2 alkalic unit), pegmatites, and late crosscutting alkaline dikes.

The Will No. 2 alkalic unit varies from a fine-grained, light-colored rock with a weakly developed mineral fabric to a darker, biotite-rich schist with millimetre to centimetre banding. The alkaline overprinting is differentiated from the surrounding Wolverine rocks by the presence of aegirine-augite and rare earth element bearing minerals, an increase in alkali feldspar content and a decrease in quartz content. Original metamorphic banding has been enhanced by increases in concentration of mafic and rare earth bearing minerals associated with metasomatic replacement. The alkaline overprinting is cut by syenite and barren alkali pegmatite dikes. No contact between the fine-grained Mount

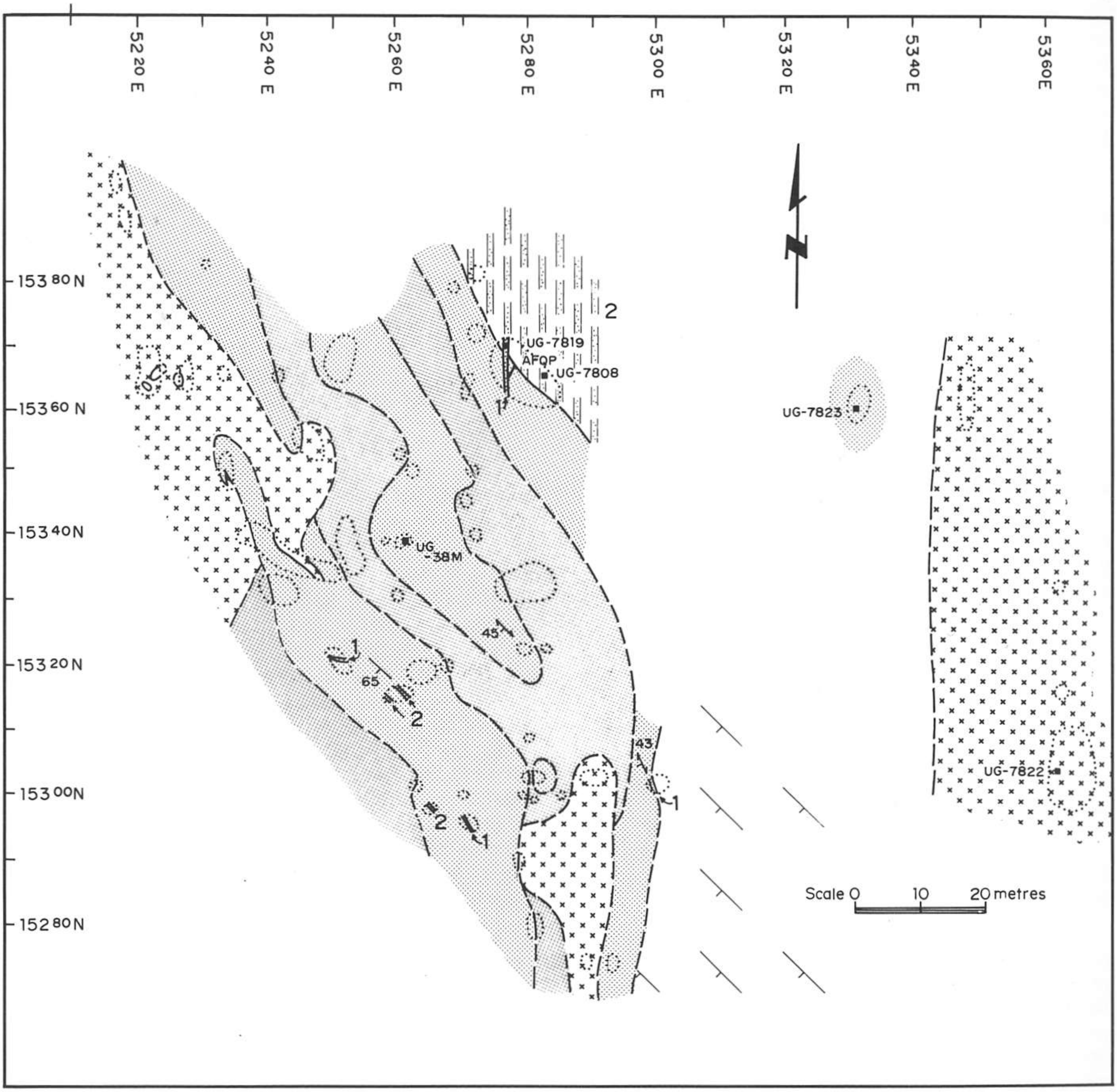

Figure 2-9-4. Detailed geological map for the Will No. 2 property. Lithologic and symbol legends are given in Figure 2-9-2a.

Bisson intrusions and the Will No. 2 alkalic unit was observed; however, angular xenoliths of metasomatized Wolverine rocks rarely occur within the intrusions.

Syenite dikes comprise two mineralogically distinct rock types: an aegirine-augite-rich (>70 volume per cent) dike and a fine-grained equigranular alkali feldspar (>90 volume per cent) dike which crosscut both the alkalic unit and pegmatites.

URSA

The Ursa showing (Figure 2-9-1) is a mylonitized, gneissic pegmatite 10 metres long and 1 to 2 metres wide. The rock comprises potassium feldspar, quartz, albite, monazite and traces of biotite and sphene. The biotite is partly altered to chlorite. This unit occurs within fine-grained phlogopite-bearing calcsilicate gneisses of the Wolverine suite. To the west, the pegmatite is truncated by a fine-grained felsic intrusion.

DESCRIPTIVE PETROLOGY

ALKALIC DIKE ROCKS

SYENITE

Three types of syenite dike occur within the Mount Bisson alkaline complex: alkali feldspar rich dikes (on the Will No. 2); aegirine-augite rich dikes (on the Will No. 1 and No. 2) and rare earth element enriched dikes (on the Will No. 2).

The alkali feldspar dikes contain 90 per cent potassium feldspar rimmed by plagioclase and only 10 per cent disseminated mafic minerals, principally aegirine-augite.

The aegirine-augite dike consists of 40 to 60 per cent, inclusion-filled aegirine-augite 1 to 15 millimetres long; 35 per cent perthite; 3 per cent sphene with rare allanite inclusions; 1 per cent euhedral apatite; and traces of magnetite, chalcopyrite, malachite and allanite. The mafic phases are concentrated in millimetre-wide bands.

The rare earth enriched dikes consist of 80 per cent intergrown, inclusion-filled aegirine-augite; 8 per cent potassium feldspar, 5 per cent apatite, 3 per cent allanite, 2 per cent sphene with traces of calcite and biotite (as inclusions in aegirine-augite). The rare earth elements reside in the allanite which occurs as intergrowths with sphene and aegirine-augite.

PEGMATITES BARREN OF RARE EARTH ELEMENTS

Quartz-feldspar pegmatites and aegirine-augite pegmatites (Figures 2-9-2, 2-9-4) are both characterized by the absence of rare earth elements. The contact relationships between them are not known; however, while the rock types commonly outcrop together, no lithologic contacts have been observed. Large xenoliths of Wolverine amphibolites commonly occur within the pegmatite bodies.

The quartz-feldspar pegmatite consists of 5 to 10-millimetre polycrystalline quartz, potassium feldspar, plagioclase (An_{27}) and trace to minor magnetite, biotite, chlorite, zircon, monazite and opaques. Rare, euhedral, zoned monazite is the only rare earth bearing mineral in this lithology.

The coarse grained aegirine-augite pegmatite comprises zoned, antiperthite (An_{23}), aegirine-augite, minor perthitic potassium feldspar, occasional elongate quartz crystals and late fracture-filling epidote. Subhedral aegirine-augite is broken and reorientated and includes plagioclase (An_{32}), euhedral sphene, hornblende and biotite.

PEGMATITES BEARING RARE EARTH ELEMENTS

The allanite pegmatites (Figure 2-9-2) are the only pegmatites with rare earth concentrations above geochemical detection limits; in several samples the allanite is so abundant that it becomes a rock-forming mineral rather than an accessory phase.

The mineralogy consists of perthite, up to 35 per cent allanite, 5 per cent sphene, plagioclase (An_{25} to An_{27}), apatite, minor to trace aegirine-augite, polycrystalline quartz, traces of pink pleochroic zircon and opaques. Subhedral to euhedral, zoned, green to brown-pleochroic, 0.3 to 20 millimetre allanite occurs with sphene and euhedral, grey-pleochroic apatite, sometimes as intergrowths. Sphene can be divided into two types: anhedral intergrowths with allanite and apatite, and euhedral crystals up to 1 centimetre long, found within the allanite mineralized zones. The latter also occur as smaller individual subhedral crystals in the feldspar-rich phases.

SECONDARY ALKALIC ROCKS

The Will No. 2 and Laura alkalic units (Figures 2-9-2 and 3) are secondary alkalic rocks and are identical except that: Will No. 2 variety has predominantly plagioclase in the felsic rock types whereas the Laura alkalic unit has potassium feldspar. Additionally, the Laura alkalic unit has better developed mineral banding.

These alkalic units are fine to medium-grained crystalline rocks with regular banding on a millimetre to centimetre scale. Commonly the original Wolverine fabric is still evident. The dark bands comprise aegirine-augite, sphene, allanite and apatite with minor hornblende (in the Laura alkalic unit). Light bands are dominanted by plagioclase (Will No. 2) or potassium feldspar (Laura unit). Euhedral sphene, apatite and allanite occur together, at times as aggregates, in the mafic bands. Minor apatite grains are also found in the felsic bands. Rare polycrystalline quartz occurs as interstitial filling.

MOUNT BISSON INTRUSIONS

This unit includes all of the fine-grained, light-colored, massive, fresh-looking aplitic intrusions; many of which have been examined only in handspecimen. The constituent minerals include quartz, plagioclase (An_{22} to An_{24}), potassium feldspar, biotite, chlorite, traces of magnetite, allanite, apatite and zircon. Plagioclase occurs as cloudy microphenocrysts and as smaller grains which exhibit undulatory extinction. Biotite is locally altered to chlorite and in some satellite bodies (*e.g.* Will No. 2) defines the rock fabric.

ECONOMIC GEOLOGY

The primary economic significance of the Mount Bisson alkaline complex is the occurrence of the light rare earth elements, cerium, lanthanum, neodymium, samarium, and praseodymiun in the minerals allanite, cerorthite, and monazite. The minerals are often very coarse (5 to 10 millimetres). Heavy rare earth elements are found in concentrations of hundreds of parts per million. Niobium is found at concentrations as high as 0.8 per cent. At this time two major rare earth deposit-types have been found within the complex: pegmatites, 1 to 4 metres wide and over 30 metres long, enriched in rare earth elements (combined 0.3 to 14.0 per cent, Table 2-9-1) and the Will No. 2 and Laura alkalic units with concentrations of 0.07 to 0.64 per cent light rare earth metals over widths of 1 to 2 metres and tens of metres in area (Table 2-9-1). In addition, rare earths have been found in a syenite dike (0.80 to 4.26 per cent, Table 2-9-1) and a mylonized gneissic pegmatite (2.1 per cent, Table 2-9-1).

The alkali syenite environments are characterized by the highest overall light rare earth content. Independent light rare earth element minerals are formed more frequently here than in other igneous rocks. This is very important for recovery of the rare earths as some of them are contained in the apatite, zircon, pyroxenes and other rock building minerals.

TABLE 2-9-1
RARE EARTH ANALYSIS SAMPLING DETAILS

Unit	Sample type	Per cent REM*	Description
AP	representative	0.30 to 0.64	samples of main pegmatite bodies (5 samples)
AP	representative	5.50 to 14.50	samples of mineralized zones in pegmatites (3)
SA	representative	0.14 to 0.55	numerous grab samples (6)
SA	Chip	0.13 to 0.64	Chip samples across 1.6 metres and 1.0 metre (2)
RED	representative	0.80 to 4.26	a sample of the main body and of the mineralization
URSA	representative	2.10	a sample of the monazite mineralization

AP = allanite pegmatite
SA = secondary alkalic unit (Will No. 2 and Laura)
RED = rare earth enriched dike
URSA = mylonitized pegmatite with monazite
REM* = cerium, lanthanium, neodymium, samarium and praseodymium
Preliminary data from Halleran (1988). Analysis by ACME using ICP.

CONCLUSION

The alkalic rocks on Mount Bisson include primary crosscutting dikes, pegmatites, and a secondary metasomatic replacement of Wolverine amphibolite gneisses. Where the original Wolverine gneisses have been been metasomatized, the alkalic overprinting is recognized and made a mappable unit by an increase of aegirine-augite, sphene, allanite, apatite and feldspar, and a decrease in quartz, hornblende and biotite. This secondary alkalic overprinting may represent a preferential replacement process associated with a deep seated, unexposed alkalic intrusion.

The youngest igneous events are represented by a single alkali feldspar syenite dike (included in the Alkaline Dike Rocks, No. 1) and the Mount Bisson intrusions which clearly intrude the secondary alkali units and crosscut the other dikes. In general, however, the temporal relationships of the dikes and pegmatites to the secondary alkalic unit are unclear. The emplacement of the dikes may have occurred before or after the metasomatic alteration, or been in part responsible for the alkalic replacement. The rotated amphibolite xenoliths in the pegmatites indicate magmatic injection into a more or less solid country rock.

The pegmatites are unique in that coarse allanite is a major constituent associated with high concentrations of light rare earth elements. Otherwise they contain similar rare earth element bearing minerals to the secondary alkalic units.

Major, minor, trace and rare earth element concentrations are being determined for 25 representive rock samples. Future work will involve electron microprobe studies of all phases to determine the nature of rare earth element zonation. This information will be used to interpret the crystallization history of the alkalic bodies, to characterize the rare earth element mineralization process, and develop a metallogenic model for the deposits.

ACKNOWLEDGMENTS

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STRATIGRAPHIC AND STRUCTURAL SETTING OF THE SHASTA Ag-Au DEPOSIT (94E)

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KEYWORDS: Economic geology, Shasta Au-Ag deposit, Toodoggone volcanics, Jock Creek volcanics, dacite dome, structure, alteration, exploration model.

INTRODUCTION

The Shasta deposit is in north-central British Columbia, approximately 300 kilometres north of Smithers (Figure 2-10-1). The property is reached by a four-wheel-drive road from the Sturdee airstrip, 9 kilometres to the west or from Fort St. James via the Omineca mining road and the Cheni mine road.

The objective of this study is to define the stratigraphic and structural setting of the deposit. Toward this goal 1:10 000 mapping of a 15 square kilometre area around Jock Creek and detailed (1:1000) mapping of the mineralized area were carried out in 1988 and 1989. Marsden and Moore (1989) give an account of the geology outlined by mapping during the 1988 field season. This report presents some additions and revisions to that stratigraphy, integrates field observations with petrographic work and literature research done during the winter of 1988-89, and provides a model for the setting of the Shasta deposit. This model can be used to guide exploration on the property and at a regional scale in exploring for similar geologic environments elsewhere in the Toodoggone district.

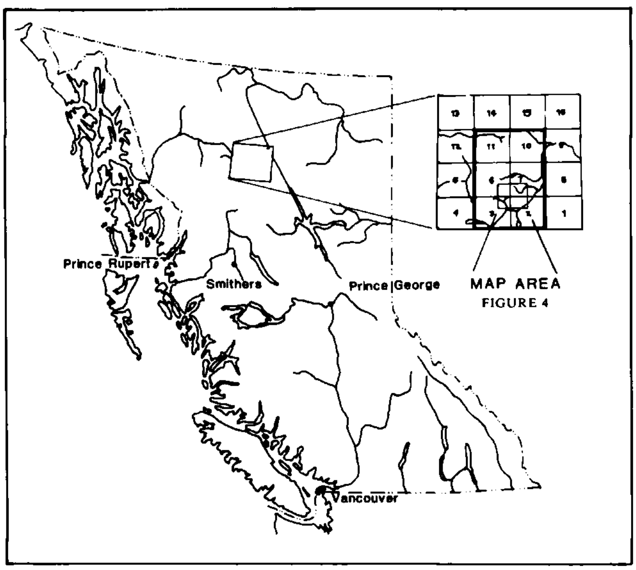

Figure 2-10-1. Location map of the study area.

HISTORY

The Shasta property was discovered in 1972 and has been explored intermittently since 1983. Esso Minerals Canada worked on the property in 1987 and 1988 and Homestake Mineral Development Company Ltd. is currently active on the property after acquiring the interests of Esso in the spring of 1989. The 1987 and 1988 exploration programs (Holbek and Thiersch, 1988; Holbek, 1989) included geological mapping, geochemical and geophysical surveys, trenching and over 13 000 metres of diamond drilling. By the end of 1988, Esso had outlined possible geologic reserves of 537 000 tonnes of 8.7 grams gold equivalent per tonne (70 grams Ag = 1 gram Au) or 1.02 million tonnes of 5.7 grams gold equivalent per tonne (Holbek, 1989).

The current program is a joint venture between Homestake and International Shasta Resources Ltd. that is operated by Homestake. International Shasta is also independently running a small open-pit operation on the property and plans to mine 100 000 tonnes of ore from the JM and Creek zones.

REGIONAL GEOLOGY

The project area lies within the Stikine Terrane along the eastern margin of the Intermontane Belt of the Canadian Cordillera. Stikinia is an allochthonous assemblage of Paleozoic to Jurassic island arc volcanics and associated basinal sediments. The oldest rocks in the terrane, the Stikine assemblage, are Paleozoic mafic volcanics, marine sediments and Permian limestones. These are overlain by tholeiitic arc rocks of the Stuhini Group and Lower to Middle Jurassic arc rocks of the Hazelton Group. This accretionary collection of arc-related rocks is overlain by post-accretion sediments of the Middle to Upper Jurassic Bowser assemblage and the Cretaceous Sustut Group.

The Hazelton Group is exposed around the perimeter of the Bowser basin (Figure 2-10-2). Tipper and Richards (1976) defined it as an island arc sequence deposited in the Hazelton trough between Sinemurian and Callovian time.

Hazelton Group volcanics in the Toodoggone River area are traditionally divided into a western felsic facies called the Toodoggone volcanics (informally named by Carter, 1972)

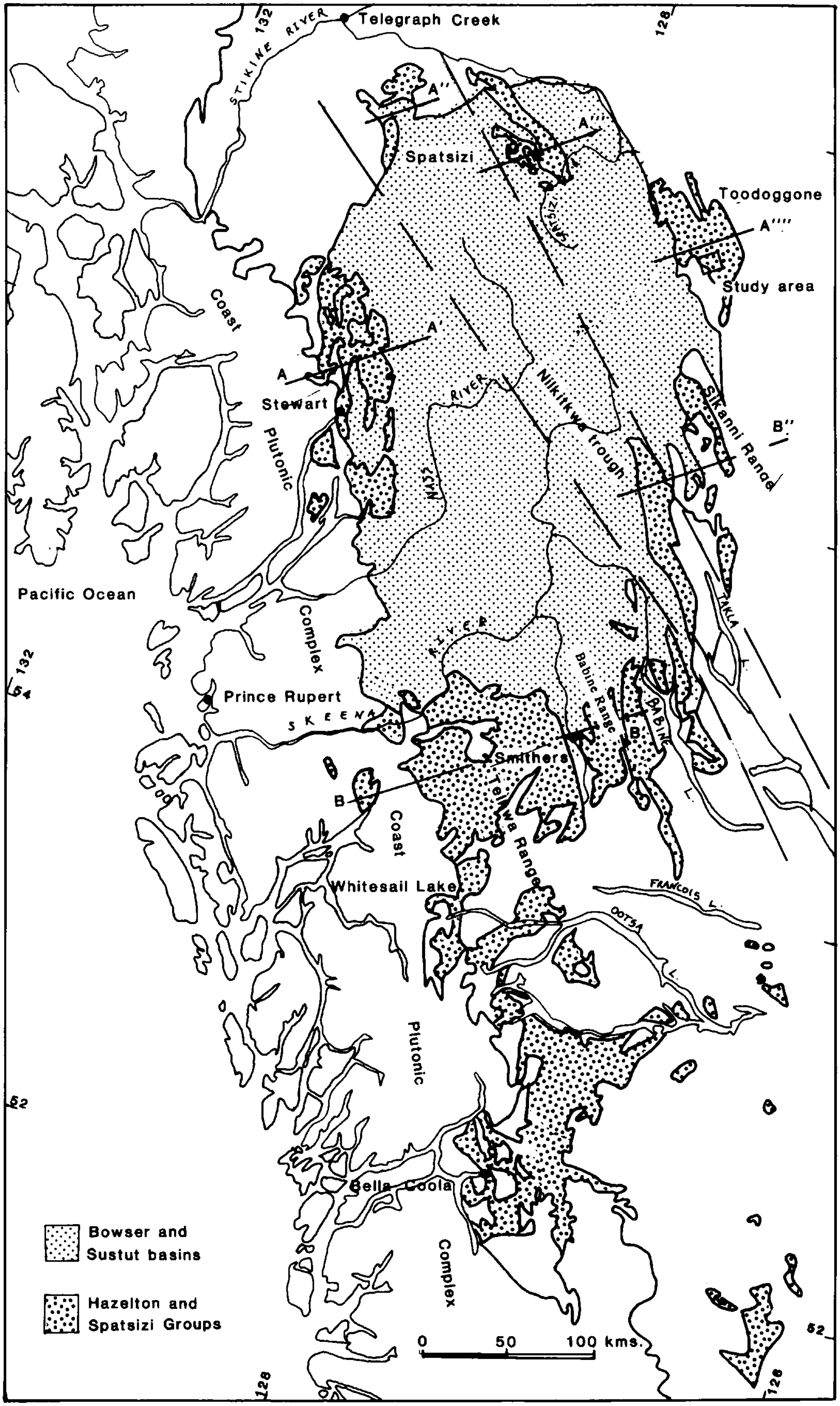

Figure 2-10-2. Distribution of the Hazelton and Spatsizi groups in the northern Intermontane Belt. Rectangle indicates study area.

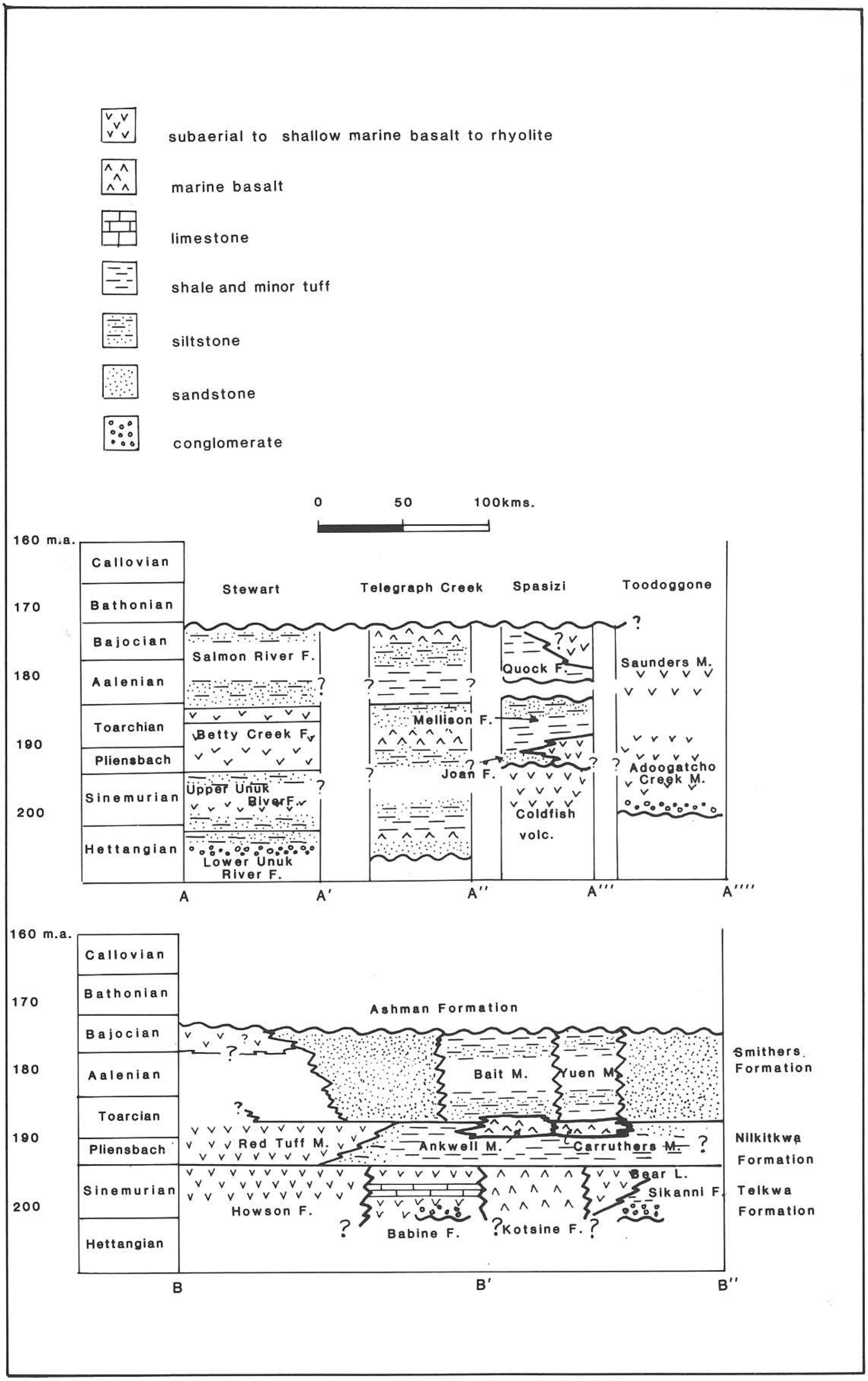

Figure 2-10-3. Time-stratigraphic sections across Figure 2-10-2.

and an eastern facies of predominantly intermediate volcanics. Detailed mapping by Daikow *et al.* (1985) has defined the Toodoggone volcanics as a predominantly calc-alkaline andesitic to dacitic subaerial succession that ranges from Toarcian to Aalenian in age.

A pattern has evolved from the mapping in the Hazelton Group during the last decade. The stratigraphic relationships of the various formations and facies of the Hazelton Group are summarized in Figure 2-10-3, two schematic time-stratigraphic sections across the Jurassic section of the northern Intermontane Belt. Two predominantly subaerial volcanic chains, exposed in the Telkwa-Stewart area and the Toodoggone-Coldfish area are separated by a marine basin 150 kilometres wide. The two volcanic chains may represent a true island arc (the Telkwa-Stewart chain) a marine back-arc (the Nilkitwa trough) and a continental back-arc (the Toodoggone-Coldfish volcanics) similar to the continental back-arc system currently active in New Zealand (Stern, 1985). There is not yet a sufficient database to determine if the two volcanic chains are geochemically distinct and representative of different tectonic environments. De Rosen-Spence and Sinclair (1988) used a very limited collection of data from several sources to suggest that the Telkwa volcanics are calcalkaline and more iron-rich than the alkaline to calcalkaline volcanics of the Nilkitwa trough and Toodoggone-Coldfish arc.

VOLCANIC STRATIGRAPHY

This study concentrates on a small area within the Toodoggone volcanics. Daikow (personal communication, 1988) recognizes three main members within the Toodoggone succession. The Adoogacho Creek member, the oldest, 201 to 204 Ma, is comprised of explosive dacitic volcanics. The middle member, the Metsantan member, is exposed only in the central part of the volcanic belt and is characterized by andesitic to basaltic stratovolcanoes dated at 197 to 200 Ma. The Saunders member is a 182 to 183 Ma massive section of dacitic ash-flow deposits exposed throughout the southern part of the Toodoggone belt. In this paper we propose that there is a fourth member in the Toodoggone area, exposed only to the east of the Saunders fault, that is younger (Bajocian?) than the Saunders member and is, at least in part, correlative with the undivided Hazelton volocanics exposed immediately to the east of the Toodoggone volcanics.

The Toodoggone volcanics are cut by at least two major right-lateral faults. Daikow *et al.* (1985) have documented the Saunders fault, a major feature of the current study area, and Vulimiri *et al.* (1983) have shown that the Attorney fault, exposed in the Cheni mine, is a right-lateral fault that can be traced to the north and south of the minesite.

The study area is transected by the Saunders fault (Figure 2-10-4). The Shasta deposit (Figure 2-10-5) is located on the west side of the fault and the stratigraphy exposed in this area is not the same as that to the east of the fault. This is partly a function of level of exposure; higher stratigraphic levels are exposed on the east side of the fault. West of the Saunders fault, the stratigraphy can be divided into three major parts: the Stuhini Group (Late Triassic), the Jock Creek volcanics (informally named in this paper) and the Saunders grey dacite, dated in this map area at 182 ± 8 Ma. The Jock Creek volcanics are not correlative with any of the main members described above and are correlated with rocks exposed south of the Finlay River (Daikow *et al.*, 1985) that gave a K-Ar date of 189 ± 6 Ma. Very limited observations by the authors at the Cheni mine indicate that the host volcanics in that area are mineralogically similar to the Jock Creek volcanics and occupy a similar position in the stratigraphy. The stratigraphy on the west side of the Saunders fault was described in detail by Marsden and Moore (1989); some revisions and additions to that stratigraphy are outlined below.

The oldest rocks exposed within the map area are mafic volcanics and associated sediments of the Upper Triassic Stuhini Group (Unit 1). The section is dominated by submarine basalts with well-developed pillows and hyaloclastite. The volcanic rocks are overlain by thin to thick-bedded volcanic sediments. The fine-grained sediments are well bedded and commonly graded, with small scale crossbedding and loading structures. These sedimentary rocks were probably deposited from debris flows and associated turbidity currents in a marine basin adjacent to a volcanic edifice. These marine deposits are locally overlain by subaerial or very shallow marine mafic flows. These amygdaloidal augite-feldspar-phyric flows have massive chilled bases and reddened, scoriaeeous flow tops. Overlying and lateral to the flows is a sequence of coarse purple and green epiclastic deposits (Unit 2). The base of this unit is always marked by several metres of coarse conglomerate with subrounded cobbles of an intermediate, equigranular intrusive rock. This distinctive deposit is overlain by up to 100 metres of predominantly coarse clastic rocks with green to purple pyroxene-phyric fragments derived solely from the Stuhini Group. The youngest deposit in this sequence is a distinctive purple volcanic conglomerate with cobbles of a fine-grained hornblende feldspar porphyry. These deposits mark the onset of tectonic activity associated with the beginning of Lower Jurassic volcanism.

The Toodoggone volcanics exposed between the Saunders fault and the Black Lake stock (Figure 2-10-4) can be divided into two parts that can each be broken into several units, described in detail by Marsden and Moore (1989).

At the base of the volcanics in the Jock Creek area is a sequence of quartz-hornblende-biotite-feldspar-phyric ash-flow deposits, a felsic dome and volcanic sediments, predominantly coarse laharic deposits. The lowest rocks (Unit 3) consist of thin ash-flow units and abundant purple and green volcanic conglomerates. These rocks, from the earliest episode of Jurassic volcanism, are very small volume deposits that were reworked in an active sedimentary environment.

Unit 3 is overlain by the Shasta dome complex comprising a dacitic dome with steep flanks (Unit 4a), associated epiclastic sediments (Unit 4b) and ash-flow deposits (Unit 5). The dome is a homogenous dark green to purple-green hornblende-biotite-quartz-feldspar porphyry with unbroken phenocrysts in a recrystallized matrix. The outer 25 metres of the dome is flow layered, with fine laminar banding apparent on weathered surfaces. The flow layering is usually parallel to the edge of the dome but is locally contorted. The dome is not present in all parts of the area mapped in detail (Figure 2-10-5) and where absent its stratigraphic position is taken by

Figure 2-10-4. Local geology. (*see* Figure 2-10-3 for legend.)

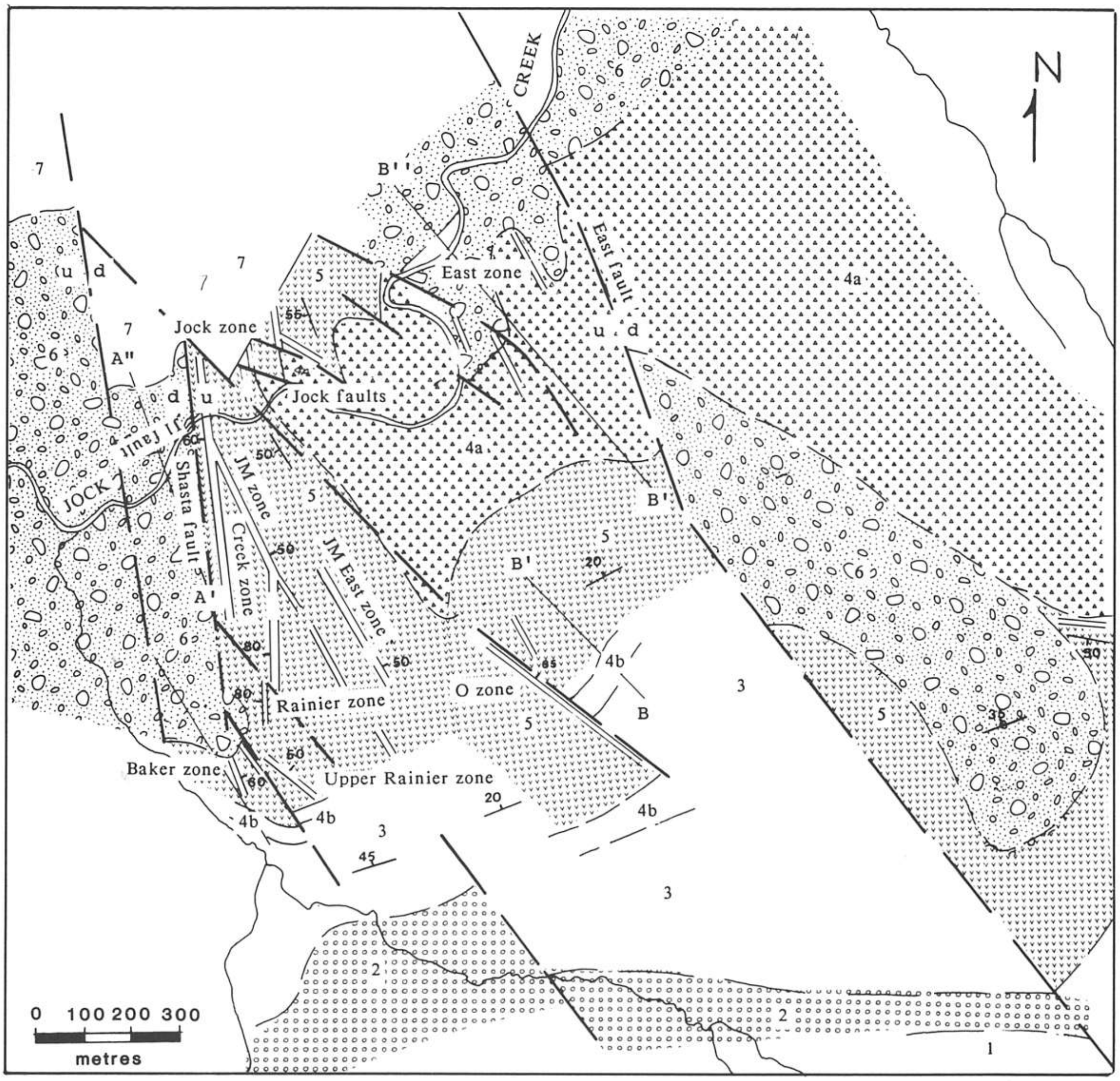

Figure 2-10-5. Geology of the Shasta property. (*see* legend on Figure 2-10-3.)

very coarse epiclastic deposits consisting of angular blocks of welded tuff up to several metres in diameter, of moderately welded tuff and local concentrations (channels) of purple to green volcanic siltstone and sandstone (Unit 4b). Ash-flow deposits (Unit 5) are exposed on the flank and beside the dacite dome. Where the ash flows were deposited on the flank of the dome they dip 50° to the southwest, whereas the rest of the stratigraphy and the ash-flow deposits away from the dome, dip gently to the north (Figures 2-10-6 and 7). The anomalous orientation of the ash flows can be partially attributed to their deposition on the steeply sloping flank of the dome but the 40–50° degree slope required is well beyond the angle of repose for loose material. These deposits must have been deposited on the slope of the dome and then oversteepened by continued dome growth; the dome and the ash flows are cogenetic and coeval. The ash-flow deposits can be divided into a lower section of welded ash tuff with highly flattened essential lapilli and an upper subunit that lacks flattened lapilli and includes more accidental frag-

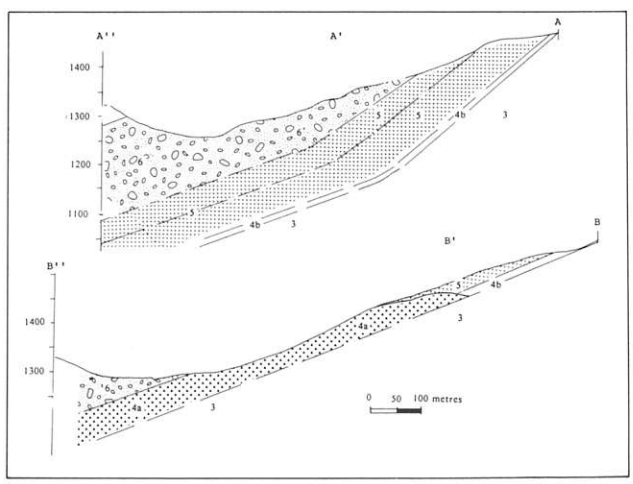

Figure 2-10-6. Cross-sections A-A″, B-B″. (*see* locations on Figure 2-10-5.)

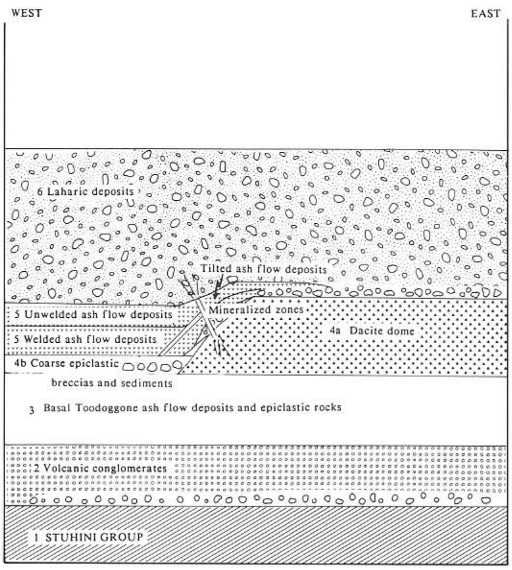

Figure 2-10-7. Schematic section of the Shasta property.

ments. The base of the ash flows is locally marked by green siltstones, a carbonaceous tuff with charred wood fragments and by local concentrations of subangular to subrounded cobbles of the dacitic dome within the ash-flow deposits.

The Shasta dome complex is buried beneath 200 metres of epiclastic deposits (Unit 6) that consist of numerous lobes of coarse, very poorly sorted laharic deposits and minor volcanic sandstone and siltstone. The fine sediments are typically crossbedded and graded and represent minor stream deposits developed on top of individual debris flows. These deposits fill a paleotopographic low adjacent to the dacitic dome and onlap the top of the dome. They all dip moderately to the north; they were deposited after dome growth ceased. Where the sediments are observed overlying the flank of the dome there is a prominent apparent unconformity between the tilted ash-flow deposits and the overlying laharic deposits (*see* Figures 2-10-6 and 7).

The rest of the Jock Creek volcanics consist of a strongly welded chloritic lapilli tuff (Unit 7) that is locally interbedded with the laharic deposits, laminated maroon crystal ash tuffs (Unit 8) and a complex sequence of quartz-free biotite-hornblende-feldspar-phyric volcanic breccia and spherulitic lapiili tuff (Unit 9). This unit is capped by a massive purple-brown to grey hornblende-biotite-feldspar porphyry that was interpreted by Marsden and Moore (1989) as a flow. It is macroscopically similar to subvolcanic intrusive rocks located north of the east end of Black Lake (Figure 2-10-4), and may be a sill (as originally mapped by Daikow *et al.*, 1985).

The Jock Creek volcanics are overlain by the Saunders grey dacite ash flow, a unit described in detail below.

Mapping in the northeastern corner of the study area (Figure 2-10-4) has outlined a previously unrecognized stratigraphy. In this area the oldest rocks that can be positively identified are part of the Saunders grey dacite, a unit that caps the succession in all areas west of the fault.

The Saunders grey dacite is divided into three distinctive subunits. The dominant unit (11a) is a single cooling unit, over 400 metres thick, of dark grey lapilli tuffs containing quartz, biotite, hornblende and feldspar crystals, moderately flattened feldspar-phyric lapilli and accidental intrusive fragments that are contained both within the lapilli and the matrix. The matrix consists of flattened and welded glass shards. The lapilli-rich deposits grade upwards into green crystal-ash-tuff deposits (Unit 11b) with the same phenocryst mineralogy as Unit 11a, rare feldspar-phyric lapilli and small but prominent intrusive fragments. The top of the ash flow deposits is marked by a purple volcanic sandstone to pebbly sandstone (Unit 11c) that consists solely of reworked material from the underlying ash flows.

Overlying the purple sandstones is a heterogeneous sequence of grey-green hornblende-pyroxene-biotite-feldspar-phyric flows, tuffs, subvolcanic intrusives and associated sediments (Unit 12). Medium-grained, hornblende-biotite-feldspar porphyritic subvolcanic intrusive bodies dominate much of the section and on the north ridge of Mount Todd (Figure 2-10-4) one of these is observed in intrusive contact with the Saunders dacite. The second most abundant rock type is coarse volcanic conglomerate with subrounded cobbles identical to the intrusive rocks. On the east ridge of Mount Todd these conglomerates directly overlie purple volcanic sandstones of the Saunders dacite. The conglomerates are usually associated with fine-grained, pastel-coloured purple to green volcanic siltstone, sandstone and mudstone. These deposits are very thin bedded and commonly graded, suggesting deposition from small turbidity currents in a shallow aqueous environment, possibly related to debris flows that deposited the volcanic conglomerates. Higher in the section the sediments are interbedded with fine to medium-grained, weakly amygdaloidal pyroxene-hornblende-feldspar-phyric flows and minor lapilli tuffs. Although Unit 12 everywhere rests on top of the purple volcanic sandstone, the contact is not planar, indicating considerable paleotopography, presumably as a result of synvolcanic faulting. These rocks are intruded by an elongate intrusive body, mapped by Daikow *et al.* (1985), that defines the eastern limit of the current project area. This body is actually an intrusive complex dominated by coarse-grained, subcrowded hornblende-biotite-feldspar porphyry (Unit A) and lesser fine to medium-grained equigranular granodiorite (Unit B). The porphyritic rocks are very similar to subvolcanic rocks in Unit 12 but are slightly coarser grained and less crowded. Daikow *et al.* (1985) mapped all of the Hazelton rocks east of this intrusive body as "undivided Hazelton". These rocks may be at least in part correlative with Unit 12 of this report and therefore in part younger than the Saunders grey dacite (Unit 11).

STRUCTURE

All the rocks within the map area are gently tilted and lack any evidence of ductile deformation. The distribution of the stratigraphy has been disrupted by numerous brittle faults developed at a high level in the crust. The Saunders fault interpreted by Daikow (personal communication, 1987) as a right-lateral fault with approximately 5 kilometres of displacement indicated by the offset of a small plug exposed south of the Finlay River. The fault zone strikes 150° and consists of numerous steeply dipping fault strands within a wide zone of altered and highly fractured rock. The south-western corner of the map sheet (Figure 2-10-4) is transected by another major fault structure that places the Black Lake stock and part of the Stuhini Group against the Jurassic volcanics. The Black Lake stock does not intrude the Jurassic stratigraphy and there is a small sliver of Triassic basalt between the stock and the Toodoggone volcanics; the stock has been faulted into its current position relative to the Toodoggone volcanics. This Black Lake fault may be part of the Attorney fault system.

Within the project area there are four faults striking 160° to 190° that splay off of the Saunders fault (Figure 2-10-4). Three of them are exposed on the west side of the Saunders fault; they all exhibit east-side-down displacement. The displacement decreases from north to south suggesting that the faults initiated adjacent the Saunders fault and propogated southward.

Smaller scale faults observed on the property are all moderately dipping; normal or normal-oblique faults. The actual slip vector cannot be uniquely determined in most cases, although a potential slip vector can be determined from a stereonet plot of conjugate fault sets; the assumed slip vector is perpendicular to the line of intersection of the fault set (Figure 2-10-8).

The faults on the property are all postmineral structures that offset the mineralized zones. Several of the faults have similar attitudes to the mineralized zones and they reflect a similar structural setting; they are all small-scale extensional structures.

MINERALIZATION

The mineralized structures are quartz-carbonate stockwork zones within larger areas of moderate to strong potassic feldspar alteration. The zones of strong alteration are tabular, with numerous anastamosing quartz-carbonate veinlets in which at least three distinct episodes of veining can be seen (*see* Thiersch and Williams-Jones, 1990, this volume). The veinlets, especially a late carbonate-dominated episode, carry native silver, gold and electrum in association with argentite, galena, sphalerite, pyrite and chalcopyrite. The zones grade outward into less-altered and weakly veined wallrock. The JM and Rainier zones (Figure 2-10-5) are characterized by irregular, steeply dipping central breccia veins. Individual veinlets within and adjacent to each zone, and the central breccia veins, usually have a similar strike to the overall zone, but a steeper dip.

The mineralization is not confined to well-developed stockwork zones; there are numerous mineralized areas on the property and some of the best grades yet obtained are from fracture fillings in weakly altered rock. The main stockwork zones did not control ore deposition but they have localized economically significant widths of precious metal mineralization.

Eleven mineralized zones were recognized by the end of the 1988 exploration program (Holbek, 1989). The 1989 exploration program has extended the known strike length and dip extent of the Creek zone (Figure 2-10-5) and has

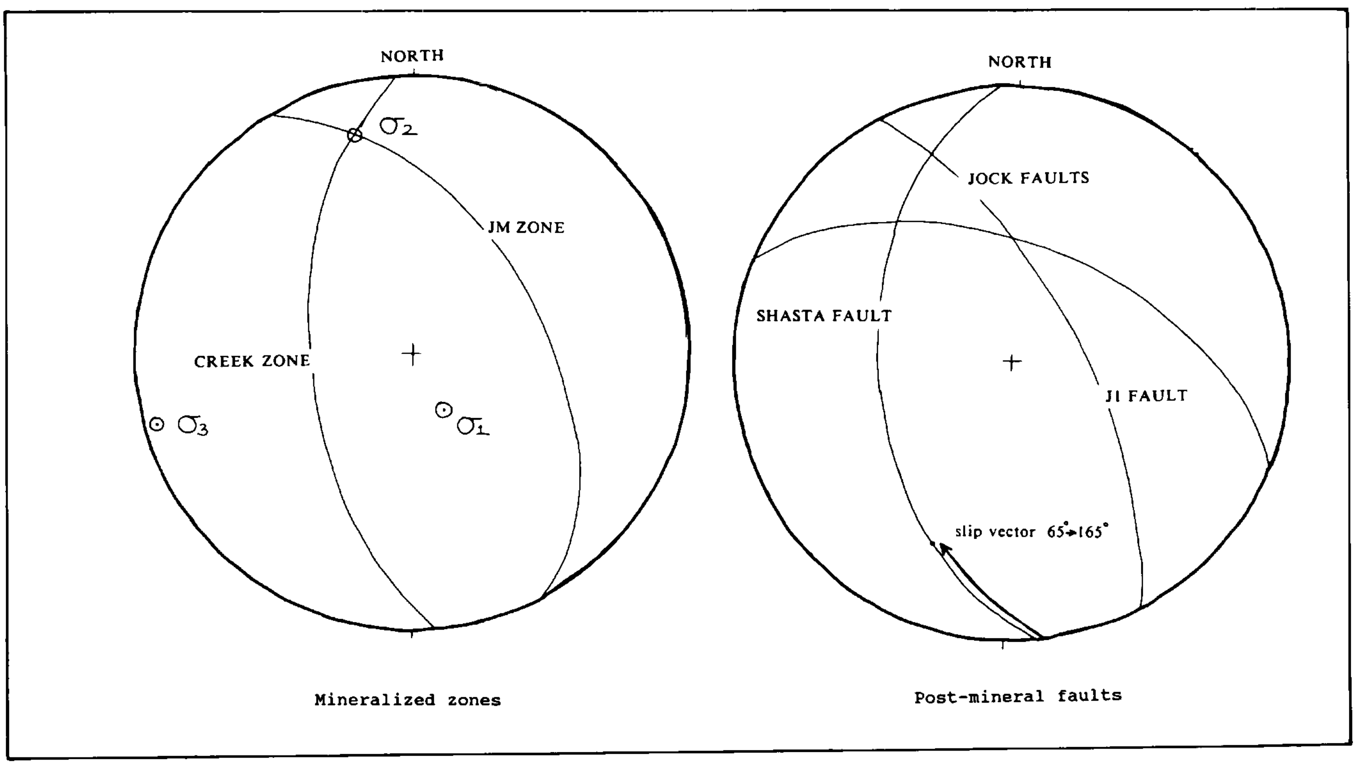

Figure 2-10-8. Equal-area plots from the Shasta property.

shown that the Rainier, Cayley, Baker, Upper Rainier, JM and Creek zones are all segments of a continuous vein sytem.

The core of the deposit is made up of the Creek and JM zones. The Creek zone strikes 170° and dips 60° west; the JM zone strikes 330° and dips 50° to the northeast. The two zones join in Jock Creek. They subtend an angle of 80° and their line of intersection plunges 25° toward 350° (Figure 2-10-8). If the mineralization formed before tilting of the stratigraphy then rotation of the stratigraphy and its contained mineralization back to the horizontal gives a subhorizontal line of intersection and a subvertical maximum compressive stress. As mentioned above, veinlets within the stockwork zones typically have a preferred orientation parallel to the overall strike of the zone but are generally more steeply dipping. This is consistent with minor extensional movement along the two zones (Figure 2-10-9). This pattern has important implications for exploration. Well-mineralized pods resulting from: (1) the intersection of the two main mineralized orientations, or (2) from areas of preferential extension localized by rolls in the fault plane, or (3) as extension gashes within the zone, will all be rod-shaped bodies that plunge to the north, parallel to the line of intersection of the two zones (Figure 2-10-8).

Longitudinal sections along the JM and Creek zones (Holbek, 1989) show that ore-grade mineralization does indeed bottom out along a gently north-plunging line. In the case of the Creek zone this is probably a structural control, but in the JM zone this line coincides with the base of the pyroclastic rocks (Unit 5). Where drill holes intersect the JM zone within the dacitic dome the zone is weakly developed, with no significant width of mineralization. The entire area underlain by the dacite dome between the JM and East zones (Figure 2-10-5) is sporadically mineralized and some spectacular grades have been obtained from minor veinlets, but no significant structure has yet been defined. Alteration has been identified in the overlying laharic deposits (Unit 6) but drilling on the Upper Rainier vein system has shown that the mineralization pinches out in the basal stratigraphy (Unit 3). All of the significant mineralization defined to date is hosted by the pyroclastic rocks (Unit 5).

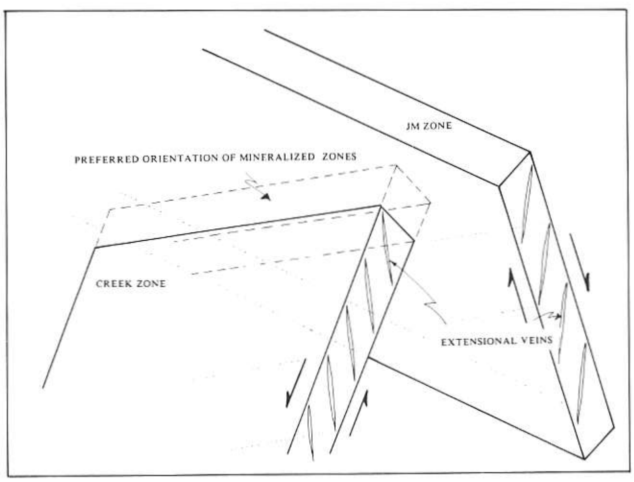

Figure 2-10-9. Structural model for the JM and Creek zones.

SUMMARY

The Shasta silver-gold stockwork deposit is an epithermal precious metal deposit spatially related to a dacitic dome of lower Middle Jurassic age. The mineralization consists of native metals in association with argentite and base metal sulphides within a quartz-carbonate stockwork with potassic alteration envelopes, an assemblage typical of epithermal precious metal deposits formed at moderate depths (several hundred metres). Significant zones are hosted by pyroclastic deposits that were deposited on the flank of a coeval dacite dome.

Economic widths of mineralization are primarily hosted by two predominant vein orientations; 170°/60° west and 330°/50° northeast. This pair of structures is a set of conjugate extensional fractures that intersect near Jock Creek. The geometry of the zones is such that the richest ore-shoots are rod-shaped elements that plunge gently north.

These zones are not within a major fault structure and did not form in response to a high applied stress. They may be preferentially located in the pyroclastic rocks because this unit was an aquifer and the high pore-fluid content helped initiate rock failure and dilation, concentrating fluid flow and resulting in mineral precipitation.

This exploration model can be used on the property in three ways. The dip of an individual zone can be quickly estimated based on its strike and the orientation of contained veinlets. Poorly developed structures such as the O zone and the East zone (Figure 2-10-5) can be traced into the pyroclastic rocks of Unit 5 where they may be better developed. The stratigraphic throw and an estimated slip vector on the Shasta fault can be combined to predict the location of a blind offset of the JM zone.

Although ages for the dacite dome and the mineralization are not available it is likely that there is a close genetic relationship between the two. Previous dating in other parts of the Toodoggone region (Daikow, 1985; Clark and Williams-Jones, 1988) has shown that the mineralization is close in age to the volcanics, a typical scenario for most active hydrothermal systems. The dome and its source area may not only have provided the heat that drove hydrothermal convection, but the oversteepening of deposits on the flanks of the dome may have created an inherent instability that encouraged extensional failure. Further exploration in the southern part of the Toodoggone belt should concentrate on locating similar dacitic flows and tuffs in stratigraphy of equivalant age (*ca.* 190 Ma). Within suitable host stratigraphy, additional favourable exploration criteria are anomalous stratigraphic orientations indicating possible domal or graben structures (*see* Vulimiri *et al.*, 1983) and pink altered feldspars (associated with elevated K_2O values) within dacitic tuffs or flows.

ACKNOWLEDGMENTS

The ideas presented in this paper are a compilation of the efforts of several individuals. Peter Holbek has been supervising the Shasta project for three years. Peter Thiersch has also been involved in the project for the last three years and is currently preparing an M.Sc. thesis on the mineralization. Margaret MacPherson has been mapping and producing core logs that are a joy to work with for the last two years. Critical appraisal by Ron Britten has kept the work on track and the ideas reasonable.

We would like to express our gratitiude to Homestake for offering generous support and exceptional freedom for the final year of this project. A research grant from the British Columbia Ministry of Energy, Mines and Petroleum Resources will allow the research to be concluded during the present academic year.

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PARAGENESIS AND ORE CONTROLS OF THE SHASTA Ag-Au DEPOSIT TOODOGGONE RIVER AREA, BRITISH COLUMBIA (94E)

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KEYWORDS: Economic geology, Toodoggone, Shasta, epithermal, stockwork, breccia, gold, silver, paragenesis, ore controls.

INTRODUCTION

The Shasta deposit is an epithermal multiphase quartz-carbonate stockwork/breccia vein deposit containing significant silver and gold mineralization. Previous exploration work on the property has been hampered by the apparent lack of correlation between economic grades of mineralization and the degree of alteration or stockwork intensity. The current study was initiated to address this problem and to identify other characteristics of the deposit that could be used to guide exploration and development of the property.

This report summarizes fieldwork and preliminary results of the study. Fieldwork during July and August consisted of detailed logging of mineralized intersections in drill core, and sampling of core for petrographic, fluid inclusion, stable isotope and geochemical studies. This work forms part of an M.Sc. thesis underway at McGill University by the senior author.

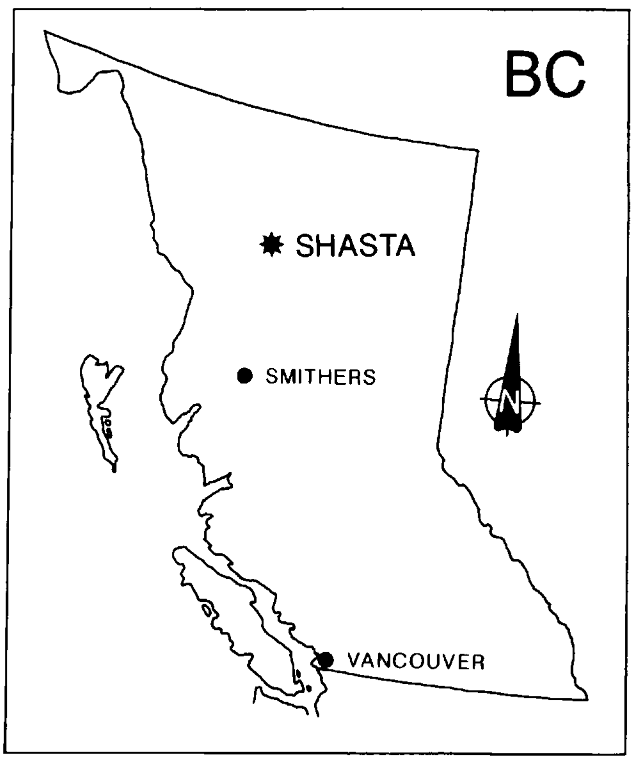

Figure 2-11-1. Location map of the Shasta deposit.

The Shasta property is located in the Toodoggone River area of British Columbia (Figure 2-11-1), on NTS map sheets 94E/2, 3, 6 and 7 at latitude 57°15′ north, longitude 127°00′ west. The property is accessible by gravel road from Sturdee airstrip (approximately 300 kilometres by air from Smithers) a distance of 9 kilometres, or from Fort St. James, some 675 kilometres via the Omineca mine road.

The property was first staked in 1972 for Shasta Mines and Oils Limited (now International Shasta Resources Ltd.). Subsequent work led to the discovery of the Main (Rainier) zone gold-silver-bearing quartz-carbonate stockwork. The property was worked intermittently until 1983 when it was optioned by Newmont Exploration of Canada Limited, which mounted the first integrated program of geology, geochemistry, geophysics and diamond drilling. This work led to the discovery of the Creek zone.

In 1987, Esso Minerals Canada optioned the property and conducted a comprehensive program of geological mapping, soil and rock geochemistry, VLF-EM and induced polarization surveys, trenching and about 6000 metres of diamond drilling. This work delineated several new mineralized zones, including the JM zone, which together with the Creek zone comprise the bulk of the deposit. Reserves have been estimated by Holbek (1989) at 1.02 million tonnes of 5.7 grams per tonne gold equivalent with a 3.0 gram per tonne cutoff (using 70 grams silver = 1 gram gold).

Esso Minerals Canada was divested by its parent company, Imperial Oil, in 1989 and the Shasta option was acquired by Homestake Mineral Development Company, which has continued exploration on the property. International Shasta Resources Ltd. is concurrently conducting a small, high grade open-pit mining operation on the Creek and JM zones.

REGIONAL SETTING

The Toodoggone River area lies within the Stikine Terrane on the eastern margin of the Intermontane Belt, in the Cassiar-Omineca Mountains. The regional structure is dominated by major dextral strike-slip faults on the margin of the Intermontane Belt.

The oldest rocks in the area are Permian limestones of the Asitka Group, which generally occur in thrust contact with Late Triassic Stuhini Group volcanics and as roof pendants within Omineca intrusions. Stuhini Group rocks are dominantly alkaline to subalkaline, submarine, mafic volcanics. Unconformably overlying the Stuhini are Lower to Middle

Jurassic Hazelton Group rocks representing a probable island arc sequence of volcanics and associated sediments.

The Toodoggone volcanics (Carter, 1972; Gabrielse *et al.*, 1976) represent a distinctive quartz-bearing facies of the Hazelton Group, and comprise dominantly calcalkaline, intermediate to felsic subaerial volcanics (Schroeter, 1982; Panteleyev, 1982, 1983; Diakow, 1984; Diakow *et al.*, 1985). Clark and Williams-Jones (1987, 1988) have proposed that Toodoggone volcanism can be divided into two depositional stages. They point out that all known epithermal gold-silver deposits are restricted to (Stage I) Toodoggone volcanics, underlying the (Stage II) Saunders grey dacite (Diakow *et al.*, 1985), and suggest that regional mineralization occurred during the waning of Stage I or during a hiatus between Stages I and II.

The youngest rocks in the area are Tertiary to Cretaceous Sustut Group sediments, which unconformably overly the Toodoggone volcanics. Late Triassic to Early Jurassic Omineca intrusions of granodiorite and quartz monzonite intrude the Toodoggone and Stuhini Group rocks.

PROPERTY GEOLOGY

Marsden and Moore (1989 and 1990, this volume) provide a detailed treatment of the geology of the area; only a brief summary is presented here.

The Shasta property is underlain by two distinct lithologies within the Toodoggone crystal ash tuffs or Attycelley tuffs (Diakow *et al.*, 1985). They were informally named the pyroclastic series and the "epivolcaniclastic" series by Holbek and Thiersch (1988). The pyroclastic series unconformably overlies pyroxene-feldspar-phyric basalt flows and breccias of the Stuhini Group. In the central part of the property, the pyroclastics consist of dacitic feldspar-quartz crystal tuffs, chloritic and heterolithic lapilli tuffs and an underlying feldspar-quartz-biotite porphyry flow (Marsden and Moore, 1989). These units all contain characteristic orange-weathering plagioclase feldspars. The epivolcaniclastic series consists of green to maroon feldspar-phyric tuffs, heterolithic agglomerates, lahars and ash tuffs. These strata overly the pyroclastic series, but are typically seen in fault contact with them, as explained below.

The structure of the deposit area is dominated by north to northwest-trending normal and/or dextral faults. These are cut by minor east to northeast-trending cross-faults. Strata underlying the area generally dip gently northward to northwestward, coinciding with the regional attitude, except in a central fault-bounded panel of pyroclastic series rocks which dips steeply southwest. The north-trending Shasta fault bounds one side of this rotated fault block (Marsden and Moore, 1989), separating epivolcaniclastic from pyroclastic series rocks. This fault also forms the hangingwall of the Creek zone stockwork.

Mineralization and alteration are essentially restricted to the pyroclastic series and underlying Stuhini Group rocks. The overall lack of alteration and mineralization in the epivolcaniclastics suggests that these rocks were deposited, or displaced by faulting, after the mineralizing event. The absence of alteration at the hangingwall contact with the Creek zone supports this interpretation. However, the recent discovery of small isolated veins and alteration zones in these rocks, some distance into the hangingwall, suggests that the epivolcaniclastic rocks may, after all, have been deposited prior to mineralization.

MINERALIZATION AND ALTERATION

The following sections are based on a detailed study of core from the Creek and JM zones, and limited work on the Rainier, East and O zones. General discussions of mineralization and alteration are presented first, followed by specific treatment of individual zones. Plate 2-11-1 shows the location of these zones. Several other showings are not well defined by drilling and hence are not discussed here.

MINERALIZATION

The Shasta deposit consists of multiple overlapping quartz-calcite stockwork/breccia systems that display generally similar characteristics. They occur as narrow (<1 metre) curvi-planar breccias that pinch and swell within wider (>10 metre) sections of variable alteration and veining intensity. Quartz and calcite gangue occur individually in single-stage veins, as multistage banded veins and breccias, and also intimately mixed in a single stage (Plate 2-11-2). Both gangue minerals display open-space-filling textures in banded veins and rare drusy vugs. Calcite is dominantly late, commonly occurring in the centre of earlier quartz veins and as the matrix in quartz vein and silicified wallrock breccias.

Silver and gold mineralization occurs erratically within quartz and calcite stockworks and breccias. Grades of mineralization appear to be independent of the intensity of alteration or brecciation. However, some of the highest silver values occur in late-stage calcite breccia. Gold to silver ratios vary unsystematically from 1:10 to 1:100, with a deposit average of about 1:45 (Holbek, 1989).

Silver-gold mineralization is associated with finely disseminated grey sulphides and coarser grained pyrite. The main sulphide phases are pyrite, sphalerite, galena and minor chalcopyrite, in decreasing order of abundance. Two distinct types of pyrite are recognized: disseminated euhedral crystals occurring in altered wallrock (PY I); and disseminated subhedral to irregular, fractured grains, with inclusions of galena, occurring in quartz and calcite gangue (PY II), (*see* Plates 2-11-3 and 4). The latter type is commonly associated with other base metal sulphides.

The fine-grained grey sulphide is dominantly sphalerite, which occurs as irregular worm-like grains (SP I) interstitial to quartz and calcite, and also as larger grains in contact with pyrite and/or galena (SP II). Some pyrite grains appear to be corroded and replaced by sphalerite (SP II). Most sphalerite contains abundant fine inclusions of exsolved chalcopyrite.

Galena occurs as subrounded inclusions in pyrite (GL I), or fills fractures in pyrite and forms discrete irregular grains, interstitial to other sulphides, quartz and calcite (GL II). Chalcopyrite occurs as exsolved inclusions in sphalerite (CP I), or interstitially between other sulphides and as free grains (CP II).

Scanning electron microscope analyses by Holbek (1989) identified native gold and silver, electrum and acanthite. The gold and silver minerals occur as inclusions (AG-AU I) in base metal sulphides, as rims (AG-AU II) around sulphide

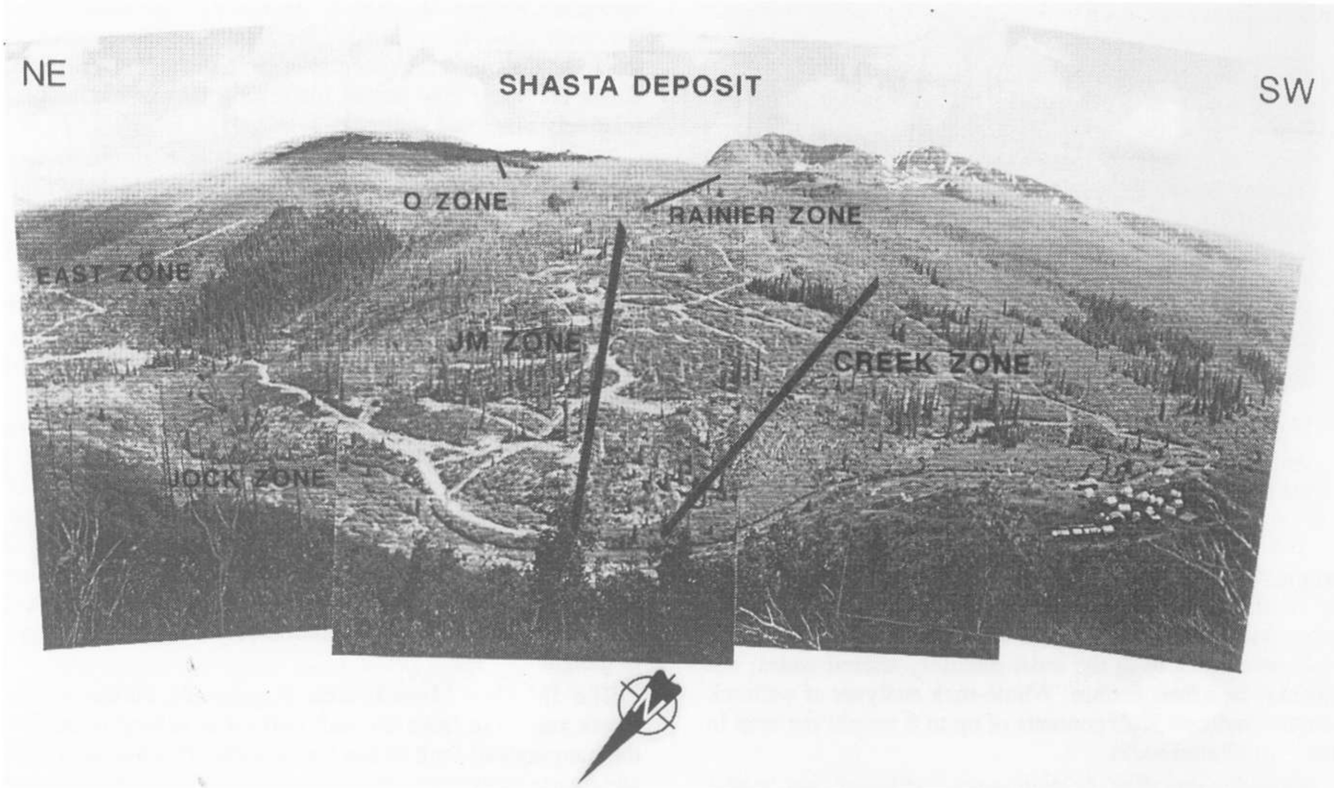

Plate 2-11-1. Panoramic photograph of the Shasta deposit, facing southeast. Solid black lines are traces of ore zones. Approximate vertical extent of zones is 350 metres. Camp in lower right hand corner.

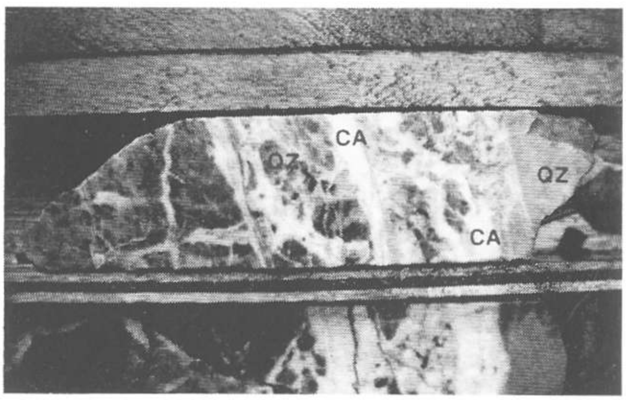

Plate 2-11-2. Drill core (ddh 88-6, 104.5 m) showing multistage quartz and calcite veining. Core diameter 4.5 cm. WR = wallrock, QZ = quartz, CA = calcite.

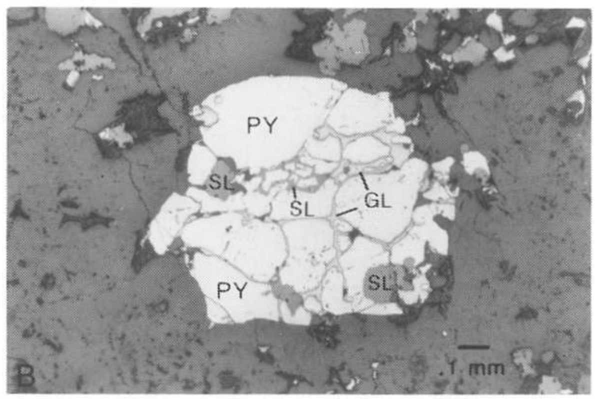

Plate 2-11-3. Photomicrograph of fractured PY II pyrite grain (white), infilled by later GL II galena (light grey) and SP II sphalerite (dark grey).

grains and as minute free grains (AG-AU II). Acanthite coexists with native gold, but not with electrum or native silver. Our precursory SEM analysis reveals that appreciable silver is also contained in galena.

ALTERATION

Wallrock alteration can be divided into four types: potassic, chloritic, phyllic and propylitic. Only potassic and chloritic alteration are directly associated with mineralization.

Propylitic alteration occurs on a regional scale and is characterized by an assemblage of chlorite + pyrite ± carbonate, mainly replacing mafic phenocrysts and lapilli fragments in pyroclastic series rocks.

Potassic alteration is associated with early quartz veins, and characterized by pervasive silicification and potassium metasomatism, resulting in distinctive orange-pink bleaching of stockwork zones, up to tens of metres wide. This alteration may also occur as narrow envelopes, less than 1 centimetre across, around quartz veins. Generally, increasing intensity of potassium metasomatism has resulted in the

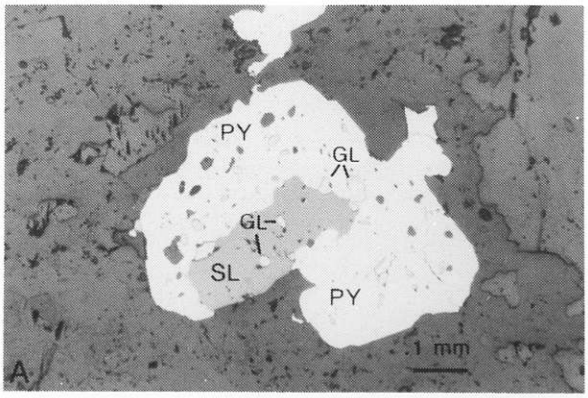

Plate 2-11-4. Photomicrograph of coprecipitated PY II pyrite (white) and SP II sphalerite (dark grey), both containing abundant inclusions of GL I galena (light grey).

progressive replacement of plagioclase phenocrysts, chloritic lapilli fragments and groundmass, in that order, by potassium feldspar and quartz. Minor sericite and clay minerals occur locally in the most intensely altered zones, but this may be a later feature. Whole-rock analyses of wallrock samples indicate K_2O contents of up to 8 weight per cent in the most altered rocks.

Chloritic alteration is dominantly, although not exclusively, associated with late-stage calcite veins and is represented by the assemblage chlorite $\pm$ epidote $\pm$ hematite. It occurs disseminated within veins and replacing wallrock fragments. Hematite and epidote also occur independently, in early to late fractures. This alteration has a more restricted distribution than potassic alteration.

Phyllic alteration is relatively uncommon, irregular and restricted. It is the latest event, overprinting both potassic and chloritic alteration, and usually destroying primary textures. It consists of pervasive sericite and finely disseminated pyrite, and does not appear to be spatially related to any particular features of the rock.

CREEK AND JM ZONES

The Creek and JM zones host most of the known reserves and outcrop between the 1260 and 1360-metre elevations, over strike lengths of 350 and 500 metres, respectively. The Creek zone strikes 180° and dips moderately westward. The JM zone trends 150° and dips steeply northeastward. These zones thus appear to merge to the north, although their intersection, if any, has not been identified. The Shasta fault forms the hangingwall of the Creek zone at surface, but appears to diverge from the zone as the fault attitude flattens with depth. The Creek and JM zones are hosted by two similar units of feldspar-quartz crystal lapilli tuff (one with heterolithic fragments, the other with dominantly chloritic fragments) which do not appear to exert lithological control on mineralization or alteration. It is notable, however, that silver-gold mineralization decreases sharply between 70 and 100 metres below surface, although stockwork zones may persist to greater depths.

The Creek zone is characterized by a well-defined stockwork system with strong silicification and coeval potassic alteration. Phyllic alteration occurs locally. Late calcite veining, with associated chloritic alteration, is abundant lower in the zone. Calcite veinlets continue through the footwall of the Creek zone and may persist for several tens of metres into relatively unaltered wallrock.

Silver and gold mineralization occur in both quartz and calcite veins, usually intimately associated with blebs of fine-grained pyrite. In calcite veins, pyrite is commonly associated with chloritic alteration of wallrock fragments. At higher levels in the zone, sulphide and silver-gold mineralization in quartz veins is commonly fine grained, and calcite veins are typically barren. At lower elevations, however, mineralization in calcite veins is usually coarse grained and quartz veins are sparsely mineralized.

Quartz and calcite breccias occur irregularly throughout the stockwork zone. Typically, wide (>10 centimetres) single-stage veins of either gangue mineral are barren or poorly mineralized, whereas narrower veins, multistage veins, and particularly mixed quartz-calcite-stage veins tend to carry abundant mineralization. It is also notable that sulphide and silver-gold mineralization tends to precipitate at the margins of veins and at contacts between different stages of gangue.

The JM (Just Missed) zone is generally similar to the Creek zone, but lacks the well-defined structural control of the hangingwall fault of the Creek zone. Potassic alteration and quartz stockworks appear more pervasive and somewhat stronger, while calcite veins and chloritic alteration are more restricted. Hematite and epidote alteration is locally abundant in the wallrock.

Early, narrow (<2 centimetres) quartz veins are commonly grey to green and fine grained. Late calcite breccias are usually wider (10-30 centimetres), and contain a coarse white calcite matrix that is usually barren. Gold and silver mineralization occurs in both quartz and calcite veins, but more commonly in narrow (<1 centimetre) veinlets toward the footwall of the zone. Two isolated occurrences of silver-gold mineralization in intensely silicified and potassically altered wallrock were noted in drill core, suggesting that early stages of quartz may be mineralized.

O ZONE

The O zone, situated 500 metres southeast of the JM zone, between the 1500 and 1550-metre elevations, strikes 130°, dips steeply northeast and is hosted by feldspar-quartz crystal lapilli tuffs and an overlying polymictic agglomerate.

Alteration and mineralization in the O zone is markedly different from that of the Creek and JM zones. Early alteration is characterized by strong, pervasive epidote, chlorite and hematite that is overprinted by moderate potassic alteration and weak silicification. These superimposed assemblages produce striking colour variations in drill core, particularly in the altered polymictic agglomerate.

Intense stockworks and breccias form well-defined zones with sharp boundaries, but are poorly mineralized. However, in rare mineralized sections, gold:silver ratios indicate significant enrichment in gold relative to other zones. Veins are quartz dominant, and although calcite is present, it is not intimately mixed with quartz as in the Creek or JM zones. Late fractures with narrow (<1 centimetre) potassium feldspar and quartz envelopes are commonly filled by calcite with abundant epidote and pyrite.

EAST ZONE

The outlying East zone, 750 metres northeast of the JM zone at the 1250-metre elevation, outcrops in the feldspar-biotite-quartz porphyry unit. The zone strikes approximately northwest.

East zone mineralization and alteration also differ from the Creek and JM zones. Pyrite is more abundant in quartz veins and is locally semimassive. Chalcopyrite is also more common and occurs with galena along hairline fractures in moderately potassically altered wallrock. Silver-gold mineralization occurs in both quartz and calcite veins, but is concentrated in pyrite-rich quartz veins. Late potassic alteration is superimposed on strongly epidote-altered wallrock. A distinctive alteration type confined to the East zone is the conversion of plagioclase to dark green, translucent sericite (?) accompanied by pervasive phyllic alteration.

RAINIER AND JOCK ZONES

The Rainier zone, 300 metres south of the Creek zone at the 1400-metre elevation, is hosted by feldspar-quartz crystal lapilli tuff and trends roughly north with a subvertical dip. The zone has been subjected to extensive faulting and its morphology is unclear. The Jock zone, outcropping immediately northeast of the JM zone above the 1260-metre elevation, is also poorly defined due to complex faulting.

The Rainier and Jock zones were not examined in detail and will not be discussed at length. It is apparent however, that the mineralization and alteration styles of the Rainier and Jock zones are similar to the Creek and JM zones, respectively. This observation suggests that they may be extentions of the Creek and JM zones.

PARAGENESIS

Silver-gold mineralization of the Creek and JM zones can be divided into five stages: Quartz I, II, and III, and Calcite I and II (Figure 2-11-2). Quartz I comprises silicification and associated potassic alteration, and is widespread but only rarely mineralized. Quartz II consists of fine-grained, grey to clear quartz in narrow veins that are frequently well mineralized. Quartz III is fine-grained, dark grey to green, commonly forms wide breccias, and is generally barren. Calcite I is white to green, associated with chloritic alteration, is commonly well mineralized and frequently occurs with Quartz II as a single intimately mixed stage or as breccia matrix. Calcite II is white or cream coloured, very coarse grained and generally forms barren late-stage veins.

Textural relationships indicate a close temporal association between sulphide phases and silver-gold mineralization. All sulphides share mutually interlocking grain boundaries, and silver-gold minerals occur as both rims around and inclusions in various sulphide phases. Sphalerite, galena, chalcopyrite, silver, gold, electrum and argentite (now acanthite) appear to have been precipitated almost contemporaneously; only pyrite exhibits a wider temporal range. Features such as fractures and corroded embayments in pyrite indicate early deposition and subsequent replacement.

Sulphide and silver-gold mineralization occur in several stages of vein filling, but are particularly abundant in the intimately mixed Quartz II and Calcite I stage. This coprecipitation of quartz and calcite indicates that a single, common mechanism controlled the deposition of both these gangue phases and silver-gold mineralization. The occurrence of silver in galena also requires a common parent fluid or precipitation control for galena and silver. The observation that rich silver-gold mineralization commonly occurs in narrow (<3 centimetres) veins suggests that low fluid:rock ratios may also have been a significant factor in deposition.

CONTROLS OF SILVER-GOLD DEPOSITION

The wide occurrence of hydrothermal breccias in the Shasta deposit is strong evidence for boiling of the mineralizing fluid. Boiling and brecciation could also explain several other features of the deposit. The widespread silicification and fine-grained nature of Quartz I suggest that early hydrothermal fluids were supersaturated with respect to silica. This is consistent with the rapid decreases in temperature and pressure that are associated with boiling (Fournier, 1985a). The observation that there were repeated episodes of brecciation can be explained by sealing of the system with rapidly deposited silica, and hydraulic fracturing due to the resulting build-up of pressure.

Boiling of the hydrothermal fluid would also have led to a loss of dissolved H_2S and CO_2 to the vapour phase. The loss of H_2S was probably responsible for the deposition of gold, as gold solubility is dominantly controlled by complexes

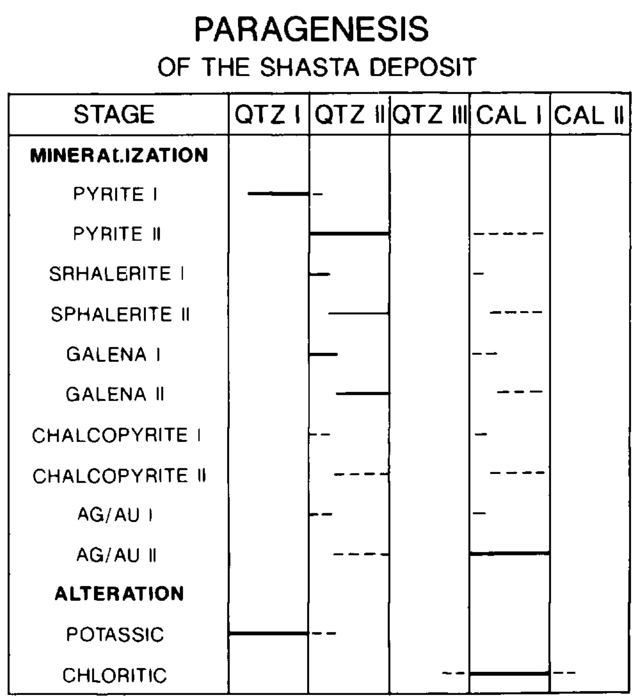

Figure 2-11-2. Paragenesis diagram showing general relationships between vein stages, sulphides and silver-gold mineralization. Solid lines indicate significant mineralization, thick lines up to 5 per cent; thin lines 1 to 2 per cent; dotted lines, trace mineralization.

involving H_2S (Seward, 1984). Partitioning of H_2S and other acid-forming volatiles into the vapour phase would also have increased the pH of the liquid, and led to the precipitation of calcite (Fournier, 1985b). The deposition of silver and base metal sulphides may have occurred in response to either this increase in pH, or the rapid decreases in temperature and pressure that accompanied boiling.

It is thus likely that a boiling process was responsible for the near-contemporaneous deposition of base metal sulphides and silver-gold mineralization, and the intimate mixing of Quartz II and Calcite I. The concentration of silver-gold mineralization at the contacts between different gangue stages and at vein margins is also an expected consequence of boiling and brecciation. The limited vertical extent of mineralization in the Creek and JM zones is likewise a predicted effect of boiling of the mineralizing fluid.

SUMMARY

Detailed examination of drill core from the Shasta silver-gold deposit reveals five stages of quartz and calcite vein filling in the Creek and JM zones. Narrow veins of fine-grained, clear to grey Quartz II, not directly associated with potassic alteration, and white to green Calcite I, associated with chloritic alteration, contain the bulk of silver-gold and sulphide mineralization. Gold, silver, electrum and acanthite were precipitated almost contemporaneously with second stage sphalerite, galena and chalcopyrite, suggesting a common mineralizing fluid and/or precipitation control. The simplest explanation for the observed relationships between gangue, base metal sulphides and precious metals is that of deposition in response to episodic boiling and brecciation.

Future work will involve fluid inclusion, stable isotope and geochemical studies to test the validity of a boiling model for the Shasta deposit, and to determine the physiochemical conditions of silver-gold deposition. These studies should contribute to a better understanding of the genesis of the deposit and provide useful guidelines for the continuing exploration and development of the property.

ACKNOWLEDGMENTS

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NOTES

TEMPERATURE AND COMPOSITION OF FLUIDS IN THE BASE METAL RICH SILBAK PREMIER Ag-Au DEPOSIT, STEWART, B.C. (104B/1)

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University of Western Ontario

KEYWORDS: Economic geology, Silbak Premier, epithermal, paragenesis, fluid inclusions, homogenization temperatures, salinities, Raman spectrometry, deposit models.

INTRODUCTION

The Silbak Premier mine is in northwest British Columbia, 21 kilometres north of Stewart (Figure 2-12-1), on the east side of the Salmon River Valley. The original discovery of a high-grade vein in 1910 culminated in an underground mine that from 1918 to 1953 produced 4.3 million tonnes of ore grading 14.6 grams gold and 304 grams silver per tonne from sulphide-bearing stockwork veins and breccia zones. Current open pit reserves are 5.9 million tonnes grading 80.23 grams per tonne silver and 2.16 grams per tonne gold (Randall, 1988).

Chemical composition of ore-forming fluids has been estimated by measuring the temperature of homogenization and freezing point depression of fluid inclusions; mass spectral analysis of gaseous species using a Raman spectrometer; and microprobe x-ray maps of precipitates from fluid inclusions to define nonvolatile components. These fluid inclusions are in quartz which can be reliably placed in the paragenetic sequence. The change in temperature and chemical composition of fluids before, during, and after ore deposition are used to interpret chemical constraints on gold and base metal deposition. These constraints assist development of an exploration model.

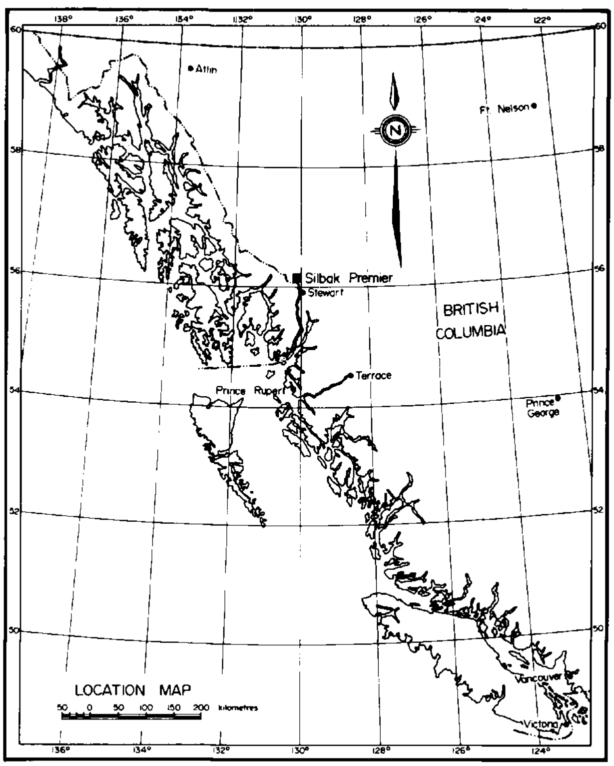

Figure 2-12-1. Location of Silbak Premier deposit near Stewart, B.C.

REGIONAL GEOLOGY

The Silbak Premier deposit is at the west margin of the Intermontane Belt, adjacent to the Coast Belt. It is hosted by north-northwest-trending volcanic rocks of the Hazelton Group that are unconformably overlain by sedimentary rocks of the Bowser Group to the east (Alldrick, 1985, 1987). The Hazelton Group comprises Late Triassic to Early Jurassic andesite flows, breccias and tuffs overlain by Early Jurassic dacite lapilli tuffs, dacite flows and volcaniclastic rocks. Rhyolite breccia and tuff are a regional marker horizon at the top of the Hazelton Group (Brown, 1987). Three intrusive episodes are represented by porphyritic dacite of the Early Jurassic Texas Creek plutonic suite, Eocene Hyder quartz monzonite and granodiorite, and Oligocene to Miocene biotite lamprophyre dikes.

The regional structure is a north-northwest-striking system of open to tight folds with moderate to steeply dipping west-southwest cleavage. Northerly trending, right-lateral strike-slip faults segment regional stratigraphy. Metamorphic grade is lower greenschist facies with a thermal peak of 330°C (Alldrick *et al.*, 1987).

GEOLOGY OF SILBAK PREMIER DEPOSIT

The Silbak Premier deposit consists of sulphide-bearing, parallel and stockwork veins and breccia zones marginal to and crosscutting irregular to tabular plugs and dikes of Jurassic potassium feldspar porphyritic dacite. Discrete and conjugate merging lobes of dacite intrude or are conformable with andesite flows, breccias and tuffs and dacite flows of Late Triassic to Early Jurassic age (Figure 2-12-2 and 2-12-3). Eocene granodiorite, quartz monzonite and microdiorite dikes traverse andesite-dacite volcanic and intrusive rocks.

Layering in dacite flows has a north to northwest strike with a variable westward dip. Cleavage developed during the first fold generation has a northerly strike with a variable dip to the west. At the deposit, this broad open fold is an asymmetric syncline with a steep east limb. The second foliation is a spaced fracture cleavage which strikes east and

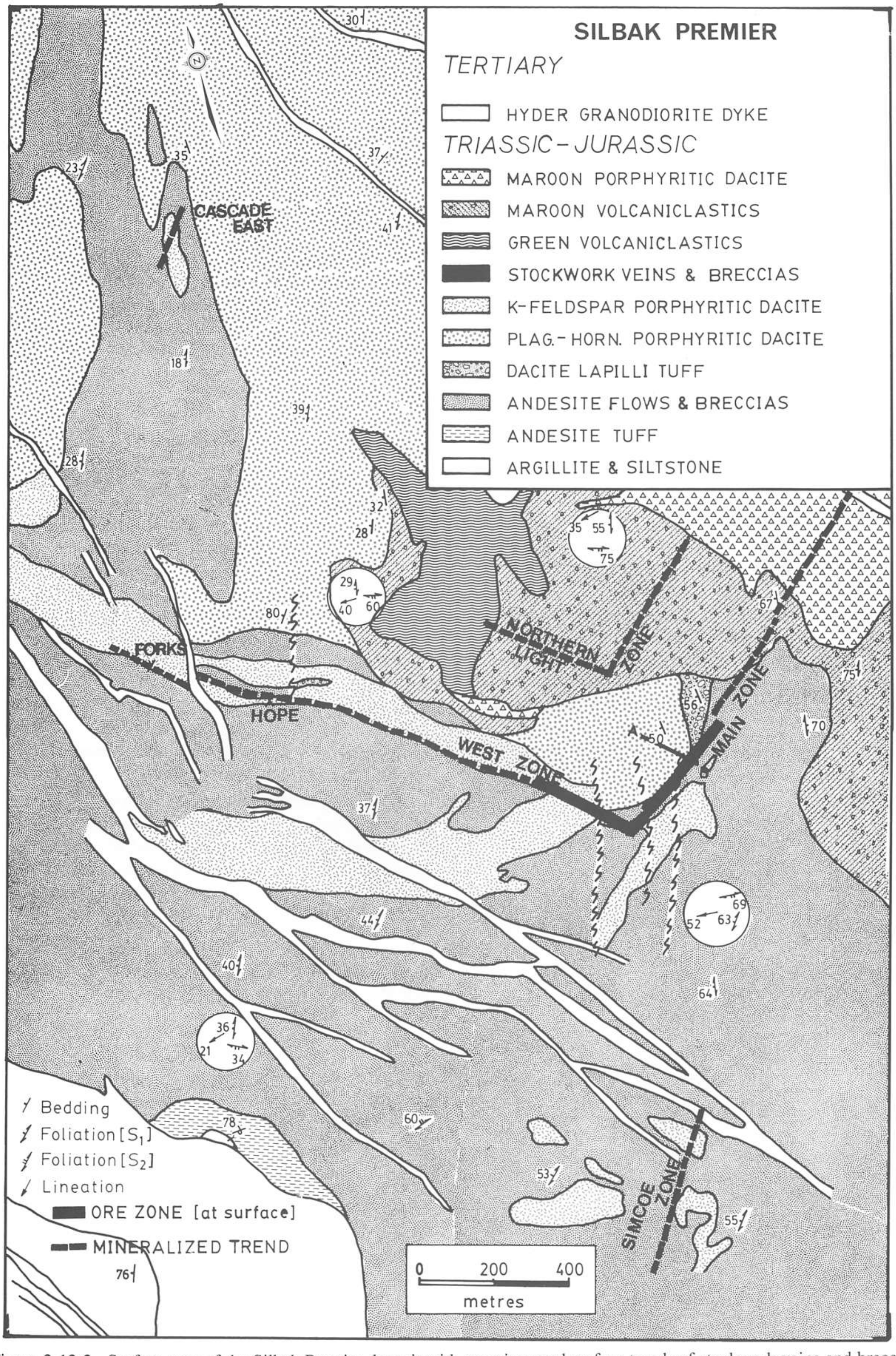

Figure 2-12-2. Surface map of the Silbak Premier deposit with superimposed surface trends of stockwork veins and breccias. Cross-section 2340 N (Figure 2-12-3) is defined by line A-B through the Main zone. Portions modified from Brown (1987).

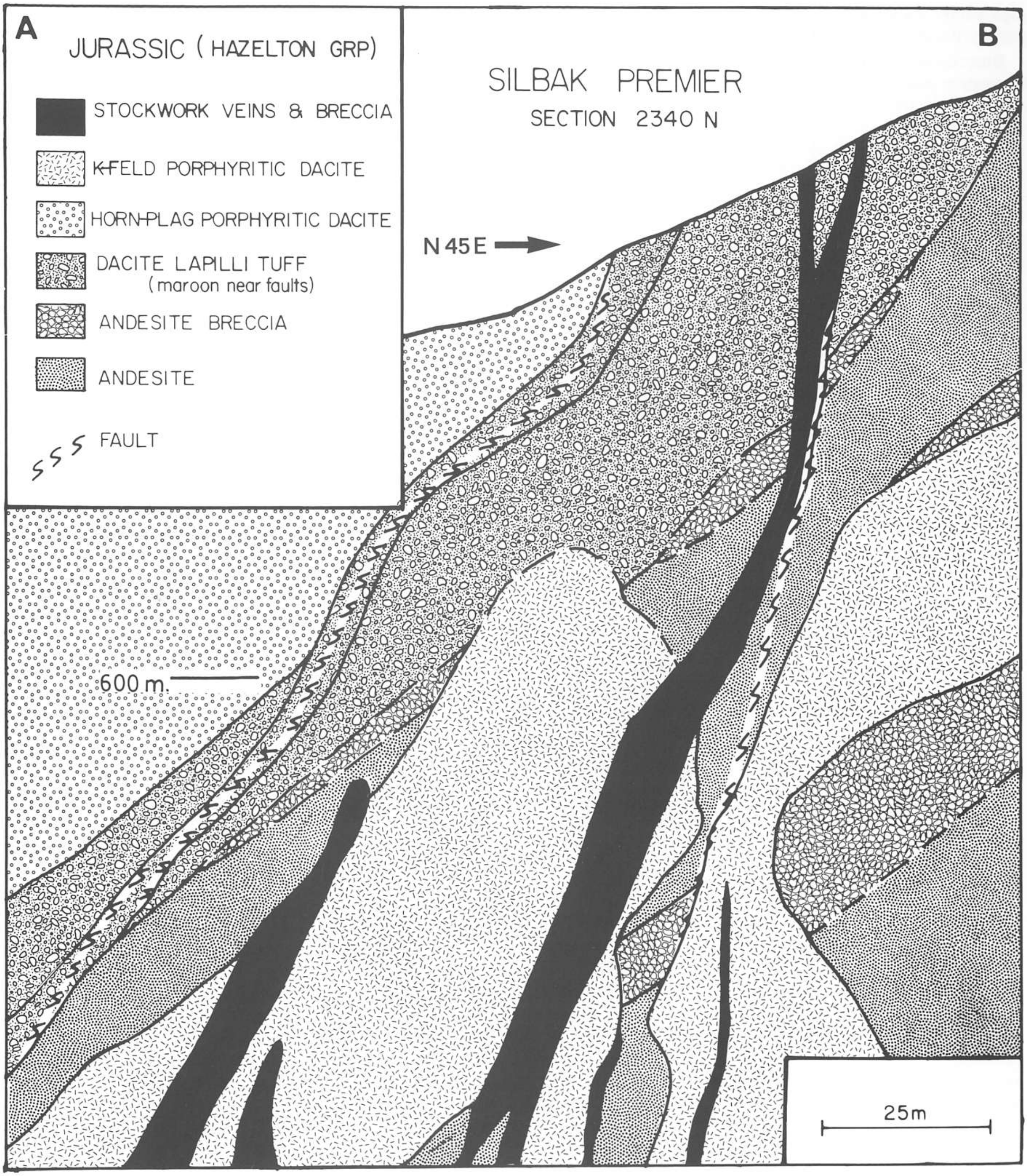

Figure 2-12-3. Geological cross-section 2340 N. Sharply defined veins in stockworks and matrix-to-breccia contain metallic mineral assemblages and follow a number of subparallel trends within and marginal to a potassium feldspar porphyritic dacite. The porpyritic dacite and metal-bearing veins and breccias exhibit both conformable and crosscutting relationships with andesite and dacite country rocks of Jurassic age.

dips south. Subsequent conjugate faults are the locus for porphyritic dacite intrusions, breccias and stockwork veins.

Breccias and veins formed before, during and after ore deposition and are referred to as early, middle and late-stages, respectively. Early-stage breccia is a crackle or *in situ* breccia occuring as a 5 to 15-metre zone of siliceous, rounded to angular andesite fragments enclosed in a quartz, calcite and pyrite matrix. These breccia zones are both coeval with and subsequent to the intrusion of porphyritic dacite and form within andesite peripheral to it. Early-stage veins are primarily banded quartz and chlorite with pyrite accummulations along the margins. They are subparallel, and crosscut early-stage crackle breccia, but are cut by middle-stage stockwork veins.

Middle-stage stockwork veins and matrix to the breccia contain most of the metallic minerals and the greatest amount of silver and gold. They have distinct precious and base metal rich mineral assemblages where base metal abundances increase and precious metals decrease with depth. Orthogonal fractures are infilled with quartz, pyrite and minor amounts of electrum, polybasite, pyrargyrite, argentite and native silver. Late, coarse-grained quartz-chlorite veins occupy vertical and horizontal fractures and cut stockwork veins in en echelon pattern.

Early-stage breccia with propylitic alteration minerals chlorite, pyrite, calcite and sericite developed within andesite along the upper margins of the porphyritic dacite. Middle-stage propylitic alteration is overprinted by potassic alteration minerals potassium feldspar and sericite in the wallrock adjacent to veins and breccias. Narrow envelopes of propylitic alteration occur on the margins of late-stage veins.

Both precious and base metal rich veins and mineralized breccia matrix show a paragenetic sequence from sulphide-rich minerals to sulphosalts and native minerals (McDonald, 1988). In both types of veins, pyrite is earliest and is coincident with quartz. Subsequent sulphide and sulphosalt minerals occur with quartz and later potassium feldspar.

STUDY OF FLUID INCLUSIONS

DESCRIPTION

Fluid inclusions are small, generally microsopic, volumes of fluid trapped in irregularities within crystals. Although typically trapped as a homogeneous fluid at the temperature of growth, they are normally multiphase (solid, liquid, vapour) at room temperature. The sealing off of such irregularities in crystals may occur during the growth of the surrounding crystal to make primary fluid inclusions, or by recrystallization along fractures at a later time forming pseudosecondary or secondary inclusions (Roedder, 1984).

There are two optically distinct varieties of quartz in all veins examined: grey to milky translucent quartz and transparent quartz which either engulfs or occurs along microfractures in milky quartz. Milky quartz is usually turbid and has many isolated clusters of inclusions and tiny grit-like particles without birefringence or crystal form. The minute size and high density of inclusions in milky quartz make them unsuitable for study.

Fluid inclusions in clear quartz typically have a liquid phase and vapour bubble. Daughter minerals, halite and sylvite, are noticeably absent. Primary, pseudosecondary and secondary inclusions are discriminated by size, shape and distribution with respect to fractures in quartz.

TYPES OF INCLUSIONS

Primary inclusions are identified by their solitary location unrelated to fractures (Plate 2-12-1A) or in zones parallel to and within zoned crystal growths (Plates 2-12-1B and 2-12-1C). Two populations of primary fluid inclusions are present:

Type 1 primary inclusions are elliptical to irregularly shaped, 5 to 15 microns in diameter, and in places necked. These two-phase, liquid-vapour inclusions account for 90 per cent of primary inclusions with relatively constant 20 to 30 volume per cent vapour and no daughter crystals.

Type 2 liquid-vapour, primary inclusions are rare, rounded to elliptical and 15 to 20 microns in diameter (Plate 2-12-1D). The vapour to liquid volumetric ratios for Type 2 inclusions are fairly uniform, with the vapour phase commonly occupying less than 20 per cent of the inclusion volume. Type 2 inclusions are most common in vuggy quartz and late-stage veins. Inclusions in both base and precious metal, middle-stage veins have similar degrees of fill, however, late-stage veins appear to have more volume of fill.

Pseudosecondary inclusions are planes of inclusions that traverse growth zones, but do not cross grain boundaries (Plate 2-12-1E). Dense clusters of fluid inclusions are called pseudosecondary because one cannot be certain they are primary. Pseudosecondary inclusions closely resemble primary inclusions in size, but are commonly necked. The volume of liquid and vapour is variable from 20 to 35 per cent. No daughter crystals have been identified.

Secondary inclusions occur in planar arrays that traverse growth zones (Plate 2-12-1F), tend to be 1 to 5 microns in diameter, and elliptical to flattened. They are more common than primary inclusions and normally occur along fracture planes. A few secondary inclusions are liquid only, however, volume of gas bubbles typically varies from 10 to 30 per cent.

HOMOGENIZATION TEMPERATURES

Fluid inclusions were studied by standard heating and freezing techniques on a Th 600 Linkam gas-flow heating and freezing stage (MacDonald and Spooner, 1981). The stage was calibrated using small samples of reagent-grade compounds with known melting points for the temperature range of −30° to 350°C. Homogenization measurements were assessed with melting point standards and were reproducable within 1°C for temperatures in the range −20° to 250°C and as much as 3°C for higher temperatures. This study is based on 410 measurements on 21 samples from three different phases of veining (Table 2-12-1, Figure 2-12-4).

Homogenization temperatures of primary inclusions in early-stage veins ranged from 179° to 240°C with an average of 204.6°C. Primary inclusions in quartz from precious metal rich, middle-stage veins homogenize from 187° to 287°C and average 225.4°C. Inclusions from base metal rich, middle-stage veins have slightly higher homogenization temperatures that range from 180° to 292°C and average 228.8°C.

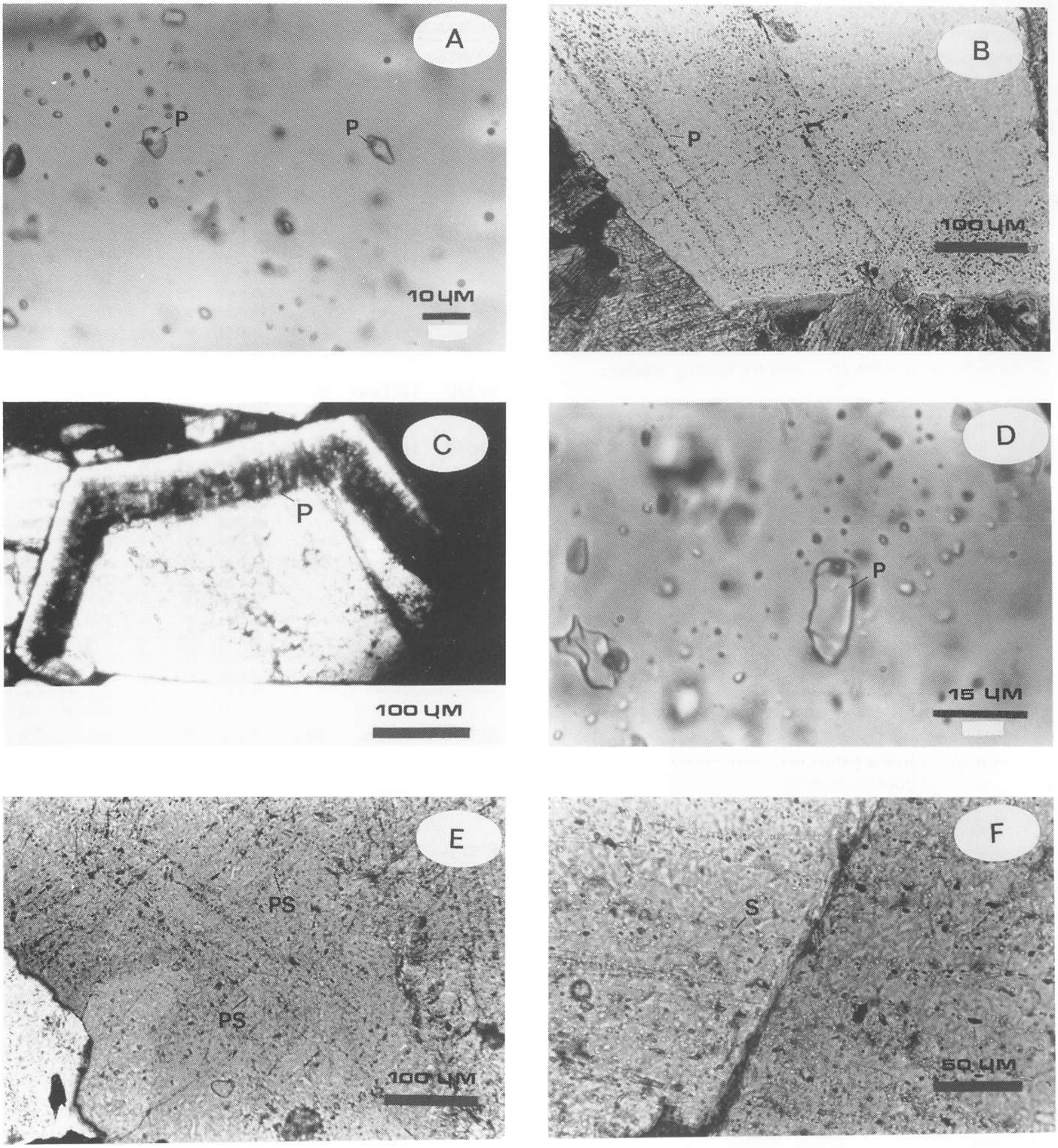

Plate 2-12-1.

A. Photomicrograph of a two-phase primary inclusion (P); water solution plus vapour bubble. Polished section SP-221.

B. Photomicrograph of milky quartz crystal cut perpendicular to the C-axis. Crystal growth was not continuous, but episodic, as evidenced by the band of primary inclusions (P). Polished section SP-114-A.

C. Photomicrograph of concentric band of primary inclusions (P) in the interior of a hexagonal quartz crystal. Polished section SP-47-U.

D. Photomicrograph of Type 2 aqueous primary inclusions (P) with vapour bubbles. Polished section SP-10.

E. Photomicrograph of intersecting pseudosecondary inclusions (PS). The planar array of inclusions does not cross grain boundaries. Polished section SP-80.

F. Photomicrograph, taken with partly crossed polars, of planes of secondary inclusions in quartz from gold-quartz veins. Note that many of the planes crosscut grain boundaries, outlining former throughgoing fractures. Polished section SP-181.

Type 2 primary inclusions have similar homogenization temperature ranges as Type 1 inclusions in all vein stages. Homogenization temperatures for pseudosecondary inclusions in each stage of veining are typically 10° to 20°C lower than their primary equivalent. Homogenization temperatures from secondary inclusions are similar in all stages of mineralization and range from 140° to 216°C, averaging 167°C.

Freeing Point Depression Temperature

The salinity of fluid inclusions is estimated from depression of the freezing point of ice using a gas-cooled freezing stage. The results are expressed as equivalent weight per cent NaCl, because NaCl is found to be the major salt component in the inclusions. Freezing point determinations from each vein stage (Table 2-12-2 and Figure 2-12-5) were measured on the same fluid inclusions used for heating studies.

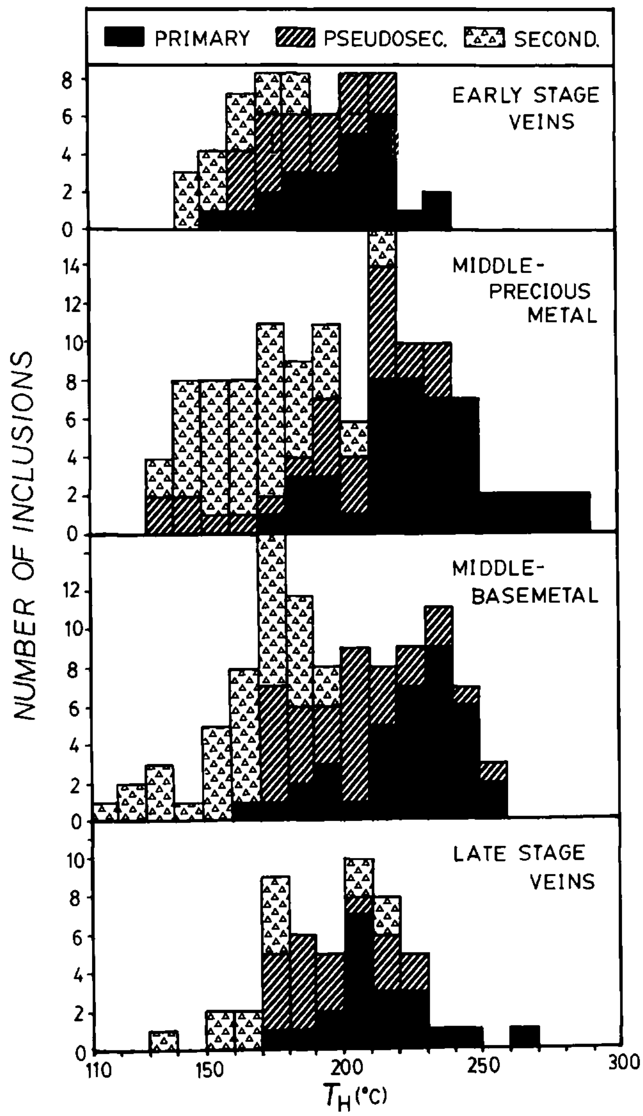

Figure 2-12-4. Histogram of homogenization temperatures for fluid inclusions. Dark areas are primary inclusions in quartz, banded areas are pseudosecondary and stippled triangled areas are secondary. Both early and late-stage veins are barren of sulphides.

TABLE 2-12-1
HOMOGENIZATION TEMPERATURES OF FLUID INCLUSIONS

SAMPLE #	Location	Elevation	Stage*	Type	No. Inc	Range Temp	Mean Temp
SP-165	DH 85-69	615.76	I	P	10	182-240	209.1
			I	PS	8	162-214	186.3
			I	S	4	151-189	165.2
SP-166A	DH 87-248	652.42	I	P	9	179-220	199.6
			I	PS	9	168-211	185.7
			I	S	5	147-181	165.8
SP-50S	Surface	664.66	II-P.M.	P	8	207-227	218.2
			II-P.M.	PS	7	195-246	224.7
			II-P.M.	S	5	156-216	178.6
SP-97S	Surface	675.76	II-P.M.	P	8	189-216	204.9
			II-P.M.	PS	5	176-198	182.8
			II-P.M.	S	5	156-216	184.2
SP-229	DH 86-79	600.04	II-P.M.	P	11	198-228	212.4
			II-P.M.	PS	7	186-221	201.5
			II-P.M.	S	5	152-204	178.8
SP-59U	110 Level	581.82	II-P.M.	P	10	200-235	221.4
			II-P.M.	PS	6	146-191	163.7
			II-P.M.	S	4	164-191	177.1
SP-51U	110 Level	583.33	II-P.M.	P	8	229-287	242.6
			II-P.M.	PS	6	205-225	215.4
			II-P.M.	S	5	163-211	177.4
SP-207A	2-Level	530.31	II-P.M.	P	8	212-241	229.1
			II-P.M.	PS	5	194-218	206.3
			II-P.M.	S	4	144-166	155.5
SP-47A	2-Level	530.45	II-P.M.	P	8	217-243	231.4
			II-P.M.	PS	5	201-231	216.9
			II-P.M.	S	6	149-194	167.1
SP-204A	3-Level	478.79	II-P.M.	P	10	206-263	240.6
			II-P.M.	PS	6	206-242	221.2
			II-P.M.	S	4	179-205	188.8
SP-213A	5-Level	312.12	II-P.M.	P	11	187-256	218.4
			II-P.M.	PS	6	163-211	192.8
			II-P.M.	S	5	109-141	152.2
SP-5U	6-Level	234.24	II-P.M.	P	9	193-273	238.1
			II-P.M.	PS	6	178-215	193.9
			II-P.M.	S	6	138-167	149.2
SP-136A	Surface	666.67	II-B.M.	P	9	196-241	219.6
			II-B.M.	PS	6	187-228	205.3
			II-B.M.	S	5	147-182	169.2
SP-158	DH 85-74	633.34	II-B.M.	P	10	187-231	214.6
			II-B.M.	PS	6	156-193	176.4
			II-B.M.	S	5	132-178	155.5
SP-48A	DH 86-114	564.24	II-B.M.	P	11	198-244	228.7
			II-B.M.	PS	7	184-220	202.4
			II-B.M.	S	5	153-204	179.2
SP-203A	3-Level	478.79	II-B.M.	P	10	217-271	244.6
			II-B.M.	PS	6	173-237	205.9
			II-B.M.	S	5	142-171	155.3
SP-214A	5-Level	312.12	II-B.M.	P	9	212-292	249.3
			II-B.M.	PS	7	187-227	199.8
			II-B.M.	S	6	148-210	172.3
SP-200A	6-Level	234.24	II-B.M.	P	10	180-242	217.2
			II-B.M.	PS	9	182-207	189.1
			II-B.M.	S	5	112-182	155.2
SP-118	DH 84-35	647.58	III	P	10	179-219	208.4
			III	PS	5	174-217	206.6
			III	S	4	141-171	154.5
SP-14A	DH 87-255	667.58	III	P	9	184-220	204.1
			III	PS	7	182-201	186.1
			III	S	6	151-206	168.4
SP-112A	DH 87-235	602.73	III	P	9	197-236	213.1
			III	PS	8	176-212	206.5
			III	S	5	140-179	164.4

* I— EARLY-STAGE
II— MIDDLE-STAGE
BM— BASE METAL RICH
PM— PRECIOUS METAL RICH
III— LATE-STAGE

The melting point of ice in H_2O-dominant inclusions from early-stage veins range from 0.0° to -0.3°C, corresponding to salinities of about 0 to 2 equivalent weight per cent NaCl. Inclusions from precious metal rich, middle-stage veins have salinities ranging from 0.5 to 3.0. Salinities of base metal rich, middle-stage veins are slightly greater than the precious metal rich, ranging from 0.4 to 4.5. Late-stage veins have salinities from 0 to 1.5 NaCl. Type 1 inclusions have slightly greater salinities than Type 2 inclusions.

During freezing of inclusions, no unusual solid phases such as CO_2 hydrate were observed. Because the most abundant solute in the inclusions is normally NaCl, the commonest hydrate observed is hydrohalite ($NaCl \cdot 2H_2O$) The low Tm_{ICE} values indicate the presence of minor hydrohalites in the inclusions.

Laser-Excited Raman Microprobe Analysis

The laser-excited Raman spectrometer is a generally non-destructive, *in situ*, semiquantitative analytical technique in which the laser beam is focused on a doubly polished wafer to give pinpoint analysis of individual fluid inclusions (Delhaye and Dhamelicourt, 1975; Rosasco *et al.*, 1975).

The laser-excited Raman Spectrometer at Surface Science Western is a Dilor OMARS 89, equipped with an optical multichannel analyzer. As a light beam passes through a medium the vibration spectra are used to characterize the nature and structure of compounds in individual inclusions. A small part of the photon scattering that results from this interaction is inelastic, Raman scattering, and gives rise to radiation of displaced frequencies, or wavenumbers, that are characteristic of the scattering substance.

Inclusions from all three periods of veining are predominantly H_2O with minor, variable CO_2 (Figure 2-12-6). Inclusions from middle-stage veins also have trace abundances of N_2 and CO, however, high H_2O background makes them difficult to detect. The abundances of volatile components for each vein type are given in Table 2-12-3.

TABLE 2-12-2
FLUID INCLUSION SALINITIES

SAMPLE #	Location	Elevation	Stage	Type	No. Inc	Saln. Range	Mean Saln.
SP-165	DH 85-69	615.76	I	P	4	0.1-0.9	0.55
SP-166A	DH 87-248	652.42	I	P	4	0.3-1.5	0.65
SP-50S	Surface	664.66	II-P.M.	P	5	0.8-2.2	1.5
SP-59U	110 Level	581.82	II-P.M.	P	5	0.6-2.2	1.2
SP-207A	2-Level	530.31	II-P.M.	P	5	1.0-3.0	1.45
SP-204A	3-Level	478.79	II-P.M.	P	4	1.0-2.8	1.75
SP-136A	Surface	666.67	II-B.M.	P	5	1.0-3.5	2.1
SP-203A	3-Level	478.79	II-B.M.	P	5	0.8-3.0	2.05
SP-214A	5-Level	312.12	II-B.M.	P	4	1.2-4.4	2.3
SP-200A	6-Level	234.24	II-B.M.	P	5	1.5-4.1	2.4
SP-118	DH 84-35	647.58	III	P	4	0.2-0.8	0.6
SP-112A	DH 87-235	602.73	III	P	4	0.4-1.4	0.75

* I— EARLY-STAGE
II— MIDDLE-STAGE
BM— BASE METAL RICH
PM— PRECIOUS METAL RICH
III— LATE-STAGE

The ability to estimate mole fractions of species in a multicomponent fluid inclusion is based on theoretical calculations (Wopenka and Pasteris, 1986, 1987). Taking the intensity calibration of the spectrometer into acccount, these formulae reduce to the expression:

$$Xi = (A_i/\sigma_i f_i)/(\Sigma A_i/\sigma_i f_i)$$

where: A_i = average area of the Raman peak measured by planimeter, multiplied by the intensity (counts sec^{-1});

i = effective relative cross-section of Raman scattering; and,

f_i = instrument correction factor.

In practice one makes several simplifying assumptions and uses the area of the Raman peaks and the scattering cross-sections, a measure of the scattering efficiency of a particular vibrational mode for those species, to calculate their relative

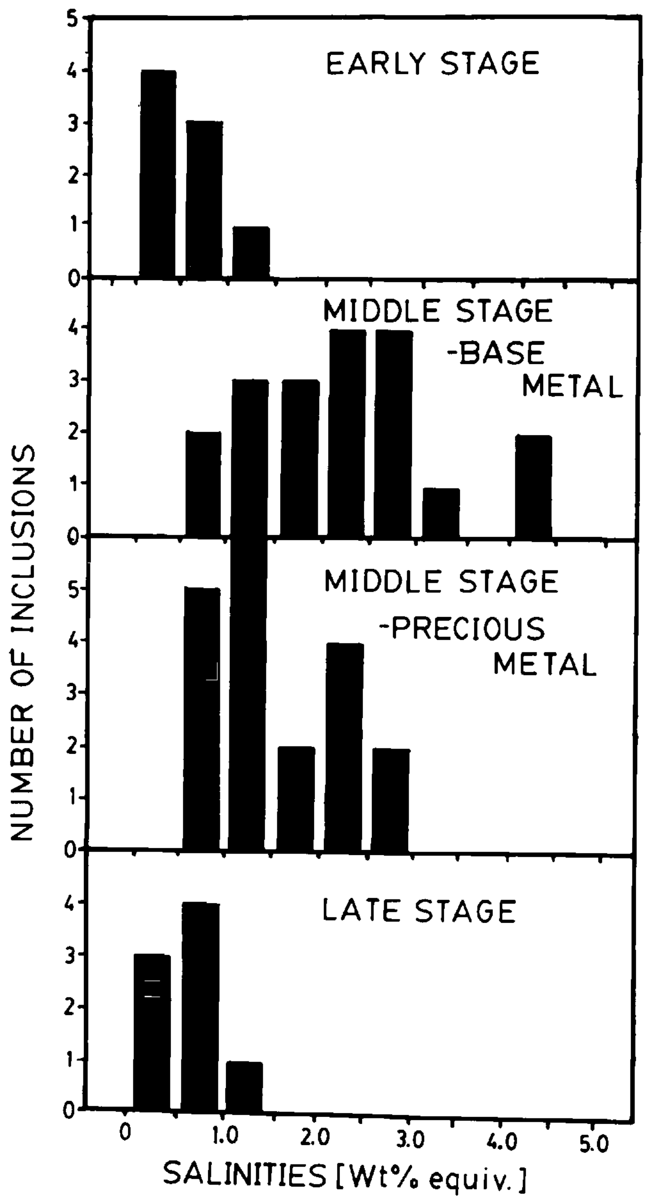

Figure 2-12-5. Histogram of salinity for primary fluid inclusions from all stages of veining.

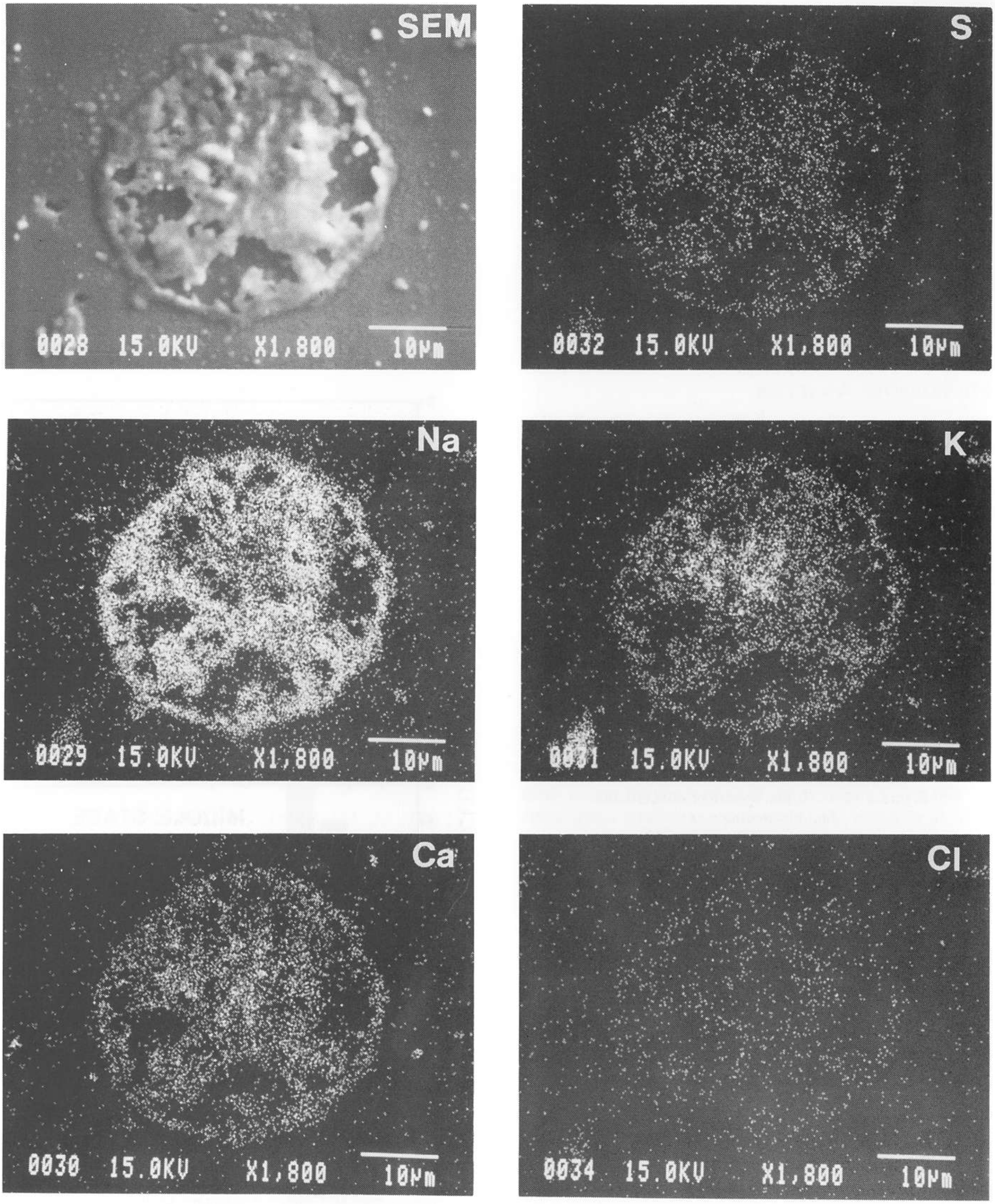

Plate 2-12-2. SEM photomicrograph of a decrepitation precipitate from a precious metal-rich, middle-stage vein. Sample SP-214-A. X-ray dot maps with the EDS system for S, Na, K, Ca, Cl.

molar proportions. A CO_2 peak from Sample SP-14, a base metal rich, middle-stage vein has the following values:

$A_{(CO2)} = 97$;

$\Sigma A = 229\ 534$;

$\sigma_i = 1.21$ (Schrotter and Klockner, 1979).

The instrument correction factors are not required because the Raman spectrometer has a multidiode array. Thus:

$$Xi = (A_{CO2}/\sigma_i f_i)/(\Sigma A_i/\sigma_i f_i),$$
$$= (97/1.21)/(229{,}534/1.21)$$
$$= 4.22 \times 10^{-4} \text{ moles.}$$

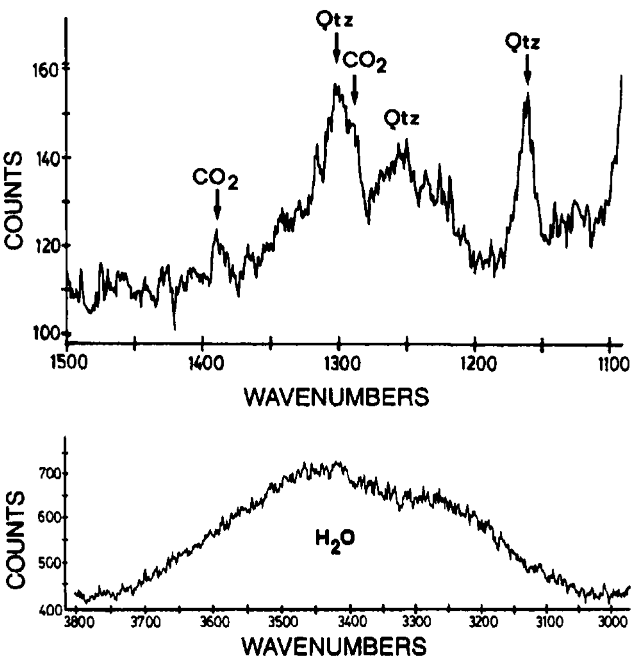

Figure 2-12-6. Raman spectra of middle-stage, sulphide-bearing veins.

A. The presence of minor CO_2 in a H_2O-dominant inclusion. CO_2 peaks are at 1292 and 1387 cm^{-1}. Sample number SP-14.

B. Pure H_2O inclusion in quartz. The small, narrow peaks are extremely weak and have not been identified. Both spectra are from a multichannel scan, 10 points per second, points spaced 0.3 cm^{-1}. Sample number SP-50-S.

TABLE 2-12-3
LASER RAMAN MICROSCOPE — SPECIES OF GASES

	GAS SPECIES — VOLUME %						
	SO_2	CO_2	CO	N_2	H_2S	CH_4	H_2O
EARLY-STAGE	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	<99
MIDDLE-STAGE							
PRECIOUS METAL RICH	N.D.	<1	N.D.	TRACE	N.D.	N.D.	99
BASE METAL RICH	N.D.	1-2	TRACE	TRACE	N.D.	N.D.	98
LATE-STAGE	N.D.	N.D.	N.D.	N.D.	N.D.	N.D.	99

* N.D. — Not detected.

Therefore the CO_2 mole fraction in a precious metal rich, middle-stage vein is 4.22×10^{-4}. This low CO_2 abundance should have little affect on pressure corrections of homogenization temperatures and depth determinations. Salts such as NaCl, KCl and $CaCl_2$ are not appreciably Raman-active either in pure crystalline form or in aqueous solution, so electron microprobe analysis of fluid inclusion precipitates has been attempted.

ELECTRON MICROPROBE

After heating and freezing temperatures have been measured, the temperature of the heating stage is raised at a rate of 15° to 20°C per minute to 570°C. The inclusion is ruptured and the decrepitating fluid forms a precipitate on the surface of the slide. The decrepitation precipitates generally have a yellowish brown colour and a shape similar to that of a volcanic crater (Plate 2-12-2).

Semiquantitative analysis of individual fluid inclusions can be obtained by analyzing the precipitates, which are the bulk of the nonvolatile compounds in the inclusion. The Jeol JXA-8600 electron microprobe, in the energy dispersive mode (EDS) with an accelerating voltage of 15 kilovolts has been used for scan analysis of the decrepitation precipitates (Table 2-12-3).

A scanning microscope was used to define the chemical constituents within the precipitates using a conventional x-ray dot map in scanning mode (Plate 2-12-2). For all stages of veining, sodium, calcium and sulphur scans have the most intense patterns with dispersed patterns for potassium, chlorine and manganese. Both early and late-stage veins, have no concentrations of chlorine, manganese and fluorine. In both base and precious metal rich veins, silver precipitate is defined by electron backscatter and intense x-ray maps (Plate 2-12-3).

VERTICAL THERMAL GRADIENTS

Homogenization temperatures of quartz in middle-stage veins and breccias are contoured as 210°, 220°, and 230°C isotherms and superimposed on a longitudinal section of Silbak Premier mine workings (Figure 2-12-7). These isotherms define a broad thermal gradient that parallels the plunge of mineralized zones. There is a systematic decrease in temperature with depth, however, there is a low-temperature zone between 5 and 6-level. Homogenization temperatures have a greater variation lower in the deposit.

PRESSURE CORRECTIONS AND PALEODEPTH ESTIMATES

Where hydrothermal fluids have not been boiled, fluid inclusion homogenization temperatures require a pressure correction to obtain the trapping temperature. Most inclusions have trapped fluids at a pressure-temperature combination above the liquid/vapour curve (Figure 2-12-8). A bubble does not occur in the inclusion until the pressure and temperature have dropped to the liquid/vapour curve. Because an inclusion has a constant density, the cooling path follows an isochore corresponding to its density.

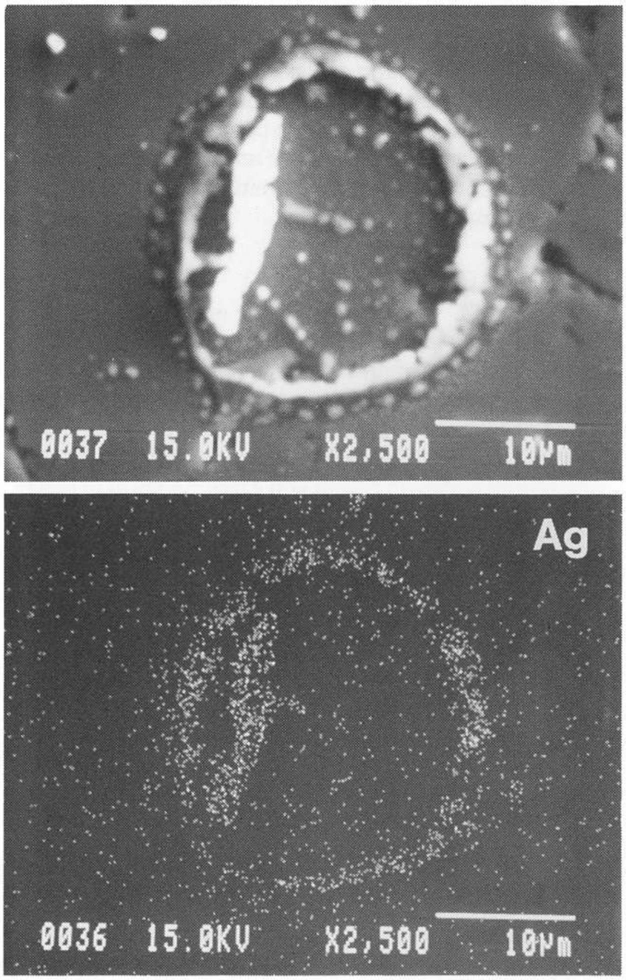

Plate 2-12-3A. Decrepitation precipitate of a precious metal rich, middle-stage vein in the backscatter electron scanning mode (A). Sample SP-47-U. The backscatter image provides a map of the spatial variation of average atomic number. Silver has an intense white backscatter image. B. X-ray dot map of silver in the precipitate.

In the case of homogenization by disappearance of vapour bubbles, the minimum pressure is read directly from the corresponding isopethal PT section of the H_2O-NaCl system (Figure 2-12-8). Using the intersecting-isochore method, a trapping temperature of 226°C (T_1) has a pressure (P_1) of less than 10 bars. An independent pressure estimate, using degree of fill and temperature (Roedder and Bodnar, 1980) indicates a true confining pressure (P_2) of 35,000 kilopascals (350 bars). Consequently, the pressure-corrected trapping temperature (T_2) is 255°C for middle-stage base metal rich veins. Trapping temperatures for other stages of veining are given in Table 2-12-4.

Identification of even small amounts of dissolved gas such as CO_2, CO, N_2, H_2S and CH_4 in inclusions can significantly affect the pressure estimates (Pasteris *et al.*, 1986). The addition of even small amounts of CO_2 to H_2O-NaCl significantly raises the vapour pressure and concomitantly, the calculated depth of formation. Because inclusions are H_2O-dominant in all stages of veining, with only trace amounts of

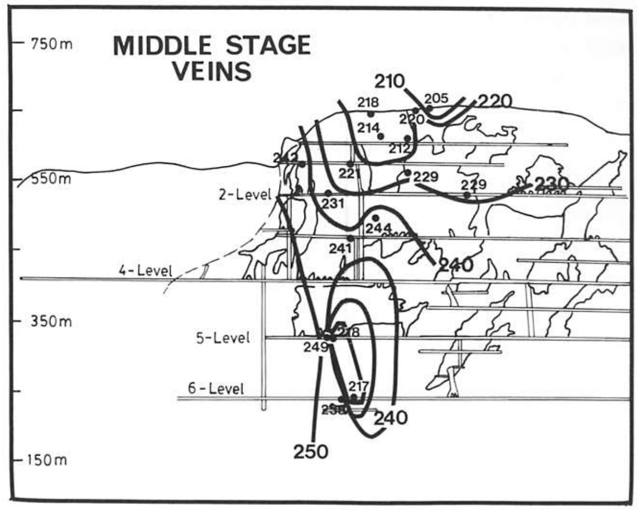

Figure 2-12-7. Longitudinal section, parallel to the Main zone trend (Figure 2-12-2) of the Silbak Premier mine workings (Grove, 1971). Shaded areas are shrinkage stopes. Contoured homogenization temperature isotherms of fluid inclusions, from both precious and base metal, middle-stage veins, are superimposed on the section. Isotherms are skewed updip, parallel to the stopes.

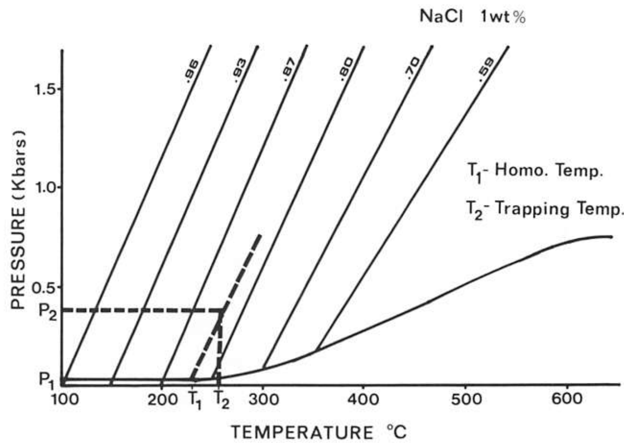

Figure 2-12-8. P-T section for 1 weight per cent NaCl aqueous solution. Density (g/cm³) of aqueous NaCl solutions contoured on pressure-temperature (P-T) coordinates. Isochores are lines of constant volume and hence constant density. The degree of filling indicates the density of the ore-forming fluid is 0.83 (Roedder and Bodnar, 1980).

TABLE 2-12-4
CONVERSION OF PRIMARY HOMOGENIZATION TEMPERATURES TO TRAPPING TEMPERATURES.

	HOMOGENIZATION TEMPERATURE °C	TRAPPING TEMPERATURE °C
EARLY-STAGE VEINS	204.6	239.8
MIDDLE-STAGE VEINS:		
PRECIOUS METAL RICH	225.4	256.3
BASE METAL RICH	228.81	258.2
LATE-STAGE VEINS	208.5	234.1

CO_2, N_2 and CO, temperature and pressure corrections are small.

Paleodepth estimates determined from fluid inclusions are based on the pressure-temperature-salinity relationships (Haas, 1971) and related to a maximum depth for an incipient boiling NaCl brine. The temperature-pressure-salinity data from Silbak Premier indicate a fluid pressure corresponding to a depth of 600 metres in a slightly saline fluid. Estimates determined by this method can have large uncertainties because of the presence of water vapour bubbles, carbon dioxide and other gases and changing hydrostatic versus lithostatic load conditions.

INTERPRETATION

Self-sealing and rebreaking of hydrothermal breccias in active geothermal systems is well described from several locations (Henley, 1985). The wide range of homogenization temperatures of primary inclusions in quartz may reflect a continuum of several hydrothermal episodes rather than one specific event. Many of the quartz veins examined in this study are repeatedly fractured by recurrent movements of the veins, resulting in abundant planes of pseudosecondary and secondary inclusions.

Fluid inclusions in early-stage veins have low homogenization temperatures and salinities and are spacially and temporally associated with intrusive porphyritic dacite. There is a significant increase in homogenization temperatures and salinities associated with base and precious metal deposition. Late-stage veins are produced in extension zones during the waning stages of hydrothermal activity.

The inclusions are aqueous and have dissolved sulphur, sodium, calcium, potassium, manganese and ore metals, such as copper and silver, which form in the decrepitation precipitates. Large concentrations of sulphur in calcium-rich fluid inclusions suggest the fluids may have provided the necessary sulphur to precipitate silver as a sulphide or sulphosalt.

The mixing of meteoric waters with hot, metal-bearing fluids, is the most likely mechanism for base and precious metal deposition, because there is no coexistance of liquid-rich and vapour-rich inclusions to indicate boiling. Decreasing temperature, and increasing pH with mixing, causes bisulphide complexes to release metals and precipitate sulphide minerals (Spycher and Reed, 1989). Aqueous sulphide may be sufficiently depleted by sulphide precipitation to cause gold to precipitate from $Au(HS)^{-2}$.

Reaction with the wallrocks could have changed the pH or decreased the H_2S activity of fluids, but this is likely to have affected the vein system evenly, resulting in widespread gold in wallrock rather than the observed spotty deposition of high-grade ore shoots. The restriction of high-grade ore to late fractures in the veins also suggests that the ore-forming fluid had limited interaction with the wallrock.

In contrast, separation of the ore fluid into immiscible phases, accompanied by the escape of CO_2 vapour, is more localized and abrupt and may have formed much of the vein ore. These effects contribute to the "telescoping" or vertical changes in base and precious metal abundances in epithermal deposits. The same effects could also produce sulphide-rich, precious metal-rich and barren stages in ore formation.

If salinity is relatively uniform, vertical gradients in T_h indicate density gradients and hence the direction of fluid movement (Roedder, 1984). The change from base to precious metal rich assemblages and declining sulphide abundances coincide with temperature decreases. Declining temperatures in the lower part of the deposit may also indicate a bottoming of the hydrothermal system. Unfortunately, not enough fluid inclusion data are available to determine if there is a correlation between the variation in the ore grade and the composition of the associated inclusions.

The temperatures and salinities of base and precious metal vein stages at Silbak Premier are reminiscent of adularia-sericite epithermal deposits (Heald *et al.*, 1987), but in contrast have large proportions of base metals and spatial association with intrusions. The Silbak Premier deposit has common characteristics with silver-gold deposits in the Iskut and Unuk River valleys and may represent a distinct epithermal metallogenic district.

APPLICATION IN EXPLORATION

With increasing sophistication of analytical techniques and availability of fluid inclusion data, the quantitative chemical evolution over a range of depths can be used as a guide to ore and provide a way of comparing ore deposit models. Because halos of fluid inclusions are more widespread tban correponding halos of alteration minerals (Bodnar, 1981), they are larger targets for mineral exploration provided appropriate techniques are used. Homogenization temperatures of fluid inclusions in hydrothermal systems increase as new inclusions form closer to the ore (Roedder, 1984). Reconstruction of isotherms may therefore hold promise as an exploration tool in similar silver-gold vein deposits. Increasing concentrations of salt, carbon dioxide and other gases also indicate proximity to mineral occurrences.

CONCLUSIONS

The incorporation of fluid inclusion data with surface and underground mapping leads to the following geologic reconstruction of the Silbak Premier orebodies:

(1) Conjugate fault sets are the locus of a series of porphyritic dacite intrusions into andesite flows, breccias and tuffs. Hydrothermal fluids followed emplacement of porphyritic dacite, initially into diffuse areas of crackle breccia along the margins of the dacite, subsequently into crosscutting conjugate fault sets. Base and precious metal precipitates along these faults were sucessively sealed and broken. Fluids barren of metal and sulphur were emplaced along tensional features in the last stage of the hydrothermal event.

(2) The close proximity, in both time and space, of the Texas Creek granodiorite to porphyritic dacite and metal-bearing veins and breccias suggests a partial magmatic source for heat, fluids and metals. These magmatic fluids may mix with meteoric water or, to a lesser extent, seawater.

(3) Vein-hosted base and precious metals were deposited at pressure-corrected temperatures of 250° to 260°C at a minimum depth of 500 metres. Typically, these veins formed in hydrous fluids with salinities less than 4.5

equivalent weight per cent NaCl. A limited number of inclusions contain liquid carbon dioxide, with lesser nitrogen and carbon monoxide gas. Sodium, calcium, sulphur and silver are the dominant dissolved constituents with lesser amounts of chlorine, potassium, manganese, iron and magnesium. Metal transport and deposition in hydrothermal fluids coincided with thermal peaks and fluids of greatest salinity. Both early and late-stage veins have lower temperatures and salinities.

(4) Where low-temperature hydrothermal fluids have elevated sulphur concentrations, thiosulphides are the dominant complexing agent. The mixing of magmatic, H_2S-rich hydrothermal fluids with meteoric waters decreases temperatures and increases pH and oxygen content of the hydrothermal fluids causing bisulphide complexes to precipitate sulphides, sulphosalts and finally native metals. Restricted lateral distribution and changes from barren to base metal rich, then precious metal rich assemblages with elevation, coincide with a moderating thermal and salinity gradients. These gradients, in combination with declining sulphide abundances, reflect the declining solubilities first of base metals then of precious metals.

ACKNOWLEDGMENTS

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NOTES

GEOLOGY OF NICKEL MOUNTAIN AND THE E&L NICKEL-COPPER PROSPECT (104B/10E)

By Kirk D. Hancock

KEYWORDS: Economic geology, Nickel Mountain, Iskut River, Hazelton Group, gabbro, pentlandite.

INTRODUCTION

The E&L deposit, the second largest known nickel resource in British Columbia, is located on Nickel Mountain in the Iskut River district north of Stewart. The deposit consists of pyrrhotite, pentlandite and chalcopyrite hosted in an olivine gabbro stock that intrudes Lower Jurassic sediments and volcanics. Exploration has identified 2.9 million tonnes grading 0.80 per cent nickel and 0.62 per cent copper with anomalous values in gold, silver and platinum group elements (Quartermain, 1987; Sharp, 1968). Fieldwork for the present study was carried out in 1988 and 1989 as part of an ongoing regional mapping project in the Iskut-Sulphurets area.

Nickel Mountain is situated in the headwaters of Snippaker Creek (Figure 2-13-1), 27 kilometres east-southeast of the Bronson Creek airstrip and 5 kilometres east of the 950-metre Snippaker Creek airstrip. Access to the property is either by helicopter or on foot.

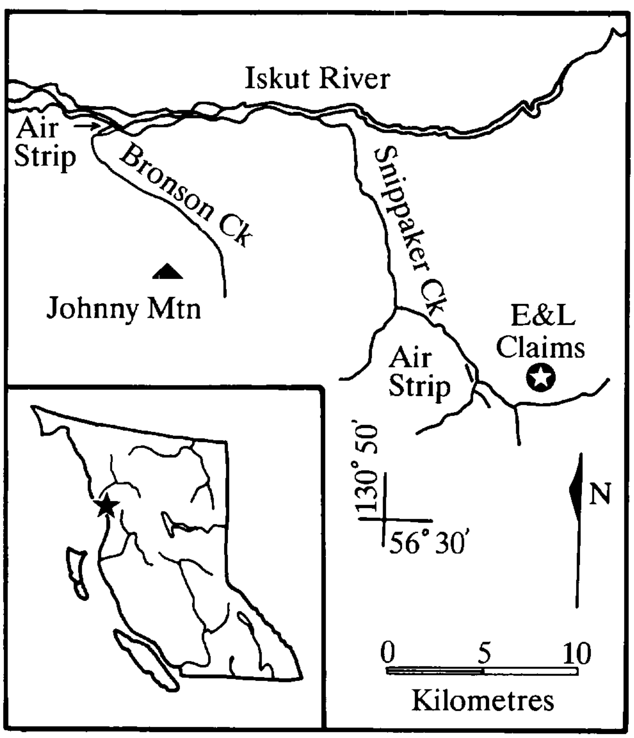

Figure 2-13-1. Location of Nickel Mountain and the E&L claims.

The Nickel Mountain stock crops out at 1850 metres elevation along the crest of a steep ridge which slopes south toward Snippaker Creek and continues northward as a series of razorback ridges separating glaciers and snowfields. A large ice-filled cirque abuts the northeast edge of the showings. The upper slopes are generally talus fans free of vegetation, slopes below 1100 metres are well timbered.

EXPLORATION HISTORY

Nickel Mountain was initially prospected in 1958 by Ed and Lela Freeze for the BIK syndicate (Silver Standard Mines Limited, Kerr-Addison Gold Mines Limited and McIntyre Porcupine Mines Limited.) The E&L 1 and 2 claims were staked at that time followed by the E&L 3 to 28 claims in the winter of 1964-65. Geological mapping, geochemical sampling, hand trenching and x-ray drilling were carried out in 1965 (Hedley, 1965). Further mapping, trenching and drilling were done in 1966 and a small airstrip was built 5 kilometres downstream on Snippaker Creek (Jeffery, 1966). A tote road provides access to the property.

Sumitomo Metal Mining Corporation optioned the E&L claims in 1970 and began an underground exploration program. A 450-metre adit was collared 390 metres below the surface showings and driven toward the mineralized zone (Hirata, 1972). Nine underground diamond-drill holes tested the downward extent of mineralization.

Subsequent activity on the property has been minor. In 1986 and 1987 ground magnetometer as well as airborne magnetic and VLF electromagnetic surveys were conducted by Western Geophysical Aero Data Ltd. to outline mineralization beneath the cirque to the northeast (Hermany and White, 1988). In 1986 selected grab samples were analyzed for platinum group elements.

Work on the E&L claims has identified three zones of nickel-copper mineralization exposed at surface and three additional zones underground. Published reserves (Anonymous, 1976; Quartermain, 1987; Sharp, 1968) are presented in Table 2-13-1.

Platinum group element values ranging from less than 50 to 400 ppb platinum and from less than 5 to 415 ppb

TABLE 2-13-1
INDICATED AND INFERRED RESERVES

Category	Tonnes (000's)	Ni %	Cu %	Au g/t	Ag g/t
Trench and drill-					
Indicated	1734	0.80	0.62	0.34	6.8
Inferred	1194	0.80	0.62	0.34	6.8

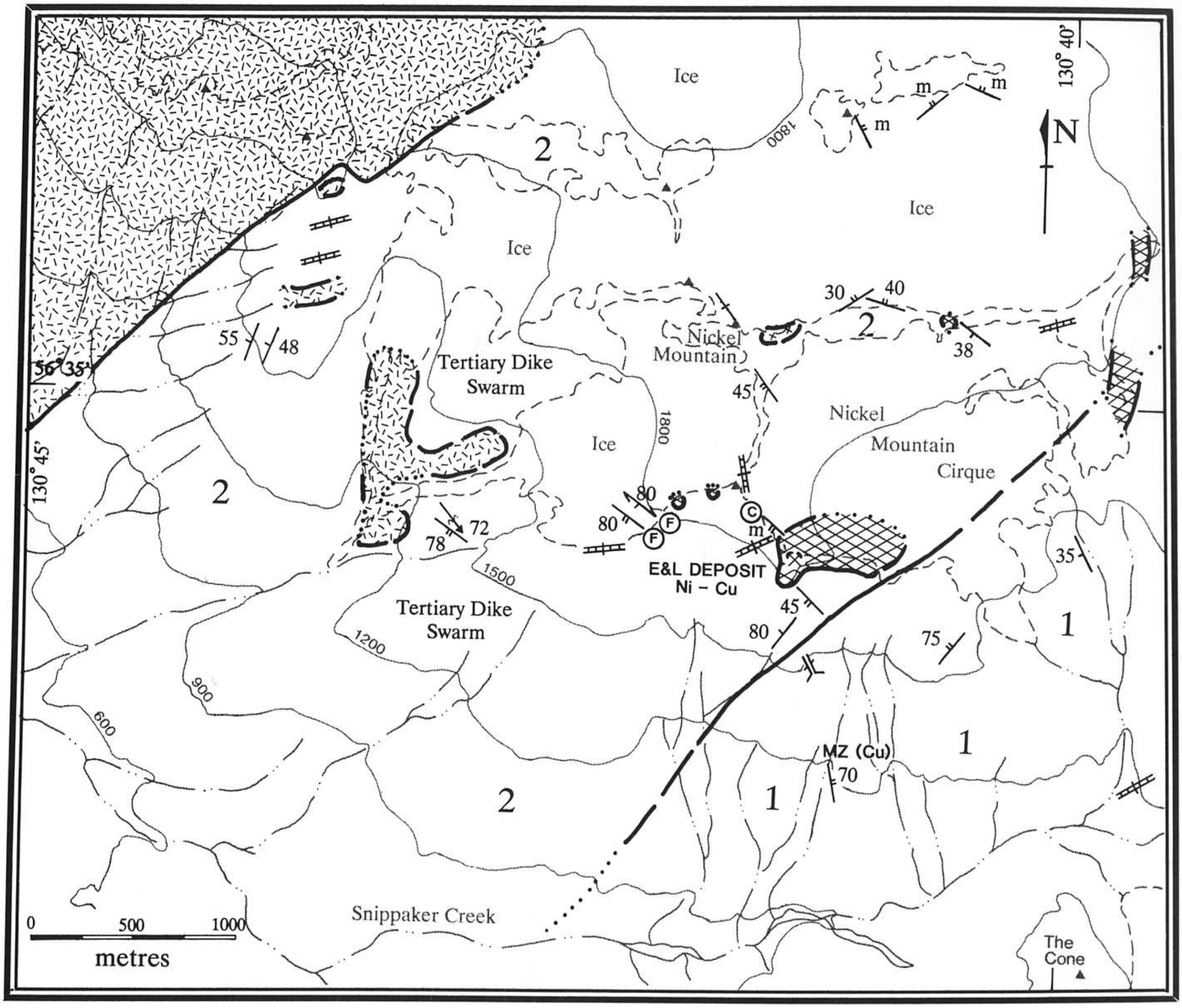

Figure 2-13-2. Interpreted geology of the Nickel Mountain area and the E&L nickel-copper showings.

palladium were obtained from grab samples collected in trenches by Consolidated Silver Standard Mines Limited in 1986 (Quartermain, 1987) and ministry geologists in 1988.

GEOLOGICAL SETTING

Regionally, strata trend northeast with gentle to moderate northwest dips. The Nickel Mountain gabbro intrudes a thick sedimentary and volcanic sequence of the Lower Jurassic Hazelton Group. A large monzodiorite pluton intrudes the volcanosedimentary package 3 kilometres northwest of the deposit. Regional deformation postdates the pluton. Late postdeformation mafic dikes crosscut all rocks in the area.

Sedimentary strata hosting the mineralized gabbro stock are assigned to the Lower to Middle Jurassic Salmon River formation based on lithology, stratigraphic position and fossils. The rocks are black, evenly laminated, rusty weathering, very fine grained argillaceous sandstones, siltstones and mudstones. Small lenses of limestone-chip breccia occur locally and consist of black mudstone matrix with limestone chips up to 4 centimetres across. Lenses are usually less than 1 metre thick over lengths of up to 15 metres. The basal calcareous grit and fossiliferous limestone member of the Salmon River formation type section (Alldrick, 1985; Alldrick *et al.*, 1987) has not been identified in the Nickel Mountain area. However, due to the presence of the limestone-breccia lenses and proximity to the underlying volcanic package, it is inferred that the sediments at Nickel Mountain are lowermost Salmon River formation (Figure 2-13-2).

Ammonoid fossils are preserved in the black argillites. Samples collected in 1965 were identified as *Hildocerataceae*, some similar to *Haugia*, suggesting a Toarcian age (GSC Location 86273; Grove, 1986). The samples were poorly preserved and additional samples were collected in 1989.

LEGEND

TERTIARY

Medium-grained, grey diorite dikes and small plugs (?), 1 to 10 metres across.

MIDDLE JURASSIC TO MIDDLE CRETACEOUS

NICKEL MOUNTAIN GABBRO

Medium to coarse-grained olivine gabbro; composed of plagioclase, pyroxene and olivine. Orbicular textures are present in both pyroxene and plagioclase.

JURASSIC

LEHTO PORPHYRY

Medium to coarse-grained quartz monzodiorite; composed of white plagioclase, pink potassium feldspar, grey quartz, black hornblende and minor biotite. Locally, the potassium feldspar phenocrysts are up to 3 centimetres long. The border phase is commonly a grey/green diorite.

LOWER TO MIDDLE JURASSIC

SALMON RIVER FORMATION

2 — Rusty weathering, black, well-laminated, very fine grained argillaceous sandstones, siltstones and mudstones. Locally contains small lenses of limestone-chip breccia. The basal calcareous grit and limestone are not present. Ammonoid fossils are present but not abundant.

BETTY CREEK FORMATION

1 — Felsic to intermediate volcanics; light to medium green, dacitic ash tuffs and lapilli tuffs, commonly plagioclase porphyritic. Contains interbeds of black siltstone and very fine grained sandstone. Also contains some hematitic clastic layers.

SYMBOLS

Contact (known, approximate, assumed)
Bedding (tops known, tops unknown, vertical)
Mineral prospect
Mineralized showing — MZ (Cu)
Macrofossil sample location — F
Conodont sample location — C
Adit
Stream
Contour (in metres) — 1500
Peak
Permanent snow/ice boundary

A thick sequence of felsic to intermediate volcanics and thin interbedded sediments underlies the Salmon River formation. The package consists primarily of dacitic ash tuffs and lapilli tuffs, commonly plagioclase porphyritic. These rocks are light to medium green with some dark green andesitic layers. Tuffs are fine to medium grained and massive to locally well bedded. Interbedded sedimentary members are usually black, thin-bedded fine sandstone and siltstone. These thin sedimentary units are distributed randomly throughout the volcanics. This volcanic sequence can be correlated with the Lower Jurassic Betty Creek formation; however, only minor hematitic clastic sedimentary units characteristic of the Betty Creek type section, are present on Nickel Mountain.

The volcanic strata are dominated by dacitic units in this area and cannot be usefully subdivided. Thus, no distinction has been made between the Betty Creek and Mount Dilworth formations in this study and some of the formational divisions established to the east and south may not be valid in the Nickel Mountain area.

The Nickel Mountain gabbro is a unique lithology in the Stewart-Iskut district. The gabbro intrusions consist of four small plugs less than 100 metres wide at surface, one large stock approximately 800 metres across and a dike swarm approximately 250 metres wide, all occurring along a 3 kilometre northeast trend. The large stock and dike swarm may be connected as they are separated by a large ice-filled cirque.

Petrographic study shows that the dominant mineral phases are plagioclase, pyroxene and olivine. Orbicular textures are common in gabbro adjacent to the mineralizaton (Plate 2-13-1). The presence of fractures and sheared mineral grains indicates the gabbro has undergone deformation. The stratigraphic and structural evidence suggests the intrusion of the gabbro postdates the Lower to Middle Jurassic sediments and predates the mid-Cretaceous deformation. This brackets the age of intrusion at 185 to 110 Ma.

A large stock of porphyritic quartz monzodiorite, the Lehto porphyry, truncates sedimentary strata of the Salmon River formation north and northwest of Nickel Mountain. Lehto porphyry is interpreted as a Jurassic pluton, based on alteration, deformation and presence of hornblende and potassium feldspar megacrysts; a potassium-argon analysis is in progress. The rock is typically medium to coarse grained with white plagioclase, pink potassium feldspar, grey quartz, black hornblende and lesser biotite. Locally the potassium

Plate 2-13-1. Orbicular textures in the Nickel Mountain gabbro.

feldspars occur as phenocrysts up to 3 centimetres long. The margin of the pluton is fine to medium-grained diorite. The contact with the surrounding sedimentary rocks varies from sharp, with hornfelsed country rocks, to irregular with included stoped sedimentary blocks.

Medium-grained diorite dikes crosscut all other units in the area and are most probably Tertiary in age. They are typically rusty weathering, dark grey diorites, 1 to 10 metres wide with marked variations in strike. They are extensive and continuous but generally trend northeast with subvertical dips. Fresh, light green, fine-grained dikes with needle-like hornblende phenocrysts have also been identified. One crosscuts sedimentary rocks and a gabbro plug west of Nickel Mountain.

Regional deformation has been dated at approximately 110 Ma in the Stewart area (Alldrick *et al.*, 1987). At Nickel Mountain there is a general shortening along a northeast axis. Sediments have taken up most of the stress in open, cylindrical folds. Stereonet plots (Figure 2-13-3) indicate one phase of folding with a fold axis of 15°/305° and an axial plane of 126°/80° southwest. Weak penetrative axial planar cleavage is present in the fine-grained sediments. Volcanic units are block faulted with individual blocks generally undeformed. Interbedded sediments show small-scale folding. Tertiary northwest-southeast extension controlled intrusion of the diorite dikes.

MINERAL DEPOSITS AND ALTERATION

Nickel and copper sulphide mineralization occurs exclusively within the central gabbro body. At surface there are three major mineralized zones. The Northwest and Southeast zones are the most significant; both are roughly triangular with dimensions of 60 by 45 by 45 metres (Sharp, 1965). The East zone is considerably smaller and less continuously mineralized than the other two. Surface and underground drilling indicate an irregular pipelike, possibly interconnected, form to the three zones at depth. Structural data collected by Sumitomo Metal Mining Corporation indicate a 70° southwest plunge to the mineralized pipes (Hirata, 1972) (Figure 2-13-3). Vertical extent of the mineralization has been proved to a depth of 210 metres and the zones remain open laterally and to depth.

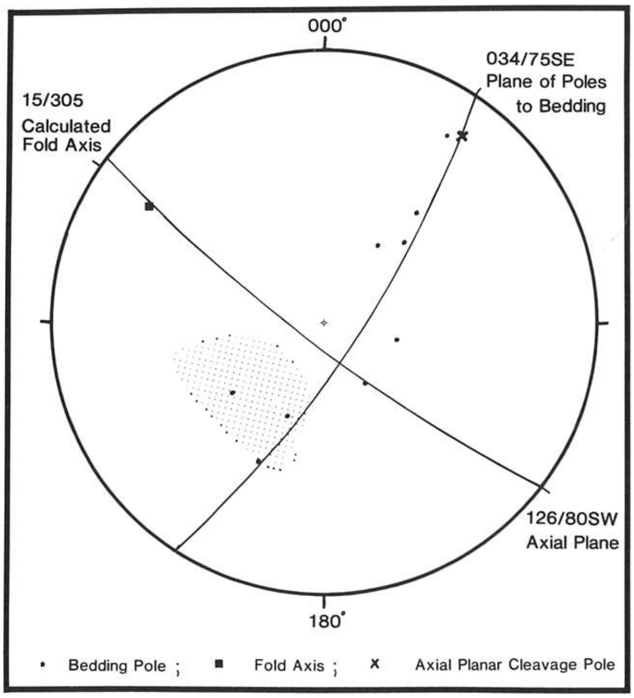

Figure 2-13-3. Stereographic projection of structures in the Salmon River formation, Unit 2. Shaded area is contoured zone of projected plunge of mineralization in the Nickel Mountain gabbro (after Hirata, 1972).

Mineralization is localized along the margins of the intrusion as irregular pipelike zones of veins, disseminations and massive lenses. The mineral textures and spatial relationship of the sulphides to the gabbro indicate that the mineralization is magmatic. Pyrrhotite, pentlandite and chalcopyrite are the dominant sulphides with minor amounts of pyrite, magnetite and "siegenite". Nickel occurs predominantly in pentlandite but it is also present in a secondary nickel sulphide with a composition between siegenite $(Co,Ni)_3S_4$, and violarite $(Ni, Fe)_3S_4$. Chalcopyrite shows minor supergene alteration where covellite locally forms rims around the chalcopyrite and occasionally completely replaces it. Trace amounts of cobalt, noted in assay results, probably occur in both the pentlandite, replacing iron, and the siegenite (Cabri, 1966).

The central gabbro plug is massive with some local shears and faults. The unmineralized part of the intrusion has plagioclase compositions of An_{30-60}, andesine to labradorite; mineralized zones have plagioclase compositions of An_{85-95}, bytownite to anorthite (Hirata, 1972). Hirata concluded that mineralization was restricted to zones where the plagioclase compositions were An_{85-95}. Large lenses of massive, undeformed sparry calcite, 10 to 50 centimetres wide, are present along the margins of the gabbro and tail off into thin stringers.

Gabbro within and around mineralized zones shows extensive alteration; olivine grains are partially or totally altered to serpentine, most plagioclase is altered and abundant chlorite, amphibole, biotite, carbonate, epidote and prehnite occur throughout the matrix (Hirata, 1972).

Alteration of the host sediments is limited to an aureole, less than 20 metres wide, of intense bleaching to a light green colour and partial loss of textures. Previous mapping identified these thermally altered sediments as either chert, siliceous tuffs or metadiorite.

CONCLUSIONS

The Nickel Mountain area is the site of the second largest nickel deposit in British Columbia. The host rock gabbro and its nickel-copper deposit were emplaced during mid-Jurassic to mid-Cretaceous time and are therefore unrelated to the main Lower Jurassic and mid-Tertiary plutonic suites of the region. This suggests that the extensive Jurassic Bowser basin stratigraphy to the east is prospective terrain for similar deposits.

ACKNOWLEDGMENTS

Mr. R. A. Quartermain of Consolidated Silver Standard Mines Limited allowed review of company files. His cooperation and interest are gratefully acknowledged. Geological mapping was completed by J.M. Britton, S.N. Hiebert and the author in 1989 and this manuscript also contains data from I.C.L. Webster and C.W.P. Russell, 1988.

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NOTES

GEOLOGY OF THE COLAGH PROSPECT, UNUK MAP AREA (104B/10E)

By Mary E. MacLean

KEYWORDS: Economic geology, Betty Creek formation, Hazelton Group, Iskut River, epigenetic massive sulphide, gold.

INTRODUCTION

The Colagh prospect (MINFILE 104B 352) is a recently discovered polymetallic occurrence located in the Iskut-Sulphurets area of northern British Columbia (Figure 2-14-1). The prospect is 7 kilometres east of the Snippaker airstrip and directly south of the Copper King Glacier. At 1370 metres elevation in a glacial cirque, the area is free of snow only in late summer. Outcrop is scattered and glacial debris is ubiquitous. Vegetation is restricted to heather and scrub-brush on the adjacent slopes.

The prospect was discovered in August 1988 by C.P.W. Russell during routine mapping as part of the ongoing Iskut-Sulphurets project. The area was visited by the author for three days in July and August 1989, when detailed mapping and sampling were carried out.

Omega Gold Corporation and Ecstall Mining Corporation jointly own the Macgold claims covering the Colagh prospect, and to date have established a grid, completed induced polarization and geochemical surveys, and chip sampled in three trenches. Most of this work has been done since the author's last visit in late August.

STRATIGRAPHY AND INTRUSIVE ROCKS

Approximately 3 square kilometres surrounding the prospect were mapped. The rocks have been subdivided lithologically based on relative abundances of dacitic and andesitic volcanics, and sedimentary rocks (Figure 2-14-2).

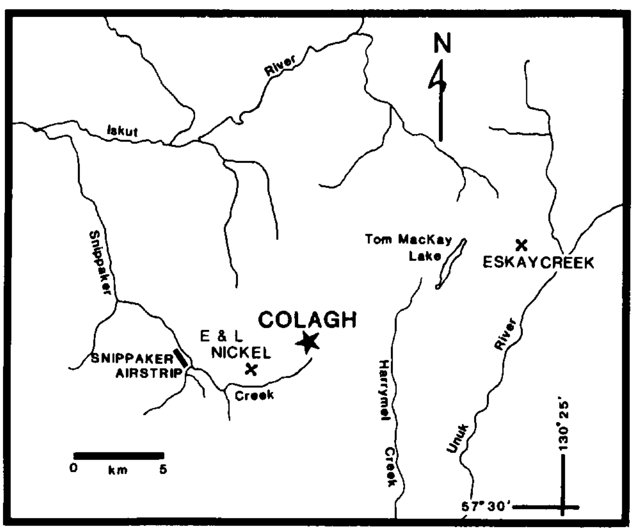

Figure 2-14-1. Location of Colagh prospect and nearby properties in Iskut River area.

The lowermost volcanic Unit (1A) consists of bedded and massive andesitic ash to lapilli tuffs, and feldspar porphyries.

Moving up-section, Unit 2D is primarily dacitic ash, crystal and lapilli tuffs, commonly well laminated with occasional graded bedding and welded structures (fiammé). A distinctive massive lapilli tuff at least 5 metres thick may be useable as a marker unit. It contains abundant, sharply angular, pink siliceous clasts which sometimes resemble potassium feldspar crystals.

Unit 2A, which hosts the Colagh showing, is a discontinous lens of massive andesitic ash and lapilli tuffs.

Higher in the stratigraphy, an increasing amount of fine to medium-grained sedimentary rock is interbedded with the volcanics. Unit 3SV contains mainly black to brown argillites, siltstones and maroon (hematitic) sandstones, intercalated with thinly bedded andesitic to dacitic ash and dust tuffs. A gradational boundary is drawn between Units 3SV and 3S, where the latter has greater than 50 per cent sedimentary rocks. Higher in the section, volcanic rocks diminish considerably.

Several gabbroic dikes intrude the sediments and volcanics. These are similar in composition to the main Jura-Cretaceous Nickel Mountain olivine gabbro to the west, which hosts the E & L nickel deposit (*see* Hancock, 1990, this volume). Other minor dikes have compositions ranging from granodiorite to andesite feldspar porphyry and pyroxene porphyry.

The western edge of the Eocene King Creek dike swarm (Britton *et al.*, 1989) crosses the map area and can be seen intruding the lowermost andesite Unit (1A). The dike swarm has no regular orientation but trends generally northeast. Compositions range from felsic to mafic.

STRATIGRAPHIC CORRELATION

The volcanics are correlated with the Betty Creek formation of the Lower Jurassic Hazelton Group. West of the Colagh prospect on Nickel Mountain, sedimentary strata are assigned to the Salmon River formation (Hancock, 1990, this volume). The increase in sedimentary material up-section in the Betty Creek volcanics reflects the gradational nature of the contact between the two formations.

The Mount Dilworth formation, which overlies the Betty Creek formation in the area east of the Unuk River (Britton *et al.*, 1989), and in the Stewart area (Alldrick, 1985), is missing from the stratigraphic section on the Macgold claims.

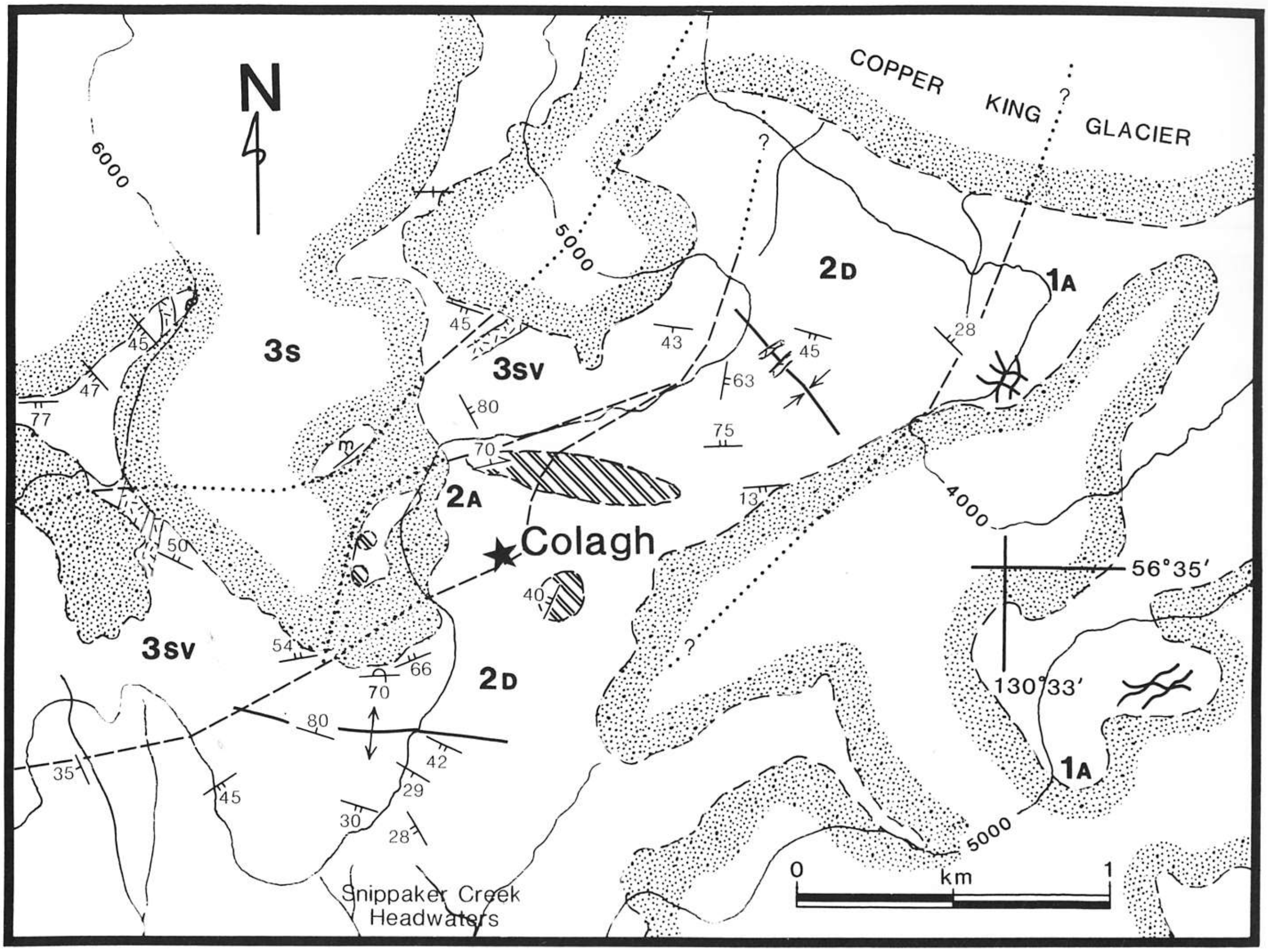

Figure 2-14-2. Geology in vicinity of Colagh prospect.

No fossils have been found in the Colagh area, although ammonoids of possible Toarcian age were collected about 4 kilometres to the west on Nickel Mountain (Grove, 1986; Hancock, 1990, this volume).

STRUCTURE

The main lithologic boundaries trend northeasterly although bedding measurements in volcanics and sediments are erratic and top indicators uncommon. Local strong foliation dips steeply north-northwest.

The axial traces of minor folds trend west-northwest and are best displayed in Unit 2D (Figure 2-14-2). This coincides with the regional orientation obtained from a stereonet; fold axes plunge 1° west (Figure 2-14-3).

MINERALIZATION AND ALTERATION

The Colagh showing consists of coarsely banded sulphide veins occupying regular northeast and northwest-trending shear zones in massive andesitic ash to lapilli tuffs.

Bornite, sphalerite, chalcopyrite and pyrite are layered in bands 3 to 4 centimetres wide. The bornite and sphalerite are closely associated, with covellite replacing bornite. Secondary minerals include hematite, azurite and malachite.

The main vein is approximately 1 metre wide and pod shaped, traceable on the surface for 8 metres. Lesser amounts of the massive sulphides occupy a thin shear for at least another 30 metres northeast along strike. The thin veins continue into autobrecciated andesitic ash and crystal tuffs laced with quartz veinlets. Other shears 2 to 3 centimetres wide, with similar mineralization, parallel the main vein in the 15 metres of continuous outcrop to the north. Cross-shears trending southwest are narrow and sparsely mineralized.

About 100 metres west-southwest of the sulphide showing, a gold and silver-bearing quartz and calcite stockwork breccia zone (known as the "Ice" showing) has been sampled in three trenches. The best assays obtained were 11.0 grams per tonne gold over a 1 metre width with a weighted average of 7.2 grams per tonne over 2 metres (C. Graf, personal communication, 1989).

LEGEND

3s Sediments/volcanics, > 50% sediments: black argillites, siltstones with inter-bedded dacitic ash tuffs, minor "pink lithic" lapilli tuff.

3sv Sediments/volcanics: black to brown argillites, minor maroon sandstone, siltstone and wackes with interbedded thinly bedded andesitic to dacitic ash tuffs.

2A Andesites: massive, locally atrongly foliated ash and lapilli tuff and tuff breccia. Often extremely calcite/ankerite altered.

2D Dacites: ash, crystal and lapilli tuffs. Often well laminated with occasional graded bedding, minor welding (fiammé). Distinctive "pink lithic" (siliceous chip) tuff.

1A Andesites: bedded ash and lapilli tuffs, feldspar porphyry.

SYMBOLS

45 Inclined bedding, tops known

45 Inclined bedding, tops unknown

45 Foliation

Geological boundary

– approximate, gradational

– assumed

Synform, antiform

Mineral occurrence

Gossanous alteration zones (pyrite, calcite/ankerite)

Gabbroic dikes

Eocene dikes (mafic to felsic) (King Creek dike swarm)

Permanent ice/snow (as of August 1989)

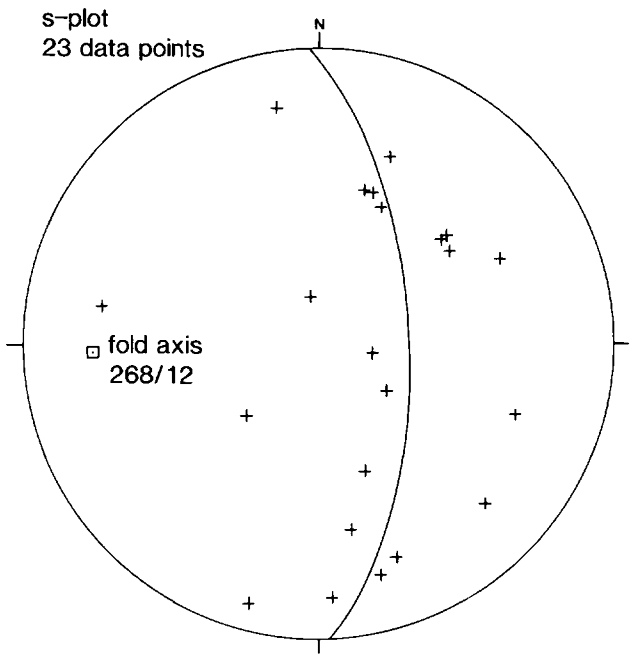

Figure 2-14-3. Stereonet (equal area) of poles to bedding (+).

The outcrop in the surrounding cirque is scattered; up to 45 metres north of the main showing vienlets of pyrite and chalcopyrite are found in andesitic ash tuffs.

The sulphide veins appear to be an epigenetic hydrothermal fracture infilling. The gold and silver-bearing quartz breccia stockwork is interpreted to be a separate (later ?) stage of mineralization (C. Graf, personal communication, 1989).

The andesites hosting the veins are unaltered, yet large gossans occur south and north of the showing (Figure 2-14-2). Alteration within the gossans consists of pyrite, calcite and ankerite. Jasper and manganese oxide also occur with minor epidote in clots and veins.

A grab sample taken from the main banded base metal sulphide lens in 1988 assayed 0.11 gram per tonne gold and 49.00 grams per tonne silver.

ACKNOWLEDGMENTS

The map was compiled from traverses by I.C.L. Webster and C.P.W. Russell in 1988, and by D.J. Alldrick, B.A. Fletcher and the author in 1989. Information concerning the Nickel Mountain area was provided by K.D. Hancock. Chris Graf and John Nicholson of Ecstall Mining Corporation are

gratefully acknowledged for supplying information, assays and samples of the Ice showing.

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A NEW RHODONITE OCCURRENCE IN THE CASSIAR AREA, NORTHERN BRITISH COLUMBIA (104P/5)

By JoAnne Nelson, Z.D. Hora and Fleur Harvey-Kelly

KEYWORDS: Economic geology, rhodonite, Sylvester allochthon, stratabound, syngenetic.

INTRODUCTION

Several occurrences of rhodonite in float and subcrop in the Cassiar map area (104P/5) were reported by Nelson *et al.* (1989). This note reports the discovery of a bedrock rhodonite deposit by Z.D. Hora during follow-up prospecting in July 1989, which has significant potential as a source of carving-quality material. The deposit is located at the headwaters of Snowy Creek 4 kilometres north of the Stewart-Cassiar Highway (Figure 2-15-1).

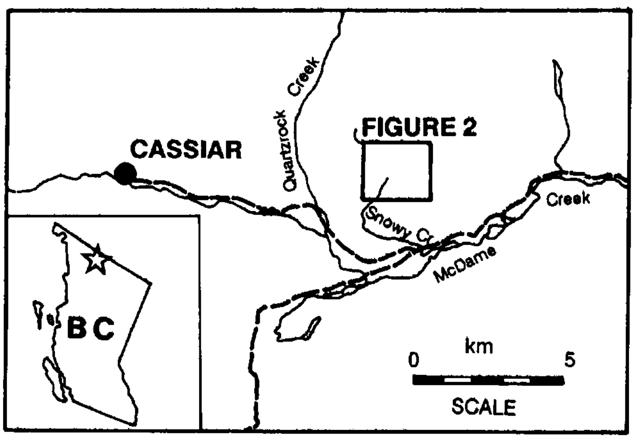

Figure 2-15-1. Location of Snowy Creek rhodonite deposit.

GEOLOGIC SETTING

The Snowy Creek rhodonite is both stratiform, albeit poddy, and stratabound. It occurs within unit IIPPvs of the Sylvester allochthon (Nelson and Bradford, 1989), a Pennsylvanian-Permian sequence of well-bedded chert and argillite, interbedded basalt and abundant diabase sills. More specifically, the rhodonite is located above a section of grey, black and pale green chert and argillite at the base of the unit [IIPPvs(1) on Figure 2-15-2] and about 50 metres below an upper section of brightly coloured, maroon, red, green and grey chert and argillite [Unit IIPPvs(2) on Figure 2-15-2]. This stratigraphic position holds over at least 4 kilometres strike length, as shown by the two cross-sections in Figure 2-15-3. The immediate host of the rhodonite is well-bedded grey to pale green radiolarian chert with argillite partings.

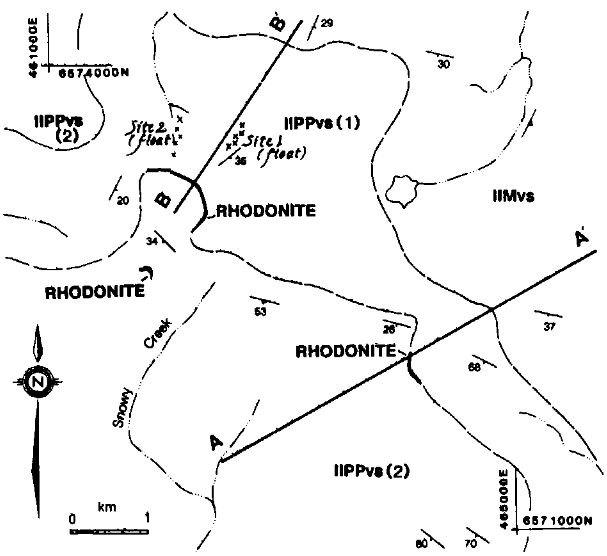

Figure 2-15-2. Simplified geologic map showing area of rhodonite horizon. IIMvs: Mississippian basalt and siliciclastic to pelagic sediments and diabase sills. IIPPvs (1): Pennsylvanian to Permian drab cherts, argillites; basalt and diabase. IIPPvs(2): Pennsylvanian to Permian brightly coloured chert, argillite; basalt, diabase.

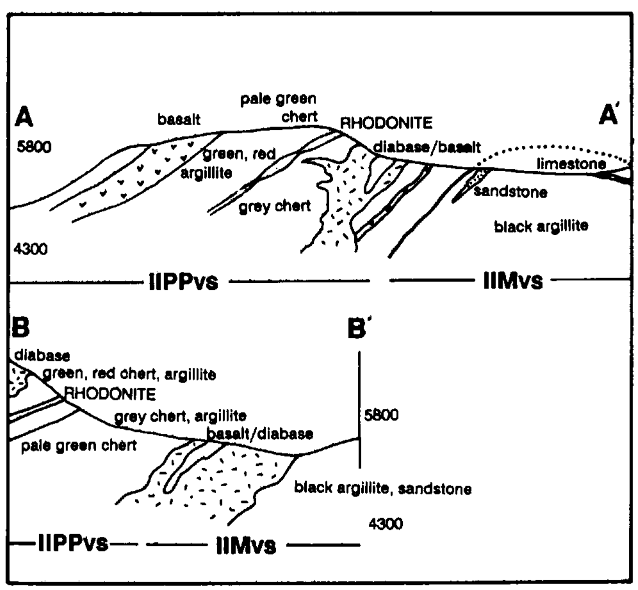

Figure 2-15-3. Cross-sections through rhodonite horizon.

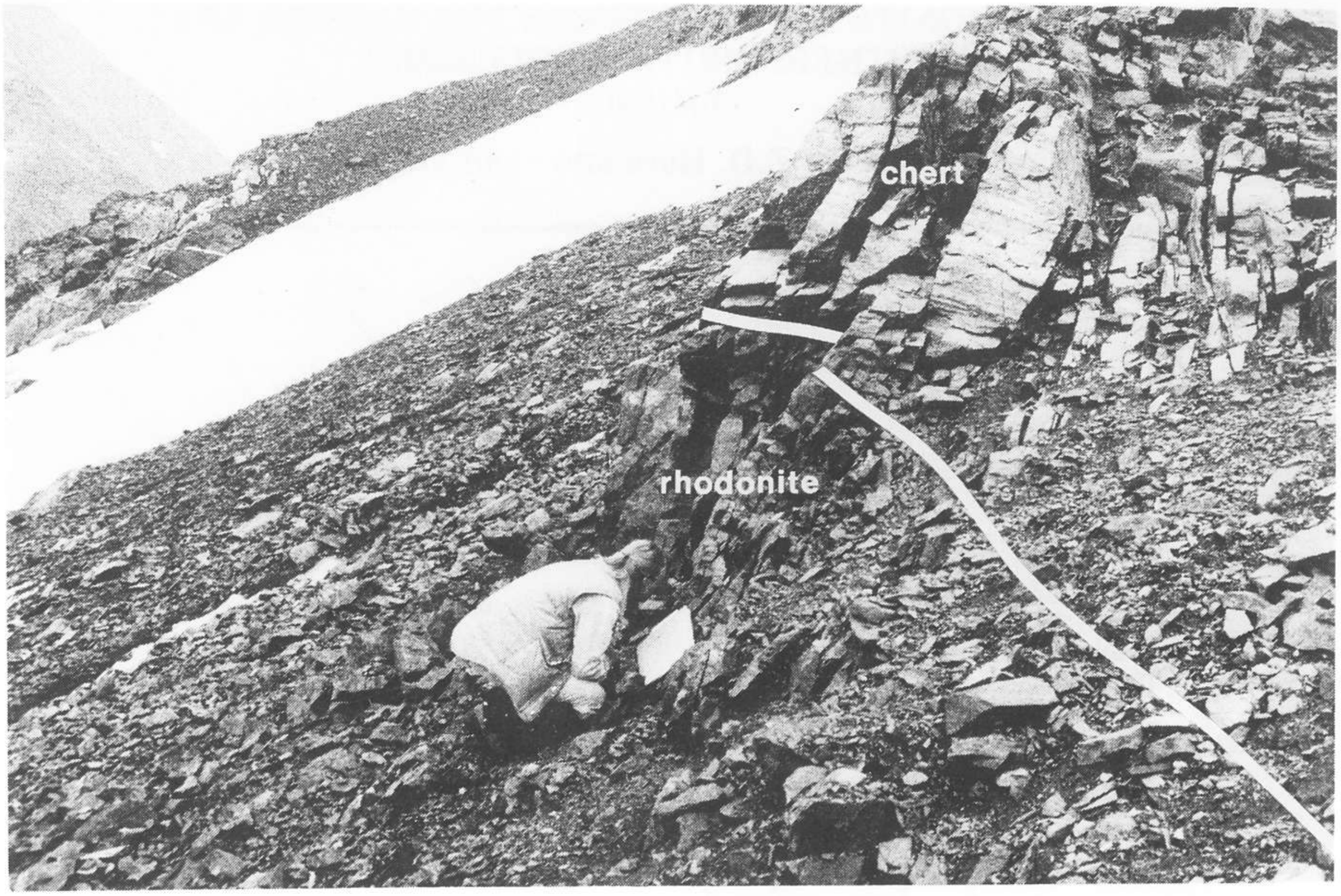

Plate 2-15-1. Outcrop of Snowy Creek rhodonite.

DESCRIPTION OF DEPOSIT

The Snowy Creek rhodonite is a bedded lensoidal sequence that ranges in thickness from zero to 5 metres. It forms a prominent soot-black-weathering outcrop that partly encircles one of the summits at the headwaters of Snowy Creek (Plate 2-15-1). A continuation of the rhodonite horizon was seen in a cliff 4 kilometres to the southeast. Between these two localities extensive diabase sills interrupt the continuity, although the possibility of making further discoveries is good.

An example of this untested potential is shown by the occurrence of rhodonite in float at Site 1 (Figure 2-15-2). There, scattered angular fragments and boulders are found along the small stream that flows northeast from the pass north of the main showing. The size of the fragments suggests a bed at least 70 centimetres thick. The fragments are scattered over 250 metres, their distribution suggesting an unrecognized source in the cliffs to the southeast. Rubble from two large rock-slide scars in the upper parts of these cliffs partly covers the rhodonite float. It is therefore assumed that the bedrock source is in the lower part of the cliffs.

The main deposit consists of three isolated outcrops each 200 to 300 square metres in exposed area, and ranging up to 100 metres of strike length (Figure 2-15-2). The outcrops consist of approximately 20 metres of grey-green chert with an enclosed rhodonite zone up to 5 metres thick, although abruptly pinching to zero thickness. Slightly different elevations of the various pods may result from a mildly transgressive style of mineralization, or from small displacements on vertical faults. Site 2 rhodonite float is derived from the main outcrop area.

The easternmost outcrop of rhodonite has a footwall and hangingwall of blood-red ferruginous chert. The rhodonite is associated with light yellow chert, which occurs as discrete layers, irregular patches and massive lenses up to 10 centimetres thick. The texture of the rhodonite varies from laminated to patchy. Massive rose-pink rhodonite alternates with delicately streaked pink and grey rock; grey chert with bright pink spots; and marbled pink, white and butterscotch-coloured varieties. The largest unfractured and massive specimens are roughly 1 metre in diameter.

MICROSCOPIC PETROGRAPHY

The predominant minerals of the Snowy Creek manganese silicate/carbonate occurrence are microcrystalline quartz interbedded with disseminated, mottled and banded rhodochrosite and rhodonite, interlayered with hematite or garnet- rich bands. Parallel and crosscutting vein mineralization is also significant. Spessartite is abundant in some specimens. Overall, rhodonite and rhodochrosite account for at least 45 per cent of the mineralization. Rhodonite corresponds to the bright pink observed in hand specimen while rhodochrosite is a lighter flesh-pink colour. Manganese oxide is invariably present lining hairline fractures and as an oxide coating, 1 to 10 millimetres thick, on the weathered surface.

Accessory minerals associated with the microcrystalline quartz-rich layers are penninite, clinochlore and biotite (possibly the manganese variety mangophyll) and traces of pyrite, chalcopyrite and sphalerite. Stilpnomelane, epidote and unidentified clay minerals are more prevalent within the hematite-rich bands. Clinozoisite and the clinoamphibole grunerite were observed in two samples; one as manganese replacement in an altered basalt exhibiting uralitization and the other associated with rhodonite in a quartz-poor section.

The hangingwall and footwall of the deposit consist of a banded metachert of interlayered hematite and microcrystalline quartz. Ghost radiolarians, infilled by microcrystalline quartz, indicate a ferruginous radiolarian chert protolith. Minor constituents of the wallrock include patchy massive carbonate, bedding-parallel thin biotite platelets and interstitial stilpnomelane.

Within the deposit, manganese silicate/carbonate mineralization is represented by a variety of textures. Rhodonite occurs as intergrowths with massive or crystalline rhodochrosite or by itself as euhedral elongate tabular crystals, stellated crystal masses or sheath-like bundles enclosed within and encroaching upon a microcrystalline quartz matrix. It has also been observed with a coarse-grained prismatic habit. This species has unusually low interference colours. The range in the interference colours of rhodonite may reflect differing calcium contents and suggest a possible solid solution with the calcium-rich manganese silicate bustamite. Rhodonite also forms spongy porphyroblasts, 500 microns in length. These are obliquely oriented in a finely laminated garnet-rich metachert. Rhodonite is seen to have grown at the expense of garnet, whereas rhodochrosite and garnet appear to stably coexist. In thin section some rhodonite samples exhibit mammillary growth textures. In other cases, prismatic rhodonite crystal overgrowths enclose interiors of botryoidal and mammillary growth, marked by dark opaque oxide along the curved surfaces. These relationships suggest that mammillary growth preceded crystal growth. Mammilary textures are indicative of open-space filling and mineralization.

Rhodochrosite ranges from massive carbonate either by itself or associated with microcrystalline quartz, rhodonite or disseminated hematite, to crystalline aggregates and intergrowths. The massive rhodochrosite is often bounded by thin layers of stilpnomelane and iron staining. It also occurs as euhedral rhombic crystals, up to 750 microns in size and, in at least one instance, is associated with euhedral zoned quartz crystals, further evidence for open-space growth.

Alumina, derived from feldspar and clay, exerts a compositional control on the distribution of spessartite. This is illustrated by the concentration of fine-grained, euhedral garnet dodecahedrons at and close to the margins of the deposit. The present mineral assemblage of quartz-rhodonite-rhodochrosite-spessartite-grunerite is of metamorphic origin, rather than origitial. The thodonite porphyroblasts are evidence for this. The deposit has been metamorphosed under pressure-temperature conditions of prehnite-pumpellyite to greenschist facies as shown by assemblages in metabasalts (Nelson, 1990, this volume). The coexistence of rhodonite, rhodochrosite and quartz implies conditions of XCO_2 and temperature approximately on the reaction rhodochrosite + quartz = pyroxmangite + CO_2 investigated by Candia *et al.* (1975). This implies very low values of XCO_2, for a total pressure of 2 kilobars and a temperature of 375°C. (Nelson, 1990).

Euhedral pyrite rhombs and cubes with pitted interiors and overgrowths of chalcopyrite and sphalerite, exhibiting chalcopyrite exsolution textures, propagate along one of the iron-stained microlaminae which represent original bedding.

A variety of veinlets and gash veinlets crosscut the deposit. Veinlet mineralization includes: microcrystalline and massive quartz, microcrystalline quartz lined with barite, quartz/carbonate and penninite intergrowths, massive and crystalline rhodochrosite, adularia and biotite, adularia and stilpnomelane, and stilpnomelane. Veinlet widths range from 10 microns to 250 microns. Network rhodochrosite veining contributes to the overall pink colour of the rock.

DISCUSSION AND CONCLUSIONS

The Snowy Creek rhodonite deposit is a small syngenetic occurrence. Field relationships, macroscopic and microscopic textures and mineralogy point to its origin as an ocean-floor hydrothermal system that produeed both exhalitive and hydrogenous mineralization at the seawater-sediment interface and apparently epigenetic effects in unconsolidated sediments. The occurrence is stratabound within radiolarian chert, yet replacement textures and open-space filling, which are epigenetic in origin, occur adjacent to delicate manganese silicate/carbonate laminations. Colloform textures may also be suggestive of accretionary growth simiiiar to that of modern marine manganese nodules. On the other hand, the presence of iron, copper and zinc sulphides, barite and adularia are additional evidence for hydrothermal precipitation and deposition, as opposed to classic manganese crust formation by hydrogenous precipitation and suboxic diagenesis.

Manganese behaviour is controlled by the oxidation state of the depositional environment; manganese is highly soluble under reducing conditions (Maynard, 1983). The preservation of a substantial unit of manganese silicate/carbonate, bounded by massive red cherts at Snowy Creek, suggests that oxidizing conditions prevailed during deposition or concentration of manganese. The precipitation of manganese oxide was probably mediated by the presence of an iron oxide substrate (the ferruginous cherts) which had the ability to adsorb appreciable quantities of ions out of seawater. Iron-oxides were deposited under pH conditions of above 6 (Maynard, 1983). Once established, under stable oxidizing conditions and a pH equal to or greater than 8.5, manganese oxide accumulated. Low-temperature diagenesis and low-grade metamorphism subsequently released silca from the radiolarian cherts which combined with the manganese oxide to form rhodonite.

The Snowy Creek deposit represents a small but interesting rhodonite resource. The apparent hardness of the carveable stone ranges from 4 to 6 depending on the relative concentrations of rhodonite, rhodochrosite and quartz. The overall colour and quality is fair to good and is a vibrant mix of light and dark pinks and greens.

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Mafic-Ultramafic Rock Studies

GEOLOGY AND PRECIOUS METAL POTENTIAL OF MAFIC-ULTRAMAFIC ROCKS IN BRITISH COLUMBIA: CURRENT PROGRESS*

By G.T. Nixon

KEYWORDS: Economic geology, ophiolite, listwanite, Alaskan-type ultramafic complex, Polaris, Wrede, Johanson, Shulaps, structure, geochemistry, gold, platinum group elements.

INTRODUCTION

The geology, geochemistry and mineral potential of mafic and ultramafic rocks in British Columbia have recently become the objectives of a number of research projects involving principally government and university workers. An evaluation of the mineral potential of ultramafic-mafic rocks in the province by the British Columbia Geological Survey Branch, funded by the Canada/British Columbia Mineral Development Agreement, was initiated in 1987 as part of a 3-year program that ends in March, 1990. This project has focused on the economic potential of Alaskan-type ultramafic complexes for platinum group elements (PGE) and gold. A more recent initiative by the British Columbia Geological Survey Branch in 1989, established a 2-year program to develop a metallogenic model for the lode gold–listwanite (carbonatized ultramafic rock) association in British Columbia. The listwanite model is particularly relevant in ophiolitic terranes. In a recent supporting project partly funded by the British Columbia Geoscience Research Grant program, university workers have also directed their attention toward a further understanding of ophiolite evolution and petrogenesis. The six papers contained in this section, and summarized below, present previously unpublished results of fieldwork conducted in 1988 and 1989 that has focused on ophiolite assemblages and Alaskan-type intrusions. Project locations are shown in Figure 3-1-1.

The ophiolite stratigraphy that represents obducted oceanic crust and uppermost mantle is rarely preserved intact. Dismembered ophiolite assemblages in British Columbia are found in the oceanic terranes of Cache Creek (*e.g.* Nahlin ophiolite), Bridge River (*e.g.* Shulaps ophiolite), and Slide Mountain (Figure 3-1-1). These tectonostratigraphic terranes represent the deformed sequences of Paleozoic to Early Mesozoic ocean basins that closed in the Mesozoic during the accretion of island arcs to the North American craton. Alaskan-type mafic-ultramafic complexes, on the other hand, represent the subvolcanic magma chambers of Late Triassic to Middle Jurassic arc volcanoes situated within these accreted terranes. They formed on continental crust where penetrative deformation during the accretion event was not nearly as severe as that which ocurred within and at the margins of the ocean basins. The subparallel belt of Alaskan-type complexes west of the Coast Mountains in southeastern Alaska is believed to be largely Cretaceous in age (Figure 3-1-1).

OPHIOLITES AND THE LISTWANITE – LODE GOLD ASSOCIATION

In the opening paper in this series, Ash and Arksey (1990a) provide an overview of the listwanite–lode gold association and examine how this model might be more extensively applied in British Columbia. Listwanites are formed where CO_2-rich hydrothermal fluids encounter serpentinized ultramafic rocks and form predominantly iron-magnesium carbonates ± quartz veins that are potentially associated with gold mineralization. Where carbonatization is accompanied by alkali metasomatism, the bright green, chrome-rich mica, mariposite(fuchsite), also occurs. Listwanitic alteration is primarily developed in fault zones that serve as channelways for circulating fluids. These zones are easily recognized in the field by their characteristically bright orange-brown weathering which stands out against a background of dark green to black serpentinite. Alteration halos are common in which talc-carbonate rocks occur in the outer envelope, at some distance from the fault zone, and quartz ± carbonate ± mariposite ± gold ± sulphides occupy the central conduits. Listwanite-generating fluids may also transport PGE leached from ultramafic rocks; the source of the gold is controversial. The development of listwanite appears to favour an ophiolitic or alpine ultramafic tectonic setting where imbrication by thrusting promotes fluid access, and serpentinization is prevalent both before and during obduction of oceanic crust and mantle sequences.

Ash and Arksey (1990b, c) go on to describe the structural controls of listwanitic alteration within the Atlin ultramafic allochthon, an imbricated package of oceanic upper mantle rocks obducted in the Mesozoic. These rocks form part of the Cache Creek Terrane, a subduction complex related to the Late Paleozoic–Early Mesozoic volcanic arcs of Quesnellia and Stikinia. The Atlin area has a long history of placer gold mining; lode gold deposits and listwanitic alteration are spatially related to major zones of southwesterly directed thrusting such as the Monarch Mountain fault (Ash and Arksey, 1990b). Rocks within these thrust packages include depleted mantle harzburgite tectonite, minor podiform dunite, metabasalts and a tectonic mélange assemblage. The listwanites are preferentially developed within serpentinized harzburgite tectonites in the hangingwall of the thrust.

In another investigation of an accreted oceanic terrane, Calon *et al.* (1990) present a preliminary report on the tectonic and petrologic evolution of the Shulaps ophiolite and a discussion of its Mesozoic accretionary history. The Shulaps, situated along the boundary of the Intermontane Superterrane, is one of the largest intact ophiolite fragments in the Cordillera (Figure 3-1-1). These authors recognize a southwest-verging, northeast-dipping imbricate thrust sys-

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

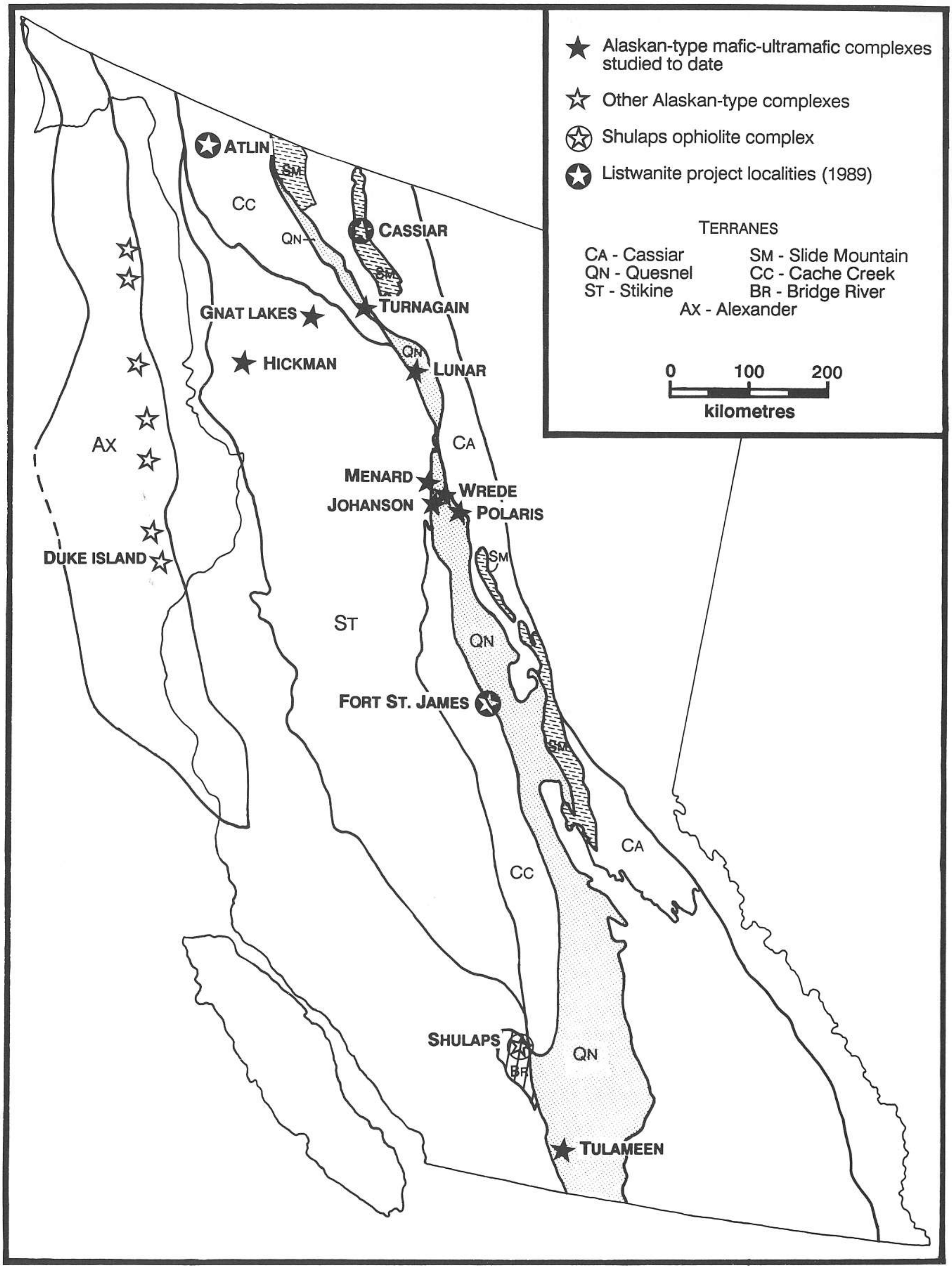

Figure 3-1-1. Map showing location of listwanite project sites, Shulaps ophiolite and most known Alaskan-type complexes in British Columbia in relation to accreted tectonostratigraphic terranes of the Cordillera.

tem comprising four main lithotectonic units. The uppermost unit, the Shulaps peridotite, is composed of depleted mantle harzburgite tectonite and minor podiform dunite, much like the upper mantle lithologies described by Ash and Arksey in the Atlin area. The peridotite unit structurally overlies a serpentinite mélange that contains variably sized, rare blocks of volcanic and sedimentary rocks, as well as more common gabbros and ultramafic cumulates. This, in turn, is thrust upon a suite of mafic plutonic and volcanic rocks. The lowermost unit comprises arc-derived siliciclastic and calcareous sedimentary rocks tentatively assigned to the Late Triassic Cadwallader Group. Fault-controlled listwanitic alteration accompanied by magnesite is known along the Yalakom fault some 12 kilometres to the northeast of the project area, and along the Relay Creek–Marshall Creek fault system 15 kilometres to the northwest, where listwanites have developed in serpentinite and are associated with tungsten and scheelite mineralization, and contain anomalous mercury, silver and gold (Schiarizza *et al.*, 1989).

ALASKAN-TYPE COMPLEXES AND THEIR PGE POTENTIAL

The Alaskan-type mafic-ultramafic complexes studied to date include all of the known major complexes (Tulameen, Polaris, Wrede Creek, Lunar Creek and Turnagain) as well as many minor bodies (Johanson Lake, Menard Creek, Gnat Lakes and Hickman; Figure 3-1-1). All such complexes in British Columbia occur within the Intermontane Superterrane and form part of the accreted volcanic arc terranes of Quesnellia and Stikinia. Only the Menard Creek, Hickman and Gnat Lakes complexes (Nixon *et al.*, 1989b) occur in Stikinia as presently defined (Wheeler and McFeeley, 1987; Wheeler *et al.*, 1988). Aspects of the geology, noble metal geochemistry and platinum mineralogy of the Tulameen, Polaris and Turnagain complexes have been documented previously (Nixon and Rublee, 1988; Nixon, 1988; Nixon *et al.*, 1989a, c, d).

The type localities for Alaskan-type ultramafic complexes are found in southeastern Alaska (Figure 3-1-1) and the Ural Mountains. Taylor (1967) has summarized many of their diagnostic features but Irvine's (1974a) classic memoir on Duke Island (Figure 3-1-1) still remains one of the most complete and detailed accounts of Alaskan-type complexes. Most of these complexes have been interpreted to represent the fractionation products of mantle-derived, crystal-poor magma or remobilized cumulates emplaced high in the crust. Their general structural and petrological characteristics include a crude outward zonation of rock types ranging from dunite in the core through wehrlite and clinopyroxenitic and/or hornblende-rich lithologies towards the periphery of the intrusion. Primary mesoscopic layering, such as that so spectacularly developed at Duke Island, is seldom observed. In many cases the ultramafic complexes are partially or completely enveloped by gabbroic to dioritic rocks, which may be genetically related. Some intrusions have well-developed contact metamorphic and metasomatic aureoles of amphibolite grade; others appear to be completely fault-bounded. Not all of the characteristic rock types are represented in each intrusion, which can make their Alaskan-type affinity difficult to establish, particularly if only the later differentiates are exposed. Early cumulate minerals are typically represented by forsteritic olivine, chromite, diopsidic clinopyroxene and rare phlogopitic mica, whereas hornblende, biotite, magnetite and plagioclase may predominate in the more evolved rock types. Orthopyroxene is characteristically absent, which has been taken to indicate an alkalic affinity.

The classification of ultramafic rocks used to characterize Alaskan-type complexes is essentially that recommended by the IUGG (Streckeisen, 1976; Figure 3-1-2). It was found beneficial in the field to arbitrarily subdivide extensive wehrlitic units into olivine wehrlite (65 to 90 per cent olivine; 10 to 35 per cent clinopyroxene) and wehrlite (40 to 65 per cent olivine; 35 to 60 per cent clinopyroxene). All of the Alaskan-type ultramafic rocks lie along the olivine-clinopyroxene join.

The Alaskan-type complexes in British Columbia are potentially important hosts for commercially exploitable platinum group metals (Rublee, 1986; Evenchick *et al.*, 1986) as well as other commodities (*e.g.* chrome, nickel, cobalt, gold, asbestos, jade). The Tulameen complex in southern British Columbia (Figure 3-1-1) ranks as the foremost producer of placer platinum in the province, reportedly having yielded over 680 000 grams of platinum nuggets between 1885 and 1932 (O'Niell and Gunning, 1934). Recent studies of the Tulameen complex have identified platinum group minerals in lode occurrences where they are associated with concentrations of chromite (St. Louis *et al.*, 1986; Nixon *et al.*, 1989a). In particular, Nixon *et al.* (1989a) were able to trace the source of platinum nuggets to chromitite horizons within the dunite core of the complex by matching the phase chemistry of gangue minerals, notably spinel and olivine, in nugget and lode occurrences. The chromitite-PGE association in Alaskan-type complexes is a promising exploration target that deserves further investigation. For example, highly anomalous PGE have recently been discovered in chromitites within the Wrede Creek complex as described below.

Alaskan-type complexes investigated during the 1989 field season include the Lunar Creek, Polaris, Wrede Creek and Johanson Lake bodies in north-central British Columbia. The former three complexes are the subjects of Open File releases (Nixon *et al.*, 1990b, d; Hammack *et al.*, 1990b) and the latter three are described in this publication (Nixon *et al.*, 1990a, c; Hammack *et al.*, 1990a).

Briefly, the Polaris complex (Nixon *et al.*, 1990c) forms a transgressive sill-like body with a minimum length of 14 kilometres and a maximum width of 4 kilometres. It intrudes Paleozoic metavolcanic and metasedimentary island arc sequences that are believed to form the basement of Quesnellia, and has a metamorphic aureole of lower amphibolite grade. All of the Alaskan-type complexes in the region are considered to be Late Triassic in age, and both coeval and cogenetic with arc-related volcanic rocks of the Takla Group (Irvine, 1974b, 1976). Kinematic indicators suggest that the Polaris complex and its hostrocks are allochthonous, having been uprooted along a westerly dipping reverse fault and transported eastwards toward cratonic North America during terrane accretion in the Mesozoic.

All of the lithologies that characterize Alaskan-type ultramafic complexes are well represented in the Polaris intrusion, although exposures of olivine-rich ultramafic rocks predominate. The complex exhibits a crude upward zonation from dunite through wehrlite and clinopyroxenite to clinopyroxene hornblendite and gabbroic rocks near the roof. The lower margin of the intrusion displays a similar variation but the zoning is much more condensed. Cumulate textures are well preserved and penetrative deformation is lacking. Chromitites are widespread in the dunite, but invariably small in size and have been remobilized shortly after deposition. Lithogeochemical analyses reveal some platiniferous chromitite horizons but they appear too few and far between to be of economic interest. Weak gold anomalies are restricted almost entirely to the contacts with metasedimentary and metavolcanic hostrocks, which suggests an external source for the gold.

The Wrede Creek complex (Hammack *et al.*, 1990a) appears to be a stock-like intrusion hosted by Takla Group volcanic and volcaniclastic rocks. Ultramafic lithologies, particularly dunite, are well represented and vestiges of an amphibolite-grade metamorphic aureole are present. In the past, the Wrede Creek complex was extensively prospected and drilled for porphyry copper-molybdenum mineralization associated with Jurassic granitoid intrusions at its southern margin (Wong *et al.*, 1985). Apparently chromitite horizons in the dunite core were not tested for their PGE potential. However, all of the chromitite samples analyzed during the course of this investigation proved to be anomalous in platinum, and some extremely so (2500 ppb). It is this same PGE-chromitite association that has been linked to the source of platinum placers in the Tulameen district (Nixon *et al.*, 1989a).

The final paper of the series describes the Johanson Lake mafic-ultramafic complex (Nixon *et al.*, 1990a), a gabbroic to pyroxenitic body spatially associated with Takla Group volcanic rocks and Jurassic dioritic plutons. The pegmatitic phases of the complex exhibit well-developed comb layering formed by acicular hornblende crystals that appear to have crystallized *in situ*. Lithogeochemical analyses for PGE do not reveal anomalous lithologies although the background for gold is distinctly higher than has been observed in other Alaskan-type complexes.

OBSERVATIONS AND FUTURE CONSIDERATIONS

The first-order spatial correlation observed between listwanite–lode gold occurrences and ophiolites or alpine serpentinites/peridotites reflects a fundamental tectonic environment. The closure of ocean basins and ensuing collisional events provide a mechanism for the emplacement of variably serpentinized oceanic crust in regions of orogenesis and crustal thickening. Imbrication of mafic-ultramafic ocean-floor stratigraphy, and subsequent high-angle faulting during later isostatic readjustment or migration of previously

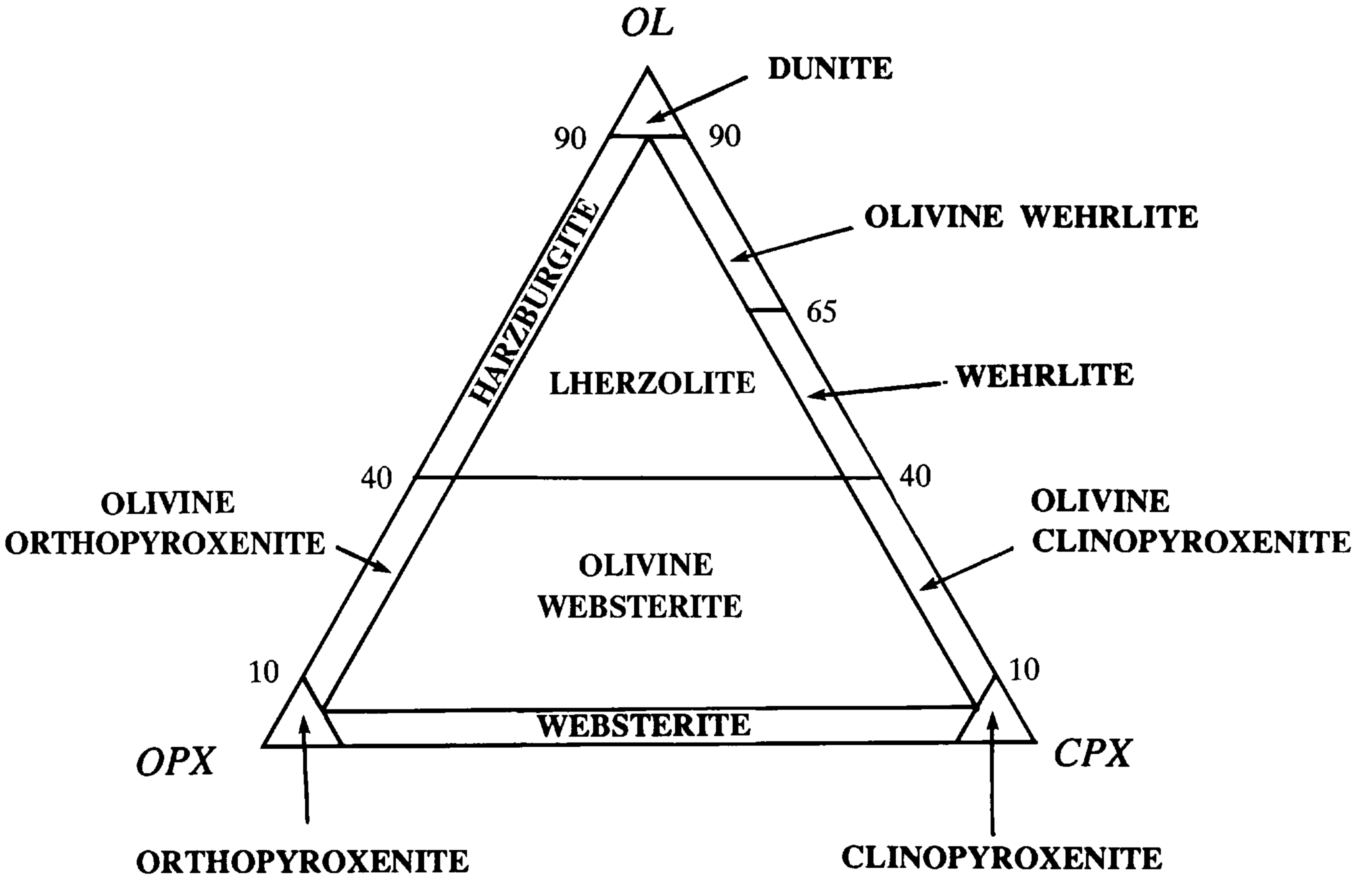

Figure 3-1-2. Classification of ultramafic rocks used to subdivide Alaskan-type ultramafic complexes (modified after Streckeisen, 1976).

accreted terranes provides adequate structural preparation for carbon dioxide rich hydrothermal fluids that promote listwanite development and deposition of precious metals. The fact that listwanite-associated lode gold deposits have been the major source of over 19 million grams of placer gold in the Atlin area, and that these zones have been shown to occur along regional-scale thrusts, should provide a firm focus for future exploration efforts (Ash and Arksey, 1990a).

That the listwanite–lode gold association appears to be *insignificant* in Alaskan-type ultramafic complexes can also be rationalized in terms of the geotectonic setting. In British Columbia, at least, many of the latter complexes have intruded at relatively high-levels in supracrustal rock sequences situated within the accreted terranes. In many cases, these complexes probably represent the subvolcanic magma chambers of Late Triassic to Early Jurassic arc volcanoes. The fact that Alaskan-type complexes have formed within continental crust in an environment relatively protected from collisional deformation, together with their limited amount of olivine-rich ultramafic lithologies and restricted degree of serpentinization, do not make them a propitious target for auriferous listwanite development. Gold-bearing fluids may have been largely spent by the time they reach these sites.

The PGE potential of Alaskan-type ultramafic complexes has not been adequately evaluated and deserves further attention by the exploration community. In northern British Columbia, many of the known intrusions form a northwesterly trending belt stretching from the Polaris complex to at least as far as the Turnagain River. Other Alaskan-type complexes will no doubt be discovered within this belt, particularly when aeromagnetic coverage in this part of the province becomes more complete. In at least two cases, Tulameen and Wrede Creek, the platinoid minerals are closely (genetically?) related to chromitites within the dunite core of these intrusions. This association should, therefore, be considered a prime target for lode occurrences and derived placers.

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THE LISTWANITE - LODE GOLD ASSOCIATION IN BRITISH COLUMBIA

By C.H. Ash and R.L. Arksey

KEYWORDS: Economic geology, listwanite, lode gold, mariposite, oceanic terrane, ophiolite, ultramafic, deposit model.

INTRODUCTION

Listwanite is a term long used by Soviet geologists working in the Ural goldfields of Russia (Goncharenko, 1970; Kuleshevich, 1984) that is now used in Europe and North America. It describes a mineralogical assemblage that results from the carbonatization of serpentinized ultramafic rocks and represents a distinctive alteration suite that is commonly associated with quartz-carbonate lode gold deposits. In British Columbia, as in the California Mother Lode deposits, listwanites are most commonly recognized within and near major fault zones cutting Paleozoic and Mesozoic oceanic and island arc accretionary terranes that have been affected by tectonism, metamorphism and plutonism.

The Listwanite Project was started in 1989 to investigate, document and develop a deposit model for the listwanite–lode gold association in the Canadian Cordillera. Aspects relevant to the development of such a model being addressed during this study include:

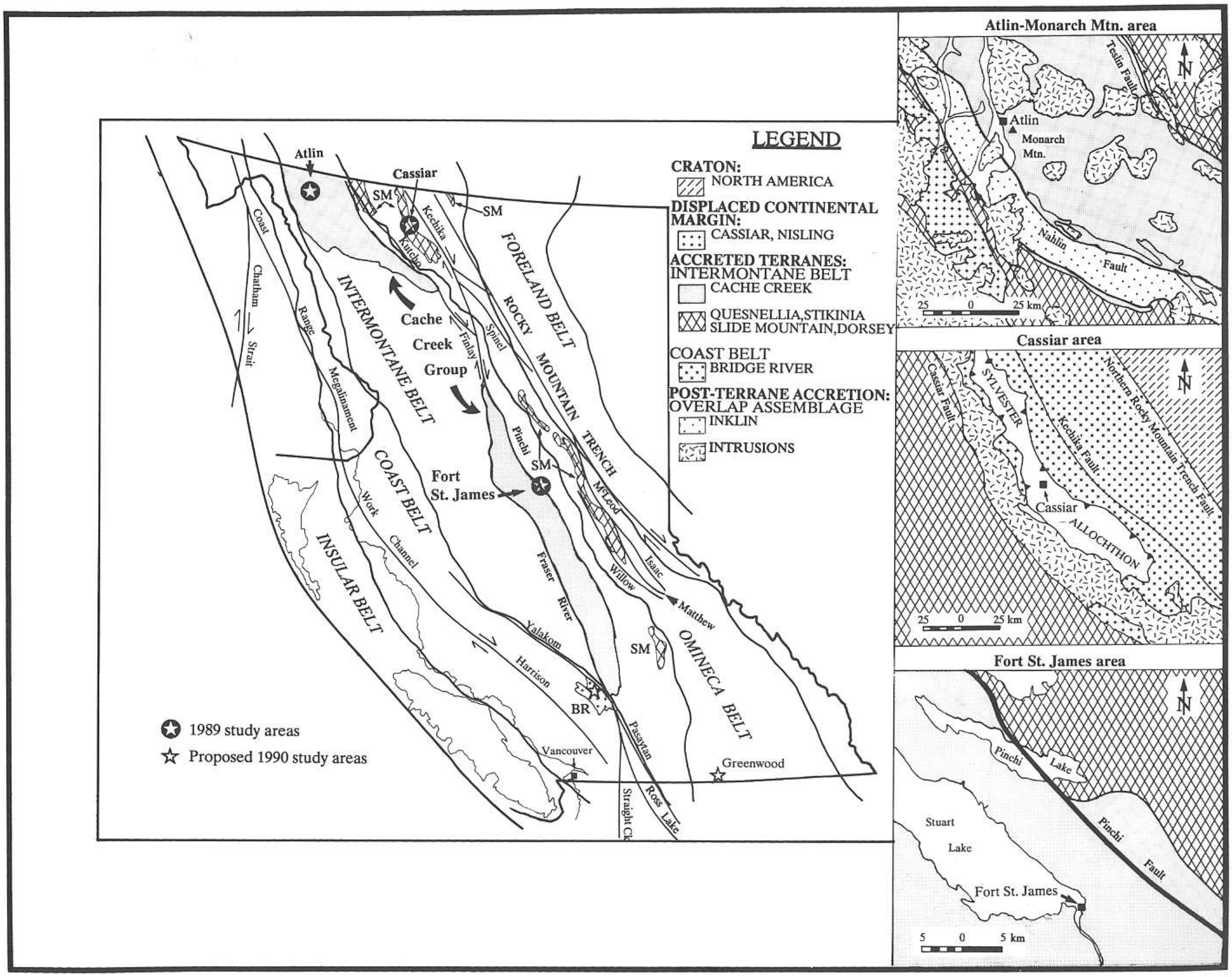

Figure 3-2-1. Location map showing simplified terrane boundaries, major transcurrent faults, the Cache Creek (CC), Slide Mountain (SM) and Bridge River (BR) terranes and the 1989 and proposed 1990 study areas.

- The setting of these deposits within the framework of tectonically dismembered ophiolitic assemblages, by combining detailed mapping, petrographic analysis, microprobe and geochemical studies.
- The age of alteration and associated mineralization using K-Ar dating of mariposite.
- Relationships between listwanitic lode gold and spatially associated alkaline intrusions will be investigated geochemically using both trace element and lead isotopic signatures.
- The potential of listwanitic alteration zones as sites of hydrothermal platinum group element (PGE) mineralization will also be investigated geochemically. As most, if not all, known hydrothermal PGE + gold occurrences are associated with sulphides, sampling focused on sulphide-bearing quartz veins or gossanous zones, generally in gabbroic and, more commonly, basaltic rocks.

This report provides a brief overview of the tectonic setting and development of the listwanite–lode gold deposit type. British Columbia examples are compared to deposits described in the literature. The Atlin, Cassiar (Erickson) and Fort St. James areas in northern British Columbia were investigated during the 1989 field season (Figure 3-2-1). Investigation of listwanite-associated lode gold deposits in southern British Columbia is planned for the 1990 season. Results of detailed mapping in the Atlin area are reported elsewhere (Ash and Arksey, 1990a, this volume, 1990b).

LISTWANITES

TECTONIC SETTING

Listwanite-associated lode gold deposits with serpentinized and carbonatized ultramafic rocks are characteristic of tectonically disrupted ophiolite sequences in accreted oceanic terranes. This tectonic setting produces thrusting and stacking of units, favorable hostrocks (serpentinite), and regional-scale reverse and normal faults to channel fluid flow.

Accreted oceanic terranes of Paleozoic to Mesozoic age, containing dismembered ophiolite packages occur along the length of the Canadian Cordillera and include the Cache Creek (CC), Slide Mountain (SM) and Bridge River (BR) terranes (Figure 3-2-1). These terranes contain oceanic uppermost mantle, crustal and sedimentary rocks, most commonly with tectonic contact relationships (Monger *et al.*, 1982). Ultramafic rocks, serpentinized to varying degrees, are represented mainly by residual mantle harzburgite (defined in Ash and Arksey, 1990a, this volume). Oceanic crustal rocks are dominated by metabasalts and are commonly referred to as greenstones. Oceanic crustal plutonic rocks, including ultramafic cumulates, gabbro, diorite and trondhjemite, may be present but are generally minor constituents. Sedimentary units include deep-water cherts and argillites as well as shallow-water limestones.

Some Cordilleran oceanic terranes are well known as gold producers, others are under-explored. Listwanitic alteration characterizes both the Bridge River (BR) camp in the south, the largest gold camp in the province, and the active Cassiar (SM) and Atlin (CC) camps in the north. The recently defined Snowbird (CC) gold prospect (X-Cal Resources Limited) located near Fort St. James in central British Columbia is also a typical listwanite–lode gold occurrence.

Other examples of listwanite-related gold deposits within accreted oceanic terranes of North America are, most notably, the many deposits throughout the California Mother Lode district (for further discussion see: Knopf, 1929; Böhlke and Kistler, 1986; Böhlke, 1989; Weir and Kerrick, 1987). The Mount Vernon deposit in northern Washington State (Gresens *et al.*, 1982) and ophiolite belts in the Dunnage zone of the Appalachian orogen in Newfoundland (Tuach *et al.*, 1988) also display characteristic listwanite–lode gold associations. Listwanite-related gold deposits are also present within accreted oceanic terranes in Saudi Arabia, Mali (West Africa), the Maritime Alps in northwest Italy, and Morocco (Buisson and Leblanc, 1985a, b).

Somewhat similar and possibly related gold deposits in the Juneau belt of southeast Alaska are also structurally controlled lodes within accreted oceanic assemblages. Goldfarb *et al.* (1988) found that the mineralizing fluids of the Juneau belt are chemically similar to ore-forming fluids in the Mother Lode deposits.

GENESIS

The dynamics of collisional orogenic belts like the Cordillera appear to play a fundamental role in the development of listwanite-associated lode gold deposits. Consuming plate margin tectonic processes instigate subduction, crustal thickening and partial melting which result in metamorphism and plutonism. The combined effects of these processes provide a number of mechanisms which may facilitate the generation of mineralizing fluids. Among these are deep seated magmatic activity, metamorphic dehydration reactions and deep circulation of meteoric waters (Böhlke and Kistler, 1986). Most importantly, brittle to ductile deformation generates fault zones which act as pathways for the altering and mineralizing fluids. Geochemical data suggest that the principal ore-forming fluids are derived primarily by metamorphic dehydration under amphibolite-grade metamorphic conditions, with possible contributions from meteoric or magmatic sources (*see* review in Kerrich, 1989). However, the source of both the fluids and the gold is a topic of current debate. Nesbitt *et al.* (1986) have suggested that mesothermal gold deposits throughout the Canadian Cordillera result from deep circulation and evolution of meteoric water in structures associated with major transcurrent fault zones.

Listwanite forms when fluids rich in carbon dioxide permeate and alter previously altered ultramafic rocks, usually serpentinite. Distinctive iron-magnesium carbonates and chromium mica (mariposite in North American terminology or fuchsite in Europe and Russia) are formed.

The importance of the serpentinized ultramafic rock is that it acts as a preferential sink for carbon dioxide from the migrating hydrothermal fluid. Carbonatization is represented by both the pervasive alteration and replacement of the ultramafic rocks and by dolomite veining. Carbonate minerals which replace the ultramafic rocks form by hydrolysis of iron, magnesium, calcium and manganese silicates to carbonates in which wallrocks donate the bivalent metal cations (Kerrich, 1983). Sericitization as a result of potassium metasomatism is commonly reflected by the formation of mariposite in which the chrome is inherited from the ultramafic hostrock as it cannot be taken up by the carbonate (Boyle, 1979)

Although the genetic significance of the ultramafic rocks remains a subject of debate, the spatial relationship between carbonatized ultramafic rocks and gold deposition appears to be consistent. Böhlke and Kistler (1986), Böhlke (1989) and Wittkopp, (1983) noted that mineralized quartz veins in the California Mother Lode deposits show a spatial association with serpentinite bodies and that the largest concentrations of free gold occur at or near the intersection of veins with the carbonatized ultramafic rocks. Pike (1976) has pointed out the association of carbonatized ultramafic volcanic rocks with the Archean quartz-carbonate lode gold deposits of Northern Ontario. Lode gold showings throughout the Atlin region (Bloodgood *et al.*, 1989; Rees, 1989) and deposits in the Erickson gold camp (Boronowski, 1988) display similar spatial relationships. Some authors argue that the ultramafic rocks are the source of the gold (Buisson and Leblanc, 1985a, b, 1987; Wittkopp, 1983) but this is far from being unaminously accepted.

MINERALIZATION

The generally accepted current hypothesis for gold deposition in and near listwanites invokes low-salinity hydrothermal fluids rich in carbon dioxide which carry gold as a bisulphide complex, $Au(HS)_2$ (Böhlke, 1989; Kerrich, 1989).

Various mechanisms recently reviewed by Kerrich (1989) have been proposed for the precipitation of gold. These include:

- fluctuations in fluid pressure in the seismic-aseismic transition zone promote carbon dioxide and hydrogen sulphide immiscibility with attendant gold deposition,
- reduction of the mineralizing fluid by graphiric rocks, or
- sulphide precipitation promoted by iron-rich lithologies.

The second mechanism was suggested by Dussel (1986) to account for gold precipitation at the Erickson mine near Cassiar, British Columbia. Buison and Leblanc (1985b, 1987) suggest that acid gold-bearing solutions precipitate silica, pyrite, arsenides and gold when entering the reducing and alkaline environment of the carbonatized rocks.

In addition to gold, other metallic mineralization associated with the gold showings investigated includes various types of sulphides (Fe, As, Pb, Cu, Zn, Ni, Co, Sb) commonly with associated arsenides and tellurides. Metal types and abundances vary on both the deposit and regional scale. Some deposits display a diversity of metals while others are selectively enriched in a specific metal or group of metals. The Atlin gold camp illustrates this local variability. The Pictou prospect located 2 kilometres east of Atlin contains a wide range of base metal sulphides, including galena, chalcopyrite, sphalerite, arsenopyrite, pyrargyrite (Ag_3SbS_3) and gersdorffite (Fe,Ni)AsS, (Ballantyne, personal communication) but sulphide mineralogy at the Yellowjacket prospect, 5 kilometres to the northeast, is dominated by pyrite, arsenopyrite and gersdorffite (Bozek, 1989). In contrast the Surprise showing, 7 kilometres to the east, has galena as the only sulphide identified in hand sample.

Silver generally displays a correlation with anomalous gold values, and was an important byproduct at the Erickson and Taurus gold mines near Cassiar (Schroeter and Panteleyev, 1985). At the Snowbird gold prospect, antimony is present in significant amounts as massive 1 to 4-centimetre grey stibnite lenses in white quartz-carbonate veins.

Although cobalt and nickel mineralization has not been identified in British Columbia listwanites, the Bon Azzer ophiolite in Morocco contains listwanite formed along faults marginal to a large serpentinite unit with quartz-carbonate lenses hosting cobalt-nickel arsenide mineralization. It represents a type of listwanitic cobalt-nickel deposit with accessory gold (Leblanc, 1986).

A number of factors combine to suggest that listwanitic alteration zones may be a potential host for hydrothermal platinum group element (PGE) mineralization. PGEs are found primarily in mafic and ultramafic rocks. Analysis for this group of elements has historically been difficult and frequently inaccurate, especially at the low concentrations in rock samples. Present day analytical techniques have overcome this problem (Theyer, 1988).

PGE concentrations occurring in hydrothermal copper and nickel sulphide ores have been reported from a number of localities including: Northwestern Ontario (Rowell and Edgar, 1986), Wyoming (McCallum *et al.*, 1976) and South Africa (Mihalik *et al.*, 1974). Theyer (1988) has recently reported anomalous platinum and palladium values in quartz veins containing 1 to 2 per cent pyrite from northern Manitoba. Further, Mountain and Wood (1988) made thermodynamic calculations on the solubility of platinum and palladium at temperatures up to 300°C. They found that bisulphide and/or hydroxide complexing are potential mechanisms for transport of these metals. Thus listwanite-generating fluids may be capable of transporting these elements.

DESCRIPTION AND IDENTIFICATION

A clearly identifiable mineralogical alteration assemblage after serpentinite consists of green chromium-bearing mica with quartz and carbonate veins in a recessive orange-brown limonitic groundmass formed by the weathering of iron-rich magnesite.

Listwanite zones form along major faults cutting or marginal to serpentinized ophiolite peridotites (Figure 3-2-2). Alteration is characteristically most intense within and above the mineralizing structure where iron magnesite/mariposite rocks are generally sheared and cut by networks of quartz-carbonate veins. The intensity of carbonatization of the serpentinized ultramafic rocks is zoned outward from the faults, producing a distinctive alteration halo (Table 3-2-1).

Listwanite alteration assemblages are highly variable in appearance depending on the relative abundance of the alteration minerals and the distribution of different fabrics produced by inhomogeneous strain. Hydrothermally altered serpentinites with minor quartz stringers and disseminated

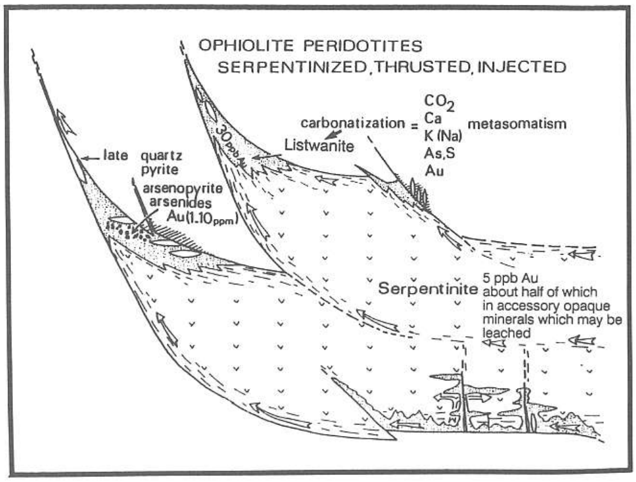

Figure 3-2-2. Generalized listwanitic alteration model which invokes carbonatization of serpentinized ultramafic rocks and the development of gold-bearing veins in and above thrust, normal and/or reverse faults (modified from Buisson and LeBlanc, 1985).

mariposite commonly range from massive to moderately deformed. Those containing high proportions of mariposite commonly form a quartz-carbonate-mariposite schist. However, rocks may vary from massive to schistose to brecciated within the complete range of alteration assemblages present.

Commonly zones marginal to or containing small slivers of carbonatized ultramafic rocks are lithologically heterogeneous and represent tectonic mélanges containing all or some portions of the oceanic assemblages described previously. Within these zones both alteration and deformation are inhomogeneously developed. Carbonatization is generally most intense in the more mafic lithologies. More massive rocks, for example volcanics, cherts and limestones, are often brecciated; fine-grained clastic sediments appear to take up a large part of the strain and are generally intensely sheared and may contain variably sized knockers of the more massive lithologies. Many of these knockers are themselves brecciated and recemented by hydrothermal carbonate, providing evidence for multiple episodes of brecciation and carbonatization. The Yellowjacket (Lefebure and Gunning, 1988), Beavis (Bloodgood *et al.*, 1989) and Mckee Creek showings in the Atlin area are representative examples of listwanitic alteration associated with tectonic mélange assemblages. Fault zones at both the Erickson gold-silver mine in Cassiar and the Snowbird gold prospect near Fort St. James are defined by structural contact relationships between metabasalts and argillaceous sediments. Slivers of ultramafic rock several metres to tens of metres in width are discontinuously distributed along low-angle fault contacts and are highly sheared and schistose.

EXPLORATION GUIDELINES

The fundamental depositional control for this deposit type is the localization of hydrothermal alteration sites along major fault zones within, marginal to, or containing ultramafic rocks. As a first approach to exploration, prominent

TABLE 3-2-1

CHARACTERISTIC ALTERATION ASSEMBLAGE OF LISTWANITE

HOST	Increase in Intensity of Alteration		FAULT
Serpentinite	Talc Carbonate	Quartz (veinlets) Talc Carbonate	Quartz Carbonate Mariposite Au, Ag Sulphides

linears outlined on topographic maps or aerial photographs within or adjacent to exposures of ultramafic rocks are sites worth prospecting. The fact that the majority of ultramafic rocks contained in oceanic terranes are mantle derived implies that their contacts must be faulted. Therefore, margins of serpentinized ultramafic bodies are also potential sites of alteration and mineralization.

Linears defined by aeromagnetic lows in serpentinite may delineate zones of carbonatization. Magnetite formed during the serpentinization of ultramafic rocks produces a strong magnetic signature. Carbonatization results in the destruction of magnetite, creating zones of reduced magnetic susceptibility. The application of aeromagnetic lows as an exploration tool in delineating zones of carbonatization in ultramafics has been discussed by Gresens *et al.* (1982). This approach has been applied by Homestake Mineral Development Company in the Atlin camp and has proven successful (D. Marud, personal communication, 1989).

Once a fossil hydrothermal system has been identified the explorationist must assess whether or not the system contains gold. Various reported geochemical pathfinders associated with listwanitic alteration systems and related to gold mineralization can aid this assessement and are summarized in Table 3-2-2. All authors indicate that arsenic displays a consistent correlation with gold, as is typical of gold deposits in general. Bozek (1989) found that arsenic and antimony showed the strongest correlation and widest dispersion halo in carbonatized serpentinite at both the Pictou and Yellowjacket prospects near Atlin.

Alkalis also show a strong association with mineralization, potassium in particular often corresponds with abundance of mariposite. Gresens *et al.* (1982) found that lithium showed the widest and most regular dispersion halo within listwanitic rocks at the Mount Vernon deposit in Washington and suggest a tentative correlation of highest lithium with highest gold values. Both potassium and sodium also show enrichment haloes but are erratic in their lateral distribution, which is attributed to localization in unevenly distributed vein minerals. At the Erickson gold mine near Cassiar, Sketchley (1986) found that both barium and potassium display a positive correlation with gold in carbonatized metabasaltic rocks.

Base metals, most commonly copper, lead and zinc, are also associated with listwanitic lode gold deposits, but tend to have a somewhat more erratic distribution. Such an association may reflect the nature of the environment in which these deposits form. Primary concentration of base metals (massive sulphides) is inherent in the process of oceanic crustal formation. Later, epigenetic mineralizing fluids channeled along major shear zones within the oceanic rocks may encounter primary base metal concentrations and become selectively enriched.

TABLE 3-2-2

POTENTIAL PATHFINDER ELEMENTS FOR GOLD IN LISTWANITE

Location	Strong positive correlation with Au	Positive and sporadic correlation with Au	Source
Atlin, B.C.			
Yellowjacket	As, Sb	Ag, K	
Pictou	As, Sb	Ag, Cd, Cu, Pb, Zn	Bozek (1989)
Cassiar, B.C.			
Erickson	As, Ba, K, B	Ag, Cu, Pb, Zn, Sb	Sketchley (1986)
Washington State			
Mount Vernon	Li	K, Na, Zn, Pb	Gresens et al. (1982)
European ophiolites	As, K	Sb, Ba, B, Bi, Ag, Cu	Buisson and Leblanc (1985b)

CONCLUSION

Listwanite (carbonatized ultramafic rock) is a distinctive alteration assemblage commonly associated with quartz-carbonate lodes that have the potential for high-grade gold mineralization. Known listwanitic lode gold deposits are:

- Structurally controlled epigenitic deposits.
- Characteristically found within accreted oceanic terranes and associated with ophiolitic ultramafic rocks that have been tectonically dismembered.
- Generally high-grade, low-tonnage deposits with erratic distribution of gold.

A model relating listwanite alteration and gold mineralization to the deformation history of their ophiolitic hostrocks throughout the Canadian Cordillera, will provide a useful guide to future exploration for mesothermal gold deposits, historically the most significant gold producers in the province (Barr, 1980).

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THE ATLIN ULTRAMAFIC ALLOCHTHON: OPHIOLITIC BASEMENT WITHIN THE CACHE CREEK TERRANE; TECTONIC AND METALLOGENIC SIGNIFICANCE (104N/12)

By C.H. Ash and R.L. Arksey

KEYWORDS: Regional mapping, Atlin, ophiolite, oceanic crust, residual mantle harzburgite, listwanite, lode gold, thrust faults, tectonic mélange.

INTRODUCTION

Quartz-carbonate lodes are the accepted source of the abundant placer gold deposits in the Atlin region of northwestern British Columbia. Gold recovered from these deposits between 1898 and 1982 totaled 19.1 million grams (Debicki, 1984) and remains the economic mainstay for the town of Atlin.

Lode gold showings in the region are all structurally controlled by fault zones and spatially (genetically ?) associated with carbonatized ultramafic rocks (listwanites). A number of variably sized ultramafic bodies are exposed throughout the Atlin area, and are in all cases associated with oceanic crustal and sedimentary rocks. These ultramafic bodies were interpreted as intrusive in origin and termed the "Atlin intrusions" by Aitken (1959) but have not been adequately studied since that time, despite their economic significance.

A major objective of the Listwanite Project (Ash and Arksey, 1990a, this volume) is to classify the geotectonic environment of formation of the listwanite–lode gold association in British Columbia. Ultramafic rocks in the Atlin area were selected for detailed geological investigation, due to the excellent road access and current exploration activity. An area of roughly 50 square kilometres located along the eastern shore of Atlin Lake, encompassing the town of Atlin and Monarch Mountain, was mapped at 1:20 000 scale (Figure 3-3-1). Following this detailed mapping a regional reconnaissance survey of the area was undertaken in order to fit the detailed results into a broader geotectonic framework.

This report provides an updated assessment of the ultramafic rocks in the Atlin area and a new structural interpretation for the region.

REGIONAL GEOTECTONIC SETTING

The map area is situated near the western margin of the Atlin Terrane in northwestern British Columbia (Figure 3-3-1). The Atlin Terrane is an allochthonous package of tectonically emplaced and internally disrupted remnants of the mainly Late Paleozoic to possibly Late Triassic Tethyan oceanic crust (Monger *et al.*, 1982). It represents the northern extension of the Cache Creek Terrane which is presently interpreted as a subduction complex related to Late Triassic arc activity on Quesnellia and Stikinia. Radiolarian cherts are present within the subduction complex, and range in age

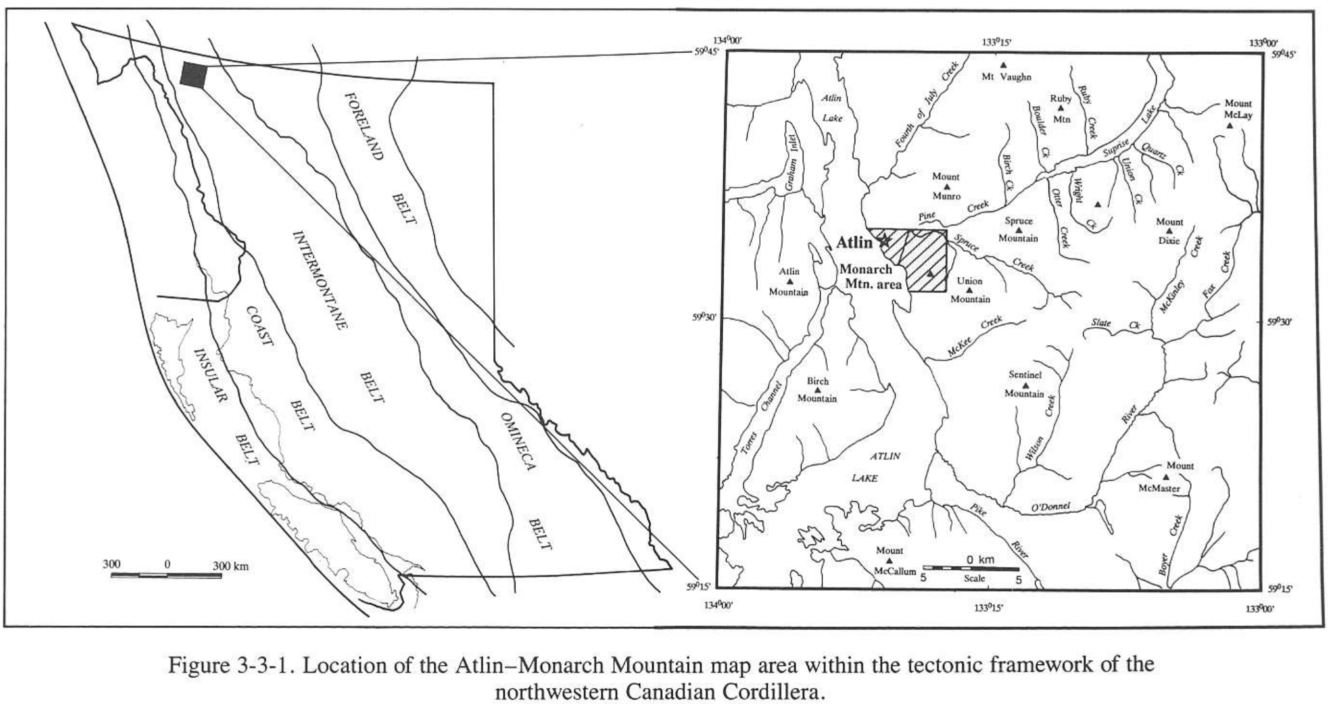

Figure 3-3-1. Location of the Atlin–Monarch Mountain map area within the tectonic framework of the northwestern Canadian Cordillera.

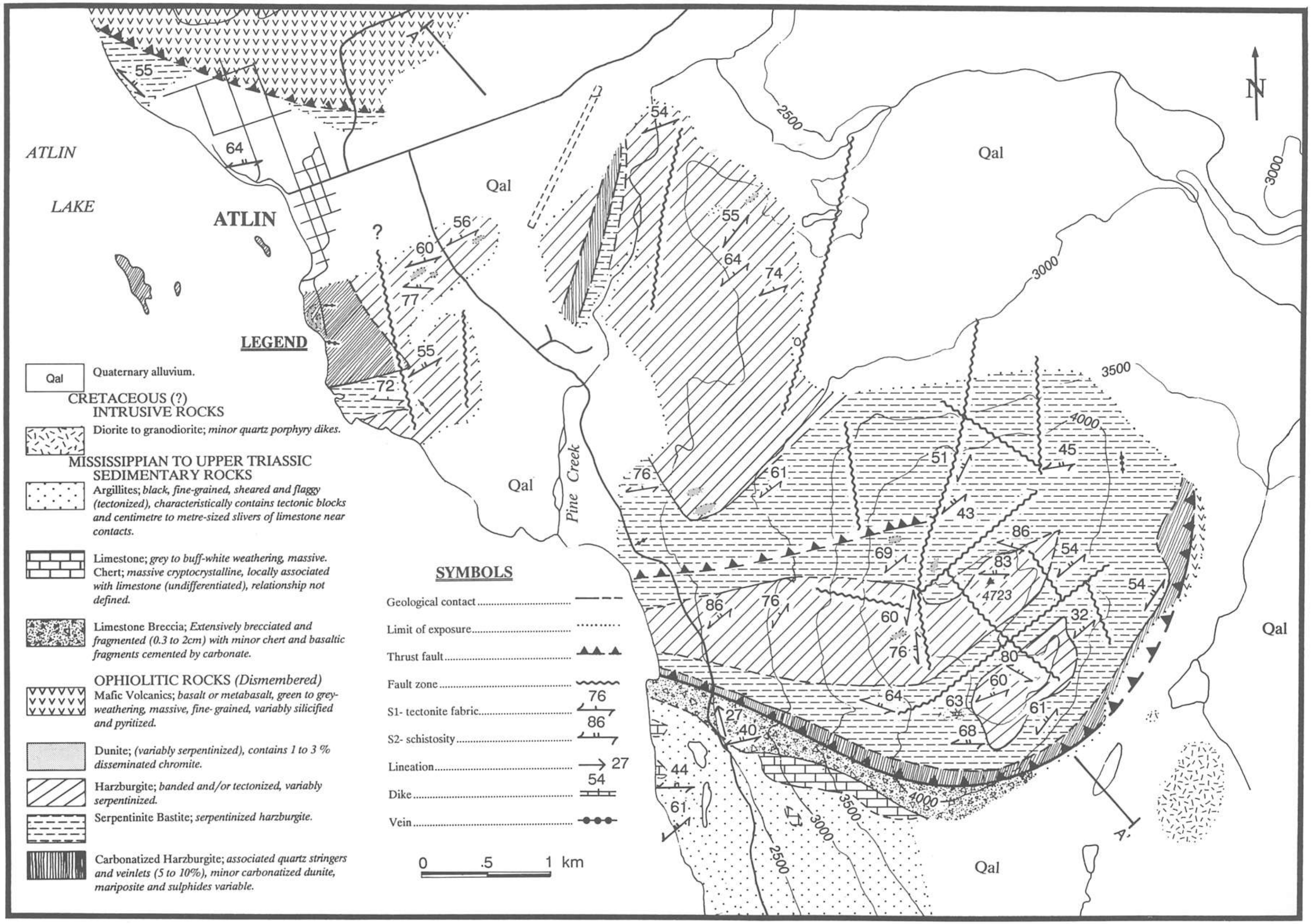

Figure 3-3-2. Simplified geology map of the Atlin–Monarch Mountain map area (*see* text for description of units).

from Permian to at least Late Triassic (Norian) (D.L. Jones *in* Monger *et al.*, 1982). This suggests that oceanic crust was forming for at least 100 million years and that the former ocean may have been quite large.

In the Late Triassic the Tethyan ocean basin was tectonically sandwiched between island arcs. This collage is Terrane I of Monger *et al.* (1982), or the Intermontane Superterrane of Gabrielse and Yorath (1989). The formation of the Omineca Belt to the east of Terrane I in the Middle Jurassic is thought to be the consequence of accretion of this composite superterrane to ancient North America (Monger *et al.*, 1982).

The combined effects of both closure of the Tethyan ocean in the Late Triassic and accretion of Terrane I in the Middle Jurassic resulted in the obduction of oceanic crust and upper mantle, represented by the Atlin "intrusions" in northern British Columbia. They are, therefore, typical Alpine ultramafic bodies.

ALPINE ULTRAMAFICS: TERMS AND CONCEPTS

Alpine ultramafic rocks are well exposed and documented in many mountain belts of the world and provide fundamental clues to their evolution. However, in the Canadian Cordillera these rocks have not received a great deal of attention and, as a result, many geologists in British Columbia still refer to them as "younger intrusions."

Originally, the term "alpine ultramafic" was used to describe the serpentinite-peridotite association occurring within orogenic belts, in contrast to the non-orogenic types or layered intrusions of stable cratons (Den Tex, 1969).

During the 1950s and 1960s, the accepted model for the origin of the alpine ultramafic rocks invoked incipient remelting of rocks from some kind of super-stratiform complex in the lower part of the crust or upper mantle during mountain building. The crystal mush produced was then injected into the upper crust under high tectonic pressure (Thayer, 1964, 1969). Aitken (1959) proposed a similar origin for the ultramafic rocks exposed throughout the Atlin region.

The geotectonic significance of the alpine ultramafic rocks was re-evaluated as a result of the evolution of plate tectonic theory. Most authors during the early 1970s came to support the hypothesis that ophiolites represent fragments of oceanic crust which were created at ocean ridges and subsequently transported to subduction zones at the convergent boundaries of lithospheric plates (Review in Coleman, 1977).

Moores (1973) redefined the term "alpine peridotite" to include ultramafic rocks occurring in linear Phanerozoic and late Precambrian deformed belts indicative of activity at accreting and consuming plate margins. Their formation involves diapiric rise at spreading centres, with subsequent emplacement and modification in subduction zones, and commonly results in mantle exposure.

Monger (1977a) used the term "ophiolitic terrane" for the Cache Creek Terrane in contrast to the non-ophiolitic, arc-related terranes of Quesnellia and Stikinia, to the east and west, respectively. The Nahlin ultramafic body in the Cache Creek Terrane of northwestern British Columbia consists predominantly of foliated peridotite (harzburgite tectonite) with minor mafic and ultramafic cumulates, trondhjemite and diabase dikes displaying chilled margins (Terry, 1977), and has been cited as a possible example of ophiolitic rocks in the Atlin Terrane (Monger, 1977a,b). Assessment of available data, combined with a limited reconnaissance survey of the body by the authors, indicates that it is indeed ophiolitic and may be more correctly referred to as the "dismembered Nahlin ophiolite".

The geotectonic significance of the ultramafic rocks representative of the type Atlin intrusions was not addressed by Monger (1977a), during his assessment of ophiolitic rocks in the Cache Creek Terrane. Subsequent workers in the Atlin region either maintained an intrusive origin for these ultramafic rocks (Christopher and Pinsent, 1979; Newton, 1985; Lefebure and Gunning, 1988, 1989) or suggested that they may represent oceanic crust but were inconclusive (Bloodgood *et al.*, 1989a).

There is clear evidence that the Atlin intrusions consist of parts of a typical ophiolite section (Coleman, 1977). They are comprised largely of mantle tectonite (described following); crustal plutonic rocks including both gabbro and diorite are also present, fitting Moores' (1973) definition of alpine ultramafics.

Mantle tectonite constitutes the basal section in most ophiolites (Nicolas *et al.*, 1980) and is now widely accepted as representing the oceanic uppermost mantle. These tectonites are comprised for the most part of foliated harzburgite (an ultramafic rock composed primarily of both olivine and orthopyroxene) with bodies of dunite, commonly in the shape of pods. Mantle tectonite represents the refractory residue from which a basaltic fraction has been extracted (depleted mantle) due to partial melting. Plutonic, hypabyssal and volcanic rocks characteristic of the typical ophiolite sequence crystallize from this basaltic melt (Coleman, 1977).

Mantle tectonites described from ophiolites throughout the world display geochemical signatures indicative of an origin by partial melting. When compared to the overlying cumulate rocks, the metamorphic harzburgites are depleted in the incompatible elements Ba, K, Rb, Sr, Zr, U, Th and rare-earth elements by one to two orders of magnitude and enriched in refractory elements Ni, Co and Cr to varying degrees (Coleman, 1977). Limited variations in both bulk rock and mineral phase chemistry throughout ophiolitic harzburgites also support a residual origin (Menzies and Allen, 1974; Malpas, 1978).

The metamorphic or tectonite fabric commonly exhibited by this rock type is developed by hypersolidus to subsolidus ductile deformation, attributed to asthenospheric flow in the mantle during and subsequent to partial melting (Nicolas *et al.*, 1980).

The term mantle tectonite is synonymous with metamorphic peridotite (Coleman, 1977), or harzburgite tectonite (Nicolas *et al.*, 1980) and may be more completely referred to as residual-mantle harzburgite tectonite. The term harzburgite tectonite will be used throughout the remainder of this report.

As is the case in the Atlin area, harzburgite tectonite exposed within Cache Creek Terrane near Fort St. James was interpreted to be intrusive in origin and was termed the "Trembleur intrusions" (Armstrong, 1949). Paterson (1973, 1977) identified the general association of ultramafics, gabbro, diabase and basalt in the area and suggested that the assemblage is most likely representative of oceanic crust and upper mantle. Ross (1977) conducted a detailed microstructural analysis of the Trembleur intrusion underlying Murray Ridge near Pinchi Lake. He concluded that penetrative F_1 and F_2 fabrics present are consistent with derivation from a mantle environment.

GEOLOGY OF THE STUDY AREA

The geology of the Atlin–Monarch Mountain area is illustrated in Figures 3-3-2 and 3-3-3. The area is underlain predominantly by variably serpentinized harzburgite tectonite which is continuous throughout, except for sporadic and volumetrically minor (2 to 3 per cent), lenticular masses of dunite enveloped by the host harzburgite. Minor pyroxenite dikes, 1 to 5 centimetres wide, occur throughout the harzburgite unit.

The division between harzburgite tectonite and serpentinite-bastite (alteration products after olivine and orthopyroxene respectively) in Figures 3-3-2 and 3-3-3 simply reflects differences in the degree of alteration of the harzburgite. Harzburgite tectonite preserves a clearly identifiable mantle tectonite mesoscopic texture. Contacts between partially altered harzburgite tectonite and the more intensely altered serpentinite-bastite are gradational.

The ultramafic rocks structurally overlie, along a northwesterly dipping thrust fault (Monarch Mountain thrust), a mélange assemblage of deep-water clastic to shallow-water carbonate sedimentary rocks to the south and massive metabasalts to the north. The sedimentary clastic and carbonate rocks are assigned to the Kedahda and Horsefeed formations, respectively (Monger, 1975). The age of the sedimentary rocks is poorly constrained but may range from upper Paleozoic to lower Mesozoic.

Metabasalts of the Nakina Formation (Monger, 1975) are interpreted to structurally overly the harzburgite tectonite unit along the Beavis fault zone near the northern margin of the map area.

The ultramafic rocks are cut by two distinct intrusions. The oldest intrusive phase is represented by granodiorite which forms a small, medium-grained, equigranular plug outcropping near the summit of Monarch Mountain. Porphyritic dike-equivalents of the granodiorite, 0.5 to 1.5 metres wide, intrude at several isolated locations. Potassium-argon isotopic ages from a small granodiorite body outcropping in the valley south of Monarch Mountain indicate a Middle to Late Jurassic age for the intrusions. Biotite obtained from the granodiorite and muscovite from a mineralized quartz vein cutting the granodiorite yield radiometric ages of 167.2 ± 4.7 Ma and 160.6 ± 7 Ma, respectively (K. Dawson, personal communication, 1989). These ages are consistent with the timing of granitic plutonism related to the collision of Terrane I with the North American craton.

Younger, fine-grained, occasionally plagioclase-porphyritic mafic dikes, 0.5 to 1.0 metre wide, were mapped at several locations. A mafic dike exposed in a trench on the Anna claims, near the summit of Monarch Mountain, crosscuts the carbonatized ultramafic rock and appears to postdate the listwanitic alteration. A tentative correlation with Recent olivine-phyric vesicular basalts which outcrop as small cinder cones (?) near Ruby Mountain is suggested. The basalts display spectacular columnar jointing where the unit overlies Tertiary to Quaternary pay-zone placer gravels along the banks of Ruby Creek.

ULTRAMAFIC ROCKS

HARZBURGITE

Harzburgite displays a relatively homogeneous modal distribution of both olivine and orthopyroxene (olivine: orthopyroxene ratio averages 7:3) or their altered equivalents, serpentine and bastite. Preferential orientation of the orthopyroxene imparts a weak to moderate foliation fabric upon most of the harzburgite. The rock is medium to coarse grained with a xenomorphic granular fabric.

Weathering gives rise to uneven, mottled surfaces on which lustrous brown to bottle-green pyroxenes and black chromite crystals are visible within a matrix of rusty brown olivine (Plate 3-3-1). Fresh surfaces are dark green and massive with the orthopyroxene grains readily distinguishable due to their lustre.

Three types of harzburgite have been recognized on the basis of structural style. In decreasing order of abundance these are: (1) foliated with no compositional banding, (2) massive, and (3) foliated with parallel compositional banding. Both the foliated and massive harzburgite are relatively

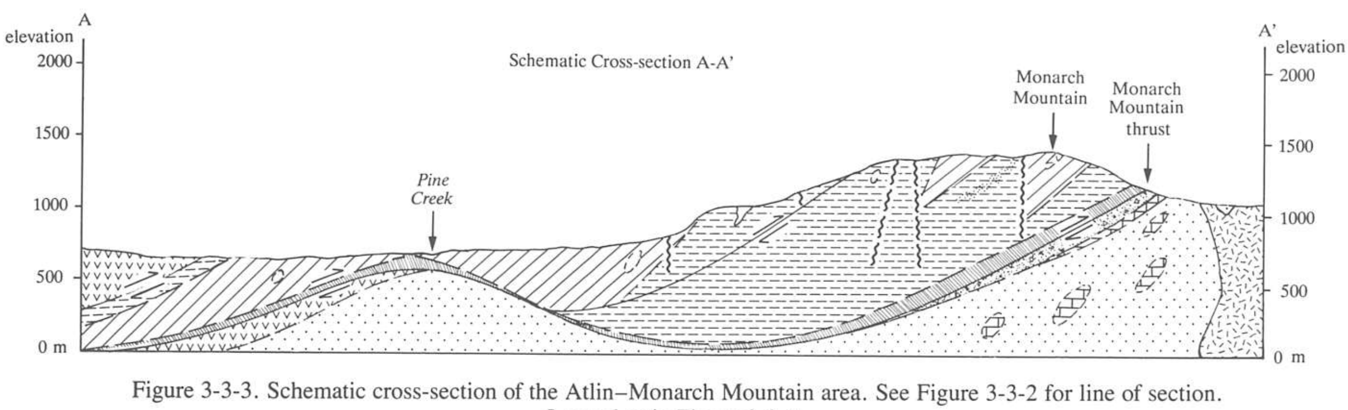

Figure 3-3-3. Schematic cross-section of the Atlin–Monarch Mountain area. See Figure 3-3-2 for line of section. Legend as in Figure 3-3-2.

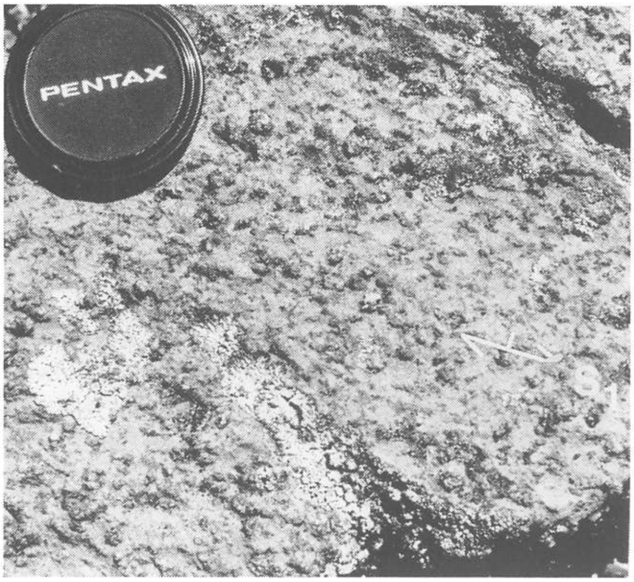

Plate 3-3-1. Weathering appearance of harzburgite with studded orthopyroxene in recessive dunite.

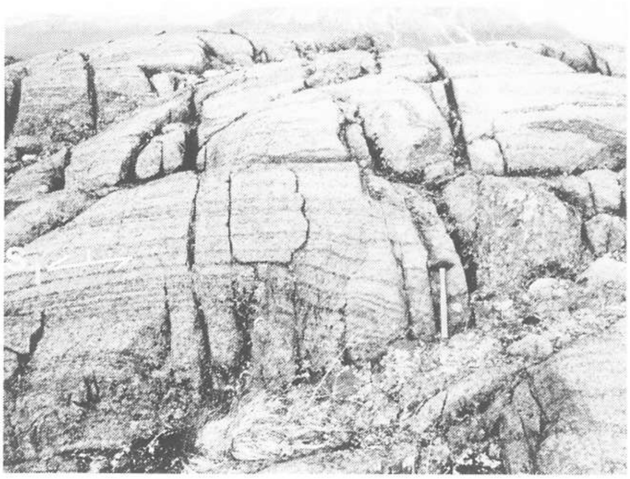

Plate 3-3-2. Tectonite banding (S_1) defined by alternating orthopyroxene-rich and poor layers. Hammer is 35 centimetres long.

homogeneous in modal composition, containing 65 to 85 per cent olivine and 15 to 35 per cent orthopyroxene uniformly dispersed with trace to accessory amounts of chrome spinel. Clinopyroxene may also be present as an accessory phase.

Areas of compositional banding generally consist of alternating pyroxene-poor and pyroxene-rich zones, on a 1 to 2-centimetre scale. Banding is commonly laterally discontinuous and disrupted. Well-developed banding is best exposed on a north-trending knoll 500 metres southwest of the summit of Monarch Mountain (Plate 3-3-2).

Texturally the harzburgite is medium to coarse grained porphyroclastic (as defined by Nicolas *et al.*, 1980). In thin section, chrome-spinel is typically dark red-brown and has an irregular habit. Larger orthopyroxene grains commonly have clinopyroxene exsolved along cleavage and kink planes.

DUNITE

Dunite bodies, hosted by harzburgite, are generally elongate and podiform with long axes concordant with the harzburgite foliation. These bodies vary from a few metres to several hundred metres in length. Weathering of dunite produces smooth tan-brown surfaces which are easily distinguishable from the rough-textured and slightly darker weathering harzburgite (Plate 3-3-3). The rocks generally display 1 to 4 per cent (0.2 to 0.4 mm) disseminated subhedral to euhedral black chrome-spinel grains. Contacts with the enveloping harzburgite are sharply defined by dramatic changes in orthopyroxene content over distances of less than a centimetre. Smaller, 0.5 to 1-metre, dunite pods occur for several metres into the harzburgite marginal to the larger dunite bodies.

These dunite bodies are interpreted to represent open-system magma chambers in the mantle, from which olivine and chromite have accumulated during fractional crystallization of ascending basaltic melts. This interpretion is consistent with the accepted origin of such bodies in other mantle sequences (Malpas, 1978; Duke, 1983; Gregory 1984).

PYROXENITE DIKES

Pyroxenite dikes are medium to coarse grained and display sharp contacts with harzburgitic hosts. Dikes are interpreted to be intruded as liquidus melts during the deformation of the host ultramafics. Orientations of the dikes are predominantly concordant but may also be oblique to highly discordant with the tectonite fabric, indicating synkinematic to postkinematic intrusion (Plate 3-3-4).

ALTERATION

Serpentinization has produced rocks partially or wholly altered to serpentine (predominantly antigorite) and minor magnetite. The degree of serpentinization varies from 20 per cent (rare) to totally altered varieties which characteristically develop close to fault zones. Most rocks are 40 to 70 per cent altered but the relict primary mineralogy is readily discernible. Bastitized orthopyroxene forms diffuse black spots in a dull grey background of serpentinized olivine. The least-

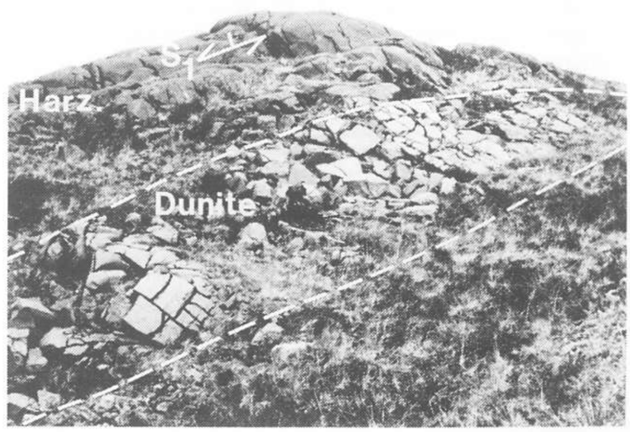

Plate 3-3-3. Typical lenticular (podiform) dunite body, with the long axis concordant with the tectonite fabric.

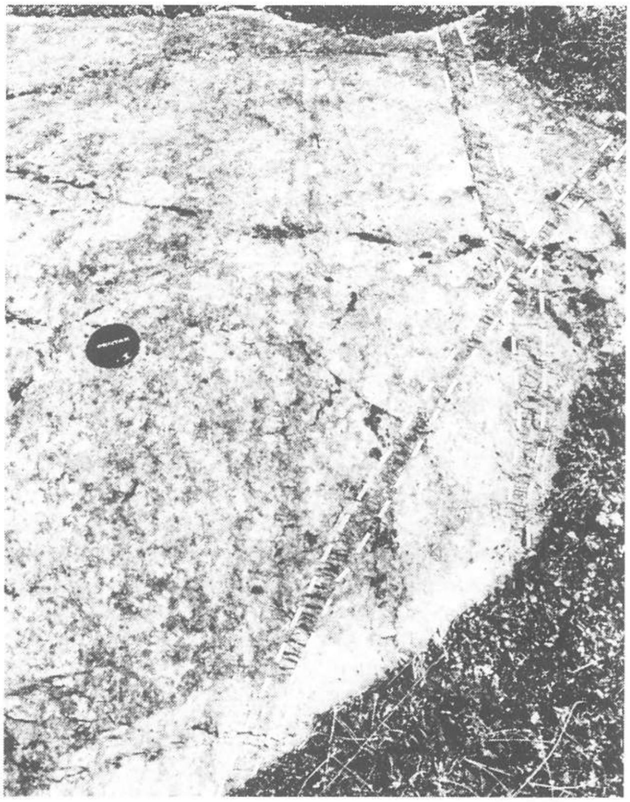

Plate 3-3-4. Concordant and discordant pyroxenite dikes in harzburgite suggest synkinematic to postkinematic emplacement under mantle conditions.

altered harzburgite exposures outcrop along a north-trending ridge approximately 60 metres southwest of the summit of Monarch Mountain.

Carbonatization (CO_2 metasomatism) of the ultramafic rocks is everywhere localized along fault zones. Alteration commonly envelopes or mantles the faults and produces a characteristic listwanitic alteration assemblage. Generally, this alteration decreases in intensity away from the fault as indicated by the inwardly zoned mineralogical variation: talc, talc + carbonate, carbonate + quartz + talc, carbonate + mariposite + quartz ± sulphides ± gold.

The onset of carbonatization of the ultramafic rocks is indicated by the appearance of talc in the serpentinized harzburgites seen in thin section. Initial replacement of bastite by talc and possibly mariposite was noted. Talc alteration of the ultramafic rocks extends from one to tens of metres away from the fault and is apparently a direct function of the scale of the structure. Intense alteration, most important economically because it hosts quartz veins that may be auriferous, is commonly concentrated in the fault zone itself.

Pervasive alteration of the ultramafics produces carbonate rocks consisting predominantly of iron-rich magnesite (Newton, 1985) and ankerite. Dolomite is common in microveinlets to veins up to 10 to 15 centimetres wide. These veins consist of white dolomite with millimetre to centimetre-scale inclusions of iron-magnesite, which may represent xenoliths of the host carbonatized ultramafite.

Quartz veining appears to be episodic and occurs on a variety of scales. Millimetre-scale quartz veinlets are common and may be preferentially oriented, however, vein networks are also present. These quartz veins may be related to desilicification of the ultramafic rocks during carbonatization. Generally later quartz veins are several centimetres to tens of centimetres wide.

Weathering of the pervasively altered ultramafics produces easily identified potentially gold-bearing targets. Outcrops are orange-brown in colour with conspicuous white dolomite and quartz veins which stand out due to differential erosion.

Chemical variations related to the alteration and mineralization of these carbonatized fault zones are currently under investigation.

Tectonic Mélange Assemblage

Massive fine-grained grey-weathering limestone characteristically forms tectonic blocks from centimetres to hundreds of metres in size. These carbonate (± chert) knockers are hosted in a fine-grained, extremely fissile and flaggy argillite unit (Kedahda Formation, Monger, 1975). The mélange is extensive throughout the southern part of the map area. The larger scale tectonic features are well exposed along the shore of Atlin Lake.

STRUCTURE

The harzburgite tectonite unit displays both an early (S_1) mantle fabric that is overprinted by a later (S_2) fabric related to thrusting which is also well developed within the argillites.

Field measurements obtained from the harzburgite unit include:

S_1: A twofold division defined by a foliation (solid state dislocation) fabric outlined by elongate grains or crystal aggregates of orthopyroxene, and compositional banding (metamorphic differentiation). Both these features are typically formed by mantle processes (Plate 3-3-2).

S_2: A serpentinite-bastite fabric, defined by black-weathered bastite (after orthopyroxene) within the lighter grey weathered serpentine (after olivine). Structures may range from partially serpentinized harzburgites which display trails of bastite around orthopyroxene cores, to highly strained and commonly more altered varieties in which a mylonitic texture is developed as thinly banded (1 to 3 mm) bastite-serpentinite with bastite porphyroclasts after orthopyroxene (Plate 3-3-5). These structures are consistent with the thrusting direction and are interpreted to be related to emplacement.

Strongly sheared and schistose grey-green serpentinite containing dark brown harzburgite knockers, 2 to 15 centimetres across, exposed along the shore near the town of Atlin, represents a typical serpentinite mélange assemblage (Plate 3-3-6). The mélange and related schistosity are S_2 deformation features.

Pronounced schistosities developed within the argillite unit are consistent with the S_2 fabric in the ultramafic rocks.

Thrust Faulting

The arcuate surface trace of the Monarch Mountain thrust fault follows the lower southern and eastern slopes of Monarch Mountain. The fault contact does not outcrop, however, its approximate location is well constrained by the contrasting lithologies and styles of deformation across it. Lithologies on both sides of the fault display evidence of intense deformation with associated hydrothermal carbonate alteration. Mantle harzburgite exposed near the basal surface of the Monarch Mountain thrust is ductilely deformed and intensely carbonatized (listwanitic) with quartz stingers oriented parallel to the thrust surface (Plate 3-3-7).

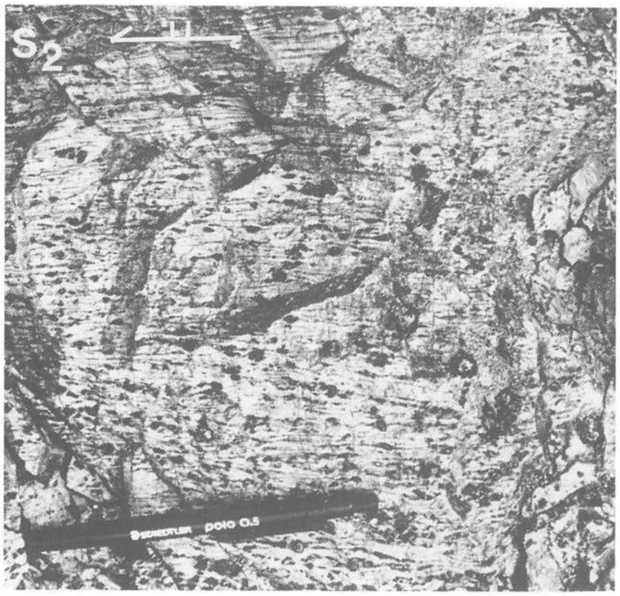

Plate 3-3-5. Well-banded mylonitic serpentinite bastite fabric (S_2) with relict orthopyroxene porphyroclasts.

Massive limestones and basalt in the footwall of the thrust yield to brittle fracture and brecciation as evidenced by the cataclastic features produced. Massive limestone exposed directly below the thrust, in a road cut along Warm Bay road, is intensely brecciated into 1 to 3-centimetre angular fragments which are cemented (flooded) by hydrothermal carbonate. Within this outcrop well-developed mullions plunge at roughly 30° to the northwest, providing further support for the concept of southeasterly directed emplacement (Plate 3-3-8).

Drill-core information from the Pictou property (MINFILE 104N 044; Homestake Mineral Development Company) defines a similar fault contact dipping at approximately 30° to the northwest. Core logs indicate that intensely carbonatized (listwanitic) harzburgite at this location sits structurally above intensely brecciated basalt cemented by carbonate.

Numerous, later high-angle faults are evident throughout the Monarch Mountain plateau. They are clearly evidenced in the field by linear topographic depressions tens of metres deep. Exposures on both sides of the depressions are increasingly altered toward the centre of the fault zone.

Structural Interpretion

A dominantly northeast structural grain characterizes the map area. S_2 fabric elements and thrust plane geometry are essentially parallel and reflect the southeasterly directed emplacement within the map area. S_1 fabric elements are more or less consistent with the northwest structural grain but also show the most deviation from it. The variation in S_1 may not reflect primary mantle geometry since the ultramafic nappe has been affected by later high-angle normal faulting. This may, in part, reflect crushing of the body during emplacement at higher structural levels. It may also be due, in part, to high-angle regional-scale transcurrent faulting which has affected the region (Monger, 1984).

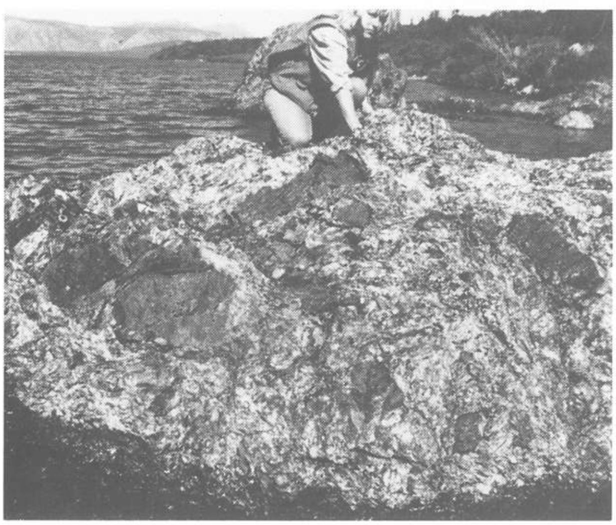

Plate 3-3-6. Serpentinite mélange with harzburgite knockers in a matrix of highly sheared serpentinite, exposed along the shore of Atlin Lake near the town of Atlin.

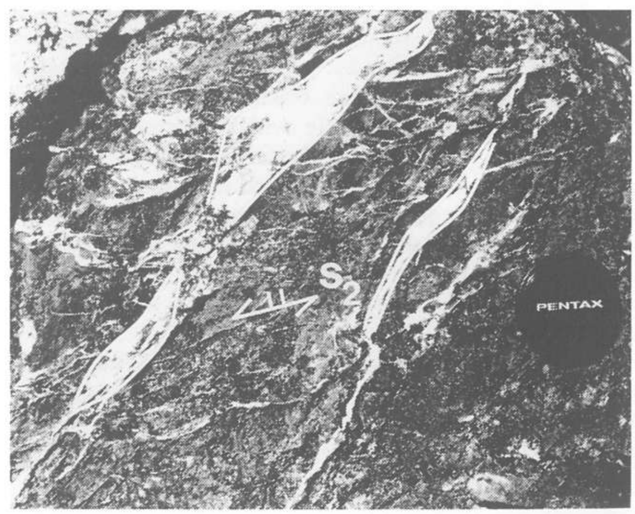

Plate 3-3-7. Quartz stringers within intensively carbonatized harzburgite near the base of the Monarch Mountain thrust. Note the dilational zones at moderate to high angles from S_2.

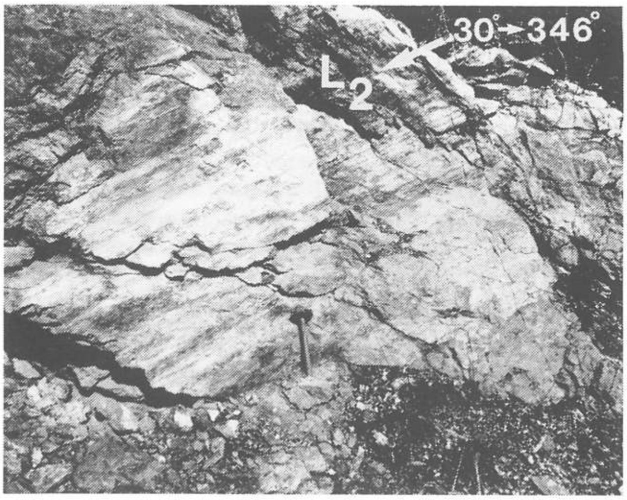

Plate 3-3-8. Northwesterly plunging mullions in brecciated limestone cemented by carbonate, exposed in road cut along Warm Bay road below the Monarch Mountain thrust.

REGIONAL GEOTECTONIC IMPLICATIONS

Analysis of the available regional geological data supplemented by new observations from aerial photography and the authors' reconnaissance suggests that the locally defined thrusting is consistent with the regional structural pattern. The 1:70 000 aerial photographs of the area display curvilinear features, on a scale of tens of kilometres, which verge toward the southeast.

A 1:20 000 geological compilation map of the Atlin area by Lefebure and Gunning (1989) depicts elongate bodies of ultramafic rocks which also define a lobate distribution pattern consistent with the structural grain seen on the aerial photographs. This map pattern is similar to that described by Monger (1975) for the Nakina Lake area, in which fault planes are marked by serpentinite slices. These strongly altered and deformed ultramafic slivers are generally contained in fault or mélange zones that may attain hundreds of metres in width. An example is the Beavis fault, described by Bloodgood *et al.* (1989a) as "characterized by brecciation and intercalation of diverse Cache Creek lithologies." Lithological diversity is clearly evident in exposures along McKee Creek, which include ultramafic, volcanic and sedimentary rocks chaotically distributed on a scale of tens of metres. Varying degrees of alteration and deformation of all lithologies by both shearing and brecciation are evident. Observation of limited outcrop exposed by placer operations on Pine Creek and Spruce Creek also suggests the presence of fault or mélange zones.

Detailed drilling by Homestake on the Yellowjacket property (MINFILE 104N 014) has demonstrated that the Pine Creek fault is a low-angle reverse fault dipping approximately 25° northwest (D. Marud, personal communication, 1989)

Recent 1:50 000 geological mapping by Bloodgood and Bellefontaine (1990, this volume) has demonstrated that southeasterly directed thrusting is a significant structural feature of the Atlin Terrane to the south of the study area.

No coherent ophiolite stratigraphy is evident in the Atlin area, however all the lithological components are represented and have been juxtaposed by tectonic imbrication.

The amount of oceanic crust emplaced or preserved in a classical ophiolite sequence is clearly a function of both the dynamics of oceanic plate consumption (obduction) and the degree of deformation that has affected that obducted crust. In the Atlin map area syn-emplacement deformation is related to the compressional tectonics during thrusting which dismembered the oceanic suite. This dismemberment is further complicated by later transcurrent faulting (post-emplacement deformation).

ECONOMIC GEOLOGY

LODE GOLD

Gold-bearing quartz-carbonate veins have been unanimously proposed by workers in the Atlin area as the source for placer deposits (Aitken, 1959; Monger, 1975; Ballantyne and MacKinnon, 1986; Lefebure and Gunning, 1988; Rees, 1989). Although placer mining has won significant quantities of gold in the Atlin region (Debicki, 1984), only minor amounts of lode gold have been recovered (Lefebure and Gunning, 1988).

Lode gold showings in the Atlin area are characteristically located within faults or fault zones and are spatially associated with carbonatized ultramafic rocks (Bloodgood 1989b; Rees, 1989), as is common with this type of deposit (*see* discussion by Ash and Arksey, 1990, this volume). As these deposits are structurally controlled, an adequate understanding of the structural environment of the area is critical to their exploration. The new structural interpretation outlined above will hopefully facilitate that process.

Interestingly, most of the current and past placer operations in the area occur along streams which follow the linear depressions that define the inferred thrust contacts. Locally the surface trace of the Monarch Mountain thrust along the southeastern flank of Monarch Mountain (Figure 3-3-2) is covered by overburden and may represent a potential zone of gold mineralization.

Aspects of the timing of carbonatization of the ultramafic rocks and related quartz veining and possible relationships between intrusive granitoids and quartz lodes are currently under investigation.

CHROMITE

Dunite bodies within the harzburgite unit represent potential hosts for podiform chromite deposits (Duke, 1983) but only minor amounts (1 to 4 per cent) of disseminated chromite were noted during our field studies.

CONCLUSIONS

Ultramafic rocks in the Atlin area represent an allochthonous unit of predominantly residual mantle harzburgite tectonite. Structural data indicate that thrust emplacement was southeasterly directed.

The "Atlin intrusions" are clearly not intrusive but rather are of ophiolitic origin for the following reasons:

- The lithological association of foliated harzburgite with pockets of dunite and pyroxenite dikes is consistent with that identified in residual mantle sequences from ophiolites throughout the world.
- The identification of synkinematic high-temperature ductile subsolidus to hypersolidus deformation features (S_1), combined with mineralogical homogeneity of the harzburgite, supports a residual mantle origin for the ultramafic rocks. Malpas (1978) demonstrated that simultaneous crystallization of both olivine and orthopyroxene from a basaltic melt is not possible by cumulate processes, due to the reaction relationship between olivine and orthopyroxene at pressures compatible with the environment of oceanic crustal formation.
- Oceanic upper mantle material is presently resting on deep to shallow-water sediments and basalts and, therefore, invokes a structural unconformity.

Mapping by Bloodgood *et al.* (1989b), indicates that tectonic imbrication has created slivers of ultramafite associated with other oceanic lithologies, displaying consistent structural relationships. Tectonic slices are separated in places by tectonic mélange that displays a range of brittle to ductile deformational features reflecting the contrasting lithological elements present and differing levels of associated deformation.

Later crustal-scale transcurrent faulting may have obscured the paleo-emplacement direction and could account for the local diversity of the structural fabrics relative to the general northwest structural grain characteristic of the Atlin Terrane (Monger, 1975, 1977b).

An understanding of the geotectonic framework is critical to the successful exploration for lode gold deposits in the Atlin area.

ACKNOWLEDGMENTS

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THE ANATOMY OF THE SHULAPS OPHIOLITE

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KEYWORDS: Regional geology, Shulaps complex, ultramafic, peridotite, ophiolite, serpentinite mélange, Bridge River Terrane, Cadwallader Group, East Liza igneous suite, Yalakom fault.

INTRODUCTION

The Shulaps ultramafic/mafic complex is exposed in the mountains of the Shulaps Range, within the Tyaughton Creek map area (Schiarizza *et al.*, 1989) approximately 50 kilometres northwest of Lillooet in the Cordillera of southwestern British Columbia (Figure 3-4-1). It forms one of the largest bodies of ultramafic rocks in the orogenic belt, underlying an area of approximately 180 square kilometres. The complex is situated along the boundary of the Intermontane and Insular superterranes (Monger *et al.*, 1982), which is marked in the area by the Yalakom fault system. Its origin and structural evolution have important bearing on unravelling the tectonic collage of suspect terranes that form this part of the Cordillera (Price *et al.*, 1985; Potter, 1986; Rusmore, 1987).

The Shulaps complex was first mapped in detail by Leech (1953), who concluded that the peridotites were part of a non-stratiform plutonic complex that was later intruded by smaller lenses of gabbro and pyroxenite. More recently, Nagel (1979) and Wright *et al.* (1982) have suggested an ophiolitic origin for the peridotite complex, interpreting it as a residual mantle tectonite section. These workers established that the western basal contact of the complex is a serpentinite mélange containing exotic blocks of sedimentary and volcanic rocks. These rocks are tentatively correlated with supracrustal rocks in the oceanic Bridge River Terrane, which structurally underlies the mélange. The mélange also contains blocks of ultramafic and mafic plutonic rocks which may represent fragments of Layer 3 of the Bridge River oceanic crust. Leech (1953), Nagel (1979) and Wright *et al.* (1982) have suggested that stratigraphic relationships exist

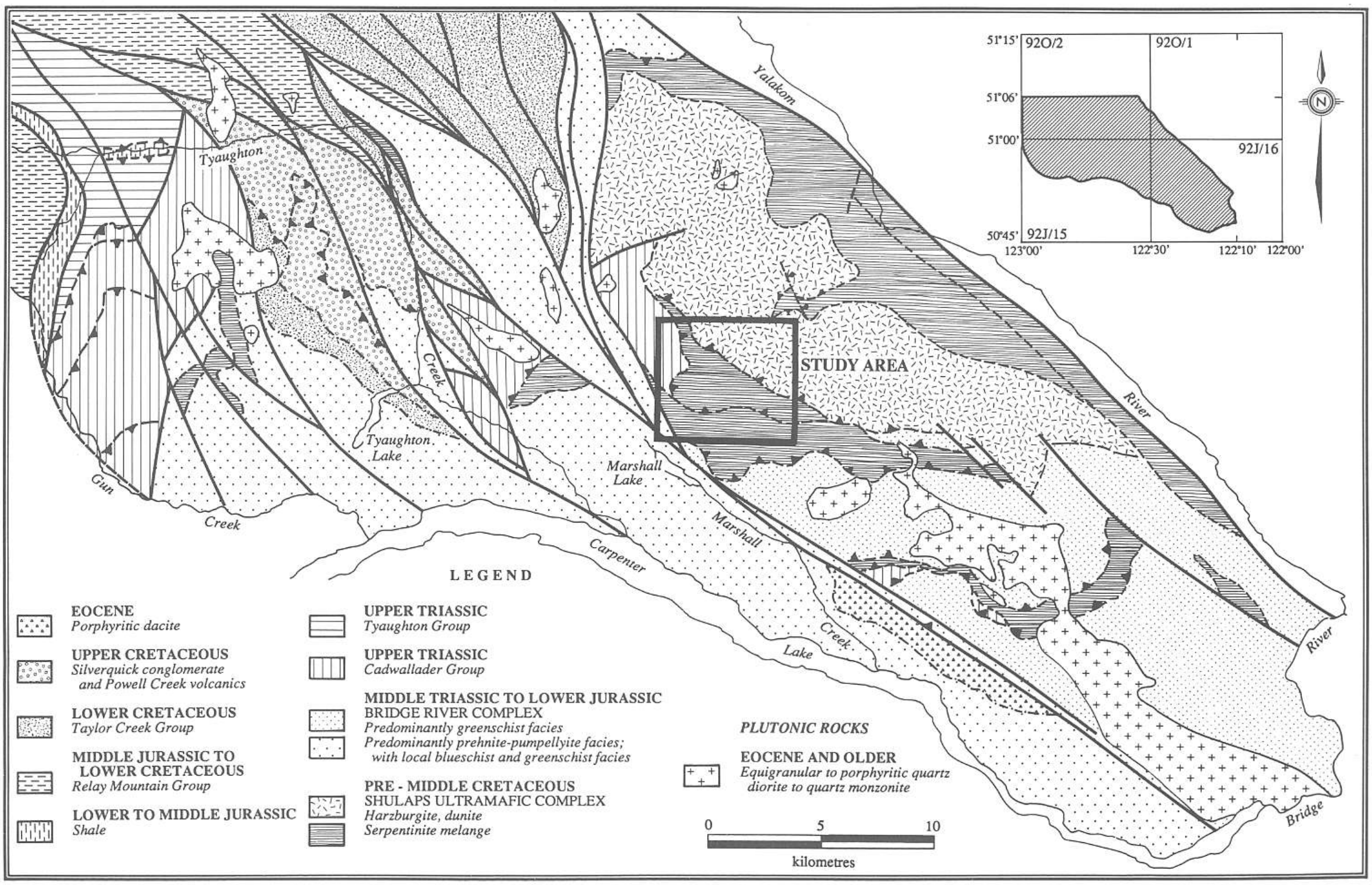


Figure 3-4-1. Regional setting of Shulaps ophiolite complex and location of the study area.

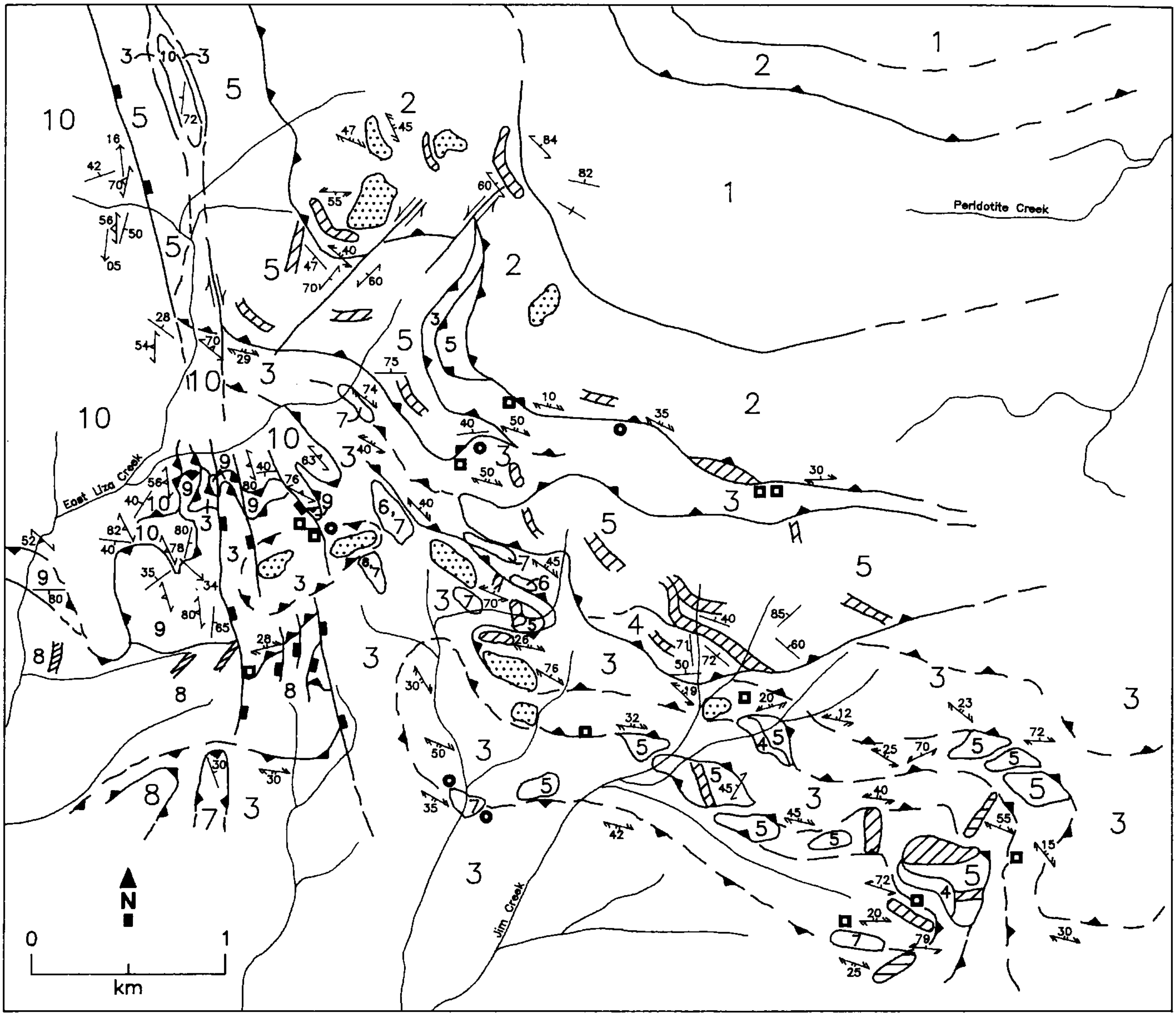
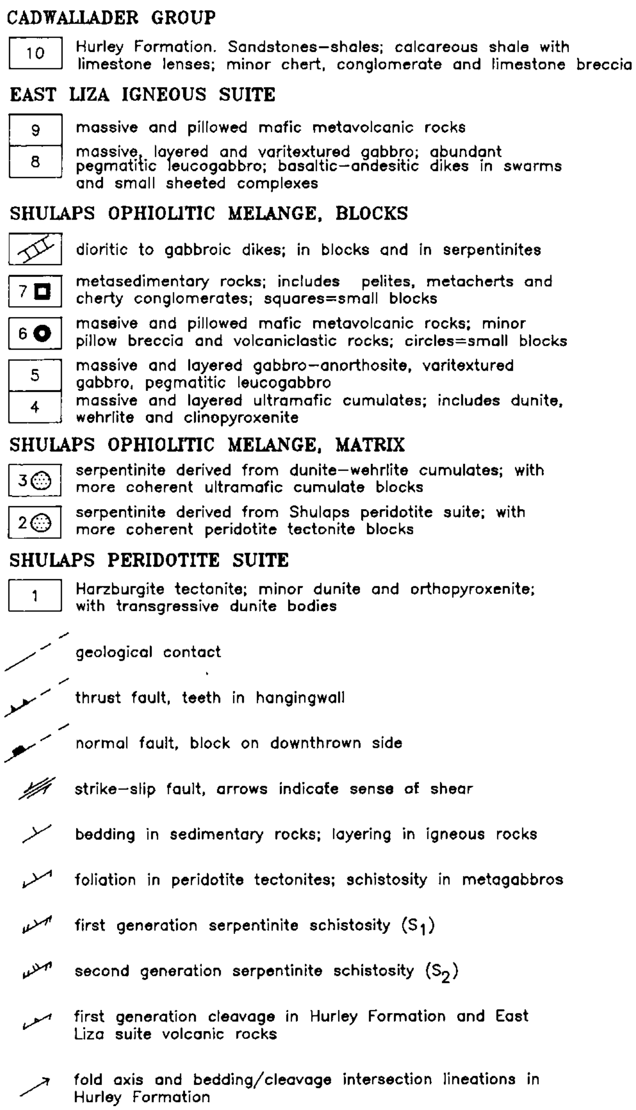


Figure 3-4-2. Schematic geological map of the study area, with legend.

between gabbroic and volcanic rocks within blocks in the mélange, implying that the basal mélange of the Shulaps complex may represent the highly dismembered remnants of a more or less complete ophiolitic suite. A segment of a Triassic sedimentary apron (Hurley Formation) of the Cadwallader volcanic arc assemblage (Rusmore, 1987) lies with fault contact directly to the west of the Shulaps complex. According to Rusmore (1987) and Potter (1986), the more or less contemporaneous Cadwallader arc and Bridge River ocean basin became juxtaposed during movement on the Shulaps sole thrust. Since this thrust has an overall westerly vergence (Potter, 1986), the arc restores palinspastically to the west of the basin. The Bridge River ophiolitic assemblage may thus have developed in relation to back-arc spreading (Potter, 1986).

To date, no modern account exists of the petrogenesis and structural evolution of the mantle peridotite section of the Shulaps complex and the underlying ophiolitic mélange. Further, the stratigraphic and structural relationships of the Bridge River Terrane, the adjacent fault-bounded Cadwallader Terrane and the Shulaps complex remain obscure. This study was undertaken primarily to establish the tectonic and plutonic evolution and petrogenesis of the Shulaps ophiolite complex and its Mesozoic accretionary history. Detailed geological mapping on scales ranging from 1:6000 to 1:14 000 was carried out in a two-week period in summer 1988 and a four-week period in the summer of 1989, covering an area of approximately 20 square kilometres centred around the upper courses of Jim Creek and East Liza Creek, along the southwestern edge of the Shulaps complex (Figure 3-4-2). This area comprises the critical transition from coherent thrusts sheets of residual mantle peridotite of the Shulaps complex to underlying ophiolitic mélange. It contains the bulk of the exotic blocks of gabbro and pyroxenite in the mélange, together with less abundant smaller blocks of volcanic and sedimentary rocks. In addition, on the western



side it covers a small segment of the contact with sedimentary rocks of the Hurley Formation in the Cadwallader Group. This report provides a detailed description of the lithological units and their structural relationships in this study area.

LITHOTECTONIC SUBDIVISION

The map units in the study area may be conveniently subdivided into four main lithotectonic units (Figure 3-4-2), which are in thrust contact with one another and display the following stacking order, from structural top to bottom:

(1) The Shulaps peridotite suite, exposed in the northern part of the study area (Unit 1).

(2) The Shulaps ophiolitic mélange, which is exposed along the southwestern margin of the peridotite suite and occupies the central and southern part of the map area (Units 2 to 7);

(3) The East Liza igneous suite of mafic plutonic and volcanic rocks, exposed in the southwestern part of the study area (Units 8 and 9);

(4) The Cadwallader Group, which comprises sedimentary rocks of the Hurley Formation and is exposed on the west flank of the Shulaps complex and East Liza suite (Unit 10).

Together, these four units comprise a complicated, polyphase, southwesterly verging, linked thrust system, which at a later stage became overprinted on its western flank by a transtensional high-angle fault system.

The Shulaps peridotite suite occupies the upper part of the thrust system. It extends far to the north and east of the study area (Figure 3-4-1) and underlies the highest peaks within the Shulaps Range. Work by Leech (1953) and Wright *et al.* (1982) suggests that it consists entirely of variably serpentinized, layered harzburgite tectonites with locally abundant dunite bodies. Preliminary field observations suggest that the upper thrust unit of mantle peridotite consists of a shingled array of moderately northeast-dipping thrust sheets of more or less coherent peridotite, separated from one another by shear zones consisting of intensely foliated serpentinite. The unit has been interpreted by Wright *et al.* (1982) as an obducted fragment of depleted oceanic upper mantle.

The Shulaps ophiolitic mélange occurs in a northwest trending belt, up to 5 kilometres wide, along the southwestern edge of the Shulaps peridotite suite. It is spectacularly exposed along the southwestern slopes of the Shulaps Range. To the northwest, it terminates abruptly against a high-angle fault system which marks the boundary between the Shulaps complex and a fragment of the Cadwallader Terrane. To the southeast, the belt extends into the Hog Creek imbricate zone delineated by Potter (1983). The mélange underlies the imbricate thrust system of the Shulaps peridotite suite with a moderately northeast-dipping structural contact. Potter (1983, 1986) has shown that the belt overlies, with gently east-dipping thrust contact, metasedimentary and metavolcanic units of the Bridge River complex directly southeast of the study area. The belt thus constitutes a partly exhumed duplex structure that defines the boundary zone between an upper plate consisting of a telescoped section of oceanic upper mantle peridotite and a lower plate of telescoped oceanic supracrustal sequences.

Internally, the mélange belt comprises a number of smaller duplexes which form an extension of the Hog Creek imbricate zone. These duplexes may be subdivided in terms of both the protolith types of the serpentinites that make up the voluminous matrix of the mélange belt, and the igneous and sedimentary lithologies that occur as abundant blocks within the serpentinite matrix. The matrix of the mélange is subdivided in two northwest-trending belts which maintain a consistent structural position in the thrust system. The serpentinite matrix of the upper belt is derived from low-temperature alteration of protoliths found only in the overlying mantle peridotite suite, whereas the protoliths of the lower belt comprise a suite of ultramafic cumulate rocks including dunite, wehrlite and clinopyroxenite. Within each belt these protoliths are locally preserved in more or less coherent blocks enveloped by intensely sheared serpentinite. The ultramafic cumulate prololiths also occur at the bases of two gabbroic blocks within the ultramafic cumulate-derived serpentinite belt.

Both mélange belts contain abundant boudins and more coherent sections of gabbroic to dioritic dikes which locally preserve chilled margins and contact aureoles of porphyroblastic olivine-talc-serpentine schists (Leech, 1953; Nagel, 1979). However, a clear distinction can be made between the two belts as regards the occurrences of blocks representing dismembered ultramafic-mafic plutonic complexes, as well as blocks of sedimentary and volcanic rocks. The mantle-peridotite-derived serpentinite mélange does not appear to contain any such blocks in the area studied. The most extensive belt of plutonic blocks is situated directly beneath the basal mantle-peridotite-derived serpentinite shear zone and preserves the thickest coherent section of ultramafic cumulates observed in the area. Moreover, blocks of plutonic complexes situated structurally lower in the ultramafic-cumulate-derived serpentinite mélange comprise generally only higher level gabbroic sections of the plutonic complexes. Sedimentary and volcanic blocks are irregularly distributed throughout the lower serpentinite belt, but are particularly prominent in number and size directly beneath the main belt of plutonic blocks.

The East Liza igneous suite forms a separate thrust unit of limited extent in the southwestern part of the map area. It is structurally overlain by the ultramafic-cumulate-derived serpentinite mélange, and overlies intensely folded units of the Hurley Formation with marked thrust contact. In the south it comprises gabbros which in all field aspects resemble those of the main gabbro blocks within the mélange. To the north, the gabbros appear to be in nonconformable stratigraphic contact with overlying volcanic rocks in a poorly exposed area dominated by abundant outcrops of dikes with screens of volcanics. The unit is equivalent to the greenstone-gabbro complex of Leech (1953).

Sedimentary rocks of the Late Triassic Hurley Formation of the Cadwallader Group occupy the western part of the map area. They comprise an upward-fining sequence of siliciclastic turbidites including some volcaniclastic rocks, associated limestone breccia and bedded chert. In the northwest, the unit is in abrupt, high-angle fault contact with a block of gabbro within the serpentinite mélange. Farther south, it occurs in a number of half windows beneath the mélange and the volcanic rocks of the East Liza suite. Rocks resembling Hurley lithologies also occur as exotic blocks within the serpentinite mélange.

LITHOLOGY AND STRUCTURE OF MAP UNITS

SHULAPS PERIDOTITE SUITE (UNIT 1)

The Shulaps peridotite suite comprises the bulk of the ultramafic rocks in the Shulaps complex as defined by previous workers (e.g., Leech, 1953; Wright *et al.*, 1982), and is exposed in the northern part of the study area. The rocks form part of a coherent basal thrust sheet of mantle peridotite that strikes northwest, dips 40° to the northeast, and is approximately 300 metres thick. The sheet is bounded at its top and bottom by serpentinite shear zones up to 500 metres thick. The boundaries are sharp structural contacts which are parallel to schistosity in the serpentinite matrix of the shear zones.

Lithologies within the peridotite sheet are dominantly layered and massive, foliated harzburgite with subordinate dunite and orthopyroxenite. Compositional layering is defined on centimetre to metre scale by modal variations in the orthopyroxene content of the harzburgite, and by parallel phase boundaries between harzburgite, dunite and orthopyroxenite. Numerous irregularly shaped pods of dunite occur within the peridotite. They are variable in size, ranging from less than a metre to several tens of metres in diameter. The dunites cut across the peridotite tectonite fabric in an irregular manner and the margins of the bodies appear undeformed. Most bodies contain abundant disseminated chromite and thin chromite stringers with variable orientations. The chromite grains range up to 1 centimetre in size and are generally euhedral.

A penetrative mineral foliation and lineation, which in most localities is parallel to the compositional layering, is observed in the harzburgite. Foliation and lineation are defined by a weak to moderate preferred orientation of orthopyroxene and spinel grains varying in size from 1 to 15 millimetres. The linear aspect of the fabric is outlined by chromite pull-apart textures. The texture of the peridotite tectonites can be classified as protogranular to mildly porphyroclastic. Mesoscopic folds of layering with associated axial planar foliation have not been observed. Layering and parallel foliation have rather constant orientation, dipping steeply to the north-northeast or south-southwest. Mineral elongation lineations are subvertical in the foliation plane. According to data from Leech (1953) and Wright *et al.* (1982), the regional attitude of layering and foliation in the Shulaps complex as a whole is similar to that observed in the study area, with steep southwesterly dips predominating. The attitude of the planar fabrics is markedly oblique to the fabric of the serpentinite shear zones in which the main serpentine foliation has a moderate northeasterly dip.

MANTLE-PERIDOTITE-DERIVED SERPENTINITE MÉLANGE (UNIT 2)

The unit forms a zone 300 to 500 metres thick that dips, on average, at an angle of 40° to the north and northeast, beneath the basal thrust sheet of the mantle peridotite suite. It structurally overlies the belt of large gabbro blocks which form the steep cliffs in the western part of the map area. Farther to the east, the unit directly overlies the ultramafic-cumulate-derived serpentinite belt (Unit 3) on the upper the upper slope of Shulaps Peak. Where the intervening main gabbro-block level is missing, the contact between the two serpentinite belts may be difficult to identify. However, suitable outcrops of matrix rocks which show transitional stages of alteration of the different protoliths are readily available on both sides of the contact in most areas.

The main structural grain of the belt is defined by a braided network of narrow zones containing an intensely schistose scaly serpentine fabric showing abundant evidence, in the form of fibrous serpentine slickensides, for reverse dip slip and oblique slip. Locally these serpentinite strands have a mylonitic aspect containing well-developed C-S fabrics which invariably indicate southwest-directed thrusting. This fabric is referred to as the second generation serpentine schistosity (S_2) in the legend of the geological map. It constitutes the younger component of a composite serpentinite fabric that can be observed in lozenge-shaped serpentinite blocks surrounded by the S_2 serpentine zones. Within such blocks an older schistosity (S_1) is generally inclined at a high angle to the main serpentinite shear zone fabric and bends gradually or abruptly into the late fabric at the edges of the blocks. Locally the first generation serpentine schistosity has itself the appearance of a fine scale C-S fabric.

ULTRAMAFIC-CUMULATE-DERIVED SERPENTINITE MÉLANGE (UNIT 3)

This component of the mélange occupies most of the eastern and southern parts of the map area. It extends eastwards into the Hog Creek imbricate zone delineated by Potter (1983). Previous workers have tacitly assumed that the serpentinite matrix of the mélange, as a whole, was derived from the overlying mantle peridotite suite. The present study, however, has revealed that this is only true for the upper part of the mélange (Unit 2). The matrix of the lower part consists entirely of serpentinite derived from ultramafic cumulates. The main lithologies that acted as protoliths are wehrlite and dunite, with lesser clinopyroxenite. The protolith types of the serpentinite matrix are identical, in all aspects, to ultramafic cumulates found as coherent sequences at the bases of two large blocks comprising segments of an ultramafic to gabbroic plutonic complex (Units 4 and 5, see following section for description).

The macroscopic structure of the ultramafic-cumulate-derived serpentinite mélange is that of a huge duplex, sandwiched between the overlying thrust system of the mantle peridotite suite and the underlying thrust system of variably deformed and metamorphosed supracrustal rocks of the Bridge River complex in the east (Potter, 1983, 1986), and the thrust stack of the East Liza suite and Hurley Formation in the west. The belt reaches a structural thickness of approximately 1 kilometre in the eastern part of the area. A number of smaller, flat-roofed, hinterland-dipping duplex structures have been mapped within the belt (Figure 3-4-2). These duplexes are focused on shingled stacks of large and smaller blocks of the ultramafic-gabbroic plutonic complex which are situated on at least three different structural levels within the belt. The roof and floor thrust zones of the duplexes are outlined by gently to moderately north to northeast-dipping zones of intensely schistose scaly serpentinite (S_2), similar in style and orientation patterns to the second generation serpentinite shear zones observed in the overlying mantle-peridotite-derived mélange belt. Within the duplexes, the first generation serpentine schistosity generally dips more steeply to the north or northeast and curves into the duplex boundaries. This schistosity often wraps around the lozenge-shaped plutonic blocks contained in the duplexes, creating the appearance that originally much larger coherent sections of the plutonic complex were telescoped along shear zones injected by serpentinite. The S_2 serpentinite strands commonly display C-S mylonite fabrics, particularly in contact zones with larger blocks in the mélange. Shear-sense indicators invariably provide evidence for southwesterly thrusting. The volcanic and sedimentary blocks are all contained within S_2 serpentinite strands and are consistently aligned with their longest dimensions parallel to the S_2 fabric.

The overall attitude of the duplex structures is remarkably flatlying in the southwestern and southeastern parts of the belt but steepens to northerly and northeasterly dips of 40° to 50° at the contact with the overlying mantle peridotite thrust system (Figure 3-4-2). Most of this change in attitude seems focused on the main belt of plutonic blocks situated along this contact, and it appears as if these big blocks acted as a footwall ramp to the overlying thrust system. A similar ramp structure, outlined by flat and steep S_2 belts in the floor thrust zone of the mélange, is created by the underlying thrust stack of East Liza suite and Hurley Formation, exposed in the half window in the western part of the area (Figure 3-4-2). In the south, the main S_2 fabric dips gently to moderately to the south, creating a broad antiformal zone along the upper reaches of Jim Creek. This structure may mark the location of a blind culmination in the footwall of the mélange, which is possibly an easterly extension of the East Liza suite–Hurley Formation thrust stack.

BLOCKS OF ULTRAMAFIC-MAFIC PLUTONIC COMPLEXES

A number of large and small coherent blocks in the lower serpentinite belt represent dismembered sections of a plutonic complex (Plate 3-4-1). They include a large variety of intrusive rocks ranging from olivine-rich ultramafic cumulates to high-level, varitextured plagioclase-rich gabbros. They display complex multiple intrusive relationships often involving a number of igneous phases, and they exhibit, at

Plate 3-4-1. View to the northwest of the eastern edge of the main gabbro block in the mélange. Lower half of section in block comprises recessively weathering screens of massive and layered ultramafic cumulates between more resistant dikes. Upper half of section comprises screens of a variety of gabbro types between dikes. The contact between ultramafic and gabbroic rocks lies along the thick dike seen in centre of the photograph. The gabbroic to dioritic dike swarm dips moderately to steeply northeast and shows complex internal geometry (resistant dikes in centre of photograph).

least locally, evidence for heterogeneous high-temperature plastic deformation associated with intrusive events.

ULTRAMAFIC CUMULATES (UNIT 4)

The ultramafic cumulates are mainly found in two blocks, where they form the present structural base of the plutonic sequences (Figure 3-4-2). Their most important occurrence is in the large block in the centre of the belt. This sequence reaches a total thickness of at least 200 metres and extends along strike for at least 500 metres. It comprises dominantly wehrlite and clinopyroxenite, with subordinate chromitiferous dunite, clinopyroxene-bearing dunite, olivine clinopyroxenite and rare olivine websterite. Plagioclase may be present, in a highly altered state, in some of the pyroxenites. The suite was previously described as the clinopyroxenite unit by Leech (1953) and Nagel (1979). Both authors clearly underestimated the total average modal amount of olivine in the suite in favour of clinopyroxene.

The present base of the suite is poorly exposed, but appears to be in structural contact with underlying serpentinite with a moderately northeast-dipping S_2 fabric. In the eastern outcrop area of the block, the top of the unit is defined by a 20-metre-thick, little-deformed diorite dike with well-developed chilled margins. This dike is immediately overlain by thinly layered gabbros. Internally, the ultramafic suite contains a number of small, fault-bounded enclaves of layered gabbro. It is also cut by a narrowly spaced diabase dike swarm that dips on average 55° north-northeast. The ultramafic cumulates thus occur as narrow screens between the late dikes (Plate 3-4-1).

The plutonic geometry of the ultramafic suite in the main block is highly variable. The rocks are generally poorly layered to massive, and have isotropic texture with anhedral olivine shapes and randomly orientated stubby diopside prisms predominating. Where phase layering is observed, it is usually nonplanar and highly discontinuous over short distances. Layering attitudes are extremely variable from screen to screen in the dike swarm. It is not clear whether this is an original plutonic feature, or an imposed feature due to rotation caused by dike intrusion. The suite comprises poorly defined cyclic sequences of phase-graded units with a basal clinopyroxene-bearing dunite layer, grading into a thick wehrlite layer followed by a thick olivine clinopyroxenite layer. Olivine-rich phases often occur as irregular, pod-like bodies within clinopyroxene-rich phases. Poikilitic textures of clinopyroxene with olivine and chromite inclusions occur occasionally in wehrlitic phases. In most rocks clinopyroxene appears to be the main adcumulus phase.

GABBROIC SEQUENCES (UNIT 5)

The gabbroic sequences constitute the most voluminous component of the plutonic blocks in the mélange, as was already noted by Leech (1953) and Nagel (1979). In all blocks complex intrusive relationships between various gabbroic phases are indicated by crosscutting phase domains, xenolith-charged margins of late intrusive stocks, structurally controlled phase boundaries including faults and shear zones, fault-bounded ultramafic cumulate enclaves in gabbroic sequences, and late intrusive dike swarms (Plate 3-4-2). The scale of these features is generally small, on average 10 to 50 metres of outcrop width. They suggest that the gabbro complexes did not evolve through crystallization in large magma chambers of simple geometrical form, but rather through spatially and temporally highly variable multiple intrusive processes.

Compositionally, the gabbroic suite ranges from clinopyroxenite and rare websterite to two-pyroxene gabbros, clinopyroxene-rich gabbros and anorthosite. Clinopyroxene and plagioclase are the dominant constituents, whereas olivine and orthopyroxene are relatively rare. Undeformed gabbros range texturally from massive to layered, and isotropic to foliated. Medium to coarse-grained pegmatitic varieties, especially leucogabbros, are common in the form of small stocks and irregularly shaped pods and veins. High-level, varitextured gabbros are the dominant component in most of the blocks in the mélange belt. They constitute the top sections of the large plutonic blocks and form the only constituent of the remaining blocks. Fine scale compositional layering, occasionally with grain-size graded or phase-graded aspect (Plate 3-4-3), is common in a number

Plate 3-4-2. Typical appearance of high-level, varitextured gabbro. It comprises sheared and isotropic gabbro, crosscutting pegmatitic leucogabbro veins and, behind the hammer, a late isotropic diabase dike.

of localities. The layering displays highly variable attitudes, even within small areas of a single block. Domains of constant layer attitude are invariably bounded by crosscutting varitextured suites, by discrete dikes of variable width, or by shear zones and faults.

Domains of penetrative plastic deformation are rare in most of the blocks. Locally, layered and massive gabbros contain a well-developed schistosity (subparallel to layering, where present) and mineral elongation lineation. These rocks have a porphyroclastic texture of coarse clinopyroxene with recrystallized rims of green hornblende, suggesting that deformation occurred under amphibolite facies conditions. The deformed gabbros are commonly cut by pegmatitic leucogabbro bodies and a variety of gabbroic to dioritic dikes. Narrow, low-temperature shear zones, characterized by chlorite alteration, are common throughout the gabbro blocks and they are generally located along macroscopic phase contacts. The shear zones are truncated by younger intrusive rocks, indicating that they developed during the magmatic evolution of the gabbroic complexes.

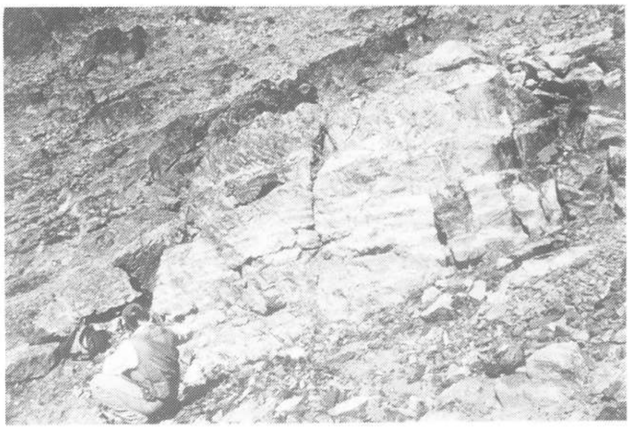


Plate 3-4-3. Fine-scale, phase-graded layering in gabbro-anorthosite sequence which forms an isolated plutonic unit within the eastern part of the main gabbro block. This gabbro sequence is in fault-controlled, intrusive contact with a large xenolith of ultramafic cumulates shown on the right-hand side of the photograph. The resistant dark-weathering unit above the layered gabbro is a crosscutting diorite dike with chilled margins against gabbro.

Blocks of Sedimentary and Volcanic Rocks (Unit 6 and 7)

The ultramafic-cumulate-derived serpentinite belt contains a number of blocks (approximately 30) of sedimentary, volcanic and volcaniclastic rocks in the study area. Such blocks have also been reported by Potter (1983) from the Hog Creek imbricate zone. Whereas the plutonic blocks in the serpentinite can reasonably be considered as indigenous to the belt, the blocks of supracrustal rocks represent a truly exotic element, justifying the use of the term mélange.

Most blocks are rather small in size, their longest outcrop dimension rarely exceeding 200 metres. Larger blocks tend to be tabular in shape with tapered edges. They are generally aligned parallel to the S_2 fabric in the surrounding serpentinites, with their longest dimension trending parallel to the regional strike of the belt. Smaller blocks, up to 10 metres in size, have a more rounded shape, and may lie inclined to the S_2 serpentinite fabric. The largest concentration of blocks is in a mylonitic serpentinite zone situated directly beneath the large plutonic blocks near the top of the belt, in the central and western part of the map area. Another conspicuous string of larger blocks lies lower in the belt in the easternmost part of the area, and appears to extend into the Hog Creek imbricate zone to the east.

The sedimentary blocks (Unit 7) comprise mainly bedded and massive chert, and thin to medium-bedded turbiditic siltstone and sandstone. In one block, bedded chert is interlayered with a unit of strongly silicified and mineralized volcanic rocks 10 metres thick. Small blocks of recrystallized limestone, limestone breccia, and cherty matrix-supported pebble conglomerate with abundant felsic igneous clasts are rare. One small block of coarse pyroclastic rock was found in Jim Creek near a block containing an upward-facing 20-metre sequence of pyritiferous laminated shale-siltstone, white bedded chert with shale partings, and massive greywacke with siltstone rip-up clasts. This sequence appears correlative with a more extensive unit of siliciclastic rocks with interbedded chert and rare volcaniclastic rocks which is exposed on the lower slopes west of

Plate 3-4-4. Detail of highly schistose, thinly layered quartz phyllite, cut by brecciated quartz diorite dike in sedimentary knocker within the serpentinite mélange in the easternmost part of the study area.

Jim Creek (Schiarizza *et al.*, 1989). The large block in the southeastern part of the map area contains an intensely foliated sequence of thinly layered pelites cut by small brecciated pods and dikes of hornblende plagioclase porphyry (Plate 3-4-4) that do not extend into the adjacent serpentinite matrix (*see* also Archibald *et al.*, 1989). Most blocks show effects of lower greenschist facies metamorphism associated with deformation in the form of cleavage development and, locally, mesoscopic cleavage folding.

Volcanic blocks (Unit 6) are less abundant than sedimentary blocks. They comprise massive and pillowed lava and pillow breccia. In some localities, the lavas show variolitic and/or vesicular texture; some contain feldspar phenocrysts and chlorite pseudomorphs presumably after primary pyroxene or amphibole. Pillow breccias locally contain lenses of chert and limestone breccia up to several metres in size. The volcanic rocks appear to range from basaltic to dacitic in composition. They are generally strongly altered due to silicification and low greenschist facies metamorphism, and show heterogeneous deformation in the form of flattening of pillows and cleavage development in the matrix of pillow breccias.

DIKES IN SERPENTINITE MÉLANGE

Numerous disrupted fragments of dikes occur within both types of serpentinite belts. They range in composition from gabbroic to dioritic; hornblende-porphyritic quartz diorite is an abundant component of the dike suite. Some large dikes in the eastern part of the area are multiple intrusive, ranging in composition from pyroxenite to gabbro and flow-banded, feldspar-porphyritic diorite (Plate 3-4-5). Many gabbroic dikes are strongly altered to either rodingite, greenschist or talc schist. On the other hand, many dioritic dikes are remarkably fresh and preserve well-developed chilled margins.

The dike fragments display a variety of shapes reflecting the degree of their deformation and related alteration. They range from small to large, rounded or lozenge-shaped boudins to rather straight and continuous dike segments, some of which extend up to 100 metres along strike. Boudins are completely surrounded by foliated serpentinite and are aligned parallel to either S_1 or S_2 as the main external fabric. They often occur in clusters with conspicuous parallel or en echelon alignment in the serpentinite fabric, reflecting boudinaged single dikes originally oriented parallel or oblique to the external fabric, respectively. Some clusters, however, are so dense that they must have resulted from boudinage of parallel dike swarms. Straight dike fragments of the dioritic suite generally preserve chilled margins against either the S_1 or S_2 fabric of the serpentinite matrix. Their contacts are, however, invariably sheared, and the fine-grained chill zones often show foliation development related to postintrusion deformation of the serpentinite matrix. The field relationships indicate that both pre-S_1 and post-S_2 dike suites are present in the mélange (*see* below). Gabbroic dikes define both early and late dike suites, whereas dioritic dikes (particularly those with preserved chilled margins) are almost all post-S_1. Dikes in the mantle-peridotite-derived serpentinite belt are relatively rare compared to the abundant occurrences in the ultramafic-cumulate-derived serpentinite

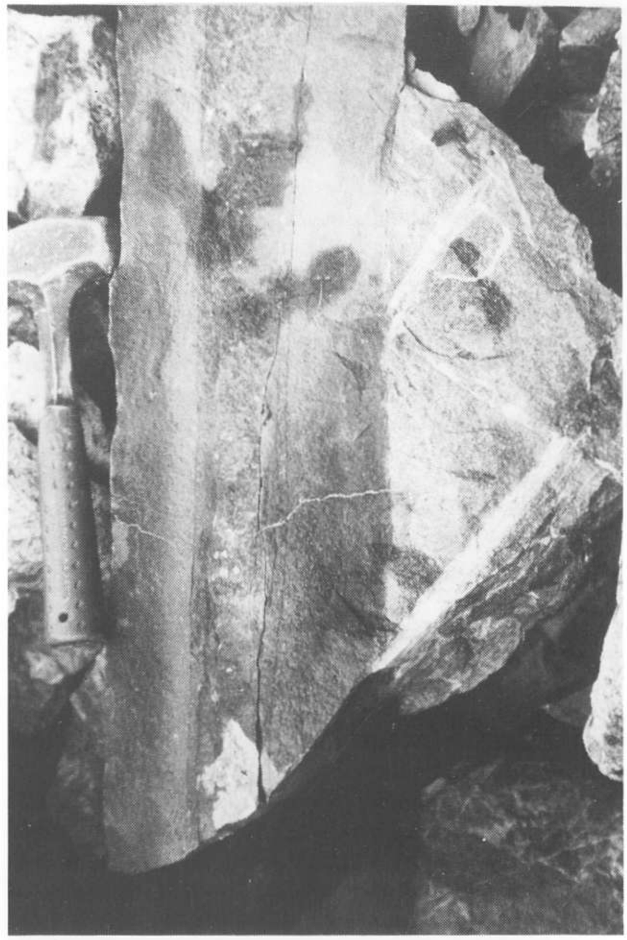


Plate 3-4-5. Detail of composite dike fragment in eastern part of the serpentinite mélange. Flow-banded quartz diorite dike, with feldspar phenocryst alignment, is intrusive into sheared and altered isotropic gabbro.

belt. They are dominantly part of the late dioritic suite intruded along S_2 shear zones near the upper and lower contacts of the belt.

Late dikes have locally imprinted contact metamorphic effects on the surrounding serpentinite. The most common contact metamorphic assemblages are olivine + serpentine $\pm$ talc, and olivine + talc $\pm$ magnesite. The rocks have a porphyroblastic texture of elongate olivine crystals in a white felted matrix. The porphyroblasts are often pseudomorphed by fine-grained, brown-weathering magnesite + talc aggregates. These rocks were first recognized by Leech (1953) and later studied in some detail by Nagel (1979), who correctly suggested that the olivine schists formed by prograde metamorphism of originally low-temperature serpentinite. He concluded that the olivine was generated at temperatures around 400°C, but ignored the effects of X_{CO_2}. This is probably erroneous in view of the common occurrence of magnesite in the assemblage; high X_{CO_2} would slightly lower the temperature range for the stability of olivine in the contact aureoles. Nagel inferred a magmatic source for the heat that caused the static metamorphism, but did not link the heat source directly to the abundant dike suites in the mélange. The present study has shown that some contact aureoles are still attached to the dike walls, whereas others have been detached from the dikes by later shearing that led to the development of the S_2 serpentine schistosity. The latest widespread dike intrusion event recorded in the mélange is thus interpreted to be syn-S_2.

East Liza Igneous Suite (Units 8 and 9)

The suite is treated as a separate lithotectonic unit, in contrast to the interpretation of Nagel (1979) who considered it to be a block in the serpentinite. It is overlain with clear thrust contact by ultramafic-cumulate-derived serpentinites along its eastern and southern margins (Plate 3-4-6). It structurally overlies sedimentary rocks of the Cadwallader Group, and nowhere in the study area can it be shown to be underlain by serpentinite. The floor thrust of the serpentinite mélange is thus drawn at the top of the unit and defines a broad easterly plunging antiform that extends eastward into the Hog Creek imbricate zone.

The unit comprises mafic to intermediate intrusive and extrusive rocks displaying complicated igneous relationships. Leech (1953) and Nagel (1979) have suggested that a transitional contact exists between gabbros and volcanic rocks. Gabbroic rocks underlie the southern part of the outcrop area and appear to be overlain with nonconformable stratigraphic contact by pillowed lavas in the north. The contact zone is poorly exposed, but local field relationships suggest that it dips gently north-northeast.

The intrusive sequence (Unit 8) consists mainly of fine-scale layered two-pyroxene gabbros with minor interlayered websterite, clinopyroxenite and anorthosite (Plate 3-4-6). These rocks show a well-developed tectonic foliation subparallel to layering, as well as discrete plastic shear zones overprinting the foliation. Layering attitudes are highly variable, as was also noted by Nagel (1979). The rocks have a porphyroclastic texture, outlined by flattened pyroxene grains with tailed recrystallized margins of brown and green pleochroic hornblende indicating deformation under amphibolite facies conditions. The deformed gabbros are cut by small, irregularly shaped stocks of isotropic, fine-grained to pegmatitic gabbros; by variably textured gabbroic veins; and by abundant fine-grained gabbroic to dioritic dikes which have highly variable orientation. This intrusive sequence resembles the high-level gabbros of the plutonic blocks in many respects.

The contact zone with the volcanic rocks is characterized by an increase in the occurrence of dikes, by frequent microgabbroic stocks, and narrow screens of intensely sheared pillowed and massive lavas between intrusive phases. Locally, dike swarms appear to have coalesced into small sheeted dike sections, but a sheeted dike complex is certainly not well developed along the contact zone. Small plugs and sills of microgabbro are found locally, higher within the volcanic succession.

The volcanic rocks (Unit 9) comprise mainly pillow lava with subordinate massive flows and pillow breccia; they are cut by fine-grained diabase dikes. In mildly deformed parts of the sequence, the pillows are small in size, ranging up to 0.5 metre in diameter, and rounded in shape (Plate 3-4-7). They are fined grained to aphanitic and locally vesicular and porphyritic; the original ferromagnesian phenocrysts now occur as chloritic pseudomorphs. Massive flows are up to 2 metres thick and locally show banding outlined by concentrations of amygdules. The rocks are strongly altered to low greenschist facies assemblages with abundant quartz, epidote and chlorite; silicification is intense and widespread. Compositionally the lavas appear to range from basalt to dacite.

The thrust contact between the volcanic rocks and underlying sedimentary rocks of the Hurly Formation is well exposed along the upper eastern slopes of East Liza Creek. In the lavas it is a zone of silicic banded mylonite and phyllonite with well-developed C-S fabrics up to 1 metre thick. At one locality, the thrust is clearly cut by a diorite dike that can be traced over some distance into the underlying sedimentary rocks. Various shear-sense criteria in the mylonites consistently indicate southwest-directed thrusting, identical to the movement pattern of the S_2 serpentinite mylonites which directly overlie the volcanic rocks in the northern part of outcrop area of the East Liza suite. Deformation associated with this thrusting has not been recognized with certainty

Plate 3-4-6. Detail of thrust contact between basal mylonite of serpentinite mélange and underlying gabbro, with steep fine-scale layering, of East Liza igneous suite. Composite fabric in serpentinite consists of an early steeply dipping schistosity (S_1) and a later, mylonitic serpentinite fabric (S_2) which is subparallel to the thrust contact.

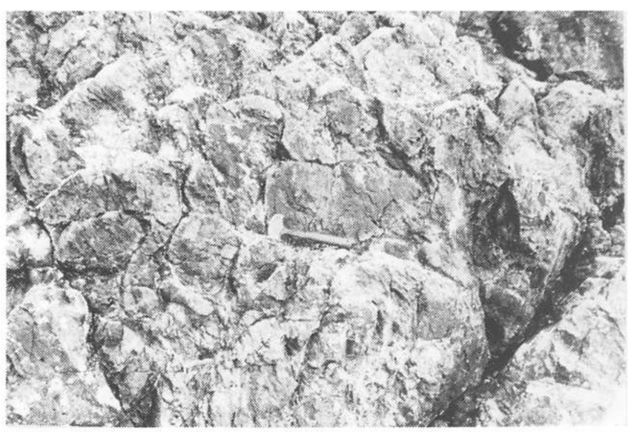


Plate 3-4-7. Mildly deformed pillow lavas of East Liza igneous suite in western part of study area.

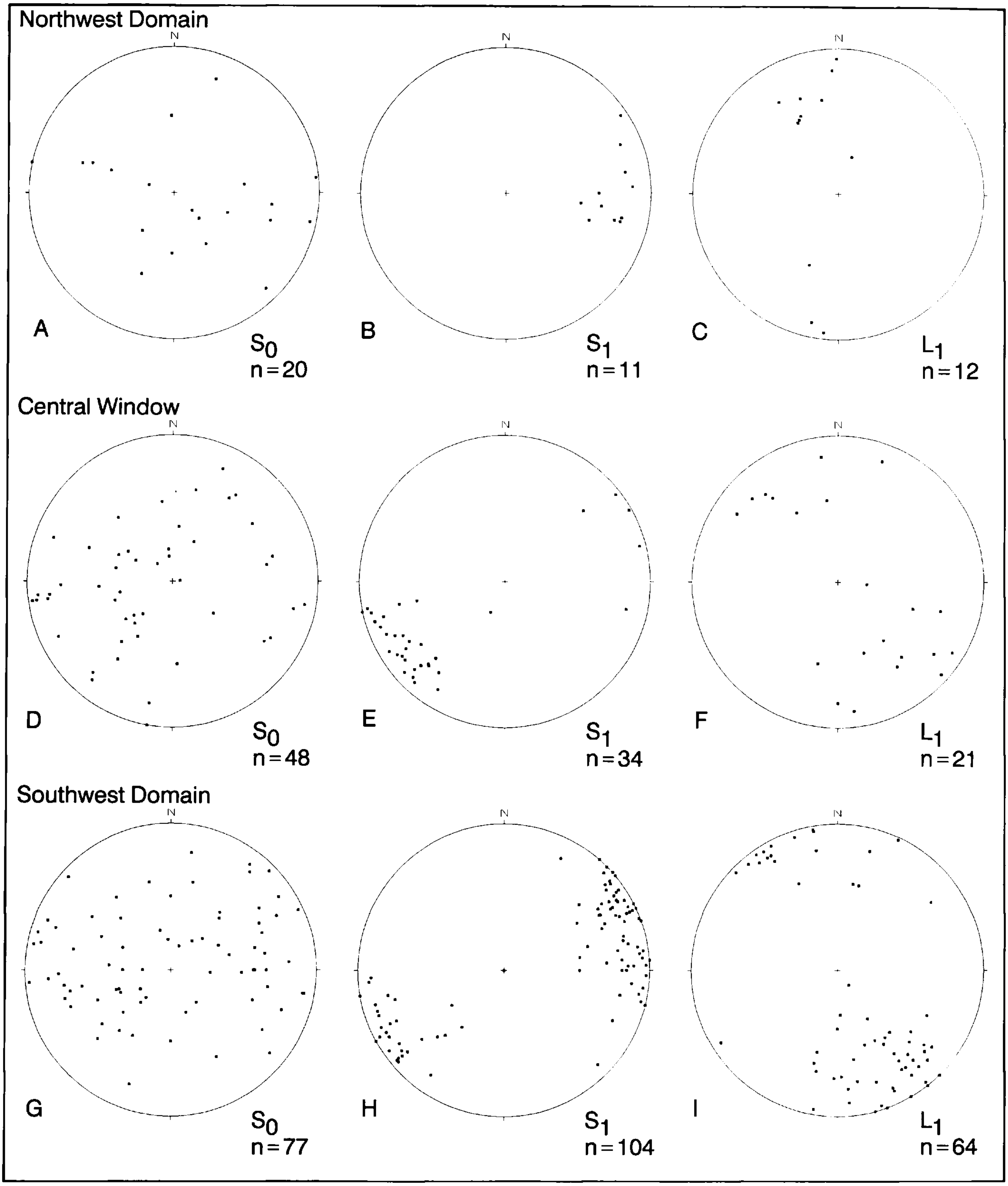


Figure 3-4-3. Orientation patterns of cleavage fold systems in three domains of Hurley Formation: (A-C) northwestern domain, (D-F) northern half-window, (G-I) southwestern domain. A, D, G are plots for poles to bedding: B, E, H are plots for poles to axial planar cleavage; C, F, I are plots of bedding/cleavage intersection lineations and axes of small-scale folds; all plots are in equal area, lower hemisphere projection.

within the volcanic sequence. The presumed thrust contact between gabbros and Hurley Formation is nowhere exposed in the study area. The contact is deformed in macroscopic cleavage folds with associated thrusts. The folding is spectacularly developed in the Hurley footwall (*see* below). It caused widespread cleavage formation with associated flattening of pillows in the volcanic sequence. This cleavage dips steeply to the southwest and obscures many of the original features of the volcanic sequence; it is only locally developed in the gabbros to the south. All structures are cut by a complex extensional fault system that controls much of the present outcrop pattern of the eastern contact of the East Liza suite with the serpentinite mélange (*see* also below).

CADWALLADER GROUP, HURLEY FORMATION (UNIT 10)

This unit comprises a variety of siliciclastic and calcareous sedimentary rocks which, on the basis of lithological correlation, are assigned to the Late Triassic Hurley Formation (Rusmore, 1987). To date, however, no biostratigraphic data are available for the unit in the map area. The most prominent sequence in the unit consists of thin to medium-bedded grey sandstones and laminated grey to black siltstones, which are turbiditic in nature displaying grading as well as convolute and crosslamination. It contains interbedded limestone, chert and pebble conglomerate, which become more abundant towards the stratigraphic base of the unit. Massive, locally fossiliferous limestone and limestone breccia occur as discontinuous lenses up to 5 metres thick. White to grey, massive chert beds intercalated with the siliciclastic rocks range from 0.5 to 3 metres in thickness. Matrix-supported pebble conglomerates contain subrounded clasts of felsic and mafic (sub-)volcanic rocks. The turbidite sequence becomes more calcareous towards its stratigraphic top (Plate 3-4-8), consisting of medium to thick-bedded (up to 1 metre), graded calcarenites, calcareous shales and rare, thin discontinuous limestone beds. The unit is cut by rare, thin dikes ranging in composition from basalt to quartz diorite.

All rocks of the unit are affected by regional folding associated with intense axial planar cleavage development (Plate 3-4-8). The fold system comprises several orders of folds, ranging from small crenulations to macroscopic antiform-synform pairs. Orientation patterns of the fold system are presented in Figure 3-4-3 for three domains in the study area. Folds in the northwestern domain, directly west of the northernmost gabbro blocks in the serpentinite mélange (Figure 3-4-2), define a plane, subcylindrical, steeply inclined system that plunges gently to the north; the folds are close to tight, asymmetric with easterly vergence. The folds in the northern half-window of Hurley Formation (Figure 3-4-2) define a plane, noncylindrical, close to tight system for which fold asymmetry is not well established. The folds are steeply northeasterly inclined and markedly doubly plunging (Figure 3-4-3). Due to severe late extensional faulting, it is not clear whether or not this folding affects the thrust contacts with the overlying serpentinites and volcanic rocks. A penetrative cleavage with identical orientation is observed in the volcanic rocks, but does not appear to affect the S_2 serpentinite mylonites. The fold system in the southwestern domain, along the eastern slope of East Liza Creek, is nonplanar noncylindrical, close to tight; it is steeply inclined to the southwest and plunges mainly to the southeast (Figure 3-4-3). It is markedly asymmetric with east-northeasterly vergence expressed by steeply west-dipping overturned short limbs and associated moderately west-dipping thrusts. the fold and thrust system clearly involves not only the Hurley Formation, but also the overlying East Liza volcanic rocks and S_2 serpentinite mylonites of the mélange. In one small area in the northern part of the domain (Figure 3-4-2), the volcanic rocks and serpentinites define an overturned antiform-synform pair that is cut on its western side by an easterly verging thrust system comprising thin sheets of volcanic and sedimentary rocks. The thrusts are outlined by thin phyllitic C-S mylonite zones consistently showing north-northeast-directed movement. The fold and thrust system effectively terminates against a late, normal fault in the east and is cut by numerous small extensional faults which are not shown on the map.

Plate 3-4-8. Limb domain of mesoscopic cleavage fold in thin-bedded calcareous shales of Hurley Formation, Cadwallader Group. This fold is a parasitic structure in the core of a southeast-plunging antiform. Note angular relationship between bedding and axial planar cleavage.

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GEOLOGY AND NOBLE METAL GEOCHEMISTRY OF THE POLARIS ULTRAMAFIC COMPLEX, NORTH-CENTRAL BRITISH COLUMBIA* (94C/5, 12)

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KEYWORDS: Economic geology, Alaskan-type ultramafic complex, Polaris, structure, geochemistry, platinum group elements.

INTRODUCTION

The Polaris ultramafic complex is one of the largest Alaskan-type bodies in British Columbia, ranked second only to the Tulameen complex in the southern part of the province. Fieldwork and geochemical sampling at Polaris were conducted from fly camps during the last three weeks of August 1988, and preliminary results were released in Open File 1989-17 (Nixon *et al.*, 1989a). The complex was revisited for one week in August 1989 in order to conduct more detailed mapping of intrusive relationships and collect additional geochemical samples. This report summarizes the results of the fieldwork and presents a compilation of over 200 previously published (Nixon *et al.*, 1989a) and new analyses for platinum group elements and gold, as well as "pathfinder" elements such as nickel, chromium, arsenic, antimony and sulphur.

The project area is covered at a scale of 1:250 000 by the Mesilinka map sheet (94C) and 1:50 000 base maps (94C/5 and 12). Aeromagnetic survey maps are also available in the smaller (Map 7777G-Fort Grahame) and larger (Maps 9074G and 9075G) scales respectively.

LOCATION AND ACCESS

The Polaris ultramafic complex (56°30'N, 125°40'W) is situated 10 kilometres northeast of Aiken Lake in the Omineca Mountains (Figure 3-5-1). Access is by dirt road stretching some 335 kilometres north from Fort St. James via Manson Creek and Germansen Landing, to reach Aiken Lake via a well-maintained gravel road leading to the Cheni mine in the Toodogonne River area. Alternatively, the area may be reached via scheduled flights from Smithers to the Sturdee airstrip situated approximately 130 kilometres northwest of Aiken Lake, and from there by helicopter. The complex underlies an area of approximately 45 square kilometres at the southern end of the Lay Range and is well exposed above treeline at altitudes between 1600 and 2200 metres. It takes its name from Polaris Creek, a tributary of Lay Creek, both of which drain the western margin of the ultramafic body and flow southwards into the Mesilinka River.

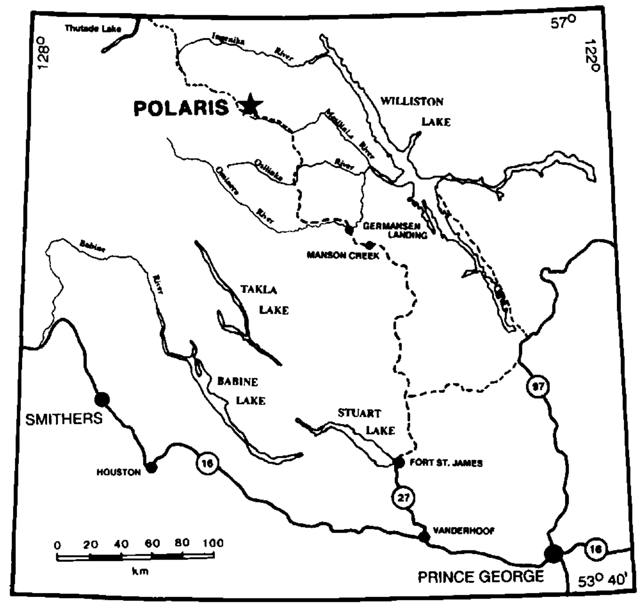

Figure 3-5-1. Location map of the Polaris ultramafic complex.

PREVIOUS WORK

The first systematic geologic mapping of the Aiken Lake area was completed by Armstrong (1946), Armstrong and Roots (1948), and Roots (1954). Earlier, Lay (1932) had examined many of the mineral prospects in the region.

The first detailed observations of the mafic and ultramafic rocks of the Polaris complex were made by Roots (1954). Modern petrologic studies and more detailed mapping of a host of Alaskan-type complexes were made later by Irvine (1974a, 1976). Prior to this paper, Irvine (1974a) and Foster (1974) presented the most complete descriptions of the Polaris complex. A revised edition of the geologic map previously released by Nixon *et al.* (1989a) is currently being prepared for publication (Nixon *et al.*, 1990).

GEOLOGIC SETTING

The Polaris ultramafic complex lies within the Omineca crystalline belt, a morphogeologic division of the Cordillera that straddles the boundary between ancestral North America and a collage of allochthonous tectonostratigraphic terranes (Quesnellia, Stikinia, Slide Mountain and Cache Creek) that

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

amalgamated to form the Intermontane Superterrane before being accreted to the margin of the North American craton in the Mesozoic (Wheeler and McFeely, 1987). The Polaris complex is the largest of a number of Alaskan-type bodies that intrude Quesnellia (Figure 3-5-2). At this latitude, the Quesnel terrane is bounded on the west by Stikinia along the line of the Pinchi-Ingenika dextral fault system (Gabrielse, 1985; Wheeler *et al.*, 1988). Its eastern boundary is marked by the Swannell fault which places Upper Proterozoic rocks of the Ingenika Group, part of the pericratonic Kootenay Terrane (Wheeler *et al.*, 1988), in thrust contact with Quesnellia (Bellefontaine, 1989).

Arc-related augite-phyric flows, pyroclastic and epiclastic rocks of the Upper Triassic Takla Group characterize the Quesnel Terrane but are also found farther west in Stikinia (Figure 3-5-2; Richards, 1976a, b; Monger, 1977; Monger and Church, 1977). With the exception of the Menard complex (Nixon *et al.*, 1989b), all of the Alaskan-type bodies lie within Quesnellia, as presently defined, and all have been considered as comagmatic and coeval with Upper Triassic volcanism (Irvine, 1976). West of the Ingenika-Pinchi fault system, rocks of the Takla Group are metamorphosed to the zeolite or prehnite-pumpellyite facies whereas correlative lithologies to the east exhibit greenschist-grade assemblages (Richards, 1976b; Monger, 1977).

Mafic and ultramafic rocks of the Polaris complex are hosted by the Lay Range assemblage, a structurally complex sequence of arc-derived, predominantly clastic and volcaniclastic rocks that have been metamorphosed to greenschist or lower amphibolite grade (Monger, 1973, 1977; Richards, 1976a, b; Irvine, 1974a). These strata have been tentatively correlated with the Upper Devonian to Upper Permian Harper Ranch Group that forms the basement of Quesnellia in southern British Columbia (Monger *et al.*, in press).

The age of the Polaris complex is not well established. Potassium-argon dates on biotite and hornblende in a peridotite have yielded Jurassic isotopic ages of 167 ± 9 (2σ) and 156 ± 15 Ma (Wanless *et al.*, 1968). These dates have been considered too young by most workers who have stressed the spatial and petrological associations of Alaskan-type ultramafic complexes in British Columbia with Upper Triassic Takla-Nicola-Stuhini volcanic rocks of Quesnellia and Stikinia (*e.g.* Monger, 1973; Irvine, 1974a, 1976; Woodsworth *et al.*, in press). Uranium-lead dating is currently in progress in an attempt to firmly establish the age of the complex.

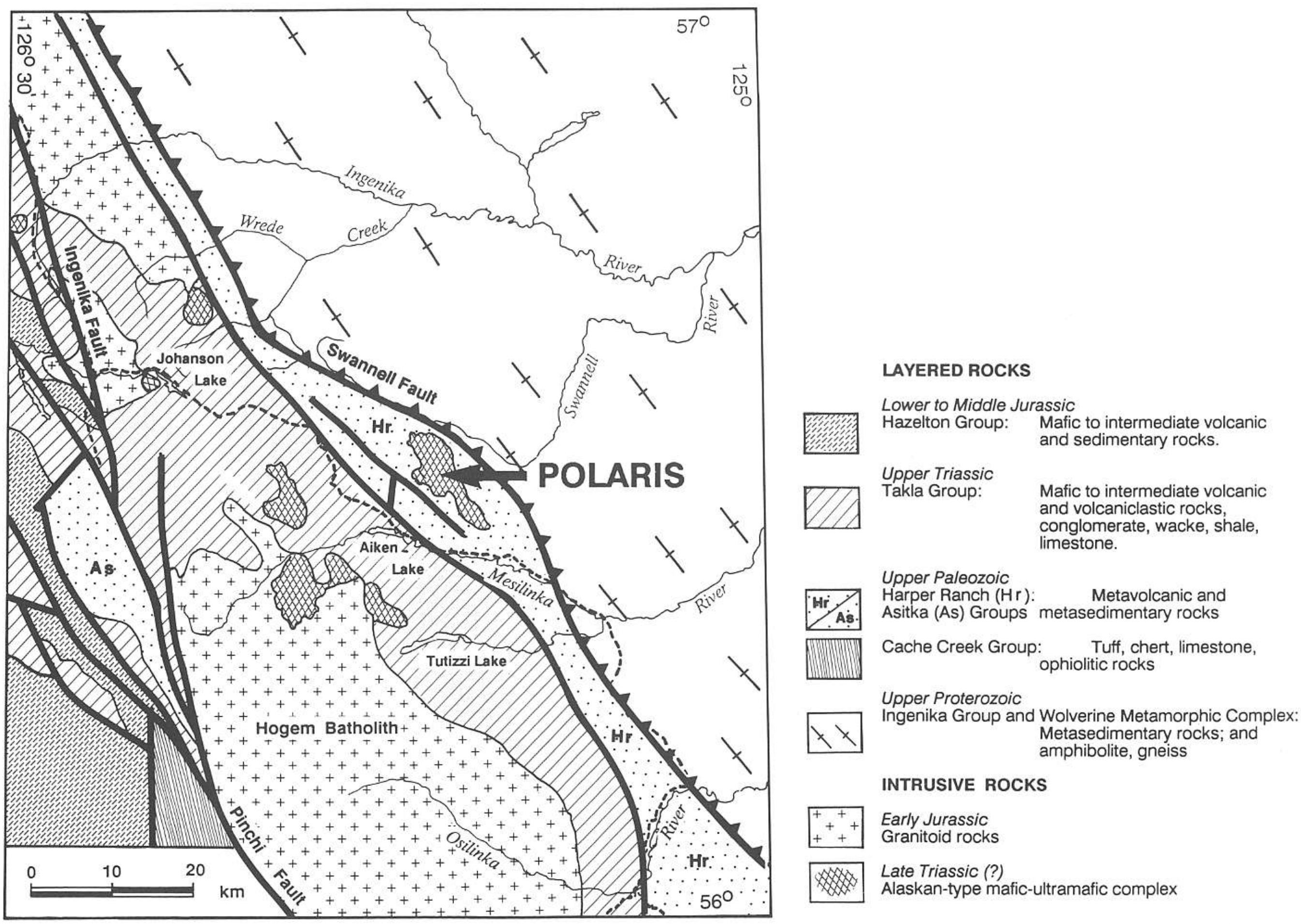

Figure 3-5-2. Geologic setting of the Polaris ultramafic complex (modified after Irvine, 1974; Monger, 1977; and Richards, 1976b).

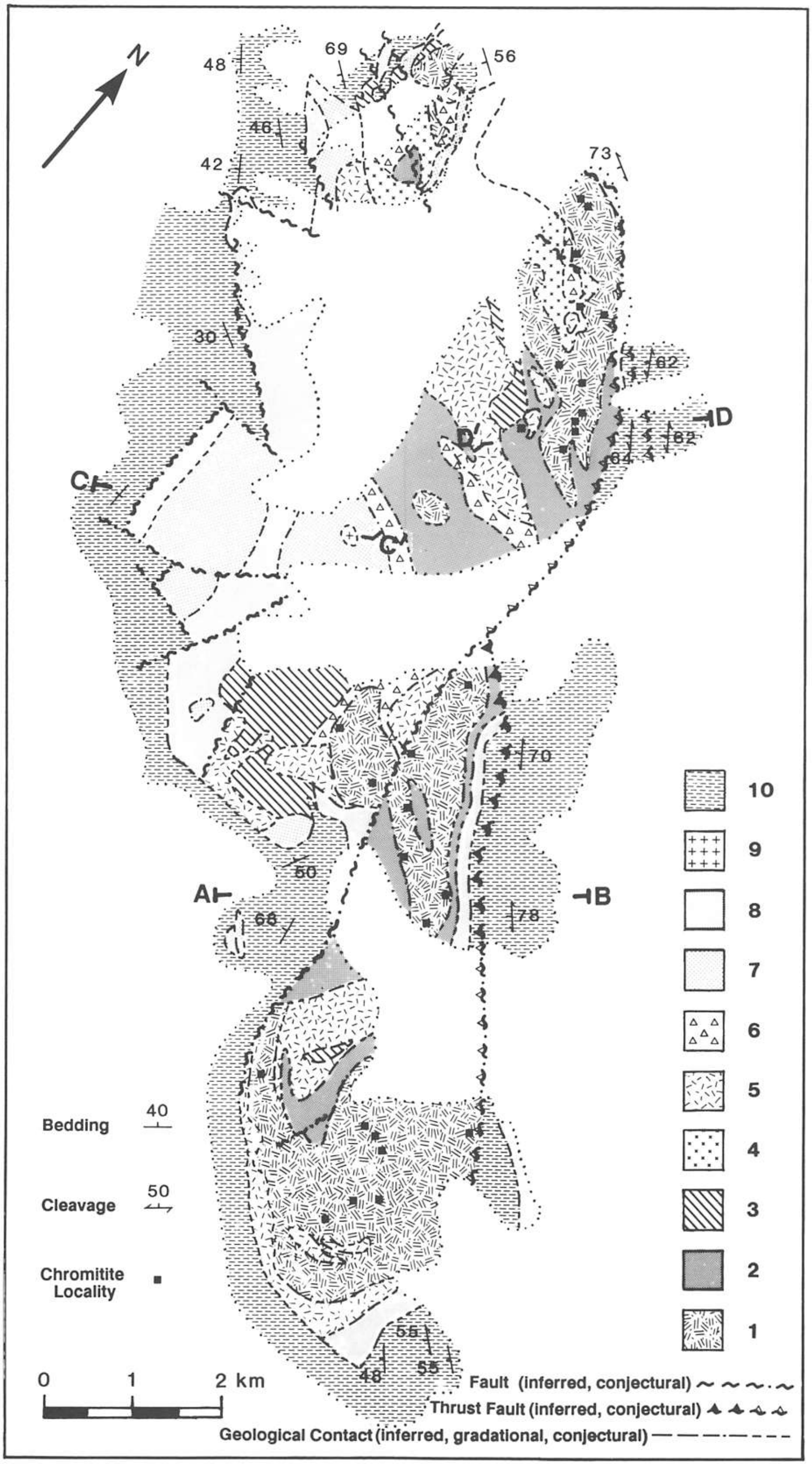

Figure 3-5-3. Generalized geology of the Polaris ultramafic complex (modified after Nixon *et al.*, 1989a). Key to map units: 1, dunite; 2, olivine wehrlite; 3, wehrlite; 4, undifferentiated olivine wehrlite and wehrlite; 5, olivine clinopyroxenite and clinopyroxenite; 6, mixed clinopyroxenitic and wehrlitic (with minor dunite) unit; 7, hornblende clinopyroxenite, clinopyroxene hornblendite, hornblendite and minor gabbro; 8, gabbroic rocks; 9, syenite/leucomonzonite; 10, metasedimentary and metavolcanic rocks of the Lay Range assemblage (Harper Ranch tectonostratigraphic terrane). Crosses indicate chromitite localities. A-B and C-C′-D′-D indicate location of cross-sections in Figure 3-5-4.

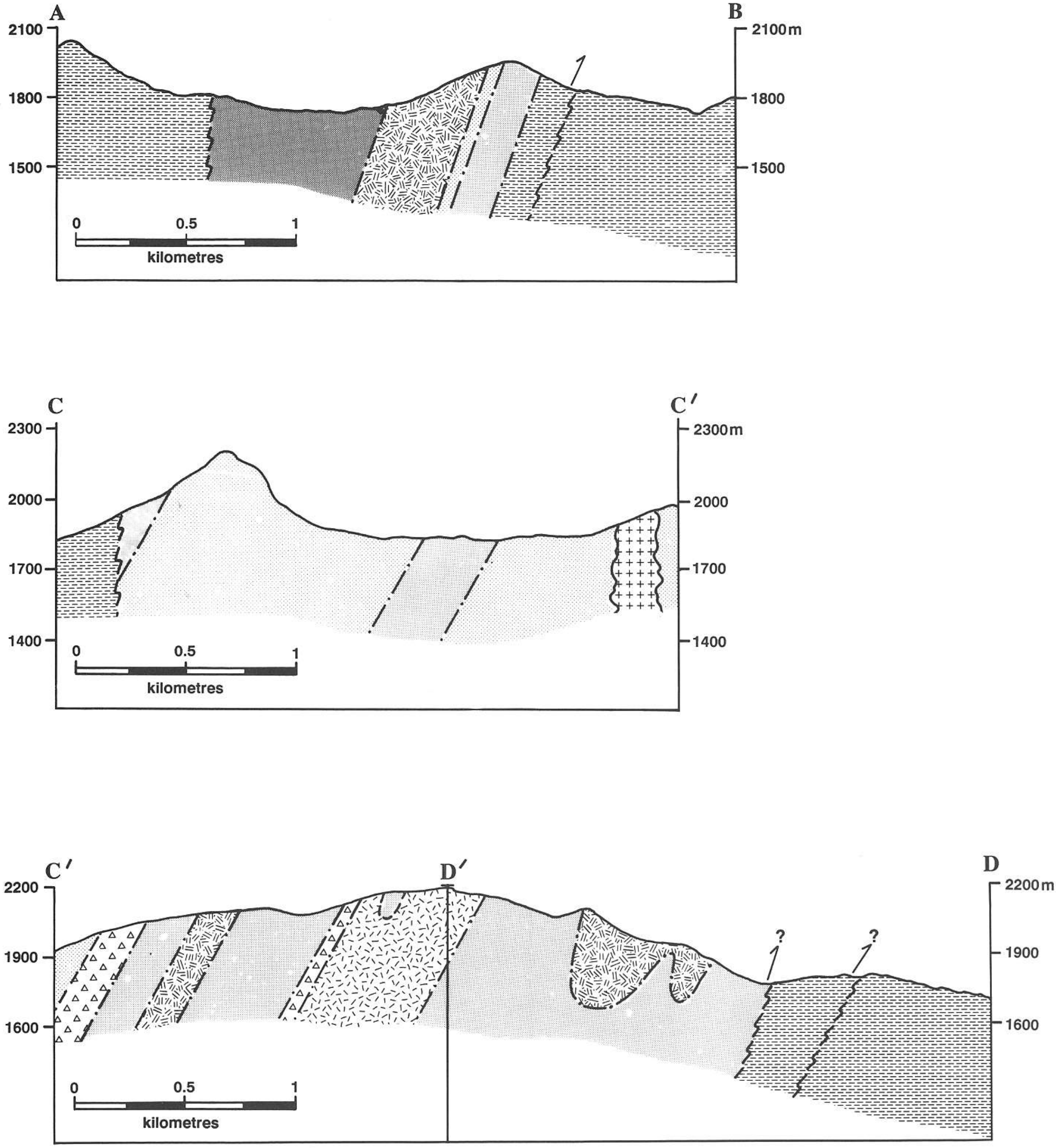

Figure 3-5-4. Schematic geologic cross-sections of the Polaris complex. *See* Figure 3-5-3 for locations. Note addition of hornblende clinopyroxenite unit in cross-section A-B.

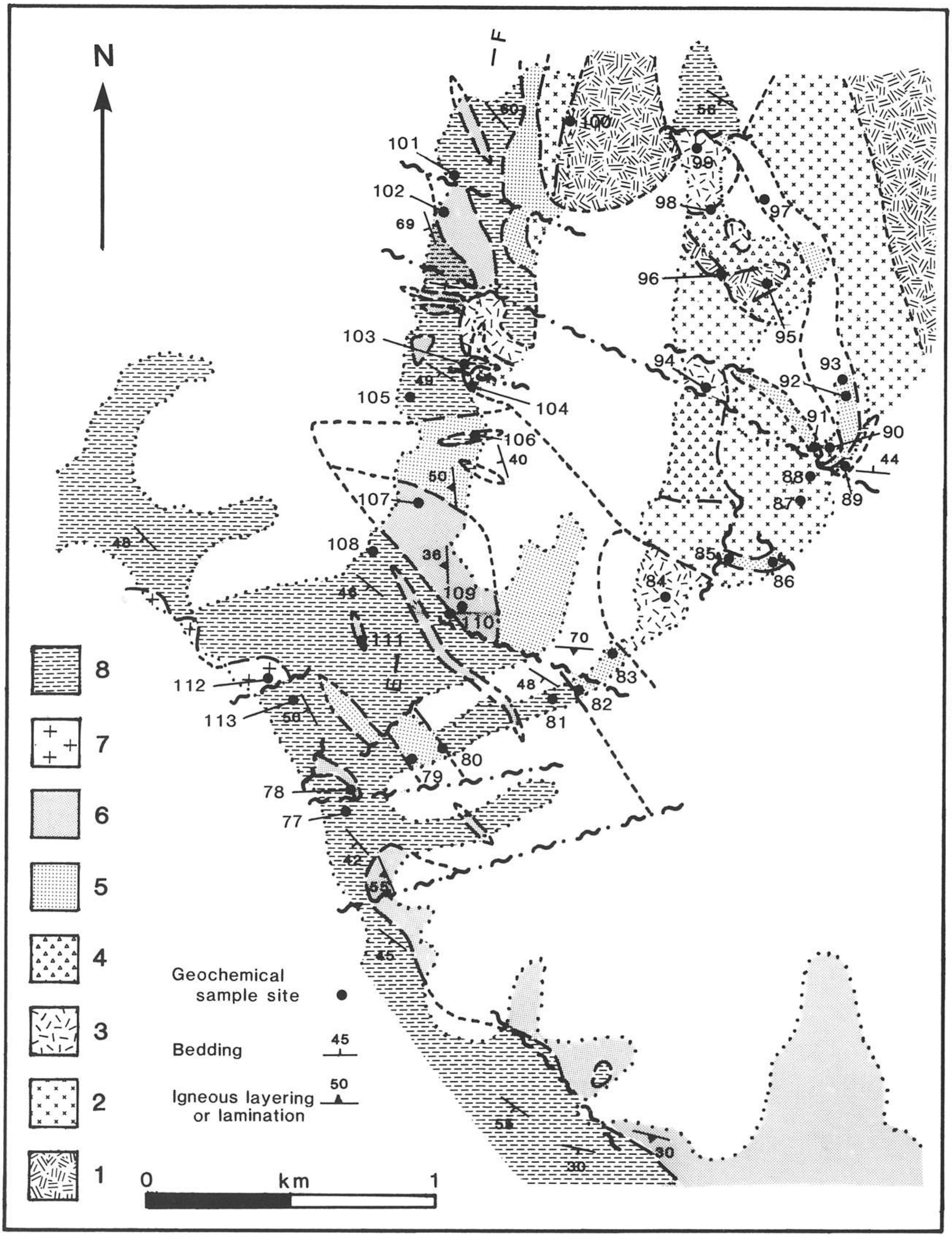

Figure 3-5-5. Generalized geology of the northwestern terminus of the Polaris ultramafic complex showing intrusive relationships and geochemical sample sites. Key to map units: 1, dunite, 2, undifferentiated olivine wehrlite and wehrlite; 3, olivine clinopyroxenite to clinopyroxenite; 4, mixed clinopyroxenitic and wehrlitic (with minor dunite) unit; 5, hornblende clinopyroxenite, clinopyroxene hornblendite, hornblendite and minor gabbro; 6, gabbroic rocks; 7, hornblende diorite; 8, metasedimentary and metavolcanic rocks of the Lay Range assemblage (Harper Ranch tectonostratigraphic terrane). E-F locates cross-section in Figure 3-5-6. Other symbols as in Figure 3-5-3.

COUNTRY ROCKS: LAY RANGE ASSEMBLAGE

Roots (1954) noted that rocks which host the Polaris complex form a moderately dipping (40°–50°) homoclinal sequence facing and inclined westwards. Monger (1973) recognized that the Lay Range assemblage is entirely fault-bounded and internally composed of a series of northwest-trending fault slices. Carbonates at the crest of the Lay Range northwest of the ultramafite have yielded mid-Pennsylvanian fossils, although younger Paleozoic or lowermost Mesozoic rocks may well exist (Monger, 1973; Monger and Paterson, 1974).

The immediate hostrocks of the Polaris complex comprise thickly to thinly bedded argillites, siltstones, sandstones, and minor carbonates, lithic-crystal tuffs and massive lava flows (Figures 3-5-3 to 3-5-6). Grey-green thinly bedded or laminated volcanogenic siltstones and fine-grained wackes are well exposed along the western margin of the complex where they form some of the highest peaks. The strata dip consistently to the west and are extensively silicified. Sedimentary features such as graded bedding, channel scours, load and flame structures, and crosslaminations indicate tops to the west. Locally, the succession contains chocolate-brown mudstones and thin (less than 0.3 metre thick), lenticular, impure carbonates characterized by brown-weathering rinds. At one locality (77, Figure 3-5-5), a thin (0 to 4 metres thick) medium grey layer of crystal-rich, non-welded ash-flow tuff contains rip-up clasts (up to 12 centimetres in length) of grey-green siltstone that are weakly imbricated and preferentially oriented within the direction of flow. The top of the deposit is reworked and the basal part exhibits normal grading and has scoured underlying siltstones. Thus, it appears to have been emplaced in a subaqueous environment.

Country rocks exposed at the northeastern margin of the complex include black fissile argillite and phyllite with interbedded fine-grained lithic tuff and pale grey carbonate. To the south, massive, aphanitic to porphyritic lavas of mafic to intermediate composition lie in sheared contact with serpentinized dunite at the southeastern margin of the complex. These lavas locally enclose concentrations of cognate xenoliths (up to 8 centimetres across) of hornblende gabbro, hornblendite and feldspathic hornblendite in a hornblende-phyric host. These rocks appear to have hornblende-phyric andesitic counterparts occurring as dikes in Lay Range assemblage strata overlying the intrusion.

POLARIS ULTRAMAFIC COMPLEX

The Polaris ultramafic complex forms an elongate body 14 kilometres long by 4 kilometres across at its broadest point. Its northwesterly trending long axis is conformable with the regional structural grain. The northern and southern terminations of the body are largely obscured by glacial drift. Farther north in the Lay Range, thin ultramafic sill-like intrusions, presumably coeval with the Polaris complex, were mapped by Roots (1954). However, the complex does not appear to have a large subsurface extension, judging from its distinctive aeromagnetic anomaly.

All of the lithologies that characterize Alaskan-type complexes are well represented in the Polaris complex (Figures 3-5-3 to 3-5-6). These include dunite, olivine wehrlite and wehrlite, olivine clinopyroxenite and clinopyroxenite, hornblende clinopyroxenite, hornblendite, gabbroic rocks, and late-stage pegmatites and finer grained feldspathic phases. A somewhat distinctive mineralogic feature is the appearance of phlogopitic mica in early cumulates, including dunite. Ultramafic lithologies are well exposed in the eastern and southern parts of the complex; gabbroic and hornblende-bearing rocks in the west.

ULTRAMAFIC ROCKS

DUNITE

The main mass of dunite forms northwest-trending ridges in the eastern half of the complex, and the floor and walls of a large cirque in the south. Dunite weathers tan to pale yellowish brown and forms smooth, blocky outcrops that typically lack a penetrative fabric. Joint planes are commonly lined with pale green to black serpentine that locally appears asbestiform. Fresh surfaces vary from dark greenish grey to black as the degree of serpentinization increases. Complete serpentinization, however, only occurs close to fault zones and, on the whole, olivines are well preserved. In thin section, the dunite is generally medium grained and composed of weakly serpentinized olivines (less than 3 milli-

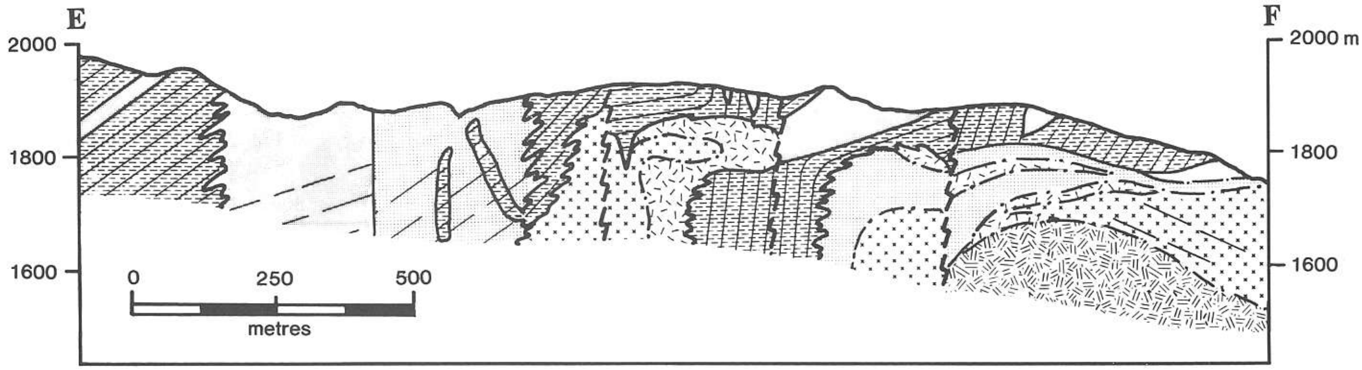

Figure 3-5-6. Schematic geologic cross-section of intrusive relationships at the northwestern end of the Polaris complex. *See* Figure 3-5-5 for location and key to map units. Attitude of bedding shown in host rocks; localized igneous layering or lamination depicted by dashed lines.

metres) and minor chromite (1 per cent by volume) that exhibit cumulate textures, accompanied by rare cumulus and intercumulus phlogopite.

CHROMITITE

Concentrations of chromite are confined to the dunite except for minor occurrences in olivine wehrlite adjacent to dunite (Figure 3-5-3). Chromitites occur as irregular pods, centimetre-scale schlieren, and millimetre-thick, planar to curved laminae. Schlieren commonly range from 6 to 15 centimetres in length and 0.5 to 4 centimetres in width; laminae can rarely be traced for more than 0.5 metre. Roots (1954) found more extensive chromitite horizons measuring almost 4 metres in length by 12 centimetres in width. Locally, coherent angular blocks of laminated chromitite up to 30 centimetres across are found juxtaposed in random orientation (Plate 3-5-1). In thin section, aggregates of chromite crystals commonly form networks or ring-like structures that partially to completely enclose cumulus olivines. Similar textures were documented by Clark (1978) from another Alaskan-type intrusion, the Turnagain River complex in northern British Columbia. The irregular geometry of podiform chromitites, pinch-and-swell nature of schlieren, and random orientation of layered chromitites are due to remobilization of previously deposited chromite-rich cumulates early in the crystallization history of the intrusion.

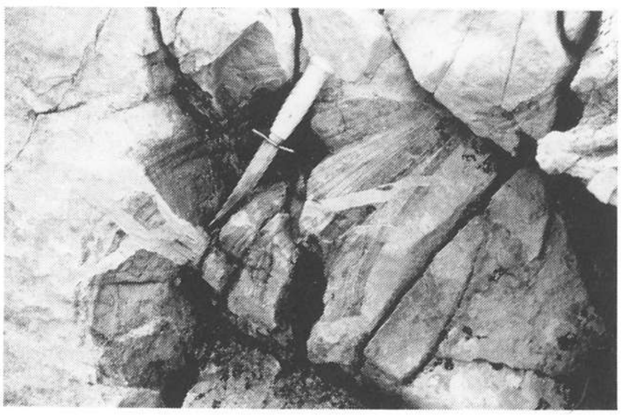

Plate 3-5-1. Disrupted block of layered chromitite in dunite cut by thin dunite dikes.

OLIVINE WEHRLITE AND WEHRLITE

Extensive outcrops of wehrlitic rocks are found in the northern and central parts of the complex and prompted subdivision into mappable units of olivine wehrlite (90 to 65 per cent olivine, 10 to 35 per cent clinopyroxene) and wehrlite (65 to 40 per cent olivine, 35 to 60 per cent clinopyroxene; nomenclature from Nixon, 1990). Wehrlitic lithologies weather pale brown to medium reddish brown and are massive to well jointed and weakly serpentinized; fresh surfaces are grey-green. These rocks commonly exhibit a knobby texture due to recessive weathering of olivine relative to clinopyroxene. In places, anhedral to subhedral megacrystic clinopyroxenes (up to 8 centimetres across) with well-developed poikilitic textures impart a distinctive lustre mottling to the outcrop (Plate 3-5-2). Olivines typically occur as cumulus crystals whereas clinopyroxenes exhibit intercumulus and cumulus textures. These primary silicates are accompanied by accessory chrome spinel and trace amounts of phlogopite. Rarely, wehrlitic mineralogy has formed by infiltration of clinopyroxene-rich magmas into an olivine-rich host, resulting in the formation of anastomizing three-dimensional networks of coarse-grained clinopyroxenite in the hostrock (Plate 3-5-3).

Plate 3-5-2. Megacrystic wehrlite exhibiting lustre mottling caused by subhedral poikilitic clinopyroxene. Magnet is 11 centimetres long. Photo by C. Nuttall.

Plate 3-5-3. Wehrlite formed by net veining of dunite near contact of clinopyroxenite and dunite bodies.

OLIVINE CLINOPYROXENITE AND CLINOPYROXENITE

Outcrops of olivine clinopyroxenite and clinopyroxenite are widely distributed throughout the ultramafic portion of the intrusion. Weathered surfaces are pale green to pale greyish green where enriched in clinopyroxene; olivine-rich areas appear rusty brown. The rocks are usually medium to coarse grained (3 to 10 millimetres) and grey-green on fresh surfaces. In pegmatitic zones, clinopyroxenes reach 8 centimetres in length and are rarely poikilitic. Locally, these zones contain olivine-rich areas or inclusions of wehrlite (less than 20 centimetres across) that are erratically distributed and render the outcrop mottled with rusty brown patches. In thin section, euhedral to subhedral clinopyroxenes are commonly schillered and exhibit inequigranular textures; most olivines appear subhedral and have cumulus or intercumulus textures. Phlogopite usually occurs as an accessory phase. Opaque oxides are notably reduced in abundance in olivine clinopyroxenites, and appear to be absent altogether in clinopyroxenites.

MIXED WEHRLITIC/PYROXENITIC UNITS

Mappable zones of intermixed olivine clinopyroxenite to clinopyroxenite and wehrlite to olivine wehrlite (and rarely dunite) are found locally near the margins of clinopyroxenite or wehrlitic bodies. They are particularly well developed in the central and northern parts of the intrusion (Figures 3-5-3 and 3-5-5). Contacts between the rock-types are generally sharp. The most common type of mixed unit comprises a chaotic assemblage of angular to subangular blocks of clinopyroxenite, ranging from less than one metre to tens of metres in size, enclosed by an olivine-rich host. This texture appears to have originated either by intrusion of clinopyroxenite magma into a semiconsolidated host, or by remobilization of zones of clinopyroxenite dike injection, similar to processes believed to be responsible for the disruption and redeposition of chromitite blocks. Another type of mixed zone is formed by concentrations of randomly oriented wehrlitic dikes cutting clinopyroxenite. The occurrence of mixed olivine-rich and pyroxenitic lithologies is also observed in the Tulameen complex (Nixon and Rublee, 1988) where they formed, in part at least, by slumping of coherent masses of dunite and/or clinopyroxenite cumulates that were plastically deformed during redeposition at lower levels in the magma chamber.

HORNBLENDE CLINOPYROXENITE AND HORNBLENDITE

This map unit comprises a gradation of rock types from hornblende clinopyroxenite through clinopyroxene hornblendite to hornblendite and feldspathic hornblendite. These hornblende-bearing ultramafic rocks are almost entirely restricted to the upper part of the complex where they are closely associated with gabbroic rocks. Hornblende clinopyroxenite occurs as a pale green to brownish green weathering, medium to coarse-grained rock studded with black hornblende crystals. It contains cumulus clinopyroxene, cumulus or intercumulus hornblende, locally abundant cumulus magnetite, accessory biotite and apatite, and intercumulus plagioclase appears in feldspathic variants. Hornblendite comprises a black, coarse-grained to pegmatitic texture with crystals reaching 8 centimetres in length. Locally, the rock exhibits a rude lineation, or a marked lamination of prismatic hornblende crystals with no directional fabric. A thin sill of megacrystic hornblende clinopyroxenite that intrudes roof rocks at the northwestern margin of the complex (Figure 3-5-5) locally displays well-developed centimetre-scale layering of cumulus subequant hornblende and clinopyroxene. Olivine-bearing hornblende clinopyroxenites are comparatively rare. One such unit occupies a narrow transition zone between dunite and hornblende clinopyroxenite at the zoned lower margin of the complex (Figure 3-5-4) and is distinguished by relatively abundant phlogopitic mica (5 per cent).

GABBROIC ROCKS

The gabbroic rocks comprise hornblende ± clinopyroxene gabbros and probably include undifferentiated dioritic phases carrying more sodic plagioclase. They are restricted to the margins of the complex and are most voluminous near the roof. They also form thin sills penetrating metasedimentary rocks of the Lay Range assemblage. Outcrops are typically lichen covered and dark grey weathering; fresh surfaces are medium grey to greenish grey depending on the degree of saussuritization of the feldspars. The rocks are usually massive, medium grained and equigranular. Centimetre-scale modal layering formed by alternating amphibole and plagioclase-rich horizons is observed locally and, near the contacts at least, is usually concordant with the attitude of bedding in the hostrocks. In thin section, the gabbroic rocks are composed of cumulus hornblende and rare clinopyroxene, cumulus to intercumulus plagioclase, accessory iron-titanium oxides, apatite and biotite, and sporadic secondary pyrite (less than 2 per cent).

Sills intruding the Lay Range assemblage have fine-grained chilled margins with rude columnar jointing or grade from medium-grained hornblende gabbro in the interior to hornblende porphyry at the contact. A narrow outer gabbroic zone at the lower margin of the complex locally displays an intense tectonic foliation that is locally mylonitic (described below) and concordant with that in metasedimentary hostrocks at the base of the intrusion.

DIKES AND VEINS

Dikes and veins of ultramafic to syenite/leucomonzonite or quartzofeldspathic composition are widespread in the Polaris complex, although there appears to be no systematic orientation to the pattern of dike intrusion. In large part, these dikes reflect the nature of, and temporal relationships between, major lithologic units.

Among the ultramafic rock types, centimetre to metre-wide dikes of olivine clinopyroxenite and clinopyroxenite are most common, and are found cutting dunite, olivine wehrlite and wehrlite (Plate 3-5-4). Dunite dikes, typically several centimetres in width, are only conspicuous where they penetrate chromitites (Plate 3-5-1); and thin olivine wehrlite to wehrlite dikes less than 0.5 metre wide transect dunite, wehrlitic and pyroxenitic units. The latter dikes locally exhibit concentrations of clinopyroxene crystals at their margins, a feature also documented at the Turnagain ultramafic complex (Clark, 1975). Thin (1 to 20 centimetres in width), medium to coarse-grained hornblendite to feldspathic hornblendite dikes have been observed cutting dunite, olivine clinopyroxenite, gabbroic rocks and metasedimentary strata of the Lay Range assemblage.

Leucocratic phases ranging from fine-grained, millimetre-width feldspathic veinlets to pegmatitic hornblende-biotite-feldspar ± quartz dikes and segregation pods several centimetres across are common within the gabbroic rocks and adjacent ultramafic units. In a ridge traverse along line D′-C′ (Figures 3-5-3 and 3-5-4), leucocratic dikes make their first appearance in dunite and wehrlitic lithologies and increase in abundance southward towards the gabbroic units. Composite dikes in the ultramafic rocks locally exhibit hornblendite margins and quartzofeldspathic cores; pegmatitic hornblende-feldspar-quartz segregation pods formed by ponding of residual liquids are also common in hornblende clinopyroxenites near the roof of the intrusion (Plate 3-5-5). The occurrence of silica-oversaturated differentiates in the Polaris complex may indicate late-stage contamination by siliceous wallrocks.

The overall sequence of dike intrusion, namely dunite, wehrlite, clinopyroxenite, hornblendite, gabbroic rocks and leucocratic residua, reflects the gross internal stratigraphy and general order of crystallization of major lithologic units in the complex. Locally ambivalent crosscutting relationships, such as those observed between olivine clinopyroxenite/clinopyroxenite and wehrlitic dikes, and hornblendite and gabbro, point to multiple intrusive events when magmas of a limited range of compositions coexisted.

Contact Relations and Intrusion Geometry

Steep contacts, rude internal zoning and the preservation of domical "roof" rocks at the northwestern end of the complex have previously been used to support a stock-like geometry for the intrusion (Roots, 1954; Irvine, 1974a; Foster, 1974). Detailed examination of contact relationships in the north confirms these crosscutting relationships (Figures 3-5-5 and 3-5-6). However, the elongate aspect of the complex, the nature of the western margin of the body where intrusive contacts are conformable with the strike and dip of metasedimentary hostrocks, the steep westward dip of zoned units at the eastern margin of the intrusion (Figure 3-5-4) and assymetrical nature of the zoning all suggest that the Polaris complex represents a high-level sill-like intrusion. Post-emplacement deformation has tilted the sill on end such that rocks forming the "roof" zone of earlier workers in fact represent wallrocks at the transgressive northern contact of the sill. Swarms of small gabbroic to pyroxenitic sills that intrude the roof zone to the west mimic the geometry of the larger intrusion.

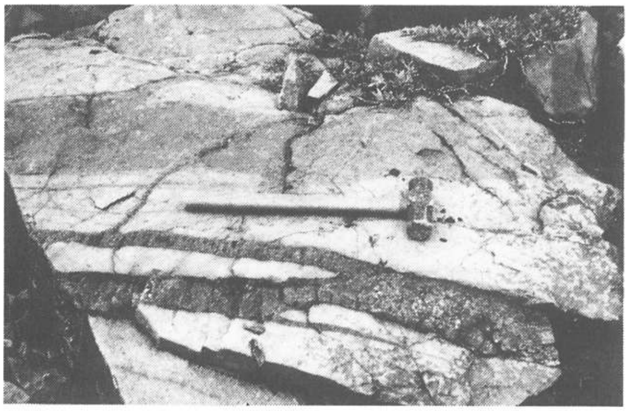

Plate 3-5-4. Bifurcating wehrlite and clinopyroxenite dikes in dunite. Note thin offshoot of clinopyroxenite dike cutting wehrlite at left of hammer handle.

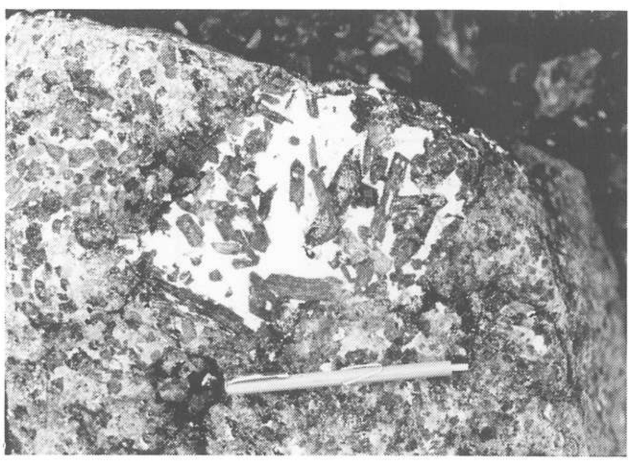

Plate 3-5-5. Pegmatitic hornblende-feldspar segregation pod in hornblende clinopyroxenite.

Internal Stratigraphy

In general, the gross internal distribution of rock types is systematically disposed about the margins. Dunite and olivine-bearing pyroxenitic rocks are concentrated near the base of the intrusion whereas hornblende-bearing pyroxenites and gabbros are well developed near the roof. In the east-central part of the intrusion, the lower margin is progressively zoned from dunite through wehrlitic and olivine clinopyroxenitic lithologies to olivine-hornblende clinopyroxenites and hornblende-rich gabbroic rocks at the contact (Figure 3-5-4). Contact relationships among the major lithologic units are sharp to gradational.

Mechanism of Emplacement

Irvine (1974a; 1976) was struck by the internal zoning (albeit complex) and evidence for disruption of layered chromitites and clinopyroxenites by olivine-rich lithologies. He suggested that these features could be explained most satisfactorily by diapiric re-emplacement of hot, thickly stratified, olivine-rich cumulates in a semi-solid state during regional tectonism. We have difficulty with this hypothesis for a number of reasons. In the first place, relationships among the wehrlitic and pyroxenitic lithologies (*i.e.* mixed units of this report) are locally quite complex and exhibit a history of multiple intrusive events. As indicated earlier, the chaotic nature of mixed ultramafic lithologies and remobilized chromitite horizons can be explained adequately by periodic, syndepositional mass flux of cumulates to lower levels in the magma chamber, events perhaps triggered by earthquakes and episodic magma recharge. Furthermore, we have found no evidence for injection of dunite and wehrlitic lithologies into gabbroic and hornblende-rich lithologies which would be expected in Irvine's model; rather, intrusive relationships dictate the reverse. Also we note the general assymetry in the development of internal zoning; the lack of gabbroic rocks at all intrusive contacts; the widespread preservation of cumulate textures, and the general lack of penetrative fabrics both within and at the margins of the body, except where faulted. Moreover, why a high density mass of olivine cumulates would migrate "diapirically" to significantly higher levels in the crust rather than sink to the crust-mantle boundary is not explained. We therefore suggest that the internal stratigraphy of cumulates developed *in situ* (*i.e.*

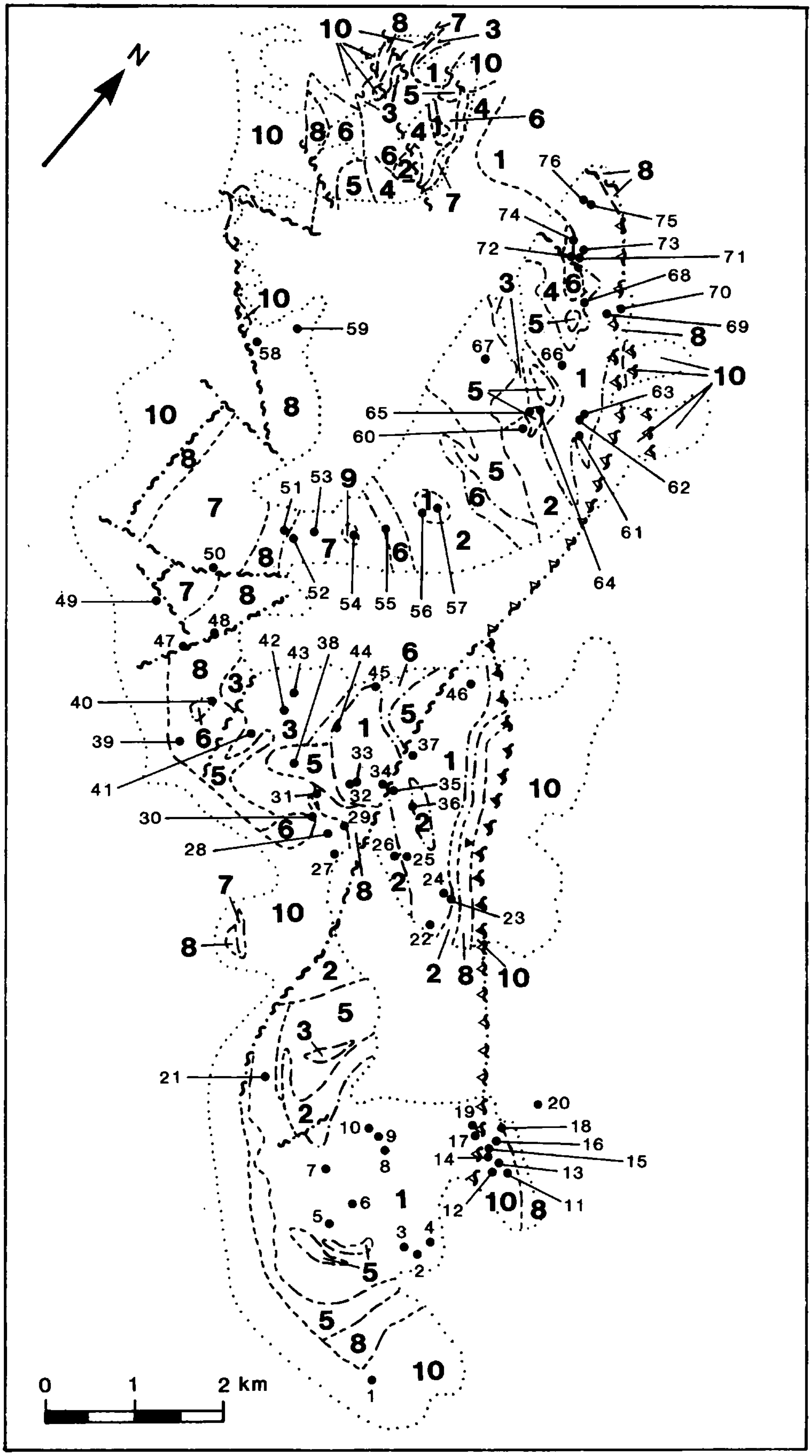

Figure 3-5-7. Location map of geochemical sample sites (also *see* Figure 3-5-5). Symbols and units as in Figure 3-5-3.

with respect to adjacent host rocks) essentially as we see it today, and probably represents the crystallization products of a subvolcanic magma chamber residing in the upper crust.

CONTACT AUREOLE

A contact aureole of amphibolite grade is well developed in rocks of the Lay Range assemblage at the margin of the intrusion. The maximum width of amphibolitic rocks is not accurately known, but has been estimated to extend on the order of 50 to 150 metres away from the contact (Irvine, 1974). The extent of hornfelsed rocks may be much greater than this, especially in the south (Roots, 1954).

In the north, metasedimentary rocks adjacent to the roof and margins of the complex have been recrystallized to an assemblage of hornblende, plagioclase, and quartz ± biotite ± potassium feldspar. The amphibolites exhibit no penetrative schistosity except where localized by faulting. Their finer grain size and preservation of relict stratification generally serves to distinguish these amphibolites from gabbroic and hornblende-rich ultramafic rocks of the intrusion.

At the southeastern extremity of the complex, serpentinized dunite lies in fault contact with black carbonaceous schists containing porphyroblasts of andalusite (chiastolite) up to 3 millimetres in length. Growth of andalusite appears to predate ductile movement in the shear zone and most likely formed in response to contact metamorphism. Andesitic lavas structurally underlying the phyllites lack a penetrative fabric but have undergone recrystallization of hornblende phenocrysts and groundmass consistent with amphibolite-grade metamorphism.

MINOR INTRUSIONS OF DUBIOUS AFFINITY

Intrusions that have uncertain relationships with the Polaris complex include a hornblende diorite sill-like body at the northwestern edge of the map area (Figure 3-5-5), a syenitic intrusion in the central part of the complex (Figure 3-5-3), and mafic dikes cutting metasedimentary rocks of the Lay Range assemblage. The dioritic body is a grey-brown weathering, fine to medium-grained rock, locally epidotized, with crystals of hornblende, plagioclase and minor potassium feldspar. The syenitic intrusion is a cream-weathering, greyish white coarse-grained rock comprising alkali feldspar (perthitic), minor hornblende, clinopyroxene and sphene, and trace amounts of biotite. It may be a late-stage differentiate of the Polaris complex that did not become silica-oversaturated for reasons yet to be determined. The dikes are dark greenish grey on weathered and fresh surfaces, and generally less than 2 metres in width, although they locally balloon to over 14 metres across. The thicker dikes have aphanitic chilled margins and hornblende-phyric interiors with up to 15 per cent phenocrysts. They are partly controlled by easterly oriented fault zones and are largely undeformed, appearing to postdate fault movement.

STRUCTURE AND METAMORPHISM

The Polaris complex is contained within a northwest-trending, southwest-facing and southwest-dipping homocline that represents a fault-bounded slice of Lay Range assemblage rocks. High-angle, northwest-trending faults are well exposed in the northern and eastern parts of the map area. They are commonly marked by schistose zones, crush breccias and quartz-carbonate alteration, and cut both the Polaris eomplex and its hostrocks. Most of these faults have displacements of unknown sense and magnitude.

Hostrocks overlying the intrusion, and wallrocks at the ends of the sill, lack a penetrative foliation. However, rocks resting structurally beneath the complex, especially lithologies adjacent to the eastern marginal fault, are generally highly schistose or mylonitic.

Metasedimentary rocks in ductile fault zones at the northeastern margin of the complex (cross-section D′-D, Figure 3-5-4), exhibit a single, well-developed slaty cleavage that is defined in thin section by biotite, muscovite, chlorite, quartz, plagioclase and minor carbonate. This mineral assemblage indicates that motion occurred during middle to upper greenschist facies metamorphism. Gabbroic rocks to the west have been inhomogeneously deformed and textures vary from massive to mylonitic with locally pronounced mineral lineations plunging steeply (60° to 70°) downdip to the northwest. In thin section, the mylonitic fabric is defined by amphibole, plagioclase, quartz, epidote, carbonate and chlorite (retrograde?) and deformation appears to have taken place under upper greenschist to lowermost amphibolite-grade conditions.

Oriented specimens were collected from metamorphosed volcaniclastic and sedimentary rocks farther south along the eastern marginal fault zone (cross-section A-B, Figure 3-5-4). C/S fabrics and shear bands observed in outcrop are defined in thin section by chlorite and biotite, and quartz and plagioclase have been dynamically recrystallized into subgrains. These textures indicate that mylonitization occurred during middle greenschist facies metamorphism. Kinematic indicators reveal that the hangingwall moved upward along southwest-dipping thrust planes. Mineral lineations plunging 65° to the northwest at the site of the C/S fabrics suggest that thrust movement, if parallel to the stretching direction, was toward the southeast. The eastern marginal fault zone continues farther south where it places dunite in thrust contact with andalusite-bearing schists that form part of the contact aureole (discussed above). A high-angle fault at the southwestern margin of the complex (cross-section A-B in Figure 3-5-4) also appears to be syntectonic with greenschist facies metamorphism.

From the preceding evidence, it is clear that major fault movements in the Lay Range took place during regional middle to upper greenschist metamorphism. The Polaris complex has been transported tectonically, together with its roof and most of its wallrocks, as an allochthonous thrust slice emplaced eastwards towards the craton, similar to other southwest-dipping, thrust-bounded packages mapped elsewhere in the Lay Range (Monger, 1973).

The eastward-verging structures in the Lay Range have counterparts in the Ingenika Group west of the Swannell fault, where they are represented by an early set of northwest-plunging, northeast-verging, tight to isoclinal folds (Bellefontaine, 1989). The Swannell fault represents a later, northeast-dipping, southwest-verging, imbricate thrust zone and associated drag folds that emplaced variably metamorphosed miogeoclinal rocks on Quesnellia (Bellefontaine, 1989). The timing of this deformation corresponds with the collision between the Intermontane Superterrane and ancestral North America which probably began in the Middle Jurassic (Gabrielse and Yorath, in press).

ALTERATION AND MINERALIZATION

Fault zones in the Polaris complex and Lay Range assemblage are commonly affected by quartz-carbonate alteration and weather to a bright orange-brown rock locally enriched in limonite, hematite, goethite and sulphides (largely pyrite). Alteration of this type develops in faults of every orientation in every lithology, but appears best developed in northwesterly and easterly trending fault zones. In addition to quartz and ferrodolomite, Irvine (1974a) noted the presence of vesuvianite, and Roots (1954) records ankerite and mariposite. Sparse quantities of asbestiform serpentine are restricted to joint surfaces and faults.

The Polaris complex appears remarkably devoid of sulphide mineralization. The only sulphides of note are exposed in several small reddish brown weathering outcrops of pyroxenitic rocks in the central part of the complex (Locality 43 in Figure 3-5-7). Here, net-textured primary sulphides, largely pyrrhotite, form immiscible blebs (up to 25 per cent) in a medium-grained clinopyroxenite. Disseminated secondary pyrite occurs locally in hornblende-rich ultramafic rocks and gabbros in amounts up to 2 per cent.

Chromitite is surprisingly sparse for the apparent size of the dunite mass. For example, in the Tulameen complex, which is a larger body but has a lower proportion of exposed dunite, chromitite is much more abundant. Likewise, the Wrede Creek complex appears to have more chromitite per square kilometre of exposed dunite than the Polaris intrusion. In both of the latter Alaskan-type bodies, platinum group elements are associated with chromitite (Nixon and Rublee, 1988; Hammack *et al.*, 1990, this volume). Evidently, the apparent size of the dunite body is no guide to its chromite or platinoid potential (discussed below). Magnetite is confined to the hornblende-bearing ultramafic and gabbroic lithologies but rarely exceeds 5 to 10 per cent of the rock and is of little economic significance.

GEOCHEMISTRY

Analytical results for gold, platinum group and "pathfinder" elements in over 130 representative samples of the Polaris complex and its hostrocks are presented in Table 3-5-1. Sample localities are shown in Figures 3-5-5 and 3-5-7. Three different analytical methods were used in two independent laboratories: inductively coupled plasma (ICP) mass spectrometry, Acme Analytical Laboratories, Vancouver; instrumental neutron activation analysis (INAA), Institut National de la Récherche Scientifique, Université du Québec; and inductively coupled plasma emission spectrometry, also Acme Analytical Laboratories. Accuracy was checked by international and in-house standards, and analytical precision (and any nugget effect) monitored by hidden duplicates and internal standards. All samples were preconcentrated by fire assay from 30 gram (ICP) or 50 gram (INAA) splits of 200 grams of rock powder (-200 mesh).

"Pathfinder" elements include sulphur, nickel, chromium, arsenic and antimony. Sulphur is generally low in abundance, and reaches a maximum value of 4.2 weight per cent in a clinopyroxenite (Locality 43, Table 3-5-1 and Figure 3-5-7) that contains net-textured sulphides and has weakly anomalous abundances of platinum and palladium. The only other sulphur-rich samples of note (greater than 1 per cent sulphur) are gabbros and metasedimentary rocks which contain secondary pyrite and exhibit no enrichment in the noble metals. Nickel and chromium have relatively high abundances in ultramafic rocks, but the latter element is particularly sensitive to the abundance of chromite in olivine-rich rocks. The highest chromium abundances reflect chip samples of high-grade chromitite; the chromitiferous dunites represent composite samples of small chromitite schlieren and host dunite. Thus, if platinum group elements (PGE) are preferentially concentrated in chromitite, as in the case of the Tulameen complex (St. Louis *et al.*, 1986), and PGE-rich and PGE-depleted chromitites exist, there will be no simple correlation between the abundanee of chromium and PGE in the sample population. The abundances of arsenic and antimony are generally low and show no correlation with PGE or gold.

The highest concentration of platinum (735 ppb) is found in chromitiferous dunite (Locality 22, Figure 3-5-3) and is accompanied by small but significant quantities of rhodium, ruthenium, iridium and osmium. Low abundances of sulphur, arsenic and antimony suggest that the PGE may be contained as discrete platinum-iron alloys as typifies the chromitite-PGE association in the Tulameen complex (St. Louis *et al.*, 1986; Nixon *et al.*, 1989c). The tenor of platinum in a duplicate analysis is over fourteen times lower, and is attributed to a nugget effect, which appears to influence other samples (*e.g.* GN-88-1039, Locality 66). Rarely, abundances of iridium and osmium seem anomalously high (*e.g.* GN-88-1055B, Locality 71) and may indicate the presence of iridium-osmium(-ruthenium?) alloys which have been documented in other Alaskan-type associations (Harris and Cabri, 1973; Cabri and Harris, 1975). Anomalously high platinum and palladium (greater than 200 ppb) occur in a clinopyroxenite dike in the central part of the complex (GN-88-4069A, Locality 33). In general, palladium remains near or below detection limits in chromitites and olivine-rich ultramafic rocks, and increases in abundance in hornblende-bearing and gabbroic rocks. This behaviour is accompanied by a concomitant decrease in the platinum:palladium ratio, similar to trends in other Alaskan-type complexes (*e.g.* Nixon *et al.*, 1989b).

Highly anomalous abundances of gold (greater than 100 ppb) occur in an andesitic dike intruding metasedimentary sequences adjacent to a fault zone (Locality 77, Figure 3-5-5); in hornblendite at the roof of the ultramafic complex (Locality 82); and in dunite near the margin of the intrusion (Locality 100). In fact, all of the gold anomalies (greater than 20 ppb) are confined to the extreme northwestern end of the complex. Almost without exception, they exhibit a strong spatial relationship with intrusive contacts, although some are close to faults, and appear to favour mafic lithologies. No correlation is evident between the abundances of PGE and gold, which is generally low within the complex, and the few quartz veins and quartz-carbonate alteration zones sampled

TABLE 3-5-1
ABUNDANCES OF NOBLE METALS AND "PATHFINDER" ELEMENTS
IN THE POLARIS ULTRAMAFIC COMPLEX AND ASSOCIATED ROCKS

Locality	Sample	S wt %	Ni	Cr	As	Sb	Pt	Pd	Rh	Ru	Re	Ir	Os	Au
			ppm				ppb							
POLARIS COMPLEX														
Chromitite and Chromitiferous Dunite														
62	GN-88-1032[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
62	GN-88-1032[2]	<0.02	1500	257955	<1.0	<0.20	<5	<5	7	<10	<5	8.70	4.8	<1.0
66	GN-88-1039[1]	—	—	—	—	—	49	5	<2	—	—	—	—	<1.0
66	GN-88-1039[2]	<0.02	2036	102308	<1.0	<0.20	121	<5	16	24	<5	33.00	18.0	<1.0
68	GN-88-1053[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
68	GN-88-1053[1]*	—	—	—	—	—	<1	<2	3	—	—	—	—	<1.0
68	GN-88-1053[2]	<0.20	1929	75107	<1.0	<0.20	<5	<5	6	10	<5	4.50	<3.0	<1.0
71	GN-88-1055B[1]	—	—	—	—	—	28	<2	<2	—	—	—	—	<1.0
71	GN-88-1055B[1]*	—	—	—	—	—	72	5	<2	—	—	—	—	<1.0
71	GN-88-1055B[2]	<0.02	1870	64710	<1.0	<0.20	75	<5	19	45	<5	88.00	51.0	<1.0
76	GN-88-1058A[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	2.0
76	GN-88-1058A[1]*	—	—	—	—	—	2	<2	<2	—	—	—	—	<1.0
76	GN-88-1058A[2]	<0.02	2036	49612	<1.0	<0.20	<5	<5	3	31	<5	3.30	<3.0	<1.0
76	GN-88-1058B[1]	—	—	—	—	—	2	<2	<2	—	—	—	—	<1.0
76	GN-88-1058B[2]	<0.02	2322	39182	<1.0	<0.20	<5	<5	4	<10	<5	2.20	<3.0	<1.0
25	GN-88-1074[1]	—	—	—	—	—	72	<2	<2	—	—	—	—	<1.0
25	GN-88-1074[2]	<0.02	2331	58402	<1.0	0.37	—	—	—	—	—	—	—	<1.0
22	GN-88-1089[1]	—	—	—	—	—	50	<2	<2	—	—	—	—	<1.0
22	GN-88-1089[2]	<0.02	1641	75803	<1.0	<0.20	735	<5	24	32	<5	43.00	19.0	<1.0
63	GN-88-1031[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
63	GN-88-1031[2]	<0.02	1660	155744	<1.0	<0.20	<10	<5	7	24	6	5.40	<3.0	<1.0
69	GN-88-1052[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
69	GN-88-1052[2]	<0.02	2080	135880	<1.0	<0.20	<10	<5	5	<5	<5	9.30	6.0	<1.0
36	GN-88-1069B[1]	—	—	—	—	—	5	5	<2	—	—	—	—	<1.0
36	GN-88-1069B[1]*	—	—	—	—	—	3	6	<2	—	—	—	—	2.0
36	GN-88-1069B[2]	<0.02	2176	27235	<1.0	<0.20	<5	<5	2	<10	<5	3.90	<3.0	<1.0
24	GN-88-1092[1]	—	—	—	—	—	4	<2	<2	—	—	—	—	<1.0
24	GN-88-1092[2]	<0.02	1896	29311	<1.0	<0.20	<5	<5	5	<15	<5	4.50	<3.0	<1.0
61	GN-88-2050[1]	—	—	—	—	—	<1	4	<2	—	—	—	—	<1.0
61	GN-88-2050[2]	<0.02	2024	224571	<5.0	<0.30	<10	<5	9	16	<5	10.00	5.0	<1.0
37	GN-88-2072[1]	—	—	—	—	—	<1	3	<2	—	—	—	—	2.0
37	GN-88-2072[2]	<0.02	2467	89573	<5.0	<0.30	<5	<5	14	12	<5	8.90	6.5	<1.0
46	GN-88-2073[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
46	GN-88-2073[2]	<0.02	2692	28022	<5.0	<0.30	9	<5	5	<5	<5	4.50	<3.0	<1.0
44	GN-88-2077[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
44	GN-88-2077[2]	<0.02	1803	211248	<5.0	0.36	<5	<5	12	39	<5	10.00	<3.0	<1.0
17	GN-88-2107A[1]	—	—	—	—	—	5	<2	<2	—	—	—	—	<1.0
17	GN-88-2107A[2]	<0.02	2282	65558	<5.0	<0.30	<10	<5	3	<15	<5	3.10	3.4	<1.0
21	GN-88-2113[1]	—	—	—	—	—	3	<2	<2	—	—	—	—	<1.0
21	GN-88-2113[2]	<0.02	2093	41928	<5.0	<0.30	<5	<5	6	<15	<5	27.00	15.0	<1.0
10	GN-88-3145[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
10	GN-88-3145[2]	<0.02	1929	143079	<5.0	<0.30	<10	<5	9	<15	<5	11.00	9.3	<1.0
75	GN-88-4068[1]	—	—	—	—	—	6	<2	<2	—	—	—	—	2.0
75	GN-88-4068[2]	<0.02	1821	261896	<5.0	<0.30	15	<10	5	<15	<5	22.0	15.0	<1.0
5	GN-88-4098[1]	—	—	—	—	—	2	<2	<2	—	—	—	—	<1.0
5	GN-88-4098[2]	<0.02	2181	107468	<5.0	<0.30	<10	<5	11	<20	<5	21.00	12.0	<1.0
9	GN-88-4102[1]	—	—	—	—	—	3	<2	<2	—	—	—	—	<1.0
9	GN-88-4102[2]	<0.02	1974	48594	6.5	<0.30	9	<5	4	<20	<5	5.70	4.0	<1.0
8	GN-88-4103[1]	—	—	—	—	—	2	<2	<2	—	—	—	—	<1.0
8	GN-88-4103[2]	<0.02	2004	113463	<5.0	<0.30	18	<5	4	<20	<5	8.60	9.4	<1.0
Dunite														
72	GN-88-1054B[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	2.0
72	GN-88-1054B[2]	<0.02	1650	1213	1.1	0.25	<5	<5	2	<5	<5	2.70	<3.0	<1.0
71	GN-88-1055A[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
71	GN-88-1055A[2]	<0.02	2228	4407	<1.0	<0.20	21	<5	<1	26	<5	5.20	<3.0	1.1
73	GN-88-1056A[1]	—	—	—	—	—	6	<2	<2	—	—	—	—	2.0
73	GN-88-1056A[2]	<0.02	1986	2339	<1.0	<0.20	<10	<5	2	25	<5	5.50	3.5	<1.0
74	GN-88-1057A[1]	—	—	—	—	—	4	<2	<2	—	—	—	—	<1.0
74	GN-88-1057A[2]	0.02	2293	2977	<1.0	<0.20	<10	<10	<1	22	<5	2.80	<3.0	<1.0
95	GN-88-2044[1]	—	—	—	—	—	<1	3	<2	—	—	—	—	<1.0
96	GN-88-2048A[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
96	GN-88-2048A[2]	<0.02	1723	3635	<5.0	<0.30	<5	<15	<1	17	<5	0.92	<3.0	<1.0
19	GN-88-2106[1]	—	—	—	—	—	5	<2	<2	—	—	—	—	<1.0
19	GN-88-2106[2]	0.03	2367	3969	5.4	<0.30	<5	<5	<1	<5	<5	1.30	<3.0	<1.0
2	GN-88-4092A[1]	—	—	—	—	—	2	<2	<2	—	—	—	—	<1.0
2	GN-88-4092A[2]	0.07	1928	2952	<5.0	0.31	<5	<5	<1	<5	<5	0.90	<3.0	<1.0
6	GN-88-4097[1]	—	—	—	—	—	2	<2	<2	—	—	—	—	2.0
6	GN-88-4097[2]	<0.02	2448	28449	<5.0	<0.30	<5	<5	2	<15	<5	3.30	3.7	<1.0
100	GN-89-6220[3]	—	—	—	—	—	<1	<2	<2	—	—	—	—	122.0

Analytical methods: [1] Inductively coupled plasma mass spectrometry, Acme Laboratories, Vancouver; [2] Instrumental neutron activation, Institutè National de la Récherche Scientifique, Québec; [3] Inductively coupled plasma emission spectrometry, Acme Laboratories, Vancouver.
Detection limits: 1 ppb for Pt and Au; 2 ppb for Pd and Rh; neutron activation detection limits vary with sample composition.
* Duplicate analysis.

TABLE 3-5-1—*Continued*
ABUNDANCES OF NOBLE METALS AND "PATHFINDER" ELEMENTS IN THE POLARIS ULTRAMAFIC COMPLEX AND ASSOCIATED ROCKS

Locality	Sample	S	Ni	Cr	As	Sb	Pt	Pd	Rh	Ru	Re	Ir	Os	Au
		wt %	ppm				ppb							
Olivine Wehrlite and Wehrlite														
72	GN-88-1054A[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
72	GN-88-1054A[2]	<0.20	2175	3661	<1.0	<0.20	17	<5	3	27	<5	2.00	<3.0	<1.0
36	GN-88-1069A[1]	—	—	—	—	—	<1	3	<2	—	—	—	—	<1.0
36	GN-88-1069A[2]	<0.02	2026	3281	<1.0	<0.20	<10	<10	<1	20	<5	0.98	<3.0	<1.0
26	GN-88-1075[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
23	GN-88-1093A[1]	—	—	—	—	—	7	<2	<2	—	—	—	—	2.0
23	GN-88-1093A[1]*	—	—	—	—	—	3	<2	<2	—	—	—	—	3.0
23	GN-88-1093A[2]	<0.02	1744	1096	2.3	<0.20	13	<5	2	<5	<5	3.00	3.8	<1.0
45	GN-88-2076[1]	—	—	—	—	—	<1	<2	3	—	—	—	—	<1.0
45	GN-88-2076[2]	<0.02	2521	99916	<5.0	<0.30	<10	<5	11	21	<5	6.90	4.8	<1.0
60	GN-88-4060[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
60	GN-88-4060[2]	<0.02	2025	94789	6.9	<0.30	<15	<5	6	<40	<5	5.20	<3.0	<1.0
32	GN-88-4070A[1]	—	—	—	—	—	7	<2	<2	—	—	—	—	<1.0
87	GN-89-8143[3]	—	—	—	—	—	2	<2	<2	—	—	—	—	3.0
73	GN-88-1056B[1]	—	—	—	—	—	3	4	<2	—	—	—	—	<1.0
73	GN-88-1056B[2]	<0.02	1199	5284	<1.0	<0.20	<20	<10	<1	23	<5	1.60	<3.0	1.2
30	GN-88-4080A[1]	—	—	—	—	—	2	12	<2	—	—	—	—	<1.0
30	GN-88-4080A[2]	<0.02	1148	2298	<5.0	<0.30	9	12	<1	<5	<5	0.56	<3.0	<1.0
7	GN-88-4101[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
7	GN-88-4101[2]	<0.02	1100	2764	<5.0	<0.30	<5	<5	<1	<10	<5	0.54	<3.0	<1.0
104	GN-89-6201[3]	—	—	—	—	—	2	<2	<2	—	—	—	—	3.0
90	GN-89-8134[3]	—	—	—	—	—	3	<2	<2	—	—	—	—	<1.0
Olivine Clinopyroxenite and Clinopyroxenite														
74	GN-88-1057B[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
74	GN-88-1057B[2]	<0.02	442	4157	<1.0	<0.20	<20	<10	<1	22	<5	0.95	<3.0	<1.0
34	GN-88-1063[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
96	GN-88-2048B[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
96	GN-88-2040B[2]	<0.02	359	1849	<5.0	<0.30	<10	<35	<1	<15	<5	<0.10	<3.0	<1.0
43	GN-88-2079[1]	—	—	—	—	—	45	22	<2	—	—	—	—	3.0
43	GN-88-2079[2]	4.24	556	726	<5.0	<0.30	53	25	<1	20	6	0.27	<3.0	3.2
88	GN-88-4042[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
64	GN-88-4052[1]	—	—	—	—	—	5	<2	<2	—	—	—	—	<1.0
67	GN-88-4063A[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
67	GN-88-4063A[2]	<0.02	249	1798	<5.0	0.46	<5	<5	<1	<5	<5	<0.10	<3.0	<1.0
33	GN-88-4069A[1]	—	—	—	—	—	239	285	<2	—	—	—	—	7.0
32	GN-88-4070B[1]	—	—	—	—	—	5	11	<2	—	—	—	—	<1.0
3	GN-88-4093[1]	—	—	—	—	—	2	7	<2	—	—	—	—	<1.0
3	GN-88-4093[2]	<0.02	529	2556	<5.0	<0.30	<5	<5	<1	<15	<5	0.20	<3.0	<1.0
98	GN-89-7130[3]	—	—	—	—	—	4	<2	<2	—	—	—	—	14.0
98	GN-89-7131A[3]	—	—	—	—	—	4	<2	<2	—	—	—	—	<1.0
99	GN-89-9173[3]	—	—	—	—	—	50	70	7	—	—	—	—	13.0
55	GN-89-9141[3]	—	—	—	—	—	<1	<2	<2	—	—	—	—	5.0
52	GN-89-9146B[3]	—	—	—	—	—	13	7	<2	—	—	—	—	8.0
94	GN-89-9169B[3]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
35	GN-88-1064[1]	—	—	—	—	—	8	13	<2	—	—	—	—	2.0
85	GN-89-8139[3]	—	—	—	—	—	8	9	<2	—	—	—	—	5.0
Hornblende Clinopyroxenite, Clinopyroxene Hornblendite, and Hornblendite														
48	GN-88-1077[1]	—	—	—	—	—	5	<2	<2	—	—	—	—	2.0
48	GN-88-1077[2]	<0.02	<150	<150	5.9	1.80	9	9	<1	7	<5	3.10	<3.0	<1.0
79	GN-88-4031A[1]	—	—	—	—	—	11	35	<2	—	—	—	—	<1.0
79	GN-88-4031A[2]	<0.02	251	732	<5.0	<0.30	12	24	1	<5	<5	0.13	<3.0	<1.0
79	GN-88-4031C[1]	—	—	—	—	—	<1	5	<2	—	—	—	—	3.0
92	GN-88-4045[1]	—	—	—	—	—	14	44	<2	—	—	—	—	<1.0
30	GN-88-4080B[1]	—	—	—	—	—	2	16	<2	—	—	—	—	2.0
30	GN-88-4080B[2]	<0.02	258	675	<5.0	<0.30	8	14	<1	<10	<5	0.15	<3.0	1.1
111	GN-89-7143B[3]	—	—	—	—	—	13	12	<2	—	—	—	—	10.0
79	GN-89-9132B[3]	—	—	—	—	—	16	20	<2	—	—	—	—	24.0
79	GN-89-9132C[3]	—	—	—	—	—	6	8	<2	—	—	—	—	22.0
82	GN-89-9167B[3]	—	—	—	—	—	48	14	7	—	—	—	—	174.0
57	GN-89-9139A[3]	—	—	—	—	—	5	5	<2	—	—	—	—	9.0
56	GN-89-9140A[3]	—	—	—	—	—	11	<2	<2	—	—	—	—	2.0
86	GN-89-8133A[3]	—	—	—	—	—	7	12	<2	—	—	—	—	<1.0
50	GN-88-1080B[1]	—	—	—	—	—	4	<2	<2	—	—	—	—	2.0
83	GN-88-2040B[1]	—	—	—	—	—	8	18	<2	—	—	—	—	<1.0
83	GN-88-2040B[1]*	—	—	—	—	—	9	15	<2	—	—	—	—	<1.0
83	GN-88-2040B[2]	0.44	<150	<150	<5.0	<0.30	14	16	<1	17	<5	0.17	<3.0	<1.0
97	GN-88-2045[1]	—	—	—	—	—	3	11	<2	—	—	—	—	<1.0
97	GN-88-2045[2]	<0.02	<150	966	<5.0	<0.30	7	<10	<1	13	<5	0.14	<3.0	<1.0
38	GN-88-4082B[1]	—	—	—	—	—	<1	14	<2	—	—	—	—	<1.0
38	GN-88-4082B[2]	<0.02	167	228	<5.0	<0.30	7	13	<1	<5	<5	<0.10	<3.0	<1.0
40	GN-88-4086[1]	—	—	—	—	—	<1	18	<2	—	—	—	—	<1.0
40	GN-88-4086[1]*	—	—	—	—	—	<1	15	<2	—	—	—	—	<1.0
40	GN-88-4086[2]	0.02	<150	287	<5.0	<0.30	6	10	<1	<5	<5	<0.10	<3.0	<1.0
93	GN-89-8147[3]	—	—	—	—	—	<1	<2	<2	—	—	—	—	2.0

Analytical methods: [1] Inductively coupled plasma mass spectrometry, Acme Laboratories, Vancouver; [2] Instrumental neutron activation, Instituté National de la Récherche Scientifique, Québec; [3] Inductively coupled plasma emission spectrometry, Acme Laboratories, Vancouver.
Detection limits: 1 ppb for Pt and Au; 2 ppb for Pd and Rh; neutron activation detection limits vary with sample composition.
* Duplicate analysis.

TABLE 3-5-1 — *Continued*
ABUNDANCES OF NOBLE METALS AND "PATHFINDER" ELEMENTS
IN THE POLARIS ULTRAMAFIC COMPLEX AND ASSOCIATED ROCKS

Locality	Sample	S wt %	Ni ppm	Cr ppm	As ppm	Sb ppm	Pt ppb	Pd ppb	Rh ppb	Ru ppb	Re ppb	Ir ppb	Os ppb	Au ppb
Gabbroic Rocks														
70	GN-88-1050A[1]	—	—	—	—	—	2	<2	<2	—	—	—	—	2.0
50	GN-88-1080C[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	5.0
59	GN-88-2094[1]	—	—	—	—	—	3	8	<2	—	—	—	—	3.0
59	GN-88-2094[2]	0.01	<150	154	<5.0	<0.30	7	6	<1	18	<5	<0.10	<3.0	1.4
58	GN-88-2095[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
58	GN-88-2095[2]	<0.02	<150	255	<5.0	<0.30	<5	<5	<1	22	<5	<0.10	<3.0	<1.0
18	GN-88-2100B[1]	—	—	—	—	—	2	<2	<2	—	—	—	—	11.0
18	GN-88-2100B[2]	2.72	<150	<150	<5.0	1.20	<5	<5	<1	19	<5	<0.10	<3.0	3.8
18	GN-88-2100C[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
18	GN-88-2100C[2]	<0.02	<150	315	<5.0	0.64	<5	<5	<1	20	<5	<0.10	<3.0	<1.0
20	GN-88-2115[1]	—	—	—	—	—	4	<2	<2	—	—	—	—	<1.0
79	GN-88-4030B[1]	—	—	—	—	—	2	20	<2	—	—	—	—	2.0
81	GN-88-4036[1]	—	—	—	—	—	2	11	<2	—	—	—	—	4.0
81	GN-88-4036[2]	<0.02	<150	184	7.9	<0.30	7	11	<1	<5	<5	<0.10	<3.0	<1.0
29	GN-88-4073A[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	4.0
102	GN-89-6165[3]	—	—	—	—	—	7	7	<2	—	—	—	—	47.0
107	GN-89-6195[3]	—	—	—	—	—	8	9	<2	—	—	—	—	55.0
109	GN-89-7164[3]	—	—	—	—	—	9	9	<2	—	—	—	—	42.0
108	GN-89-6193B[3]	—	—	—	—	—	4	<2	<2	—	—	—	—	3.0
112	GN-89-9154[3]	—	—	—	—	—	8	9	<2	—	—	—	—	5.0
83	GN-88-2040A[1]	—	—	—	—	—	5	11	<2	—	—	—	—	<1.0
83	GN-88-2040A[2]	2.29	<150	746	<5.0	<0.30	11	14	<1	18	8	0.14	<3.0	<1.0
91	GN-88-4043A[1]	—	—	—	—	—	5	21	<2	—	—	—	—	9.0
31	GN-88-4072A[1]	—	—	—	—	—	5	16	<2	—	—	—	—	<1.0
31	GN-88-4072A[2]	0.30	<150	216	<5.0	<0.30	8	10	<1	<5	<5	<0.10	<3.0	<1.0
28	GN-88-4074[1]	—	—	—	—	—	<1	14	<2	—	—	—	—	<1.0
39	GN-88-4089[1]	—	—	—	—	—	<1	10	<2	—	—	—	—	<1.0
78	GN-89-9128C[3]	—	—	—	—	—	11	8	<2	—	—	—	—	27.0
51	GN-89-9147[3]	—	—	—	—	—	2	4	<2	—	—	—	—	4.0
80	GN-89-9148B[3]	—	—	—	—	—	11	9	<2	—	—	—	—	5.0
80	GN-89-9148B[3]*	—	—	—	—	—	11	9	<2	—	—	—	—	6.0
78	GN-89-9128A[3]	—	—	—	—	—	2	3	<2	—	—	—	—	88.0
Hornblende-Feldspar Pegmatites and Felsic Dikes														
42	GN-88-2078[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
42	GN-88-2078[2]	<0.02	204	<150	<5.0	<0.30	<5	<5	<1	13	<5	<0.10	<3.0	<1.0
65	GN-88-4053D[1]	—	—	—	—	—	7	<2	<2	—	—	—	—	<1.0
29	GN-88-4073B[1]	—	—	—	—	—	2	<2	<2	—	—	—	—	<1.0
29	GN-88-4073B[2]	0.05	<150	<150	10.0	<0.30	<5	<5	<1	<5	<5	<0.10	<3.0	<1.0
27	GN-88-4076[1]	—	—	—	—	—	2	<2	<2	—	—	—	—	<1.0
41	GN-88-4085[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
41	GN-88-4085[3]	<0.02	<150	<150	<5.0	<0.30	<5	<5	<1	8	<5	<0.10	<3.0	<1.0
105	GN-89-6188[3]	—	—	—	—	—	<1	<2	<2	—	—	—	—	60.0
82	GN-89-8123B-2[3]	—	—	—	—	—	2	<2	<2	—	—	—	—	<1.0
82	GN-89-8123B-1[3]	—	—	—	—	—	10	4	<2	—	—	—	—	<1.0
53	GN-89-9145Z[3]	—	—	—	—	—	2	<2	<2	—	—	—	—	<1.0
54	GN-89-9143[3]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
LAY RANGE (HARPER RANCH) ASSEMBLAGE														
Metasedimentary and Metavolcaniclastic Rocks														
47	GN-88-2083[1]	—	—	—	—	—	<1	6	<2	—	—	—	—	<1.0
49	GN-88-2088[1]	—	—	—	—	—	2	4	<2	—	—	—	—	2.0
13	GN-88-2109[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	3.0
13	GN-88-2109[2]	1.32	<150	<150	5.5	<0.30	<5	<5	<1	<5	<5	<0.10	<3.0	1.6
11	GN-88-2110[1]	—	—	—	—	—	3	<2	<2	—	—	—	—	<1.0
101	GN-89-6162A[3]	—	—	—	—	—	<1	<2	<2	—	—	—	—	83.0
101	GN-89-6162B[3]	—	—	—	—	—	2	<2	<2	—	—	—	—	6.0
106	GN-89-6199[3]	—	—	—	—	—	9	5	<2	—	—	—	—	<1.0
1	GN-89-6225[3]	—	—	—	—	—	2	5	<2	—	—	—	—	3.0
110	GN-89-7147A[3]	—	—	—	—	—	19	20	<2	—	—	—	—	76.0
4	GN-89-8155[3]	—	—	—	—	—	2	<2	<2	—	—	—	—	12.0
103	GN-89-6202[3]	—	—	—	—	—	<1	4	<2	—	—	—	—	43.0
110	GN-89-7147B[3]	—	—	—	—	—	3	4	<2	—	—	—	—	5.0
89	GN-89-8136[3]	—	—	—	—	—	<1	4	<2	—	—	—	—	4.0
79	GN-89-9132A[3]	—	—	—	—	—	<1	<2	<2	—	—	—	—	19.0
80	GN-89-9148A[3]	—	—	—	—	—	2	6	<2	—	—	—	—	3.0
82	GN-89-9167A[3]	—	—	—	—	—	4	5	<2	—	—	—	—	4.0
Metavolcanic Rocks and Dikes														
14	GN-89-9175A[3]	—	—	—	—	—	<1	<2	<2	—	—	—	—	4.0
12	GN-89-9176[3]	—	—	—	—	—	<1	<2	<2	—	—	—	—	11.0
77	GN-89-9126C[3]	—	—	—	—	—	3	<2	<2	—	—	—	—	214.0
113	GN-89-9153A[3]	—	—	—	—	—	7	8	<2	—	—	—	—	10.0
QUARTZ VEINS AND QUARTZ-CARBONATE ALTERATION														
16	GN-88-2101[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	<1.0
15	GN-88-2108[1]	—	—	—	—	—	<1	<2	<2	—	—	—	—	2.0
84	GN-89-8126[3]	—	—	—	—	—	2	<2	<2	—	—	—	—	7.0

Analytical methods: [1] Inductively coupled plasma mass spectrometry, Acme Laboratories, Vancouver; [2] Instrumental neutron activation, Institutè National de la Récherche Scientifique, Québec; [3] Inductively coupled plasma emission spectrometry, Acme Laboratories, Vancouver.
Detection limits: 1 ppb for Pt and Au; 2 ppb for Pd and Rh; neutron activation detection limits vary with sample composition.
* Duplicate analysis.

are barren. These data suggest that circulating fluids, possibly driven by convective cooling at the margins of the intrusion, scavenged gold from Lay Range assemblage lithologies and deposited it near the contacts of the intrusion where mafic rocks acted as a chemical sink. The apparent spatial concentration of gold anomalies at the northwestern contacts is probably a reflection of sampling bias.

SUMMARY AND CONCLUSIONS

The Polaris ultramafic complex of north-central British Columbia provides an instructive example of an Alaskan-type intrusion. It represents one of the largest (45 square kilometres) of such bodies in the province, second only to the Tulameen complex in southern British Columbia. It is one of a number of Alaskan-type complexes that intrude the accreted terranes of Quesnellia and Stikinia and that are generally considered to represent the subvolcanic magma chambers of Late Triassic island arc volcanoes whose products comprise the Nicola, Takla and Stuhini Groups. Although spectacular mesoscopic layering such as that found at the classic locality of Duke Island (Irvine, 1974b) is nowhere developed, relationships with the country rocks, and between ultramafic and more differentiated gabbroic rocks, are well exposed. In addition, all typical Alaskan-type lithologies are represented, including dunite, chromitite, wehrlite, olivine clinopyroxenite, clinopyroxenite, hornblende clinopyroxenite, hornblendite, gabbros and more leucocratic pegmatitic phases. Cumulate textures are widespread and well preserved, and an interesting petrologic feature is the occurrence of cumulus and intercumulus phlogopite in early cumulates (dunite), indicative of relatively potassium-rich parental magmas. A potentially important economic consideration is the chromitite-platinoid association documented from other Alaskan-type intrusions.

Detailed mapping of the Polaris ultramafic complex and its environs has established that the body represents a westward-dipping, westward-facing transgressive sill that intruded metasedimentary and metavolcanic rocks of the Upper Paleozoic Lay Range assemblage (Harper Ranch tectonostratigraphic terrane) which forms the basement of Quesnellia. Intrusive contacts at the roof and margins of the sill are well exposed at the northwestern end of the complex. Here, minor coeval intrusions of gabbro and hornblende clinopyroxenite in the country rocks mimic the geometry of their larger counterpart.

The internal stratigraphy of the Polaris complex is well exposed in cross-section. The lower margin of the sill is zoned outward over a narrow interval from dunite through wehrlite and olivine clinopyroxenite to olivine-hornblende clinopyroxenite and hornblende gabbro. Dunite occupies much of the lower part of the intrusion, and, in a gross sense, is succeeded upward by thickly stratified wehrlitic cumulates and clinopyroxenites, hornblende-bearing clinopyroxenites and hornblendites, and hornblende-rich gabbroic rocks which are well developed near the roof of the intrusion. Contacts between the main lithologic units are sharp to gradational. On a more localized scale, evidence exists for complex multiple intrusive events between wehrlitic and pyroxenitic lithologies on the one hand, and between gabbroic and hornblende-rich ultramafic units on the other. Previously consolidated chromitite layers, and complex zones of intermixing of wehrlitic and clinopyroxenitic rocks, point to syndepositional remobilization of early cumulates by mass wasting processes which deposited them lower down in the magma chamber. The widespread remobilization of chromitite concentrations, combined with their scanty occurrence, do not bode well for commercial exploitation of chromite in these ultramafic rocks.

The Polaris complex is contained within a northwest-trending, westward-dipping homoclinal sequence of well-bedded country rocks with unambiguous sedimentary structures that face west. High-angle faults with predominantly northwesterly and westerly trends transect ultramafic and country rocks alike. A basal ductile thrust zone exposed at the eastern margin of the complex exhibits S/C fabrics and kinematic indicators which suggest tectonic transport to the east. The Polaris complex has therefore been tectonically uprooted and transported, together with its hostrocks, as an allochthonous slice, similar to the structure of rock packages documented in other parts of the Lay Range. Major faulting was active during middle to upper greenschist regional metamorphism which appears to be superimposed on a contact aureole of amphibolite grade. This deformation and regional metamorphism are probably related to the collision of accreted terranes of the Intermontane Belt, specifically Quesnellia, against the miogeocline of ancestral North America.

Lithogeochemical assays for platinum group elements, gold, and associated pathfinder elements yield some interesting but rather localized anomalies. Platinum attains a maximum abundance of 735 ppb in chromitite-bearing dunite, and is accompanied by minor iridium and osmium which reach maximum abundances of 88 and 51 ppb respectively. A sample of clinopyroxenite yields anomalously high platinum and palladium of 239 and 285 ppb respectively. The platinum group elements have no apparent correlation with the abundances of pathfinder elements suggesting that the platinoids may be contained as discrete platinum-iron or ruthenium-iridium-osmium alloys. Gold anomalies (20 to 214 ppb) show a preference for mafic lithologies at the margins of the ultramafic complex. This suggests that the mafic rocks may have acted as a chemical trap for mineralizing fluids circulating adjacent to the intrusion, and that the gold was derived from the country rocks.

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GEOLOGY AND NOBLE METAL GEOCHEMISTRY OF THE WREDE CREEK ULTRAMAFIC COMPLEX, NORTH-CENTRAL BRITISH COLUMBIA*
(94D/9)

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KEYWORDS: Economic geology, Alaskan-type ultramafic complex, Wrede Creek, structure, geochemistry, platinum group elements

INTRODUCTION

The Wrede Creek ultramafic complex is one of a series of northwesterly trending, Alaskan-type ultramafic bodies in north-central British Columbia, several of which have been described elsewhere (Nixon *et al.*, 1989a, b; 1990, this volume). The Wrede Creek ultramafite, however, is distinguished from those studied to date in the McConnell Creek and Aiken Lake map areas by anomalously high concentrations of platinum. In this respect, it is comparable to the Tulameen ultramafic complex in southern British Columbia where platinum group minerals are primarily associated with chromitites in the dunite-rich core of the complex (St. Louis *et al.*, 1986; Nixon *et al.*, 1989c).

In the past, the region surrounding the Wrede Creek complex has attracted considerable attention on account of its precious and base metal potential. The area experienced its first gold rush in 1899 with the discovery and subsequent exploitation of the McConnell Creek gold placers. Tiny platinum nuggets were reportedly found with the gold, but the placers never became as prominant a source of platinum as the rivers and creeks draining the Tulameen complex (O'Neill and Gunning, 1934; Rublee, 1986).

This report focuses on the results of geological mapping and geochemical sampling of the Wrede Creek complex completed during two weeks in July 1989. In addition, we have incorporated data, published in part by Wong *et al.* (1985), that was generated during an earlier investigation of the geology and economic potential of the ultramafic complex and its associated rocks. The reader is referred to this publication for further details concerning the geochronometry, microprobe phase chemistry and mineralization of the Wrede Creek complex.

LOCATION AND ACCESS

The Wrede Creek complex (56°40′N, 126°08′W) lies within the Ingenika Range of the Omineca Mountains, approximately 400 kilometres north-northwest of Fort St. James, and may be reached by well-travelled dirt road via Manson Creek and Germansen Landing (Figure 3-6-1). The complex is 8 kilometres north-northeast of Johanson Lake,

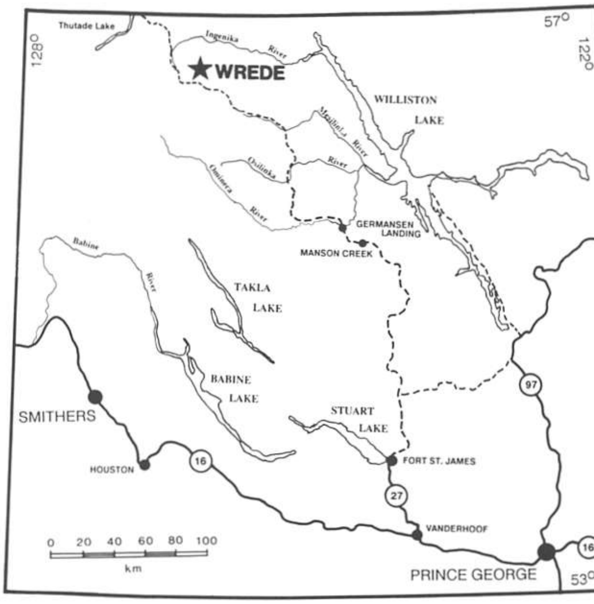

Figure 3-6-1. Location of the Wrede Creek ultramafic complex.

and is named for Wrede Creek which lies about 4 kilometres beyond its northern margin (Figure 3-6-2). The region is serviced via an airstrip at the northern end of the lake by chartered aircraft from Prince George or Smithers, or from the Sturdee airstrip some 100 kilometres to the northwest in the Toodogonne River area. Access from Johanson Lake is by helicopter, or a poorly maintained, four-wheel-drive road leading to the southernmost exposures of ultramafic rocks.

The map area is covered at a scale of 1:50 000 by NTS sheet 94D/9. Aeromagnetic survey maps are available at scales of 1:250 000 (Map 7778G-McConnell Creek) and 1:63 360 (Map 5272G-sheet 94D/9).

PREVIOUS WORK

Reconnaissance mapping (1:250 000) and geologic descriptions of the area were first completed by Lord (1948). More recently, remapping of various parts of the McConnell Creek map sheet was undertaken by Richards (1976a, b), Monger (1977) and Church (1974, 1975). These authors also describe the regional structure and stratigraphy of volcaniclastic and epiclastic rocks of the Upper Triassic Takla

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

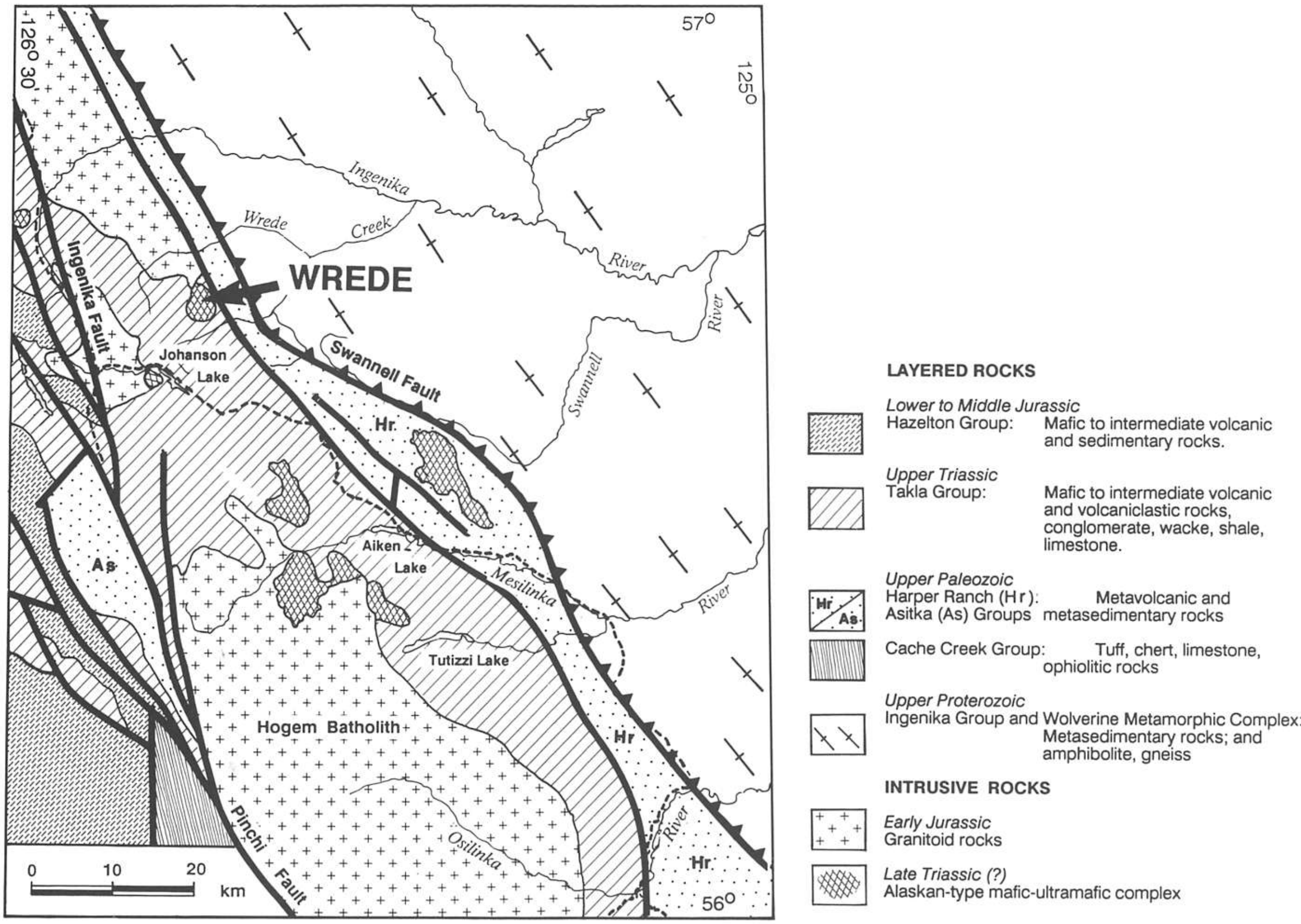

Figure 3-6-2. Regional geologic setting of the Wrede Creek ultramafic complex (modified after Irvine, 1974a; Richards, 1976b; and Monger, 1977).

Group, which hosts many of the Alaskan-type ultramafic complexes in the area (*see* also Bellefontaine and Minehan, 1988). Jurassic granitic plutonism in the region has been summarized by Woodsworth (1976) and Woodsworth *et al.* (in press). Modern studies of the Alaskan-type ultramafic complexes that occur within both the McConnell Creek and Aiken Lake map areas to the east, including the Wrede Creek complex, were made by Irvine (1974a; 1976). Wong *et al.* (1985) and Irvine (1976) presented the most detailed descriptions of the ultramafic complex.

REGIONAL GEOLOGY AND GEOCHRONOMETRY

The Wrede Creek ultramafic complex intrudes volcanic and volcaniclastic rocks of the Upper Triassic Takla Group that forms part of Quesnellia (Figure 3-6-2), one of several tectonostratigraphic terranes that make up the Intermontane Superterrane. To the west, the boundary between Quesnellia and the Stikine Terrane of the Intermontane Belt is tentatively delineated by the Pinchi-Ingenika dextral fault system (Wheeler *et al.*, 1988). Type sections for the Takla Group, as presented by Monger (1977) and Monger and Church (1977), are exposed west of the Ingenika fault in Stikinia, as is one of the Alaskan-type intrusions in this region, the Menard Creek complex (Nixon *et al.*, 1989b). East of the Wrede complex, a high-angle fault separates these supracrustal volcanosedimentary sequences from a tectonic sliver of rocks assigned to the Upper Paleozoic Lay Range assemblage (Richards, 1976a, b; Monger, 1977). This assemblage has recently been correlated with the Harper Ranch Group which forms the basement of Quesnellia in southern British Columbia (Wheeler and McFeely, 1987; Monger *et al.*, in press). Quesnellia is separated from pericratonic rocks of the Ingenika Group by the Swannell fault, an imbricated thrust zone with evidence of southwesterly tectonic transport (Bellefontaine, 1989). Deformation likely occurred in the Mesozoic, during collision of the Intermontane Superterrane with the North American miogeocline (Gabrielse and Yorath, in press).

The Takla Group west of the Ingenika-Pinchi fault boundary is little metamorphosed whereas correlative rocks to the east are characterized by greenschist-grade assemblages (Richards, 1976a, b; Monger, 1977). The predominant structural grain in the region is northwesterly. The main lithologies in the Ingenika Range west of the Wrede Creek complex include mafic to intermediate, plagioclase and augite-phyric volcaniclastic and epiclastic rocks, subaqueous and minor subaerial lava flows, and interbedded black argillites, siltstones, sandstones and minor limestones (Bellefontaine and Minehan, 1988).

Alaskan-type ultramafic complexes in the McConnell Creek and Aiken Lake areas have been considered to be coeval and comagmatic with Late Triassic volcanic rocks of the Takla Group (Irvine, 1974a, 1976). Potassium-argon dating of two hornblende mineral separates from feldspathic pegmatites in the Wrede Creek complex appear to confirm this supposition, yielding Late Triassic isotopic ages of 219 ± 10 (1 σ) and 225 ± 8 Ma (Wong *et al.*, 1985).

Granitoid rocks intrude both the Takla Group and Wrede Creek ultramafic rocks. Potassium-argon dating of hornblende has established a Middle Jurassic age of 172 ± 6 (1 σ) Ma for these intrusions in the Wrede Creek complex (Wong *et al.*,1985). This is compatible with K-Ar dates on hornblende from plutons comprising part of the Hogem batholith to the south (*e.g.* the Duckling Creek syenite dated at 171 ± 6 (1 σ) Ma; Eadie, 1976). The Fleet Peak pluton, a diorite to monzodiorite body northwest of the Wrede Creek complex (Woodsworth, 1976), has yielded K-Ar dates on hornblende and biotite of 144 ± 8 (2 σ) Ma and 156 ± 5 Ma respectively (Wanless *et al.*, 1979). This pluton apparently belongs to the Three Sisters suite of predominantly calc-alkaline intrusions that are well represented in the Stikine arch to the northwest of the project area (Woodsworth *et al.*, in press).

COUNTRY ROCKS: TAKLA GROUP

The Wrede Creek ultramafic complex lies within the predominantly subaqueous eastern facies of the Takla Group (Richards, 1976a, b). Adjacent to the complex, rocks of the Takla Group are predominantly brown to grey weathering, medium grey-green and dark grey augite and augite-plagioclase crystal tuffs, flows and volcanic breccias. These rocks are characterized by euhedral to subhedral black augite crystals up to 1 centimetre in diameter, and white to pale green, variably saussuritized plagioclase laths up to 5 millimetres in length. In thin section, clinopyroxene is observed to be completely pseudomorphed by actinolite, or at least partially altered to this mineral at crystal margins and along cleavages. Plagioclase is partially to completely altered to sericite, epidote and carbonate. Actinolite is also abundant within the fine-grained groundmass and locally takes on a weak preferred orientation. These mineral assemblages indicate a regional metamorphism of upper greenschist grade.

WREDE CREEK ULTRAMAFIC COMPLEX

In common with certain other Alaskan-type intrusions (*e.g.* Duke Island; Irvine, 1974b), the Wrede Creek ultramafic complex exhibits a rude concentric zonation of rock types that is marked by a complete gradation from ultramafic lithologies in the core to mafic phases at the margins (Figure 3-6-3). Dunite in the central part of the body grades outwards through a narrow wehrlitic transition zone into olivine clinopyroxenite and clinopyroxenite, and these units ultimately pass into hornblende-rich gabbroic rocks at the periphery of the complex. Olivine-hornblende and hornblende clinopyroxenites are present, at least locally, between clinopyroxenitic rocks and gabbroic rocks with minor hornblendite.

The complex underlies an area of approximately 10 square kilometres. The dunite core is well exposed along the crest of a northwesterly trending ridge, whereas outcrops of other map units lower down on the flanks of the ridge are much more limited and exposure in valleys is poor. In a region of little outcrop at the eastern margin of the body, the position of the contact between the ultramafite and country rocks has been estimated from the aeromagnetic map. The geometry of the intrusion is poorly constrained, but based on a limited number of short diamond-drill holes at the southern margin of the complex (Wong *et al.*, 1985), the body may represent a high-level intrusive stock.

DUNITE

Dunite (5 square kilometres) forms the dominant lithology of the Wrede Creek complex and is well exposed along a major northwesterly trending ridge at the centre of the map area. On weathered surfaces, dunite is characteristically orange-brown, yellow-orange or buff, and contains disseminated, variably magnetic, euhedral to subhedral chromite crystals (up to 1 millimetre in diameter) that weather with positive relief. The rock is generally medium grained and consists of black, glassy olivine crystals with minor chromite (1 to 5 per cent); fine-grained varieties are dark-grey.

Thin section analysis of dunite reveals an equigranular texture, or rare inequigranular fabric in which small olivine crystals (0.5 to 1 millimetre) comprising up to 80 per cent of the rock are interstitial to, and poikilitically enclosed within, coarse olivine crystals (up to 5 millimetres across). Clinopyroxene was not observed within the dunite except at the gradational contact between dunite and clinopyroxenite (well exposed at Locality 5, Figure 3-6-3). The black colour of the olivine crystals is attributed to an abundance of tiny, opaque, rod-like inclusions which range from 2 to 5 microns in length, which are evenly distributed throughout the crystal and have a preferred orientation which is crystallographically controlled. Attempts to establish the composition of these inclusions by microprobe analysis have been unsuccessful but we believe that they are an exsolution phenomenon related to oxidation during slow cooling of the dunite. The lack of these inclusions within dunite dikes which cut the dunite body suggests the diking postdates this oxidation stage.

Microfractures are prominent along crystal boundaries and within crystals, and have acted as loci for serpentinization. On average, olivine is approximately 5 per cent serpentinized, but the degree of alteration varies widely, ranging from thin envelopes surrounding microfractures, to complete serpentinization, particularly in samples collected near brittle shear zones. Altered areas are composed of antigorite, secondary magnetite, and minor brucite, talc and carbonate.

Tabular zones of bright orange weathering, medium green-grey, carbonate-quartz alteration are common throughout the dunite. Such zones are composed mainly of ankerite and minor magnesite, and have abundant closely spaced, white

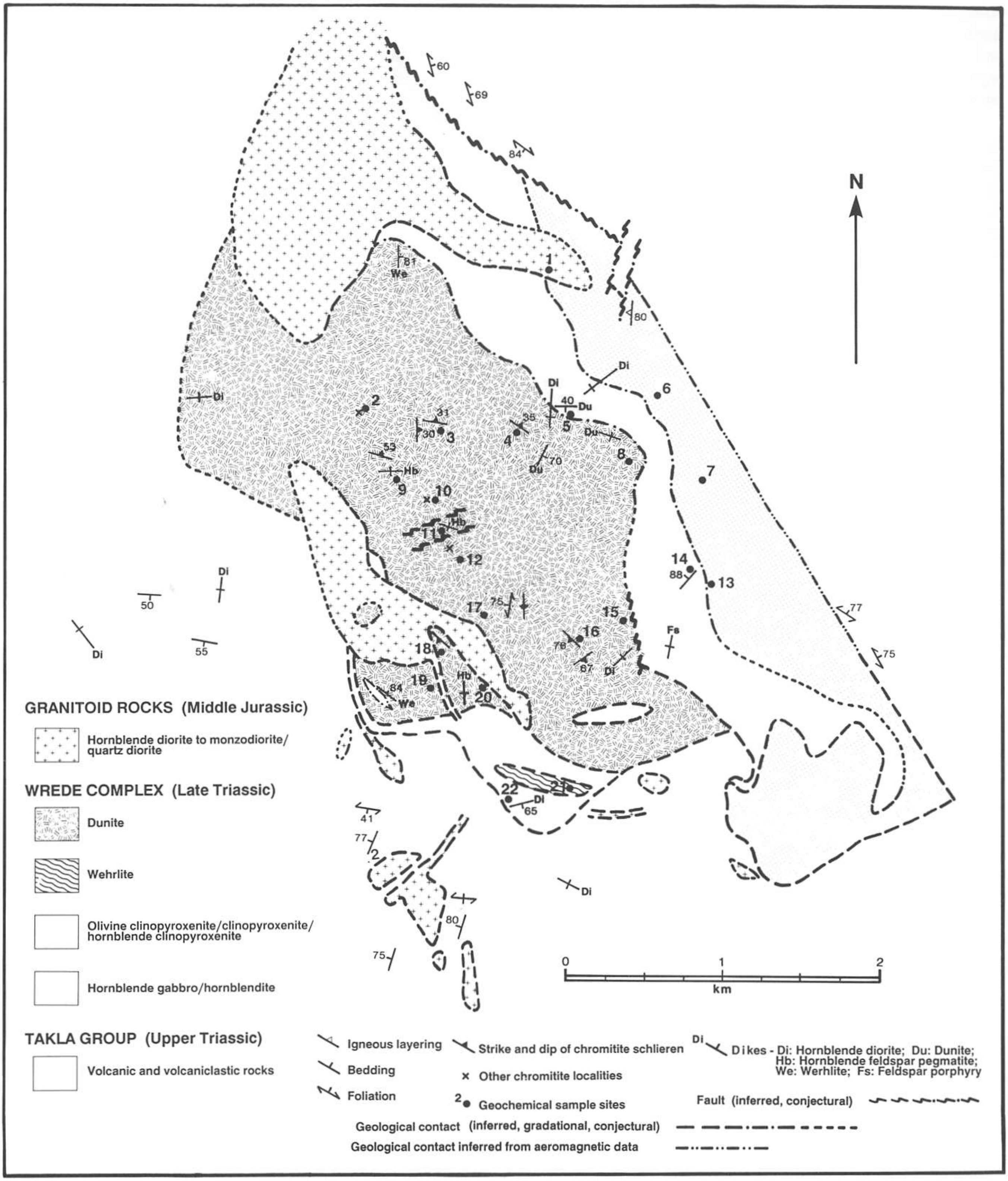

Figure 3-6-3. Generalized geology of the Wrede Creek ultramafic complex. Numbered locations represent geochemical sample sites listed in Table 3-6-1.

chalcedony veins which are commonly folded (Plate 3-6-1). Minor disseminated bornite and pyrite were observed in one of these zones at the southwestern end of the dunite outcrop (Locality 19, Figure 3-6-3). The alteration most likely occurs along faults, which possibly serve as channelways for the same fluids which introduced the sulphides so prevalent at the southern end of the complex (discussed below).

Microprobe analyses of olivines in the dunite indicate forsteritic compositions in the range of Fo_{88} to Fo_{92}. These analyses represent samples collected along an east-west traverse across the dunite core and provide little evidence of compositional zoning with respect to distance from the clinopyroxenitic rocks nearer the margin of the intrusion.

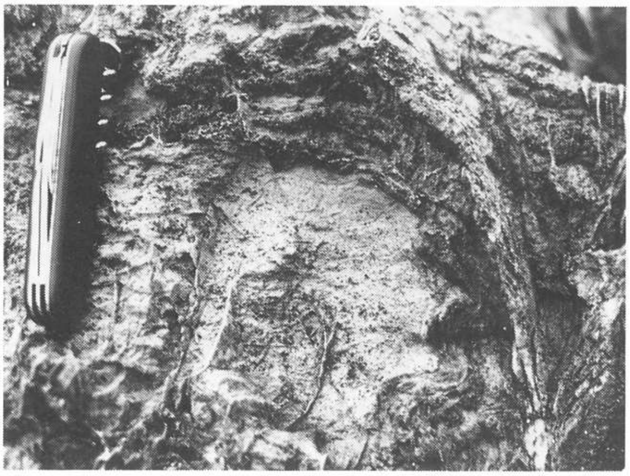

Plate 3-6-1. Quartz-carbonate alteration in dunite showing folded chalcedony veins in outcrops approximately 300 metres south of Locality 5 in Figure 3-6-3.

CHROMITITE

Although chromite is ubiquitous as tiny euhedral crystals in dunite, it is concentrated locally into irregular pods and schlieren. Chromitite schlieren range from 0.1 to 5 centimetres in width and between 5 and 40 centimetres in length; most are less than 1centimetre wide and 15 centimetres long. The schlieren are found as isolated occurrences within the dunite, or more commonly in clusters forming chromitite-rich zones several metres in width (Plate 3-6-2, Locality 2 in Figure 3-6-3). It is common to find schlieren in various orientations within a single outcrop, indicating some degree of remobilization of previously consolidated chromite cumulates. However, in rocks that exhibit a penetrative foliation (*e.g.* Localities 16 and 18) schlieren are usually oriented in the plane of the fabric.

There appears to be no structural control reflected in the spatial distribution of chromitite-rich zones, most of which are separated by broad expanses of chromitite-free dunite. Chromitite schlieren appear to be absent within 200 metres of the dunite/clinopyroxenite contact, but this may be due to more limited outcrop in this part of the complex.

CLINOPYROXENITES

The clinopyroxenite unit includes olivine clinopyroxenite, clinopyroxenite, olivine-hornblende clinopyroxenite, and hornblende clinopyroxenite. These lithologies typically form a complete gradation from olivine-rich phases adjacent to the dunite contact to hornblende-rich phases adjacent to the gabbro-hornblendite contact.

Plate 3-6-2. Zone of chromitite pods and schlieren at Locality 2 in Figure 3-6-3.

Clinopyroxenites form a semicontinuous rim around the dunite core of the complex (Figure 3-6-3). The width of the clinopyroxenite unit varies from 50 metres at the southwestern end, to 900 metres at the northern end of the complex. Along the eastern margin, clinopyroxenites are in gradational contact with clinopyroxene hornblendite and hornblende ± clinopyroxene gabbro. To the south, clinopyroxenites intrude hornfelsed volcanic rocks of the Takla Group.

OLIVINE CLINOPYROXENITE AND CLINOPYROXENITE

Olivine clinopyroxenite and clinopyroxenite are usually coarse grained, and composed of medium brown-green weathering, pale green, cumulus clinopyroxene and brown-weathering, black cumulus to intercumulus olivine. Olivine clinopyroxenite contains an average of approximately 30 per cent olivine, but modal variations range from 10 to 40 per cent. The modal abundance of olivine in clinopyroxenite averages about 5 per cent, but may vary between 0 and 10 per cent.

Olivine clinopyroxenite is most common near the dunite contact where it forms part of the gradation from dunite to clinopyroxenite. Locally, olivine forms up to 50 per cent of the rock, which is more appropriately termed wehrlite. The olivine clinopyroxenite unit varies in width from approximately 2 metres at the eastern margin of the dunite body (approximately 200 metres south of Locality 8 in Figure 3-6-3), to 250 metres at the southwestern edge of the dunite (100 metres south of Locality 20) where it is in intrusive contact with hornfelsed hostrock.

Clinopyroxenite is well exposed in the eastern part of the complex, where it is sandwiched between olivine clinopyroxenite to the west, and hornblende-bearing clinopyroxenite to the east. The clinopyroxenite zone varies in thickness from approximately 100 to 200 metres.

In thin section, equigranular clinopyroxene (0.5 to 6 millimetres in diameter) exhibits cumulus textures, and olivine occurs as cumulus and intercumulus crystals and equigranular crystal clots which range from 0.5 to 1 millimetre across. Olivine is partially to completely serpentinized. Euhedral to subhedral chromite (0.5 to 1 millimetre) makes up less than 1 per cent of the rock.

Microprobe analyses of clinopyroxene from olivine clinopyroxenite yield diopsidic compositions with relatively low alumina (1.5 to 2.6 weight per cent); olivine compositions range from Fo_{83} to Fo_{86}. Clinopyroxenes in clinopyroxenite are also diopsidic with 1 to 3 weight per cent alumina. These mineral compositions fall within the general range of silicate compositions in equivalent rock types in the Tulameen complex (Findlay, 1969; Nixon *et al.*, 1989c).

HORNBLENDE CLINOPYROXENITE AND OLIVINE-HORNBLENDE CLINOPYROXENITE

Hornblende clinopyroxenite is most extensive at the northern end of the complex where it reaches an apparent thickness of approximately 500 metres. In some exposures in the eastern part of the complex (100 metres west of Locality 6, Figure 3-6-3), a complete gradation between clinopyroxenite and hornblende-bearing clinopyroxenite is observed. In this area hornblende-bearing clinopyroxenite averages between 50 and 150 metres in width and grades into gabbro and hornblendite to the east.

Hornblende clinopyroxenite is medium brown weathering, and comprises about 20 to 50 per cent black hornblende and 50 to 80 per cent dark green clinopyroxene. Three main types are observed: a variety containing euhedral, cumulate hornblende crystals (up to 2 centimetres in length) surrounded by smaller grains (2 millimetres) of cumulus to intercumulus clinopyroxene; a coarse-grained variety comprising large crystals (1.5 centimetres) of cumulus clinopyroxene partially enclosed by intercumulus hornblende (1.5 centimetres); an equigranular coarse-grained variant of the second type with large (2 centimetres) interlocking crystals of subhedral cumulus hornblende and clinopyroxene. Locally, plagioclase appears as an intercumulus phase forming up to 5 per cent of the rock. A significant amount of both primary and secondary magnetite has made these hornblende-bearing lithologies very strongly magnetic.

Thin section analysis of hornblende clinopyroxenite reveals fresh, pale brown pleochroic hornblende and unaltered clinopyroxene. Magnetite forms up to 2 per cent of the mode and occurs as small euhedra (0.1 to 1 millimetre) disseminated throughout the rock. Apatite (less than 1 per cent) occurs as an accessory phase.

Olivine within olivine-bearing hornblende clinopyroxenite was recognized only in thin section. Where present [*e.g.* near Locality 13, and in a drillhole at the southern end of the complex (Wong *et al.*, 1985)] it may form up to 10 or 15 per cent of the rock. Typically it occurs as small (0.5 to 1 millimetre) subhedral crystals completely pseudomorphed by amphibole (cummingtonite) and magnetite and poikolitically enclosed by clinopyroxene and hornblende.

HORNBLENDE GABBRO AND HORNBLENDITE

The hornblende gabbro/hornblendite map unit includes rocks that contain variable proportions of hornblende, plagioclase and clinopyroxene, and that locally weather white to black depending on the modal abundance of feldspar. This unit occupies some 3.5 square kilometres at the eastern and southeastern periphery of the complex. Outcrop is sparse, particularly along the eastern margin of the body, and thus contact relationships are rarely seen. It appears, however, that gabbroic rocks are invariably in contact with pyroxenitic rocks toward the core of the complex, and in contact with country rocks externally. Where the contact between gabbroic and pyroxenite rocks is observed (near Localities 6 and 7), it is gradational over a few metres. This contact is typified by a decrease in clinopyroxene and an increase in plagioclase as the gabbroic unit is approached. An intrusive contact between gabbroic rocks and country rocks is observed in a stream-cut approximately 500 metres east of Locality 1. Here, both the main gabbro body and numerous gabbroic dikes intrude country rocks that have been metamorphosed to lower amphibolite grade by contact metamorphism (discussed below). Also at this locality, excellent examples of primary, centimetre-scale, rythmic layering are found within the hornblende gabbro. This texture is formed by modal variations in plagioclase and hornblende which form practically monomineralic layers from 2 millimetres to 2 centimetres thick (Plate 3-6-3).

Gabbroic rocks generally contain 10 to 40 per cent white to pale green, variably saussuritized plagioclase and euhedral to subhedral hornblende with cumulus to intercumulus textures. Locally, dark green cumulate clinopyroxene forms up to 30 per cent of the rock. Hornblendites typically have less than 5 per cent white to pale green feldspar interstitial to large (up to 2 centimetres) cumulus hornblende. In places, these rocks enclose pods of feldspathic clinopyroxene hornblendite with up to 50 per cent dark green clinopyroxene.

In thin section, the gabbroic rocks are seen to be intensely altered. Pale green, pleochroic hornblende is locally altered

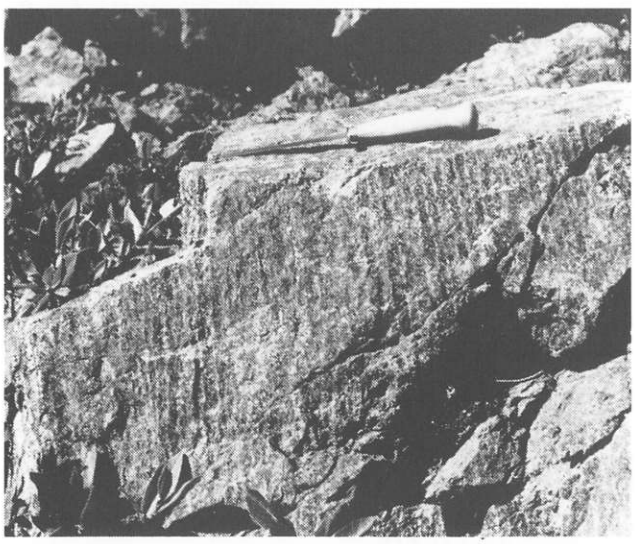

Plate 3-6-3. Magmatic, centimetre-scale layering in hornblende gabbro in a stream cut approximately 500 metres east of Locality 1 in Figure 3-6-3.

to chlorite, clinopyroxenes are partially to completely transformed to uralite, and plagioclase is almost completely saussuritized. Plagioclase relicts with albite twinning are observed rarely. Minor phases include euhedral magnetite (up to 3 per cent) and apatite (1 per cent) with cumulate textures. Epidote is found as euhedral crystals in open-space fillings as well as in granular aggregates forming part of the plagioclase alteration.

MINOR INTRUSIONS

Minor intrusions within the complex include dikes of dunite, wehrlite, olivine clinopyroxenite and hornblende-plagioclase pegmatite. These dikes are mineralogically identical to the main lithologic units of the Wrede Creek complex, which they intrude, and appear to be rooted entirely within the complex.

Ultramafic dikes composed of dunite, wehrlite and olivine clinopyroxenite, averaging approximately 10 centimetres in width, are found throughout the dunite body, but appear to be most common near the dunite-clinopyroxenite contact. Dunite and olivine wehrlite dikes that intrude olivine clinopyroxenite near the dunite-clinopyroxenite contact are particularly well exposed at Locality 5, and locally incorporate xenoliths of clinopyroxenite wallrocks. Also at this locality, olivine clinopyroxenite dikes cut earlier dikes of olivine wehrlite, attesting to a rather complex crystallization history. In areas where these dikes occur in a high concentration (*e.g.* Locality 5), there is a resemblance to the wehrlitic/clinopyroxenitic mixed units described from the Polaris ultramafic complex (Nixon *et al.*, 1990).

In thin section, dunite dikes are seen to be composed of equigranular olivine crystals (0.5 millimetre) which poikilitically enclose smaller grains (less than 0.1 millimetre) of subhedral chromite that locally form up to 5 per cent of the rock. Olivine is variably serpentinized along closely spaced microfractures.

In olivine clinopyroxenite dikes, olivine occurs as glomerocrystic aggregates (0.5 millimetre) and as single crystals poikolitically enclosed by large clinopyroxenes (up to 2 centimetres in diameter). Olivine is typically completely serpenthnized. In olivine clinopyroxenite dikes near the dunite-clinopyroxenite contact at Locality 5, olivine crystals appear to be entirely cumulate in origin, whereas olivine crystals in the host chnopyroxenite have both cumulate and intercumulate textures.

Buff-white weathering, hornblende plagioclase dikes with pegmatitic textures range from 1 to 5 metres wide and appear restricted to the dunite. They are characterized by fresh, euhedral, black hornblende crystals that measure up to 20 centimetres long and form 5 to 80 per cent of the rock. Pale greenish white, variably saussuritized plagioclase fonns the remainder of the rock, together with accessory opaque oxides, apatite and sphene. Contacts between the pegmatite dikes and the dunite are everywhere sharp. As noted earlier, hornblende separates from two of these dikes gave Middle Jurassic K-Ar isotopic dates.

CONTACT AUREOLE

Metamorphism associated with intrusion of the Wrede Creek ultramafic complex is reflected in an amphibolitic contact aureole developed in volcanic rocks of the Takla Group. The aureole is variable in width but is most extensive at the southern end of the complex where it is up to 400 metres wide. Sparse outcrop along the eastern margin of the complex shows some evidence of contact metamorphism, although the aureole does not appear to be as extensive as that to the south. Drillhole data from the southern end of the complex suggest that the contact between country rock and ultramafite dips gently to the south (Wong *et al.*, 1985). The relatively wide metamorphic aureole in this area may therefore be the surface expression of a contact at shallow depth.

In hand sample, hornblende hornfels is dark grey to black and fine grained. White-weathering subhedral feldspar and rare euhedral augite pseudomorphs help to distinguish this rock as a part of the hostrock suite. In thin section, pleochroic green to blue-green hornblende or actinolitic hornblende crystals reach 1 millimetre in length and comprises 50 to 75 per cent of the rock. The matrix comprises fine-grained granoblastic feldspar, acicular actinolite and opaque oxides. Actinolitic amphibole probably formed, in part, at the expense of hornblende as a retrograde assemblage during regional greenschist facies metamorphism.

GRANITOID INTRUSIONS

Quartz monzonite, monzonite, quartz diorite and diorite dikes of Middle to possibly Late Jurassic age intrude both the Wrede Creek complex and its hostrocks. These dikes vary from 2 to 250 metres wide and do not appear to have a preferred orientation.

Granitoid rocks in the area weather a distinctive buff-white and are white on a fresh surfaces. Typically they are medium grained and equigranular, however, a porphyritic texture is developed locally. In hand sample, black hornblende (0 to 35 per cent) occurs as euhedral to subhedral laths up to 5 millimetres long. Plagioclase forms euhedral to subhedral white to pale green crystals. Potassium feldspar and quartz form very fine grained (less than 1 millimetre), white anhedral crystals which are easily overlooked in hand sample.

Thin sections of the granitoid intrusions reveal zoning of plagioclase in some samples, particularly in plagioclase-porphyritic varieties. Plagioclase is moderately to highly sausseritized and hornblende is commonly completely pseudomorphed by chlorite, epidote and calcite, although some relatively fresh varieties were found which have a pale green pleochroism. Where quartz and potassium feldspar are observed, they are unaltered and are interstitial to hornblende and plagioclase. All minerals are overprinted with very fine acicular actinolite which probably formed during upper greenshist grade regional metamorphism.

The southern terminus of the Fleet Peak pluton, which, based on K-Ar dates detailed earlier, is latest Jurassic in age (Wanless *et al.*, 1979), lies approximately 3 kilometres to the north of the Wrede complex. This pluton is the southernmost extension of the predominantly Middle Jurassic Three Sisters plutonic suite (Woodsworth *et al.*, in press). Accordingly, the granitoid intrusions in the vicinity of the Wrede Creek complex are provisionally assigned to the calcalkaline Three Sisters suite.

STRUCTURE AND METAMORPHISM

Regionally, faulting is the dominant deformation mechanism within rocks of the Takla Group (Richards, 1976b; Monger, 1977). Folds have been observed only in the less competent lithologies, so that the attitude of primary layering within the Takla Group is presumably caused by rotation on block faults. Our limited structural observations in the vicinity of the Wrede complex shed no light on this supposition.

Faulting is common within the Wrede Creek complex, its country rocks and at the contact between them. Typically faults are manifest as zones of foliated rock from 1 to 20 metres wide. A well-developed shear foliation parallels the northern margin of the complex and appears offset by crossfaults in the south. The linear eastern boundary of the complex, inferred from aeromagnetic data, may represent a fault boundary. Northwesterly trending faults occur just beyond the western and eastern margins of the complex (Wong *et al.*, 1985). As discussed above, a well-developed contact aureole in hostrocks at the southern perimeter of the complex provides strong evidence for an intrusive contact. The grade of regional metamorphism outside this aureole attained upper greenschist facies, and promoted the formation of a relatively inconspicuous retrograde assemblage within the contact aureole.

ECONOMIC GEOLOGY

The Wrede ultramafic complex is associated with two unrelated types of mineralization. Most extensively explored is a porphyry-style sulphide mineralization at the southern end of the complex which contains anomalous copper and molybdenum. The second type of mineralization, realized in this study, is platinum enrichment within chromitite layers in the ultramafic rocks.

SULPHIDES

Wong *et al.* (1985) have described sulphide mineralization at the southern end of the complex. The sulphides are hosted by dioritic to granitic dikes which cut the complex, and also occur in the Takla Group and some pyroxenitic rocks adjacent to the contact. Mineralization is expressed as disseminations and fracture fillings of pyrite, chalcopyrite, molybdenite and bornite. The mineralized areas are covered by the NIK 1 to 9 claims which were staked in 1976 and explored by BP Minerals Limited. Exploration of the area included 1:5000-scale geologic mapping, geochemical sampling of soils, stream sediments and rock chips, magnetometer and induced polarization surveys, 2550 metres of trenching, 3050 metres of percussion drilling and 3100 metres of diamond drilling. Multi-element analysis of soils (Hoffman and Wong, 1986) delineated zones with anomalous copper and molybdenum. The most extensive sulphide mineralization was found within clinopyroxenite and hornfelsed country rocks at the southern contact of the intrusion. The sulphide mineralization appears to be structurally controlled and is most likely related to Jurassic granitoid plutonism.

Locally, disseminated bornite and pyrite are found in quartz-carbonate alteration zones within the dunite. The planar nature of these zones suggest that they are alteration envelopes surrounding faults. Here again, these zones may be formed by hydrothermal fluids with metal concentrations bearing the signature of nearby granitoid plutonism.

CHROMITE AND PLATINUM

Chromite is restricted to, and locally abundant in, the dunite core of the intrusion where it forms disseminations, pods and schlieren. The main chromitite outcrops, however, lack surface continuity but have not been tested at depth. Geochemical analyses (discussed below) indicate that these chromitites are significantly enriched in platinum. Platiniferous chromitites are now well known in other Alaskan-type bodies in British Columbia, notably the Tulameen complex (*e.g.* Nixon *et al.*, 1989c).

NOBLE METAL GEOCHEMISTRY

Analytical results for gold, platinum, palladium and rhodium in 29 lithogeochemical samples of the Wrede Creek ultramafic complex are presented in Table 3-6-1. Sample localities are shown in Figure 3-6-3. All analyses were done by inductively coupled plasma emission spectrometry at Acme Analytical Laboratories, Vancouver. Accuracy was checked by in-house standards, and analytical precision (and any nugget effect) monitored by hidden duplicates and internal standards. The noble metals were preconcentrated by fire assay from 30 gram aliquots of 200 grams of rock powder (-200 mesh).

Chromitite horizons in the core of the Wrede Creek complex are markedly enriched in platinum. Five samples of relatively high-grade chromitite ran between 120 and 2400 ppb platinum. Some of these samples also have significant abundances of rhodium; all are characterized by high platinum:palladium ratios, and one (Locality 10, Figure 3-6-3) contains some gold. Interestingly, the dunite samples, even where collected adjacent to chromitite layers, are low in platinum group elements, although two specimens of dunite (Localities 3 and 12) have anomalous gold (60 to 80 ppb). In general, pyroxenitic rocks contain low abundances of platinum group elements, hornblende-bearing varieties exhibit the lowest platinum:palladium ratios, and hornblende-plagioclase pegmatites are low in the noble metals. The highest gold abundance occurs in a gabbro near the northeastern margin of the complex (Locality 6) but another sample from the same locality is markedly less enriched (Table 3-6-1).

Apart from the fact that the platiniferous chromitites are confined to the dunite core of the intrusion, they appear to have no systematic spatial distribution within the complex. Their economic significance is obviously highly dependent upon the concentration of chromitite schlieren which, to date, remains to be more thoroughly tested.

SUMMARY AND CONCLUSIONS

The Late Triassic Wrede Creek ultramafic complex fits well into the Alaskan-type classification. Cumulate textures, igneous layering, rude concentric zonation, and gradational contacts between ultramafic and mafic lithologies are consistent with the pattern of crystallization of a differentiating primitive magma. This complex is one of only two Alaskan-

TABLE 3-6-1
NOBLE METAL GEOCHEMISTRY OF THE WREDE CREEK ULTRAMAFIC COMPLEX

Locality	Sample	Rock type	Pt	Pd	Rh	Au
2	GN-89-6006-1	Chromitite	248	<2	28	3
17	GN-89-6026	Chromitite	125	<2	6	2
16	GN-89-7027A	Chromitite	2002	5	17	4
10	GN-89-8000A	Chromitite	2388	12	72	29
3	GN-89-8002B	Chromitite	123	<2	5	<1
2	GN-89-6006-2	Dunite within chromite-rich zone	<1	<2	<2	<1
4	GN-89-6017A	Dunite within chromite-rich zone	<1	<2	<2	3
16	GN-89-7027B	Dunite within chromite-rich zone	19	<2	6	<1
10	GN-89-8000B	Dunite within chromite-rich zone	11	<2	<2	3
3	GN-89-8002A	Dunite within chromite-rich zone	14	<2	<2	63
12	GN-89-7005	Dunite	2	<2	<2	80
5	GN-89-6008-1	Dunite dike	6	<2	<2	4
19	GN-89-6024	Carbonatized dunite	5	<2	<2	8
21	G-89-8020	Wehrlite	31	<2	<2	<1
5	GN-89-6008-2	Olivine clinopyroxenite	3	<2	<2	7
8	GN-89-8011	Olivine clinopyroxenite	26	<2	<2	2
15	GN-89-9030	Olivine clinopyroxenite	30	<2	<2	5
5	GN-89-6008-3	Olivine clinopyroxenite dike	9	<2	<2	<1
22	GN-89-8018B	Clinopyroxenite	15	44	<2	2
14	GN-89-9026	Hornblende clinopyroxenite	9	<2	<2	23
6	GN-89-7007A	Clinopyroxene-hornblende gabbro	4	3	<2	9
6	GN-89-7007B	Clinopyroxene-hornblende gabbro	15	12	<2	195
13	GN-89-7032	Clinopyroxene-hornblende gabbro	8	10	<2	5
7	GN-89-7011	Hornblende gabbro	5	5	<2	10
1	GN-89-6023	Hornblendite	8	11	<2	7
9	GN-89-6004	Hornblende pegmatite dike	<1	<2	<2	7
11	GN-89-7001	Hornblende pegmatite dike	<1	<2	<2	26
18	GN-89-7041Z	Hornblende pegmatite dike	<1	<2	<2	23
20	GN-89-7043Z	Hornblende pegmatite dike	<1	<2	<2	4

Detection limits are 1 ppb for Pt and Au; 2 ppb for Pd and Rh. Sample localities are shown on Figure 3-6-3.

type bodies in the region, the other being the Polaris complex (Nixon *et al.*, 1990), that have a well-documented intrusive relationship with their hostrocks. The external geometry of the Wrede Creek complex is poorly constrained, but it may represent a stock-like intrusion. At the contact, Takla Group volcanic and volcaniclastic rocks have been hornfelsed to lower amphibolite grade, however, the contact effects have been largely overprinted by upper greenschist facies regional metamorphism.

Economic potential of the Wrede Creek complex is under explored. Enrichment of platinum group elements in chromitite pods and schlieren is encouraging and should warrant further investigation in order to determine the extent of these chromitite-rich zones. Of potentially added interest is the porphyry-style copper-molybdenum mineralization in the southern part of the complex. Renewed interest in base metal exploration may spur further exploration here.

ACKNOWLEDGMENTS

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NOTES

GEOLOGY AND NOBLE METAL GEOCHEMISTRY OF THE JOHANSON LAKE MAFIC-ULTRAMAFIC COMPLEX, NORTH-CENTRAL BRITISH COLUMBIA*
(94D/9)

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KEYWORDS: Economic geology, Alaskan-type intrusion, Johanson Lake, mafic-ultramafic complex, structure, geochemistry, platinum group elements.

INTRODUCTION

The Johanson Lake mafic-ultramafic complex is the smallest of three Alaskan-type bodies in north-central British Columbia that were investigated during the 1989 field season. It is distinguished from the Wrede Creek and Polaris complexes (Hammack *et al.*, 1990; Nixon *et al.*, 1990) by a lack of olivine-rich ultramafic lithologies (dunite, wehrlite and olivine clinopyroxenite) and characterized by predominantly amphibole-bearing clinopyroxenites and gabbros. The gabbroic rocks appear to be volumetrically dominant and contain spectacular examples of comb layering as well as the more common centimetre-scale layering observed in the feldspathic phases of other complexes. The size and nature of the Johanson Lake body are traits shared by the Menard Creek complex situated 25 kilometres to the northwest. The Menard complex, however, is distinguished by clinopyroxene gabbros containing relatively high proportions of magnetite (Nixon *et al.*, 1989).

This report summarizes the results of fieldwork conducted during July, 1989, and presents geochemical analyses for platinum group elements and gold on representative lithological samples. The project area is covered at a scale of 1:250 000 by the McConnell Creek map sheet (94D) and 1:50 000 map sheet 94D/9. Aeromagnetic survey maps are available in the smaller scale (Map 7778G–McConnell Creek) and at a scale of 1:63 360 (Map 5272G–94D/9).

LOCATION AND ACCESS

The Johanson Lake mafic-ultramafic complex (56° 34.5′N, 126°13′W) is situated in the Omineca Mountains approximately 2 kilometres southwest of Johanson Lake, for which the complex is named (Figures 3-7-1 and 3-7-2). Access to the area is by a well-travelled dirt road stretching some 400 kilometres north from Fort St. James via Manson Creek and Germansen Landing. An airstrip at the northern end of Johanson Lake is in good repair and suitable for light aircraft. The complex is situated entirely above treeline and excellent exposures are to be found in cirque headwalls and at the crests of ridges between altitudes of 1900 and 2300 metres. Talus aprons and glacial till blanket the lower slopes and valley floors.

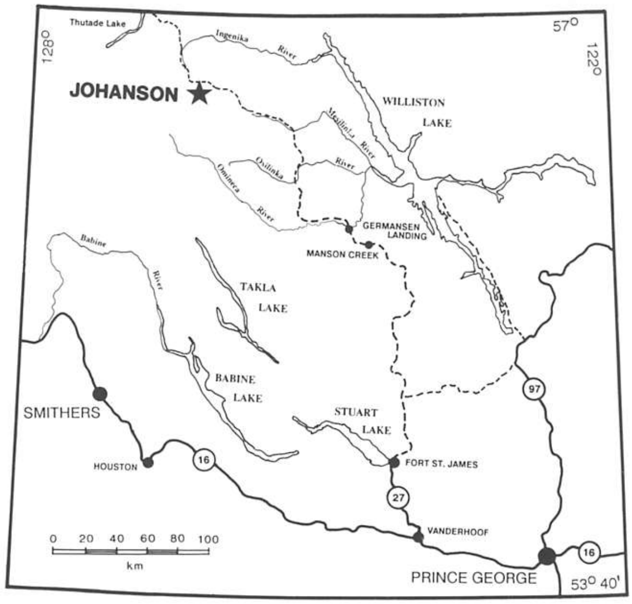

Figure 3-7-1. Location map of the Johanson Lake mafic-ultramafic complex.

PREVIOUS WORK

Geologic reconnaissance of the McConnell Creek map area was first completed by Lord (1948) who also described many of the mineral prospects in the region. Much of the map sheet was later revised by Richards (1976a, b), Monger (1977) and Church (1974, 1975). The latter authors focused on the stratigraphy and structure of the Late Triassic Takla Group which hosts the majority of the Alaskan-type complexes in the region. More recently, Bellefontaine and Minehan (1988) and Minehan (1989) published the results of geologic studies of the Takla Group in the southwestern part of the Ingenika Range approximately 8 kilometres north of Johanson Lake.

The most detailed investigations of Alaskan-type complexes in the region, including Johanson Lake, have been made by Irvine (1974, 1976). Granitoid intrusions of predominantly Jurassic age in the area have been documented by Woodsworth (1976) and Woodsworth *et al.* (in press).

GEOLOGIC SETTING

The Johanson Lake mafic-ultramafic complex lies within the Quesnel tectonostratigraphic terrane, a volcanic arc

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

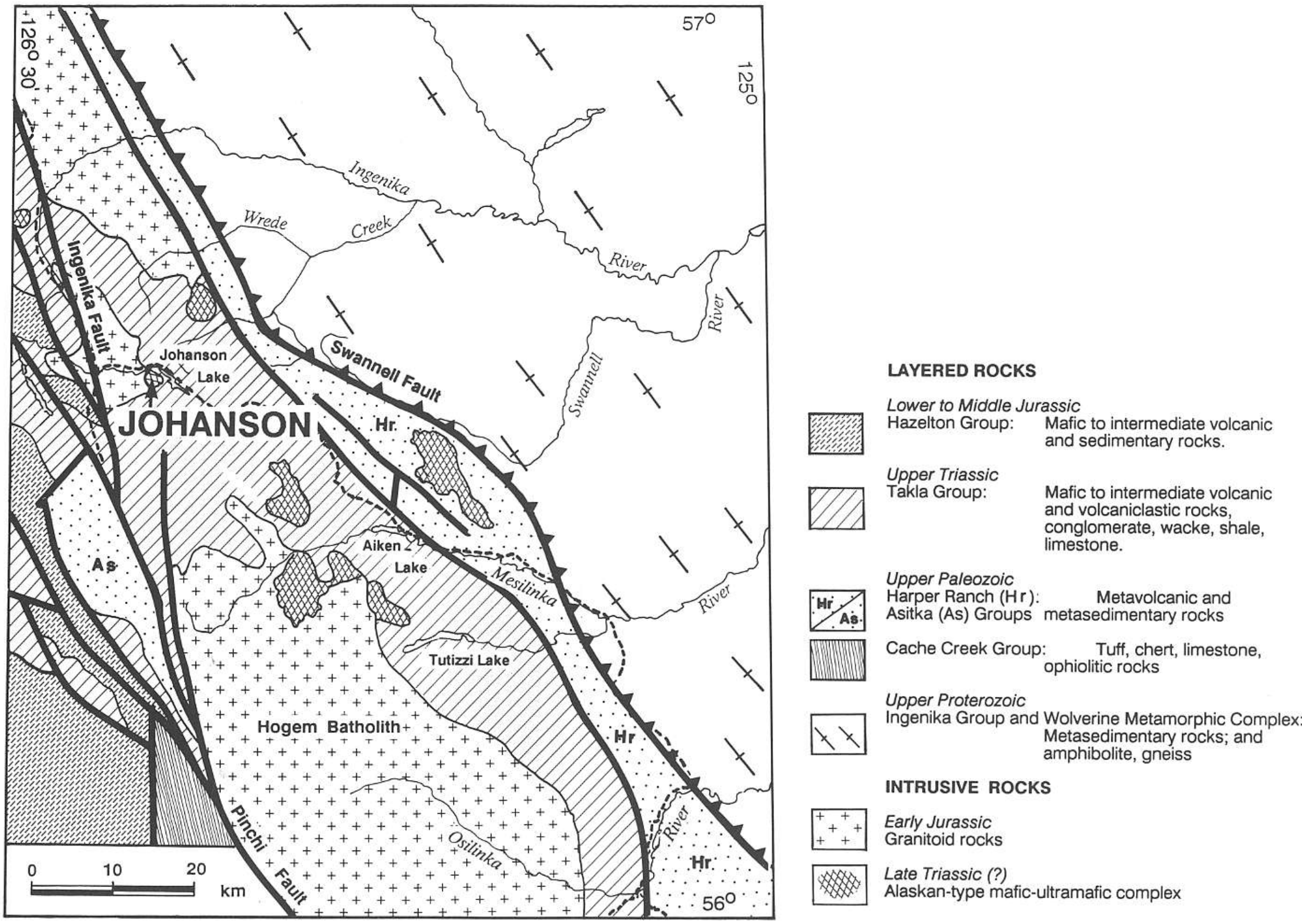

Figure 3-7-2. Geologic setting of the Johanson Lake mafic-ultramafic complex (modified after Irvine, 1974; Monger, 1977; and Richards, 1976b).

assemblage that amalgamated with other allochthonous terranes of the Intermontane Belt before being accreted to the North American craton in the Mesozoic (Wheeler and McFeely, 1987). Quesnellia is bounded on the west by Stikinia along the line of the Pinchi-Ingenika dextral fault system (Figure 3-7-2; Gabrielse, 1985; Wheeler *et al.*, 1988). Its eastern boundary is marked by the Swannell fault, a southwesterly directed thrust zone which superimposes Upper Proterozoic pericratonic rocks of the Ingenika Group on Quesnellia (Bellefontaine, 1989). Thrusting occurred in the Mesozoic during collision of the Intermontane Superterrane with the North American miogeocline (Gabrielse and Yorath, in press).

The Alaskan-type complexes are considered to be comagmatic and coeval with arc-related augite-phyric lavas, pyroclastic and epiclastic rocks of the Upper Triassic Takla Group (Irvine, 1976). West of the Ingenika-Pinchi fault system, Takla Group lithologies reach prehnite-pumpellyite grade whereas to the east these rocks have been metamorphosed to the greenschist facies (Richards, 1976b; Monger, 1977).

The Johanson Lake complex is hosted by the eastern facies of the Takla Group, an undifferentiated package of predominantly greenschist-grade, subaqueous mafic to intermediate volcaniclastic rocks interbedded with minor sedimentary material (Richards, 1976b; Monger, 1977; Bellefontaine and Minehan, 1988). These rocks are intruded by granitoid plutons of predominantly Jurassic age, the largest and possibly longest-lived of which is the composite Hogem batholith (Figure 3-7-2; Woodsworth *et al.*, in press). The prevalent regional structures are represented by northwesterly trending, high-angle brittle faults and shear zones.

COUNTRY ROCKS: TAKLA GROUP

Upper Triassic hostrocks of the Takla Group are well exposed in northeasterly trending ridges at the northern and southeastern margins of the complex, and farther south where they form a northwesterly trending ridge (Figure 3-7-3). At the faulted northern margin of the complex, the Takla Group is composed of metavolcanic and minor metasedimentary (epiclastic?) strata that dip to the north. The predominant lithologies comprise grey-green, well-cleaved, plagioclase-actinolite schists and massive, mafic to intermediate flows that vary from aphanitic to plagioclase-augite porphyries. In thin section, the latter rocks are seen to contain euhedral to subhedral phenocrysts or glomerocrysts (less

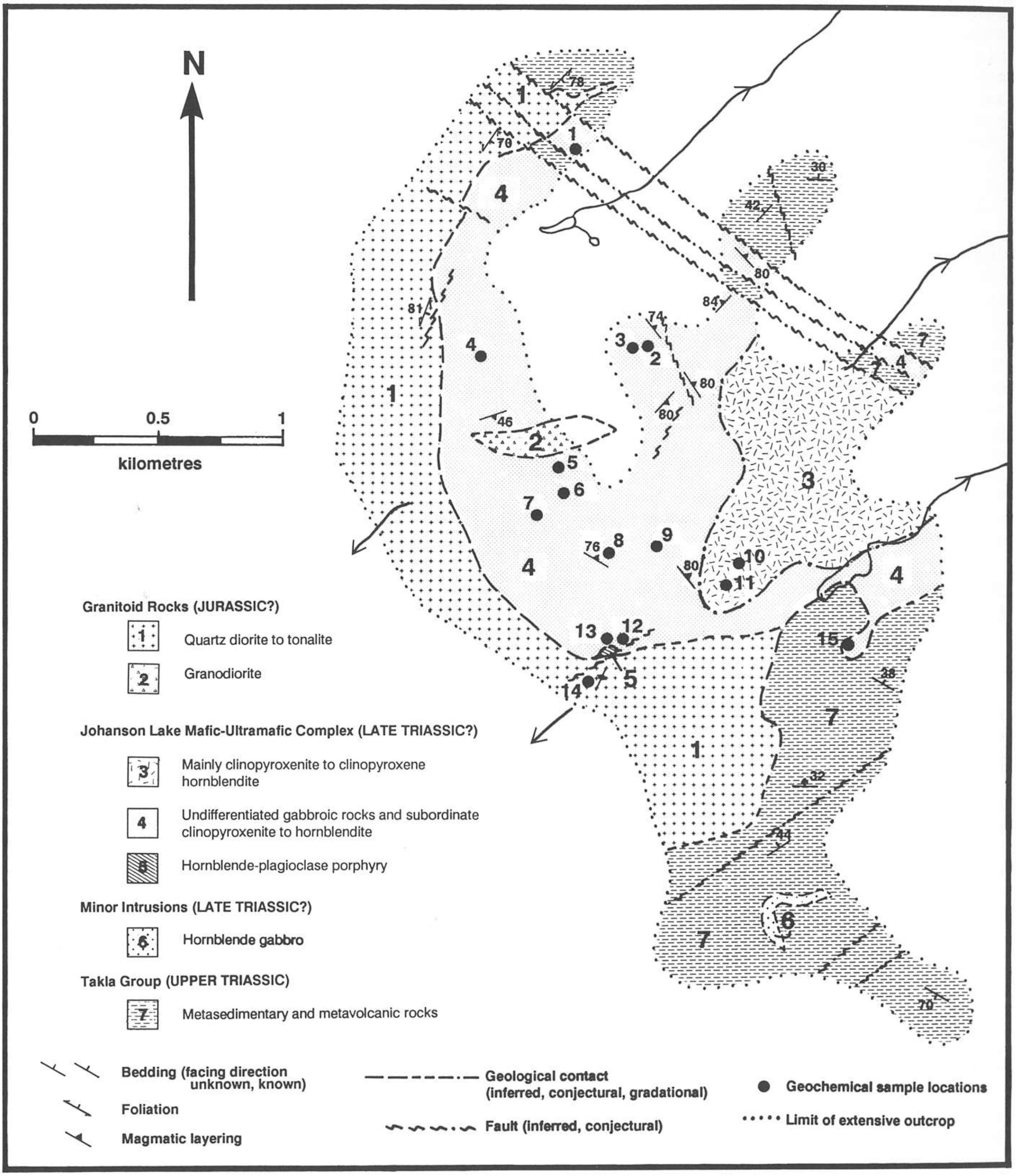

Figure 3-7-3. Generalized geology of the Johanson Lake mafic-ultramafic complex.

than 4 millimetres across) of lamellar-twinned plagioclase and actinolite pseudomorphs after augite set in a granoblastic matrix of plagioclase and actinolite (up to 50 per cent of the matrix). Textures in aphanitic flows are also granoblastic and, locally, actinolite is pseudomorphous after rare mafic phenocrysts (up to 2 millimetres in length). Relict igneous plagioclase in these rocks commonly exhibits subgrain boundaries and shows only incipient alteration to clay minerals.

In the southern part of the map area, the Takla Group contains dark grey to rusty weathering, hornblende-augite-plagioclase-phyric lavas, massive crystal and crystal-lithic tuffs, volcanic breccias, altered amygdaloidal mafic flows, and thickly bedded, greenish grey silicified siltstones and mudstones. These rocks have similar textures and metamorphic mineral assemblages to lithologies to the north. Graded bedding has been observed locally and suggests that the rocks are the right way up. The proportion of rusty weathering outcrop varies according to the abundance of pyrite, which locally reaches 10 per cent of the rock, and occurs as disseminations and in quartz veins. On the whole, the hostrocks of the Johanson Lake complex resemble Takla Group volcanic and volcaniclastic sequences north of Johanson Lake described by Bellefontaine and Minehan (1988) and Minehan (1989).

JOHANSON LAKE MAFIC-ULTRAMAFIC COMPLEX

The Johanson Lake mafic-ultramafic complex underlies an area of approximately 4 square kilometres (Figure 3-7-3). The major areas of outcrop comprise a prominent unnamed peak (2327 metres) in the southern part of the complex and several northeasterly to northerly trending ridges of rugged to gentle relief.

Previous work by Irvine (1976) established the predominantly gabbroic nature of the complex but showed a large unit of olivine clinopyroxenite occupying the southeastern portion of the body (Irvine, ibid., Figure 15.3J). This is an error since the area in question is underlain mostly by hornblende clinopyroxenite and clinopyroxene hornblendite with minor clinopyroxenite and melanocratic gabbro.

At the chosen scale of mapping, we have been able to subdivide the complex into two main lithologic units: clinopyroxene and hornblende-rich ultramafic rocks with minor melanocratic gabbros; and hornblende ± clinopyroxene gabbroic rocks with minor interlayered ultramafic lithologies. The latter map unit appears to be the most voluminous.

Intrusive contacts with the Takla Group are exposed at the southeastern margin of the complex. Internally, contacts between the major map units are usually sharply gradational, as are the contacts of interlayered minor rock types within each of the major units.

CLINOPYROXENITE

Small outcrops of medium grey-green, medium to coarse-grained clinopyroxenite occur at the northwestern extremity of the region underlain by mainly ultramafic lithologies and in gabbroic rocks at Locality 3 in Figure 3-7-3. The clinopyroxenites contain pale brown weathering, serpentinized olivine grains (2 to 5 per cent) distributed evenly throughout the rock and rare pods of olivine wehrlite to dunite up to 1 metre in length. These olivine-rich pods are elongate and irregular, and commonly exhibit a pronounced internal foliation in contrast to their massive pyroxenitic hostrocks. In thin section, subhedral to anhedral clinopyroxenes (up to 1 centimetre across) form an interlocking mosaic containing subhedral cumulus olivine crystals (5 to 10 per cent) up to 8 millimetres in diameter, and pale green, pleochroic, intercumulus hornblende (5 to 10 per cent). Olivines have been almost completely replaced by serpentine, magnetite and secondary amphibole (tremolite-actinolite). These olivine and hornblende-bearing clinopyroxenites represent the most primitive lithologies in the complex.

HORNBLENDE CLINOPYROXENITE AND CLINOPYROXENE HORNBLENDITE

A complete gradation exists among clinopyroxenite, hornblende clinopyroxenite (less than 50 per cent hornblende) and clinopyroxene hornblendite (less than 50 per cent clinopyroxene) in the ultramafic lithologies in the southeastern part of the complex. These rocks are medium to coarse grained and weather medium grey-green to dark greenish grey depending on the clinopyroxene:amphibole ratio. In thin section, hornblende crystals (up to 3.5 centimetres) poikilitically enclose clinopyroxene and appear to be replacing corroded pyroxene relicts. Iron-titanium oxides, largely magnetite, occur in small amounts (less than 5 per cent). Clinopyroxenes have been partly replaced by actinolite-tremolite, and other secondary minerals include minor calcite and epidote. These ultramafic lithologies grade through melanocratic gabbros into surrounding gabbroic rocks.

GABBROIC ROCKS

The gabbroic map unit in Figure 3-7-3 contains significant proportions of interlayered hornblende clinopyroxenites, clinopyroxene hornblendites, hornblendites and their feldspathic equivalents. Gabbroic pods are quite common within these ultramafic lithologies. The dominant rock type is medium-grained to pegmatitic hornblende gabbro or diorite that weathers pale to dark grey or grey-green, and locally contains rusty zones rich in pyrite (2 to 3 per cent). Hornblende-clinopyroxene gabbros occur locally and contain as little as 20 per cent amphibole and up to 40 per cent clinopyroxene. A complete gradation exists among these rock types.

The gabbroic unit is characterized by two types of layering. Medium-grained equigranular gabbroic rocks may exhibit centimetre-scale modal layering comprising alternating hornblende ± clinopyroxene and plagioclase-rich horizons (Plate 3-7-1). In many places, pegmatitic zones have developed spectacular comb layering defined by acicular hornblende crystals up to 20 centimetres in length (Plates 3-7-2 and 3-7-3). Almost invariably, amphibole crystals are preferentially oriented at high angles (70°–80°) to the trend of the layering. These comb-textured layers generally alternate with more equigranular, feldspathic layers (Plate 3-7-2). The boundaries between layers are curvilinear and sharply transitional, and the terminations of large prismatic hornblendes locally penetrate the adjacent layer (Plate 3-7-3). These features suggest that comb layering developed by *in situ* crystal growth under conditions where silicate melts became periodically supersaturated in amphibole.

In thin section, the gabbroic rocks contain subhedral, brown to deep green pleochroic hornblende (20 to 50 per cent), subhedral to anhedral plagioclase (30 to 60 per cent), euhedral to subhedral clinopyroxene (0 to 40 per cent), and accessory iron-titanium oxides (less than 5 per cent), apatite and minor sphene. Plagioclase is extensively saussuritized and generally occurs as intercumulus and cumulus material. Clinopyroxene invariably forms cumulus grains which are commonly partially altered to actinolite. Hornblende is generally a cumulus mineral but forms large (up to 2 centimetres)

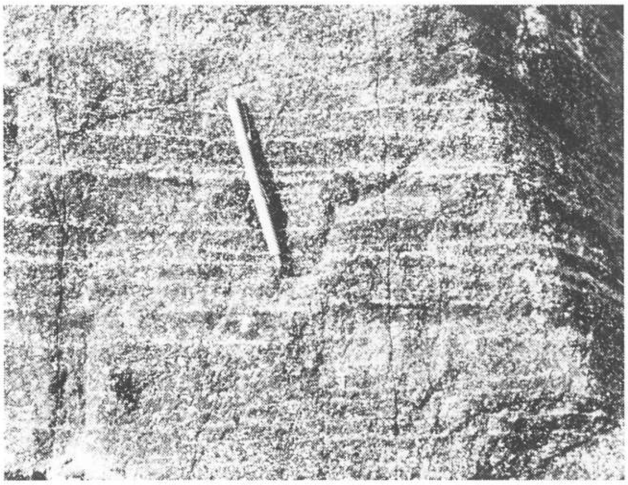

Plate 3-7-1. Centimetre-scale layering formed by modal variations in plagioclase and amphibole in hornblende gabbro. Magnet is 11 centimetres long.

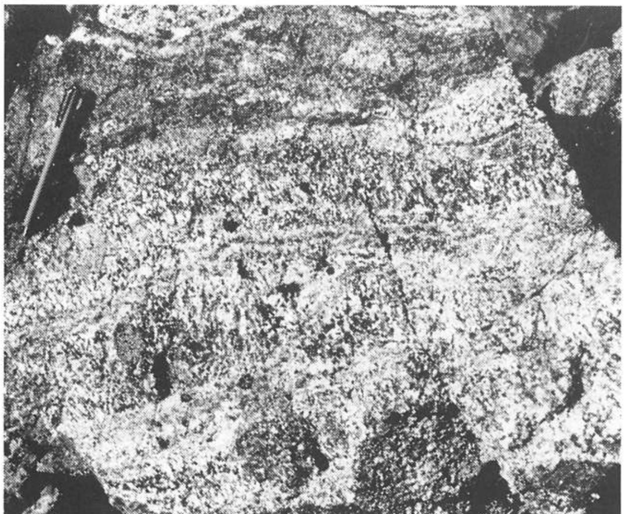

Plate 3-7-2. Comb layering in hornblende gabbro formed by acicular amphibole crystals oriented nearly perpendicular to more equigranular, centimetre-scale layers.

Plate 3-7-3. Well-developed comb layering in coarse-grained to pegmatitic hornblende gabbro. Note penetration of adjacent feldspathic layer by large amphibole crystal to right of magnet.

poikilitic intercumulus crystals in a rare variety of hornblende-megacrystic gabbro. Magnetite, apatite and sphene form euhedral cumulus crystals up to 1 millimetre in length. Quartz may occur in minor amounts (less than 5 per cent) in the groundmass of hornblende gabbros and appears to be largely primary.

Hornblende-Plagioclase Porphyry

A medium grey, fine to medium-grained, porphyritic dioritic rock is exposed at the southern margin of the complex (Figure 3-7-3). The rock contains subequant saussuritized plagioclase (40 per cent), acicular hornblende crystals partly altered to biotite and actinolite, and minor quartz (5 per cent). Locally the rock is sheared and enriched in chlorite and epidote, and cut by quartz veins. Intrusive contacts with coarse-grained gabbroic rocks are sharp and the porphyry contains irregular gabbro xenoliths (cognate?). Thin (1 to 2 centimetres) feldspathic veinlets cut the porphyry and adjacent gabbroic rocks and may represent more differentiated residual liquids derived from within the complex. The porphyry is distinguished from the adjacent quartz diorite/tonalite pluton by its more melanocratic character, and appears to represent a late marginal phase of the Johanson Lake complex.

Minor Intrusives

Narrow dikes of hornblende-rich gabbro or diorite, hornblende pegmatite, and fine-grained felsite cut mafic and ultramafic lithologies of the Johanson Lake complex. Hornblende-plagioclase pegmatite dikes up to 3 metres wide cut clinopyroxene hornblendites interlayered with gabbroic rocks. Prismatic hornblende crystals (up to 15 centimetres long) are commonly arranged haphazardly in the centre of the intrusion but may lie parallel to wallrock contacts near the margins of the dike. These rocks are compositionally identical to the comb-layered pegmatite zones of the gabbroic unit.

Dark grey-green melanocratic microgabbro or microdiorite dikes less than 1 metre wide are observed locally. They consist almost entirely of saussuritized plagioclase and hornblende variably altered to actinolite. Most textures are equigranular although a few dikes contain larger crystals of hornblende (up to 2 millimetres). These dikes are characterized by a conspicuous alignment of hornblende prisms parallel to their margins.

White to pink-weathering, aphanitic felsite dikes up to 0.5 metre wide cut gabbroic and ultramafic rocks alike. Mafic minerals form less than 5 per cent of the rock and have been replaced by epidote, chlorite and carbonate. The origin of these rocks is uncertain but they bear a strong resemblance to rather widespread feldspathic veinlets that represent leucocratic differentiates of the gabbroic rocks.

A dike-like body of grey-green, medium-grained hornblende gabbro or diorite intrudes volcanic rocks of the Takla Group south of the Johanson Lake complex (Figure 3-7-3). The rock contains euhedral prismatic hornblende (50 per cent) and subequant plagioclase laths up to 2 millimetres long set in a fine-grained recrystallized feldspathic groundmass. Amphibole crystals define a pronounced flow fabric subparallel to the contacts of the intrusion. Plagioclase is saussuritized and hornblende partly replaced by actinolite. Euhedral to subhedral magnetite and apatite occur as accessory minerals, and secondary pyrite (1 to 2 per cent) is disseminated throughout the rock. This intrusion is probably coeval, and may be cogenetic, with hornblende-rich gabbroic rocks of the Johanson Lake complex.

GRANITOID ROCKS

A large body of quartz diorite to tonalite delineates much of the western and southern margins of the Johanson Lake complex. Intrusive contacts with the Takla Group are well exposed in the northern and southern parts of the map area, but contact relationships with gabbroic rocks are not clear. We suspect that the Johanson Lake complex is older and tentatively consider the quartz diorite/tonalite to be post-Late Triassic and pre-Late Jurassic in age.

The quartz diorite/tonalite unit is a pale grey weathering, massive, medium-grained subequigranular rock containing variable proportions of anhedral quartz (10 to 30 per cent) with subgrain mosaics, euhedral to subhedral plagioclase (45 to 55 per cent), dark green pleochroic hornblende (10 to 20 per cent), minor biotite and iron-titanium oxides (less than 5 per cent). Hornblende crystals are largely replaced by actinolite at their rims and some biotite appears to be secondary. Plagioclase is partly saussuritized and disseminated pyrite is common near fault zones.

A small body of pale pinkish grey weathering, medium-grained granodiorite intrudes the western part of the complex. It contains euhedral to subhedral, partly saussuritized plagioclase (50 per cent) up to 5 millimetres in length; anhedral quartz crystals (30 per cent) up to 2 millimetres across; subhedral to anhedral potassium feldspar (20 per cent) up to 2 millimetres in length; and minor biotite and trace amounts of hornblende. The age of the granodiorite is uncertain and it has been tentatively assigned to the Jurassic.

STRUCTURE AND METAMORPHISM

The lack of distinctive marker horizons in the eastern facies of the Takla Group hampers interpretations concerning the mechanism of deformation in the area. The attitude of bedding could have resulted from folding or rotation by faulting. Northwesterly trending, high-angle fault zones bounding the northern margin of the mafic-ultramafic complex have incorporated metavolcanic hostrocks as a thin fault slice within the complex.

The grade of metamorphism throughout the map area appears to have reached middle to upper greenschist facies. A sample of crystal tuff in the Takla Group near the fault contact with clinopyroxene hornblendite at the northeastern margin of the complex has granoblastic texture and a lower amphibolite (hornblende + plagioclase ± quartz) grade mineral assemblage. Other volcanic rocks farther northwest along the same contact exhibit uppermost greenschist grade assemblages. Faulting and retrograde regional metamorphism may, therefore, have obscured a weak, and as yet poorly defined, metamorphic aureole of lowermost amphibolite grade at the margins of the complex.

GEOCHEMISTRY AND MINERAL POTENTIAL

Analytical results for platinum, palladium, rhodium and gold in 15 lithogeochemical samples of the Johanson Lake mafic-ultramafic complex are presented in Table 3-7-1. Sample localities are shown in Figure 3-7-3. All analyses were conducted by inductively coupled plasma emission spectrometry at Acme Analytical Laboratories, Vancouver. Accuracy was checked by in-house standards, and analytical precision (and any nugget effect) was monitored by hidden duplicates and internal standards. The noble metals were preconcentrated by fire assay from 30-gram splits of 200 grams of rock powder (–200 mesh).

The highest abundances of platinum and palladium, 41 and 88 ppb respectively, are found in hornblende-plagioclase pegmatites within the gabbroic sequence. The tenor of gold peaks at 41 ppb in the gabbros and rhodium is below the detection limit in all samples. In general, palladium abundances are higher in gabbroic than pyroxenitic lithologies and the noble metals are uniformly low in quartz diorites. Unlike other Alaskan-type complexes studied to date, platinum shows no systematic depletion from ultramafic to gabbroic rock types. This may be due to the relatively small number of samples analyzed or the limited range of ultramafic lithologies represented. Another notable feature is the relatively high average abundance of gold in the Johanson Lake mafic-ultramafic complex as a whole.

The distribution of noble metals bears no apparent relationship to the presence of sulphides (disseminated pyrite) or proximity to fault zones. The potential for economic concentrations of precious metals appears to be low.

SUMMARY AND CONCLUSIONS

The Johanson Lake mafic-ultramafic complex is a relatively small (4 square kilometres) Alaskan-type intrusion hosted by augite-plagioclase-phyric mafic to intermediate

TABLE 3-7-1
NOBLE METAL GEOCHEMISTRY OF THE JOHANSON LAKE MAFIC-ULTRAMAFIC COMPLEX

Locality	Sample	Rock Type	Pt	Pd (ppb)	Rh	Au
3	GN-89-9031	Ol-Hb-bearing clinopyroxenite	9	5	<2	4
9	GN-89-6034	Hb clinopyroxenite	6	5	<2	4
11	GN-89-9011	Hb clinopyroxenite	17	15	<2	15
4	GN-89-8032	Feldspathic Hb clinopyroxenite	5	4	<2	7
12	GN-89-9009	Feldspathic Hb clinopyroxenite	<1	<2	<2	26
1	GN-89-9048	Feldspathic Cpx hornblendite	5	5	<2	10
5	GN-89-8027A	Hb-Cpx gabbro	19	22	<2	41
8	GN-89-6031	Hb-Cpx gabbro	5	13	<2	14
7	GN-89-8023	Cpx-Hb gabbro	20	7	<2	41
10	GN-89-8007	Cpx-Hb gabbro	15	33	<2	22
6	GN-89-8022	Hb gabbro	7	11	<2	13
2	GN-89-9034	Hb-Plag pegmatite	8	42	<2	21
13	GN-89-9001C	Hb-Plag pegmatite	41	88	<2	28
15	GN-89-9004	Melanocratic Qz diorite	5	8	<2	<1
14	GN-89-9006Z	Qz diorite/tonalite	<1	<2	<2	<1

Abbreviations: Ol, olivine; Cpx, clinopyroxene; Hb, hornblende; Plag, plagioclase; Qz, quartz. Detection limits: 1 ppb for Pt and Au; 2 ppb for Pd and Rh. Sample localities are given in Figure 3-7-3.

volcanic and volcaniclastic rocks of the Late Triassic Takla Group that forms part of Quesnellia. The complex is dominated by hornblende-bearing gabbroic rocks with lesser proportions of clinopyroxene hornblendite, hornblende clinopyroxenite, hornblendite and clinopyroxenite. The pegmatitic gabbros exhibit spectacular comb layering that records *in situ* crystal growth from supersaturated liquids that crystallized within the central part of the intrusion.

The northeastern margin of the complex is fault-bounded and the western and southern margins are bordered by a quartz diorite to tonalite pluton of probable Jurassic age. A small body of granodiorite intrudes the western part of the complex, and other minor intrusive rocks include dikes of microdiorite to microgabbro, felsite, and hornblende-plagioclase pegmatite.

The regional metamorphic mineral assemblages within the project area, which is situated within the eastern facies of the Takla Group west of the Pinchi-Ingenika fault system, indicate middle to upper greenschist facies conditions. However, vestiges of a metamorphic aureole of lower amphibolite grade at the northeastern margin of the Johanson Lake complex may have been largely obscured by faulting and retrograde metamorphism.

The abundance of noble metals in representative samples of the mafic and ultramafic lithologies of the complex is relatively low and there appears to be little evidence for economic mineralization.

ACKNOWLEDGMENTS

Fieldwork was funded by the Mineral Development Agreement between Canada and the Province of British Columbia. We wish to thank field assistant extraordinaire, Carol Nuttall; our team of expeditors, lead by Sandy Jaycox of Jaycox Industries; and our pilots from Northern Mountain and Highland Helicopters for making our stay comfortable and safe.

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Coal
Studies

PHOSPHORUS IN BRITISH COLUMBIA COKING COALS

By B.G. Van Den Bussche and D.A. Grieve

KEYWORDS: Coal quality, coking coal, phosphorus, coal sampling, coal petrography, low-temperature ash.

INTRODUCTION

Phosphorus occurs in all coals to some extent because it is essential to plant life. Knowledge of the phosphorus content and its associations in coking coal is important from an economic point of view, as phosphorus in steel has a detrimental embrittling effect and its presence in coking coals is therefore limited to as little as 0.01 per cent.

For reference, phosphorus content of United States coals ranges from 0.0002 to 0.1430 per cent and averages 0.0185 per cent (Abernethey and Gibson, *in* Van der Flier, 1985). The world average is estimated at 0.05 per cent (Valcovic, 1983).

A study under the coal quality project was begun in 1989 to examine in detail the occurrence of phosphorus in British Columbia coking coals. The study area includes the East Kootenay coalfields and the Peace River coalfield, which together account for all of British Columbia's current coking coal and semi-coking coal production (Figure 4-1-1). The main objectives of the study are to: determine the phosphorus content of coals; determine the affinity of phosphorus in coal (organic versus inorganic); and identify phosphorus-bearing minerals in low-temperature ash. In addition, petrography, proximate analysis, sulphur and trace element contents will be determined. These data will be used to discern correlations, if any, between phosphorus content and other analytical parameters.

PREVIOUS WORK

The pioneer of trace element analysis of coal was Goldschmidt (1935). Since then a great deal of work has been done on the occurrence, distribution and associations of trace elements in coal; much of this relates to phosphorus. For examples and more detailed bibliographies refer to Gluskoter *et al.* (1977), Van der Flier (1985), Van der Flier-Keller and Fyfe (1987), and Goodarzi *et al.* (1987).

Phosphorus has been noted to have an affinity with organics (Rao *et al.*, 1951, Bogdanov, 1965, and Kuhn *et al.*, 1978, *in* Van der Flier, 1985), inorganics (Brown and Swaine, 1964, *in* Van der Flier, 1985) and a combination of the two (Francis, 1961, and Gluskoter, 1977, *in* Van der Flier, 1985).

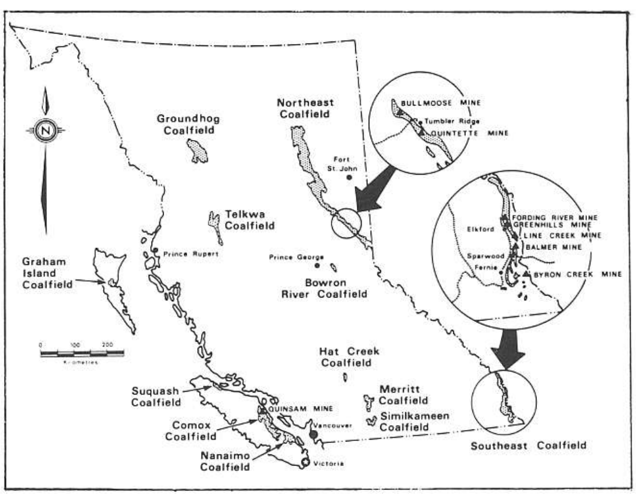

Figure 4-1-1. Sampling location map.

METHODS OF STUDY AND FIELDWORK

The project was approached in two steps. The first was the creation of a database of analyses of diamond-drill core the samples recorded in existing company assessment reports. All results included in this paper were generated by statistical analysis of these data. The second approach entailed the collection of fresh, unoxidized coal samples from each of the seven coking and semi-coking coal mines in British Columbia: Balmer, Greenhills, Fording, Line Creek and Byron Creek in the East Kootenay coalfields, and Bullmoose and Quintette in the Peace River coalfield (Figure 4-1-1).

DATABASE

Initially, a data search was performed, and a historic coal-quality database was set up using dBase III PLUS, a copyrighted database management software package. Using coal quality analysis data from British Columbia coal assessment reports, more than 1900 borehole sample records were entered into the database. The criterion for inclusion was the existance of phosphorus analyses, either in the form of P_2O_5 in coal or P_2O_5 in ash. (For this report, "P_2O_5" will refer to P_2O_5 in coal. The phosphorus contribution to the total weight of P_2O_5 is 43.64 per cent. Each record in the database contains three components: sample identification information, raw-coal data, and clean-coal data. The identification section contains location and sample-type information. The raw and clean-coal data contain the available proximate, sulphur, P_2O_5, mercury, chlorine and fluorine analyses. The clean-coal data were grouped by specific gravity fraction. To date, only raw-coal data has been studied, focusing on phosphorus versus ash relationships for coals from the Peace River and East Kootenay coalfields. Future work is planned to expand the quality program to include clean-coal data and a number of additional minor and trace elements.

Phosphorus and ash data were then grouped and copied to ASCII text files. These in turn were edited with the DOS editor, EDLIN, to the format required for GEO EAS (Geostatistical Environmental Assessment Software), a complimentary software package distributed by the United States Environmental Protection Agency. X-Y scatter plots of P_2O_5

in coal versus ash were then plotted (Figures 4-1-2 and 3; *see* Nicholls, 1968). Histograms were also constructed to help visualize the distribution of the data. Four histograms were compiled, one for each of P_2O_5 and ash for both the East Kootenay coalfields (Figures 4-1-4 and 5) and the Peace River coalfield (Figures 4-1-6 and 7). In all plots, only samples with 50 per cent or less ash were considered.

FIELDWORK

Channel samples were collected from each of the seven producing coking or semi-coking coal mines included in the study. An effort was made to obtain a sample from as many different seams as possible at each mine. In all, 68 samples were acquired. All samples from Balmer, Byron Creek, Fording, Greenhills, Quintette and Bullmoose were full-seam samples, while the samples from Line Creek were taken in 0.5-metre increments. Fording accounted for eight of the samples, representing seven seams; Byron Creek for three samples, all from the Mammoth seam; Greenhills, Balmer and Bullmoose for six samples each, from as many seams; and Quintette for two samples from two seams. Line Creek accounted for the remaining 33 samples, from four seams.

Sampling involved the removal of an approximately 8 by 8 centimetre channel of coal over the entire sample interval. This was accomplished by chipping with a geological hammer and collecting the material in a gold pan for the harder coals, and by utilizing a small scoop for the softer coals. Every attempt was made to take the samples perpendicular to bedding. The samples were then stored in labelled plastic bags for transport to the laboratory. Accurate thickness measurements were taken of all sample intervals.

Preparation and chemical analyses of the samples were done at Chemex Labs Ltd. in Vancouver, according to the flow chart in Figure 4-1-8. The raw sample was first dried and blended, then crushed to -5.3 millimetres. The material was split into three portions; one was crushed to -20 mesh, one crushed to -60 mesh, and the third set aside as reserve. A second split was taken from the material crushed to -60 mesh, with one half used at Chemex for analysis, and the other returned to the Geological Survey Branch. Some of the latter will be utilized for low-temperature plasma ashing and mineral determination by x-ray diffraction, and the remainder will be used for trace element determination by neutron activation analysis. The portion crushed to -20 mesh was returned for petrographic analysis. Chemex is responsible for proximate analysis, sulphur forms, P_2O_5 in coal, and chlorine, fluorine and mercury determinations. Results from all analyses will be reported in a later publication.

RESULTS

Data from the East Kootenay coalfield are represented by an X-Y plot of P_2O_5 in coal versus ash (Figure 4-1-2) and two histograms, one showing P_2O_5 distribution (Figure 4-1-4), and the other showing ash distribution (Figure 4-1-5). Together they represent a set of 601 data points, ranging from 0.010 to 0.240 per cent P_2O_5, and containing less then 50 per cent ash. From the plots we can see a concentration of points below 0.05 per cent P_2O_5 and between 10 and 32 per cent ash. The ash histogram shows a distribution close to normal with a mean of 24.54 and moderate positive skewness and positive kurtosis (peakedness). The P_2O_5 histogram shows a bimodal distribution. The first population is a symmetrical group of points with high kurtosis and containing less than 0.050 per cent P_2O_5, and the second is positively skewed and represents points beyond 0.050 per cent P_2O_5.

The Peace River coalfield data are also represented by similar plots and histograms (Figures 4-1-3, 6 and 7). A set of

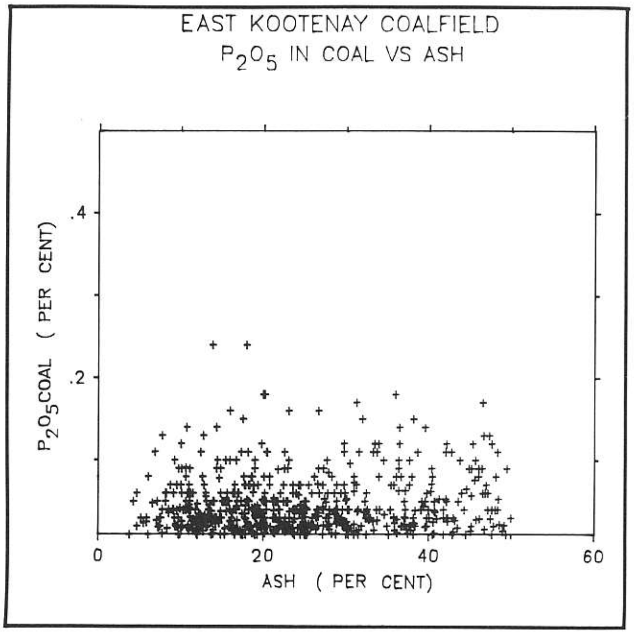

Figure 4-1-2. X-Y plot of P_2O_5 versus ash for the East Kootenay coalfield.

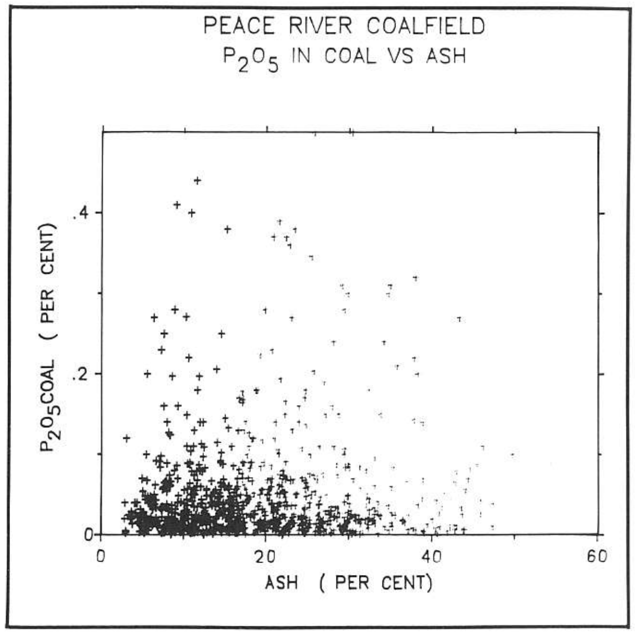

Figure 4-1-3. X-Y plot of P_2O_5 versus ash for the Peace River coalfield.

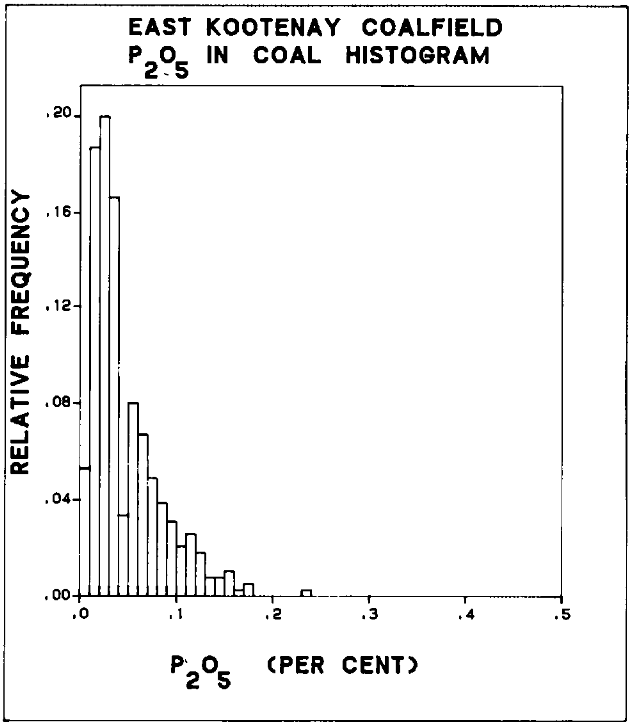

Figure 4-1-4. Histogram of P_2O_5 distribution for the East Kootenay coalfield.

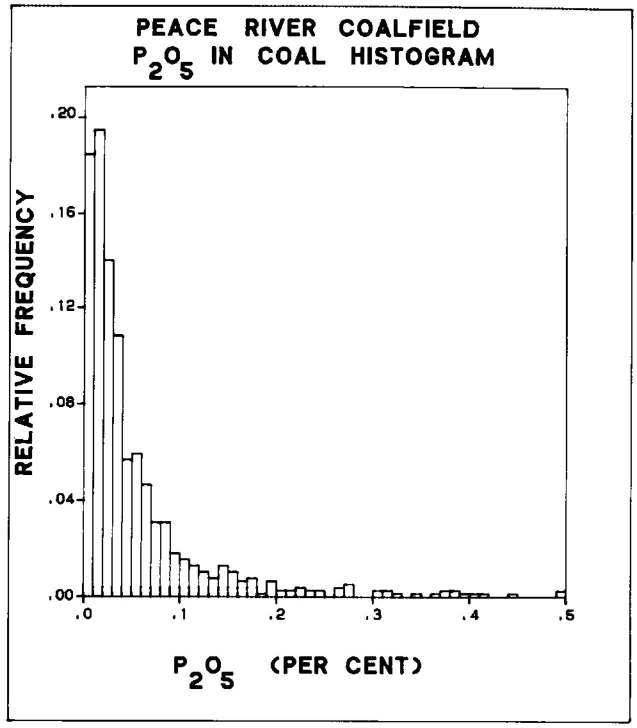

Figure 4-1-6. Histogram of P_2O_5 distribution for the Peace River coalfield.

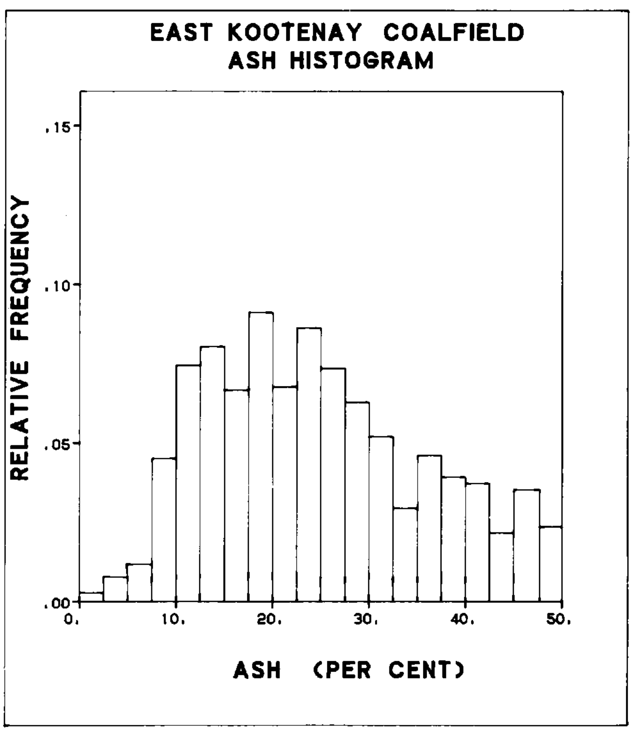

Figure 4-1-5. Histogram of ash distribution for the East Kootenay coalfield.

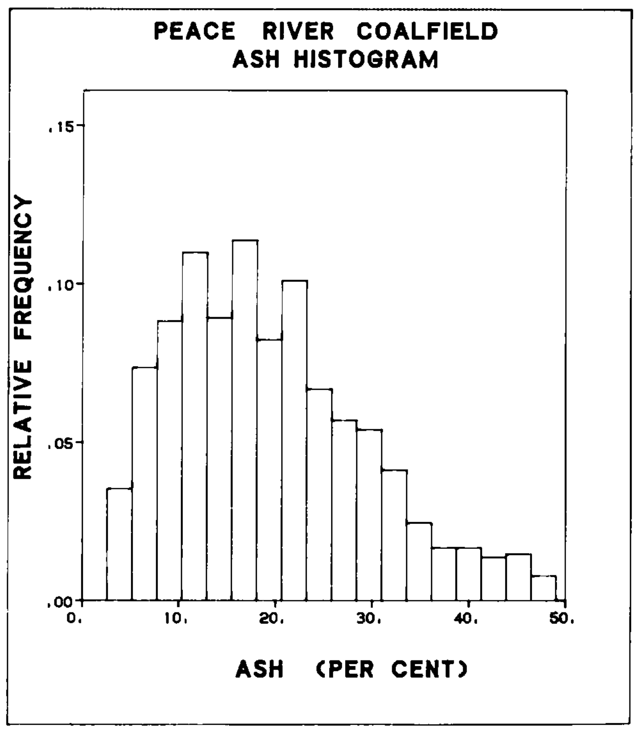

Figure 4-1-7. Histogram of ash distribution for the Peace River coalfield.

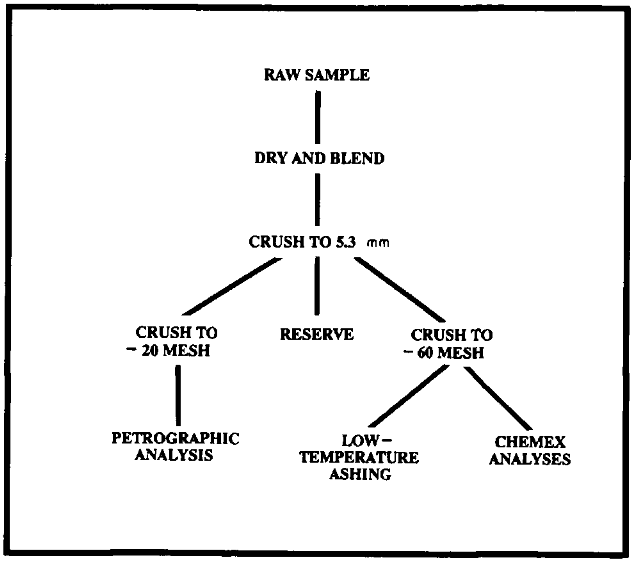

Figure 4-1-8. Simplified flowchart of sample analysis.

798 data points, ranging from 0.002 to 0.500 per cent P_2O_5 and containing less then 50 per cent ash are represented. There is a concentration of points between 5 and 30 per cent ash and below 0.050 per cent P_2O_5. The ash histogram shows a distribution similar to that for the East Kootenay coalfields, but with a higher positive skewness. The mean in this case is 18.92 per cent. Although not as apparent as in the East Kootenay coalfield, the P_2O_5 histogram for the Peace River Coalfield shows signs of a similar bimodal distribution.

DISCUSSION

The bimodal P_2O_5 distribution in both coalfields is thought to represent both organic and inorganic affinities of P_2O_5 in these coals. We can assume that all coals contain a certain amount of organic P_2O_5, since all plants need phosphorus to live. This organic phosphorus is thought to be represented by the population with less than 0.050 per cent P_2O_5. The skewed population with greater than 0.050 per cent P_2O_5 would then represent the organic phosphorus found in all coals, plus the inorganic phosphorus contributed by mineral matter. This interpretation is based on the similarity in distribution of the higher P_2O_5 population and the ash content, especially their positive skewness. The inorganic component of phosphorus is expected to vary with the amount of ash in the sample and/or the amount of P_2O_5 in the ash. Further work will be aimed at determining more precise associations and ultimately the controls on the distribution of phosphorus in coking coals of British Columbia.

ACKNOWLEDGMENTS

The authors would like to thank Sharon Chapman for compiling and entering data used in this study. Also we wish to extend special thanks to staff at Balmer, Bullmoose, Byron Creek, Fording, Greenhills, Line Creek and Quintette mines, whose assistance and cooperation during the field season ensured efficient collection of samples. Thanks to Barry Ryan for his useful discussions and suggestions, and to Jim Hunter for his assistance with illustrations.

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GEOLOGICAL INVESTIGATIONS IN THE COAL MEASURES OF THE OYSTER RIVER, MOUNT WASHINGTON AND CUMBERLAND AREAS, VANCOUVER ISLAND (92F/10, 11, 14)

By Corilane G.C. Bickford and Georgia Hoffman
Bickford Consulting Ltd.
and
Candace Kenyon

KEYWORDS: Coal geology, stratigraphy, Vancouver Island, Comox sub-basin, Nanaimo Group, Comox Formation, stratigraphic nomenclature, coalbed mineability.

INTRODUCTION

This report is part of an ongoing project begun in 1987 to update knowledge of critical geological relationships in the Vancouver Island coal deposits. Objectives of the year's study are to deliniate the extent and continuity of coal-bearing strata in the Comox Formation between Oyster River and Tsable River, and to document the structural geology and stratigraphy of the area. Also, the potential underground mineability of the coalbeds of the area will be assessed.

The study area occupies part of the eastern coastal plain of Vancouver Island, from Campbell River in the north to Cowie Creek in the south (Figure 4-2-1).The Comox sub-basin is approximately 1230 square kilometres in area and is accessible by coastal waterways, paved highways and secondary roads. Elevations range from sea level to 457 metres with fairly gentle topography. Most of the area is covered by thick underbrush, limiting coal exposures to roadcuts, and

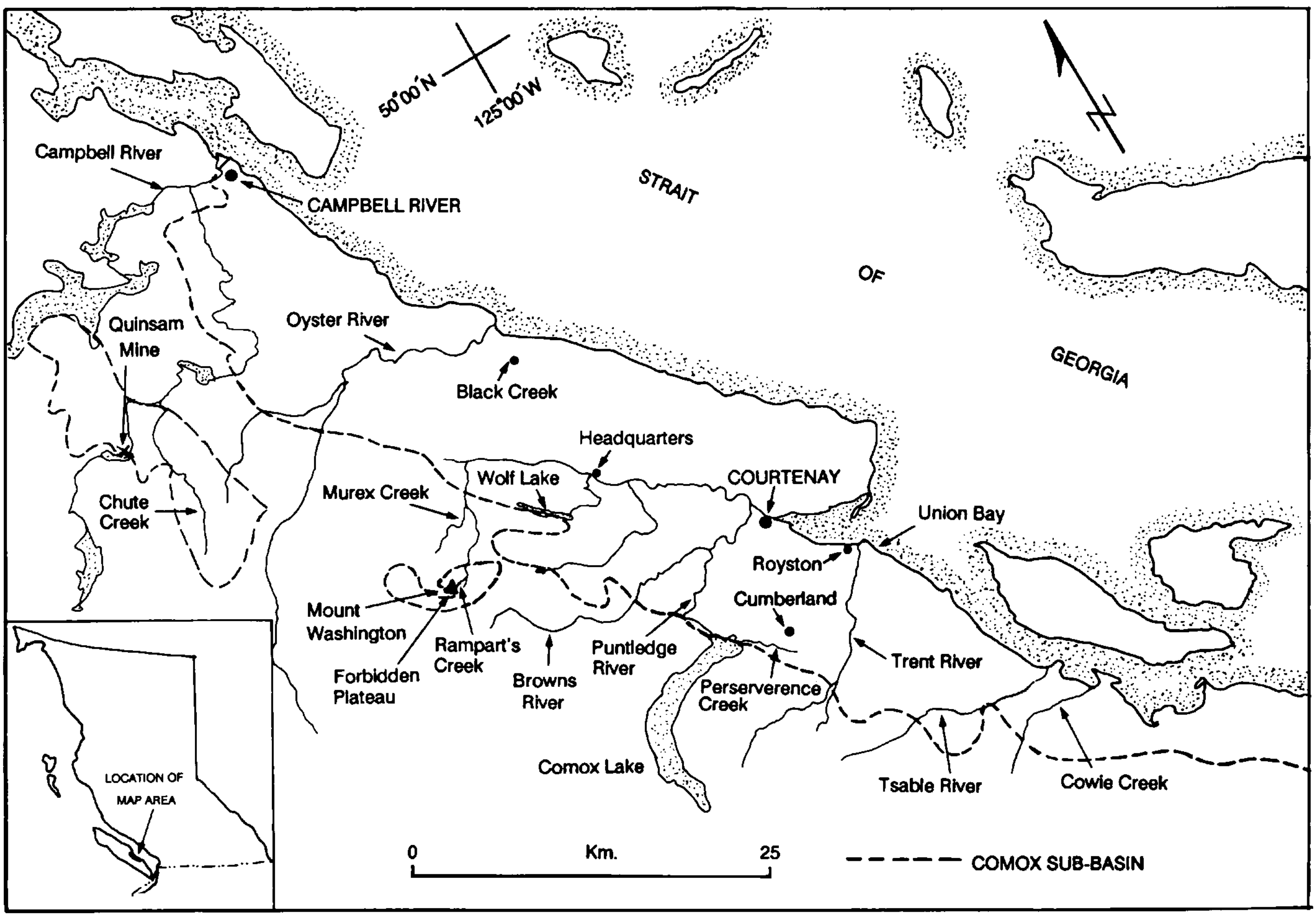

Figure 4-2-1. Location map, northern Comox sub-basin, Vancouver Island.

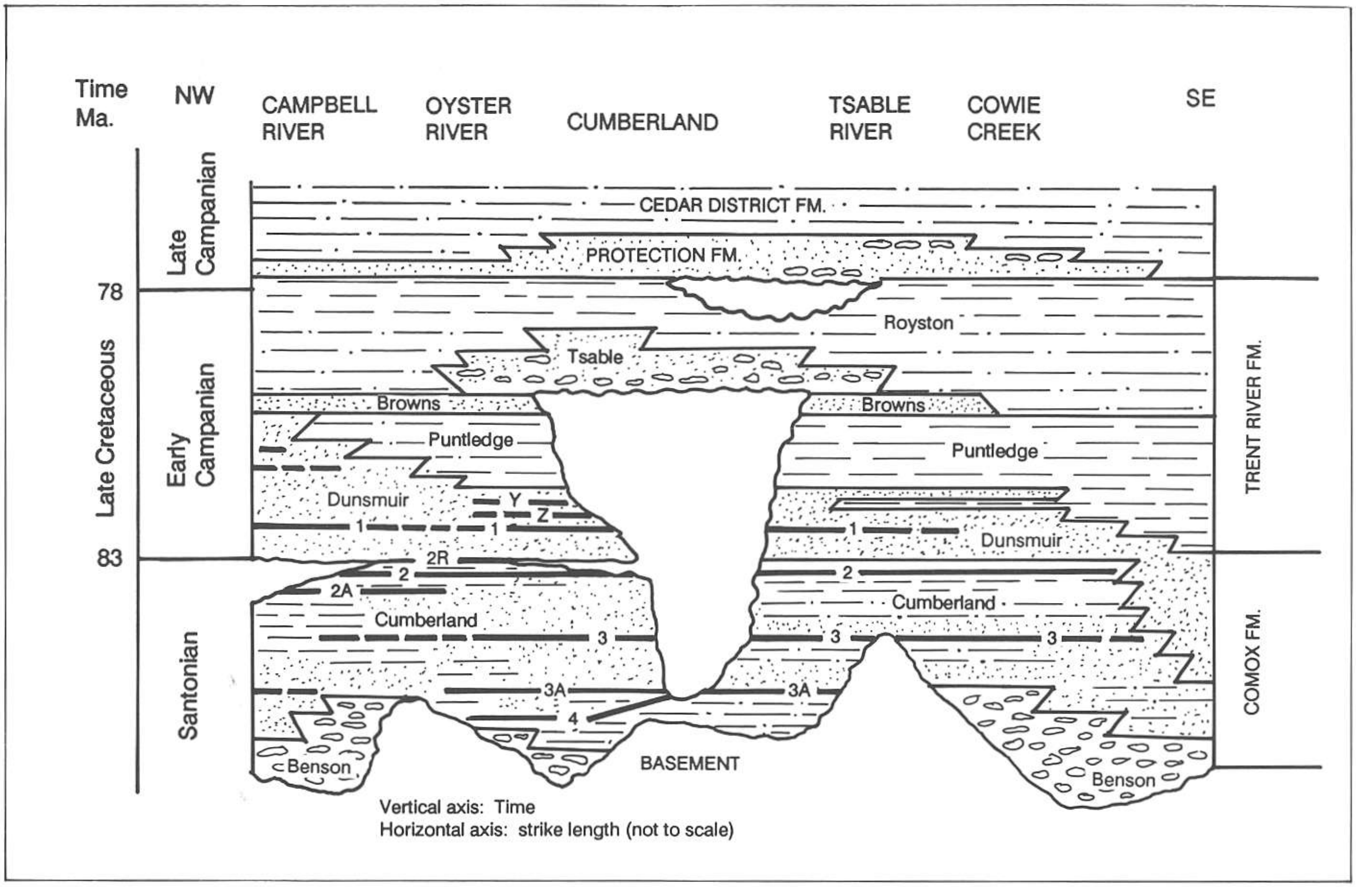

Figure 4-2-2. Chronostratigraphic diagram of the Lower Nanaimo Group.

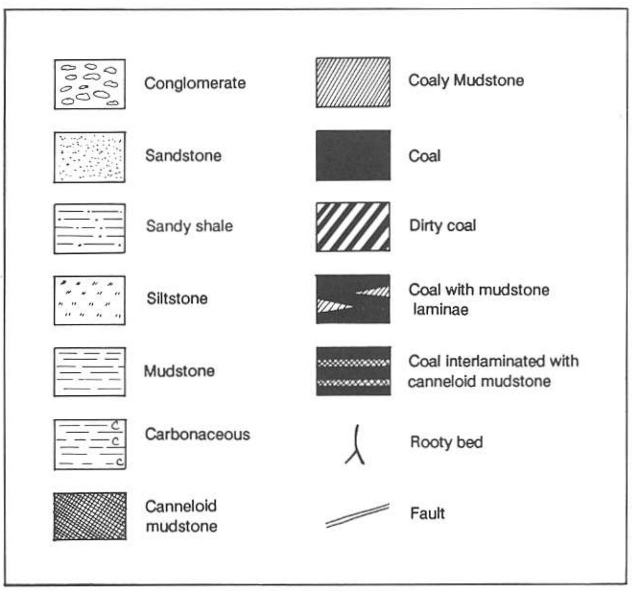

Figure 4-2-2a. Standard symbols for coal-measures sections.

creeks and rivers which drain into the Strait of Georgia. The climate is mild and humid with snow occuring only at higher elevations.

Campbell River and Courtenay are the major population centres. Small towns and resort areas are scattered along the coast. The logging and fishing industries are the economic base of the area.

FIELDWORK

Geological mapping was done using 1:20 000-scale aerial photographs to plot data which were later transferred to 1:20 000 base maps. Detailed sections of coal beds were measured on Oyster River, Ramparts Creek, Murex Creek, Perseverence Creek, Tsable River, the eastern shore of Comox Lake, in the Mount Washington area, and in several old mine tunnels near Cumberland village. Samples of coal and associated rocks were taken for petrographic analysis. Data from 36 new coal outcrops will be used to update the regional vitrinite reflectance study (Kenyon and Bickford, 1989). Fossil plants and invertebrate shells were collected at 20 locations for later identification. Trace-fossil assemblages were described where well-exposed, but were not collected owing to the impracticability of transporting large samples.

GEOLOGICAL SETTING

The study area covers the northern half of the Comox sub-basin of the Late Cretaceous Georgia basin. The coal measures occur in the Cumberland and Dunsmuir members of the Comox Formation (Bickford and Kenyon, 1988). The Comox Formation outcrops along the western edge of the eastern coastal lowland of Vancouver Island, and dips gently eastward beneath younger Cretaceous rocks and unconsolidated Pleistocene sediments. It has been traced down dip by means of drilling to the shore of Georgia Strait; its eastward submarine extent remains unknown.

In addition to the main body of Comox Formation under the coastal lowland, several outliers of the formation occur farther to the west, in the foothills of the Vancouver Island Ranges. The largest outlier contains the Quinsam and Chute Creek coal deposits (Bickford *et al.* 1989), while the smaller outliers near Mount Washington (Carson, 1960), Forbidden Plateau, (Gunning, 1931,) and Comox Lake (Daniels, 1920) contain coal showings of more limited extent. Geology maps of this area can be found in Kenyon and Bickford, 1989.

STRATIGRAPHY

Lithostratigraphic units of the lower half of the Nanaimo Group are shown in Table 4-2-1 and illustrated in Figure 4-2-2. Some formation and member names have been changed since last year's report, in light of further fieldwork during the 1989 mapping season and because of suggestions made by England (1989). England states that units at higher stratigraphic levels in the Nanaimo Group, previously mapped by Muller and Jeletsky (1970) as formations, cannot be traced from the Nanaimo sub-basin to the Comox sub-basin and these formation names therefore "contravene the rules of stratigraphic nomenclature (American Commission on Stratigraphic Nomenclature, 1970)". We recognize the validity of the Trent River Formation as suggested by England. The lithostratigraphic nomenclature suggested here represents a compromise among proposals by Muller and Jeletsky (1970), Ward (1978), Bickford and Kenyon (1988) and England (1989).

TABLE 4-2-1
LITHOSTRATIGRAPHIC UNITS OF THE LOWER NANAIMO GROUP, NORTHERN COMOX SUB-BASIN

Formation:	Member:	Lithology:
Cedar District		Mudstone and siltstone.
		—Intertonguing contact—
Protection		Sandstone and conglomerate.
		—Erosional contact—
Trent River	Royston	Mudstone and siltstone.
		—Intertonguing contact—
	Tsable	Conglomerate; minor sandstone and siltstone.
		—Erosional contact—
	Browns	Sandstone; minor siltstone.
		—Intertonguing contact—
Comox	Dunsmuir	Sandstone; minor siltstone, shale, coal and conglomerate.
		—Abrupt, locally erosional contact—
	Cumberland	Siltstone and sandstone; minor shale and coal.
		—Intertonguing contact—
	Benson	Conglomerate and red siltstone.
		—Erosional contact—
pre-Cretaceous volcanic, plutonic and metasedimentary basement.		

SUBDIVISION OF THE TRENT RIVER FORMATION

Sedimentary rocks previously mapped as the Haslam, Extension, and Pender formations are now considered to be part of the Trent River Formation in the Comox sub-basin, although Haslam, Extension and Pender are still valid names in the Nanaimo sub-basin. McGugan (1979) and England (1989) expressed concerns with the transfer of formation and member names from one sub-basin to another; we must therefore propose some new member names within the Trent River Formation in order to correctly describe these mappable units.

PUNTLEDGE MEMBER

The Puntledge member consists of dark grey to black mudstones and siltstones, 100 to 150 metres thick, which overlie the sandstones of the Dunsmuir member of the Comox Formation. This unit was previously mapped by Muller and Jelesky (1970) as the Haslam Formation. The name Puntledge is taken from the Puntledge River, along which these mudstones and siltstones are well exposed. The Puntledge member contains numerous ammonites and bivalves. Recently, a vertebrate fossil, a plesiosaur, was found in this member (Michael Trask, surveyor, B.C. Ministry of Highways, personal communication, 1989). An isolated sandstone bed, 3 to 5 metres thick, commonly occurs near the base of the member.

BROWNS MEMBER

We propose the name Browns member for a sandy unit which forms ledges in Browns, Puntledge and Tsable River, and underlies a low swamp-bounded ridge between Headquarters Village and Black Creek. It consists mainly of mixed granitic-basaltic sandstone and siltstone which is moderately to intensely bioturbated and contains abundant shell fossils, chiefly *Inoceramus*. The base of the member is abrupt, while the top is gradational and interfingering with the Royston member. Contacts between the Browns member and the laterally adjacent Tsable member are not exposed in outcrop, but borehole and mine records on file with the Ministry of Energy, Mines and Petroleum Resources suggest that the Browns member is erosionally truncated by the Tsable member.

The Browns member is 9 metres thick on Browns River and 15 to 20 metres thick at Black Creek. Locally, sedimentary structures have been almost obliterated by bioturbation, but less intensely burrowed parts of the member are usually either planar laminated or trough crossbedded. Original sorting of the sandstones is fair to good, and the less bioturbated sandstones display fair intergranular porosity. The Browns member was probably deposited as a complex of offshore bars, generally below storm wave base.

TSABLE MEMBER

The Tsable member was introduced by England (1989) for a ridge-forming conglomerate unit which outcrops between Courtenay and Tsable River. It consists of mixed basaltic-granitic conglomerate, which contains numerous blocks and cobbles of sandstone and siltstone identical to those of the underlying Dunsmuir and Puntledge members. Sandstone and siltstone interbeds are abundant in the upper half of the member, which intertongues with the overlying Royston member.

Tsable conglomerates are distinguished from the Comox Formation by their greater (although not predominant) content of granitic clasts and the ubiquitous presence of sedimentary rock clasts, which are extremely rare in the Comox conglomerates. The base of the Tsable member is erosional, truncating the underlying Browns and Haslam members of the Trent River Formation, and locally cutting down deeply into the Comox Formation. The member is 40 to 65 metres thick in the subsurface between Courtenay and Royston, but along its outcrop, east of Cumberland, it is at least 140 metres thick, while in the canyon of the Trent River, it is less than 5 metres thick. It pinches out to the northwest and southeast and does not appear to extend beyond Puntledge River in the north and Tsable River in the south.

Disorganized and chaotic sedimentary fabrics typify the basal part of the Tsable member while the upper part often displays planar graded bedding with fair to good sorting of framework grains. It was probably deposited as a submarine canyon fill; the uppermost beds were probably spilled out over a submarine fan, following filling of the main channel.

ROYSTON MEMBER

We propose the name Royston member for a unit of fine-grained sedimentary rocks which overlies the Tsable and Browns members and underlies the Protection Formation of the Comox sub-basin. The name Royston comes from the village of the same name, where this unit outcrops extensively. The Royston member replaces the Pender Formation, which is now considered to be an invalid name in the Comox sub-basin.

The Royston member consists of dark grey to olive-drab silty mudstone, siltstone and fine-grained thin-bedded sandstone, distinguishable from the Puntledge member by its greater content of sand and silt, and slightly lighter colour. The Royston member contains numerous ammonites, some rudistid and inoceramid bivalves, and occasional fish scales. It is 150 to 220 metres thick in its type locality at Royston.

STRUCTURAL GEOLOGY

According to Muller (1989), much of the Cretaceous sedimentary rock mass in the Mount Washington area is in tectonic contact with underlying basement rocks. Mapping during the field season demonstrated the validity of Muller's structural model. Some of the other Cretaceous outliers in the Mount Washington area are klippen emplaced by east-directed low-angle extensional faults. Most of the faults which affect the Cretaceous rocks tend to follow bedding near the base of the Benson member (Bickford and Kenyon,

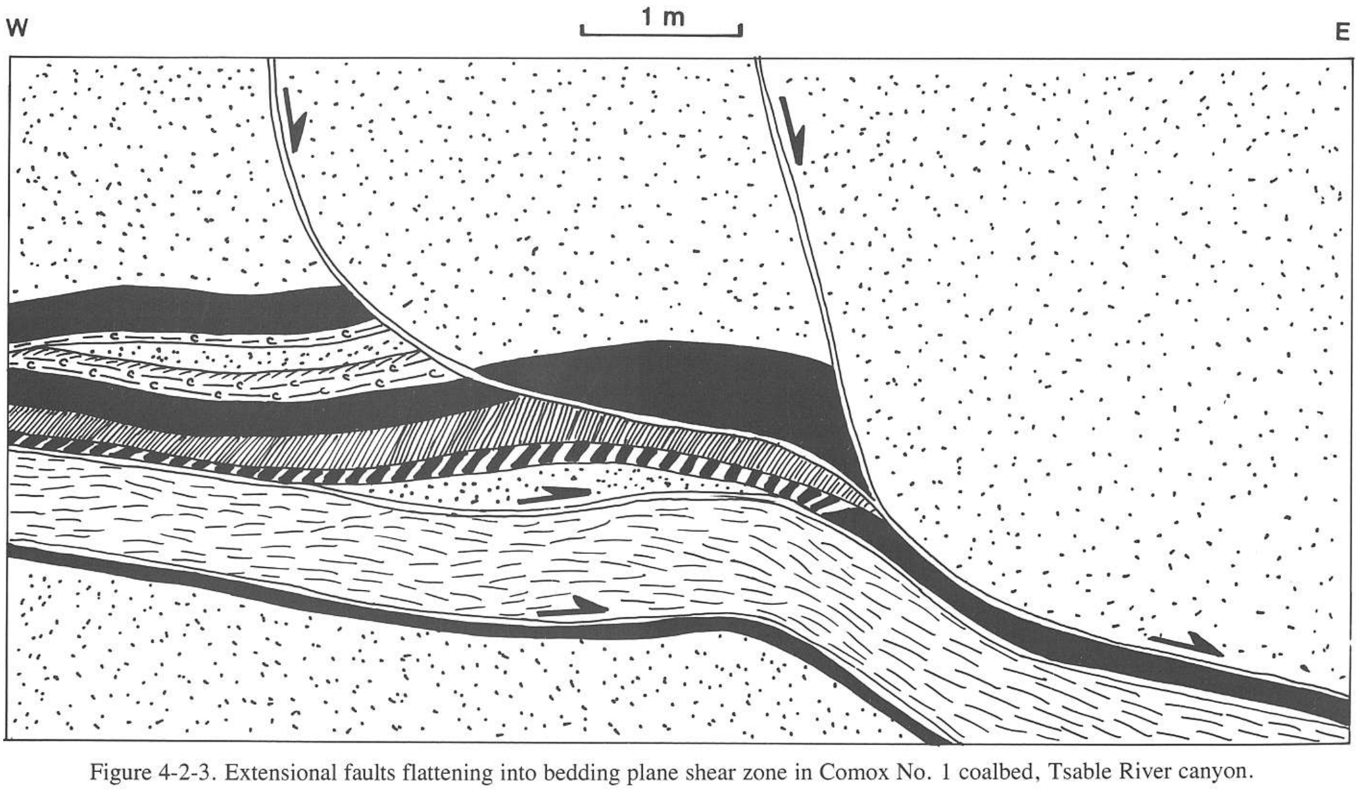

Figure 4-2-3. Extensional faults flattening into bedding plane shear zone in Comox No. 1 coalbed, Tsable River canyon.

1988). Fault movement appears to have been facilitated by the intense lateritic weathering zone which is characteristic of the basement paleosurface. The shearing exhibited by the coals of the Comox Formation, (for example, the thick coal near the top of the Cumberland member along Ramparts Creek) may have been produced during this movement.

East-directed low-angle extensional faults are also exposed along the Trent and Tsable rivers, farther to the south. On Trent River, the basal Haslam shales are intensely sheared and are in fault contact with the underlying Dunsmuir sandstones. On Tsable River, the basal Haslam shales are intensely sheared and crumpled, and contain contorted ankerite veinlets. The underlying Comox No. 1 and No. 2 coals are also the loci of bedding plane shearing (Figure 4-2-3).

PALEOCURRENT STUDIES

Paleocurrent indicators are abundant in the Benson member of the Comox Formation and the Tsable member of the Trent River Formation. Indicated paleocurrents are generally to the southwest and west in the Benson member, with a grand vector mean paleocurrent direction (based on 19 outcrops) of 237°. Benson conglomerates usually display good framework imbrication. The long axes of Benson framework clasts are parallel to bedding and perpendicular to paleoflow, indicating clast transport by rolling along a bed, consistent with a fluvial environment (Walker, 1984).

In the Tsable member, indicated paleocurrents are generally to the south and southwest. The grand vector mean paleocurrent direction (based on 14 outcrops) is 224°, slightly more southerly than that of the Benson member. Tsable conglomerates often have chaotic fabrics and, where framework imbrication is present, the long axes of the clasts dip parallel to paleoflow, consistent with transport by either mass movement or in a fluidised bed, probably under conditions of turbulent flow consistent with the interpretation of the Tsable member as a submarine canyon fill.

Further work will be required to assess whether these palecurrent directions are consistent throughout the Comox sub-basin; at present a Coast Range sediment source is possible for both the Benson and Tsable conglomerates.

ECONOMIC CONSIDERATIONS

Several coalbeds thicker than 1.2 metres have been identified in the Comox Formation during this field season. Details of these are summarized in Table 4-2-2. The UTM coordinates are included as the locations are not indicated on Figure 4-2-1. Detailed sections along the Trent and Browns rivers can be found in Kenyon and Bickford, 1989. It is likely that potential remains for additional coal discoveries in the Comox sub-basin. There are two areas that have had little exploration work; one between Oyster River and Headquarters Village and the other south of Trent River, along the coastline toward Union Bay.

The best coal showing found during the 1989 field season is on the Oyster River, where the No. 2 bed has a gross thickness of 1.73 metres and a net coal content of 90 per cent by thickness. Two other noteworthy coal outcrops are on Ramparts Creek, on the southwest side of Mount Washington, and Murex Creek, north of Wolf Lake. In both cases the coal measures have undergone low-grade contact metamorphism in the vicinity of Tertiary plutons, and the coals are of a higher rank compared to elsewhere in the sub-basin.

It is probable that the Ramparts Creek coal correlates with the Comox No.2 bed. It is intensely weathered at outcrop and it will be necessary to trench in this area to obtain an accurate section. At outcrop, the coalbed is 1.94 metres thick, with a net coal content of about 78 per cent by thickness. The Murex Creek coal is probably correlative with the Comox No.3 bed. It is only 1.31 metres thick, with a net coal content of 66 per cent by thickness, but it is of interest because it has a mean maximum vitrinite reflectance of 2.52 per cent, indicative of semi-anthracite rank (Kenyon and Bickford, 1989, sample 61). Sections of these and other potentially workable coalbeds are given in Figure 4-2-4.

Roof conditions play a major role in determining the success or failure of an underground colliery. Dark grey, variably carbonaceous mudstone forms the immediate roof of most of the Comox coals seen at outcrop. Where unsheared, this material can form a stable working roof, as has been evidenced by old workings, some more than 70 years in age. The critical roof span (Das, 1985) of these rocks is in the order of 4 to 6 metres. Where the mudstones are sheared, their performance as a working roof is inadequate. Such sheared rocks formed the roof of the Comox No.2 bed in Tsable River colliery and were difficult to support for more than a few days to weeks (Stan Lawrence, retired colliery manager, personal communication, 1989). The critical roof span of the sheared mudstones is less than 4 metres. Above the immediate mudstone roofs, many of the Comox coals have a main roof of thick-bedded to massive, strong sandstone. Where this

TABLE 4-2-2
OCCURRENCES OF POTENTIALLY MINEABLE COAL AT OUTCROP IN THE COMOX SUB-BASIN

Locality	Coordinates	Coal bed	Gross Thickness	Per cent Coal by Thickness
Woodhus Cr.	331250 E 5530000 N	Comox 2	3.19 m	69
Oyster R.	334750 E 5527060 N	Comox 2	1.73 m	90
Ramparts Cr.	335240 E 5511250 N	Comox 2?	1.94 m	78
Murex Cr.	340570 E 5518755 N	Comox 3?	1.31 m	66
Browns R.	347770 E 5506220 N	Comox 2	1.41 m	76
Cumberland (No. 3 mine)	352300 E 5497605 N	Comox 1	1.31 m	78
Perseverence Cr.	352145 E 5496830 N	Comox 3A	1.65 m	85
Allen Lk.	353035 E 5495515 N	Comox 4	1.90 to 2.19 m	58 to 73
Hamilton Lk.	349450 E 5495865 N	Comox 3A	1.70 m	83
Trent R.	354505 E 5493610 N	Comox 3	1.55 m	75
Trent R.	354360 E 5493550 N	Comox 3A	2.10 m	82

Note: Coordinates given are UTM Zone 10.

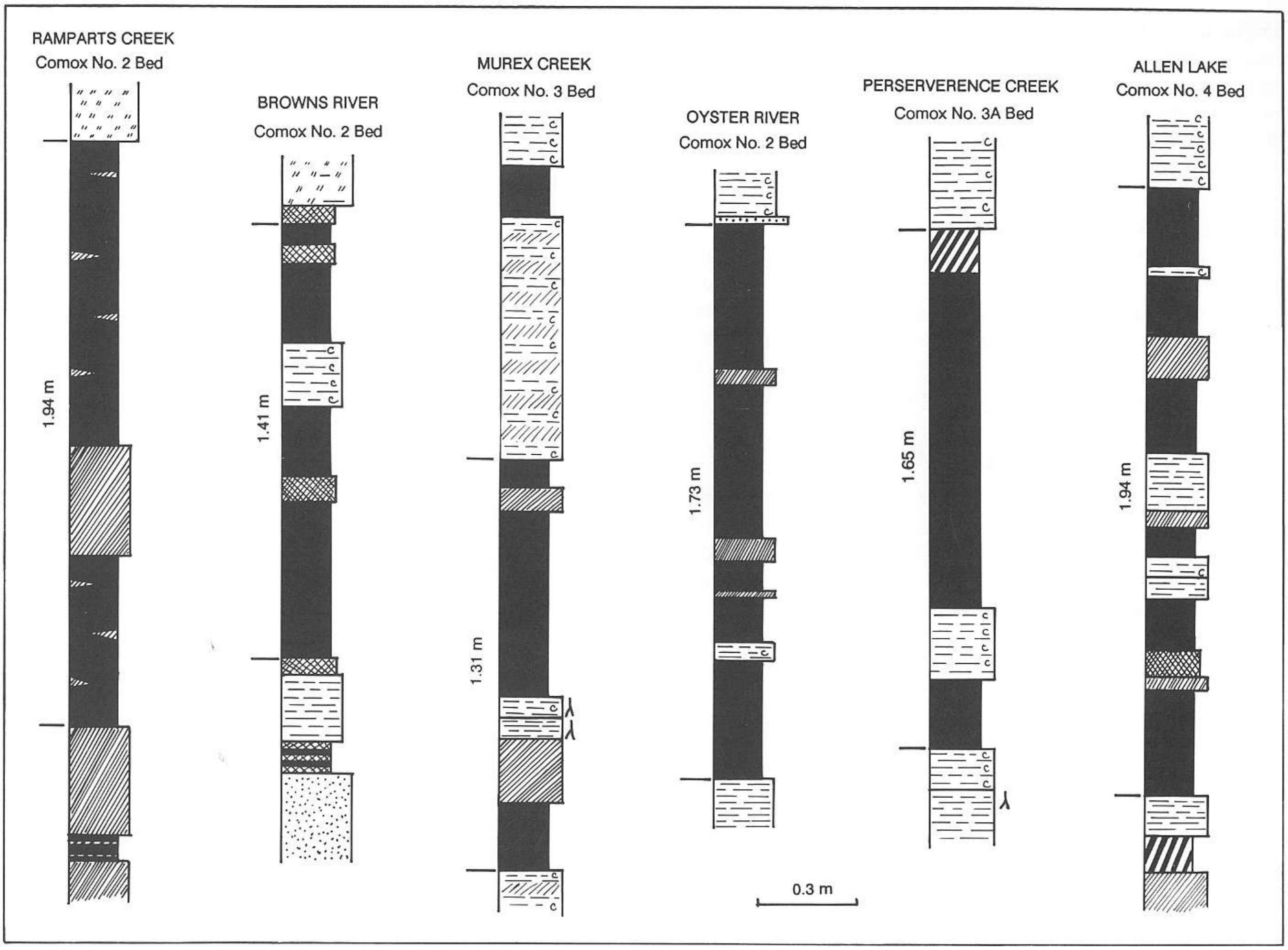

Figure 4-2-4. Measured sections of potentially mineable coalbeds in the Comox sub-basin.

sandstone is close to the underlying coalbed, it forms an excellent roof for room-and-pillar workings. For example, the hard sandstone roof of the Comox No.1 bed in the No. 3 mine at Cumberland has stood unsupported for more than 90 years.

ACKNOWLEDGMENTS

The authors would like to thank the forestry staff of Fletcher Challenge Canada Inc, and MacMillan Bloedel Ltd. for their assistance in many ways. The Campbell River Search and Rescue Society provided up-to-date logging-road maps. Michael Trask provided details of vertebrate fossil discoveries in the Trent River Formation. Stan Lawrence of Cumberland provided invaluable information regarding the Tsable River coalfield, and Cleo Pawson and Marie Norsed provided excellent field assistance.

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NOTES

SUBSURFACE COAL SAMPLING SURVEY, QUINSAM AREA, VANCOUVER ISLAND, BRITISH COLUMBIA (92F)

By Alex Matheson

KEYWORDS: Coal geology, Comox coal measures, Upper Cretaceous, Nanaimo Group, Quinsam, Vancouver Island, drilling tests, x-ray drill, packsack drill, core recovery, coal analysis.

INTRODUCTION

Since the discovery of coal in British Columbia in 1835, three major mining areas have produced substantial quantities of coal. The deposits on southern Vancouver Island were intensively mined until the middle of this century and have recently recommenced production. The Kootenay coalfields of southeast British Columbia, which came into operation at the turn of the century, are currently the largest and most productive in the province. More recently the Peace River area in northeast British Columbia has been brought into production and has potential for large future developments.

Coal deposits occur province-wide and there are many undeveloped areas, such as parts of the Telkwa, Bowron River, Merritt, Tulameen and the Queen Charlotte Islands coalfields, where there is a paucity of coal quality data. Information gathered on these and other lesser known areas would prove invaluable to future exploration projects and assist the government in its management of the province's coal resources. The least expensive method of data gathering is sampling of coal outcrops. However, there are two drawbacks to this method: first, all coal outcrops are oxidized to variable depths which affects some of the analyses; second, several areas have few surface exposures, for example, Princeton and Merritt. Any fieldwork should therefore not only incorporate outcrop sampling but selective diamond drilling as well.

The viability of such a drilling program was assessed during the 1988 field season. The main objective at that time was to examine various small drills and their capabilities with respect to the core recovery of unoxidized representative coal samples for comprehensive analysis. Several suitable drills were identified, but only two were tested, due to lack of availability. An x-ray drill was operated by Neill's Mining Company under contract, and a Packsack drill, leased from The University of British Columbia, was operated by Ministry of Energy, Mines and Petroleum Resources staff.

DESCRIPTION OF DRILLS

THE PACKSACK 4M DRILL

A portable drill driven by a 10-horsepower, two-cycle, air-cooled gasoline engine and utilizing an IEX 25.4 millimetre (1 inch) diameter, single, rigid core barrel, can penetrate to a depth of 15 metres. This hand-held unit (Plate 4-3-1) with variable throttle control is probably the lightest drill on the market. With accessories, but excluding rods, it weighs about 50 kilograms. It can be operated by one person, but two make the work considerably easier. Two holes were drilled with this machine. One was abandoned at 10 metres when the bit mudded in. The second, Hole 88-1B, reached a depth of 8.3 metres and obtained 89 per cent core recovery. The main drawback of the hand-held drill is that with a constantly changing fulcrum, the direction of pressure applied varies, resulting in deviation of the hole.

THE X-RAY DRILL

This drill (Plate 4-3-2) is a very old unit which the suppliers, JKS Boyles Ltd., replaced with the Winkie. It is powered by a 9-horsepower two-cycle, air-cooled gasoline engine, and can penetrate to a depth of over 100 metres. The unit, which is mounted, has two gears and weighs about 75 kilograms without the rods. Hole number GSB88-1 used the

Plate 4-3-1. Packsack 4M drill.

TABLE 4-3-1
AVERAGE PROXIMATE ANALYSIS VALUES ON QUINSAM DRILL-CORE SAMPLES

Coal Seam	Basis	Residual Moisture %	Ash %	Volatile Matter %	Fixed Carbon %	Calorific Value kilojoules per kilogram	F.S.I.
#2	adb	2.3	30.0	30.0	35.0	20934	0.5
	db		31.8	31.8	36.4	21164	
#1 RIDER	adb	2.3	20.0	35.0	42.0	25121	1.0
	db		20.9	36.1	43.0	25615	
#1	adb	2.5	9.5	37	50.0	28629	1.0
	db		10.0	38.3	51.7	29370	

TABLE 4-3-2
SUMMARY OF AVERAGE ULTIMATE ANALYSIS VALUES OF QUINSAM DRILL-CORE SAMPLES

Coal Seam	Basis	Residual Moisture %	Ash %	Carbon %	Hydrogen %	Nitrogen %	Oxygen %
#2	adb	2.65	28.85	51.61	3.88	0.73	12.34
	db		29.64	53.01	3.68	0.75	10.26
#1 RIDER	adb	2.3	20.0	60.18	4.34	0.79	10.57
	db		20.9	61.63	4.18	0.79	8.69
#1	adb	2.76	9.91	69.97	4.81	0.88	13.86
	db		10.19	71.95	4.63	0.91	11.73

TABLE 4-3-3
SULPHUR FORMS

Coal Seam	Pyrite %	Sulphate %	Organic %	Total %
#2	1.99	0.08	0.52	2.59
	2.04	0.08	0.54	2.66
#1 RIDER	2.87	0.16	0.66	3.69
	2.93	0.16	0.67	3.76
#1	0.23	0.01	0.33	0.57
	0.24	0.01	0.34	0.59

TABLE 4-3-4
AVERAGE VALUES OF ASH ANALYSIS ON QUINSAM DRILL-CORE SAMPLES

SEAM	$Si0_2$ %	AL_20_3 %	Fe_20_3 %	Mg0 %	Ca0 %	Na_20 %	K_20 %	$Ti0_2$ %	P_20_5 %	$S0_3$ %
#2	44.25	30.24	12.33	1.17	5.86	0.63	0.36	2.42	<0.010	4.07
#1 RIDER	36.14	21.34	19.35	0.76	10.49	0.31	0.38	1.82	0.322	9.17
#1	26.63	21.34	07.93	0.68	28.99	0.28	0.12	1.39	0.39	8.72

Plate 4-3-2. JKS Boyles Ltd. X-ray drill.

double swivel type of core barrel with an internal diameter of 35.0 millimetres (IAX); average core recovery was 96.6 per cent to depths of 34.25 and 54.25 metres. Hole GSB88-2 was drilled with a double swivel type of core barrel, with an internal diameter of 22.3 millimetres (EX). The core recovery was 89 per cent over a depth of 45.5 metres.

LOCATION OF STUDY AREA

The area chosen for the drilling test was the Quinsam coal mine which provides easy access and abundant water, and where environmental disturbance is minimal. The mine is located some 20 kilometres west of Campbell River on the east coast of Vancouver Island. The three x-ray holes were drilled on Line 82 + 50 of the Quinsam mine grid, and the two packsack holes were sunk near the pit high-wall (Figure 4-3-1).

GEOLOGICAL SETTING

The coal measures, consisting of sandstone, siltstone and mudstone, overlie the Benson basal conglomerate of the Upper Cretaceous Comox Formation of the Nanaimo Group. The sequence is moderately deformed by block faulting and tilting to the northeast. The general strike of the sediments in the vicinity of the No. 2 pit is 305° and the average dip is 7° to the northeast. The regional geology of the area is described by Bickford *et al.* (1989).

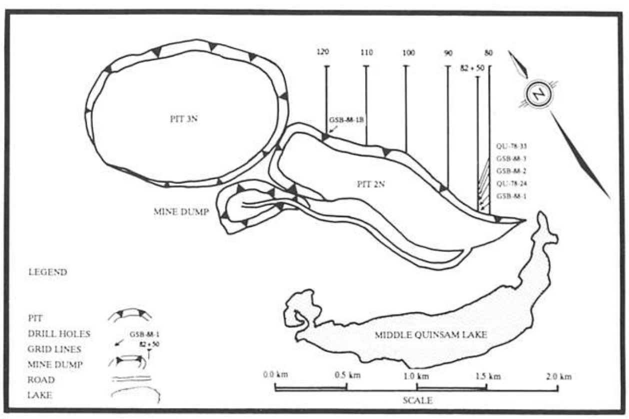

Figure 4-3-1. Quinsam mine site.

SAMPLING AND ANALYSIS

Coal was sampled in 10-centimetre increments; a few of the samples varied from 5 to 15 centimetres. The samples were crushed to -20 mesh. Petrographic rank was determined in-house by the vitrinite reflectance method and the following analyses were conducted by Chemex Laboratories of Vancouver: proximate, ultimate, sulphur forms, calorific value, free swelling index, ash analysis, chlorine, fluorine and mercury contents, and ash fusion.

DESCRIPTION OF COAL MEASURES

A typical section of coal measures, derived from the drill core, is illustrated in Figure 4-3-2. A generalized cross-section, derived from three diamond-drill holes (this study) and two exploration rotary-drill holes, is shown in Figure 4-3-3. Two coal zones, No. 1 and No. 2, occur in a succession of siltstones. The No. 1 coal zone consists of the No. 1 seam, which averages 2.3 metres in thickness, and a rider averaging 0.4 metre thick. The No. 2 zone varies in thickness from 1 to 8 metres with bands of coal up to 0.2 metre.

RESULTS AND DISCUSSION

A summary of analytical results from the 1988 Geological Survey Branch diamond-drill hole sampling program is presented in Tables 4-3-1 through to 7. Values are averages for each of No. 1 seam, No. 1 rider and No. 2 seam.

NO. 1 COAL ZONE

The fixed carbon content on a dry basis is 51.7 per cent for the No. 1 seam and 43 per cent for the rider. The volatile matter content for the No. 1 seam on a dry basis is 38.3 per cent and 36.1 per cent for the rider (Table 4-3-1). These values coupled with the calorific value, place the rank of the coal in the high-volatile bituminous B category.

The free swelling index is less than 4, placing the seam in the thermal coal category. The ash content seam is low, making the coal suitable for both power generation and cement manufacture. The volatile matter is, however, a little on the high side. The calorific value is acceptable for both applications as it is greater than the minimum of 21 000 kilojoules per kilogram (a.d.b.) required for cement manufacturing (Table 4-3-1).

The hydrogen content of the No. 1 seam, converted to a dry mineral-matter-free basis is 5.5 per cent. Given that the hydrogen range for bituminous coals is 4.5 to 5.5 per cent, clearly there has been little depletion of hydrogen which would normally result in the formation of methane. The oxygen content of the seam, on the same basis, averages 15.87 per cent, indicating that there may have been some depletion of oxygen resulting from a natural aging process. The nitrogen content of most samples ranges from 0.5 to 2.0 per cent. In this case it is reasonably low and as a result, the conversion to oxides (NO_x), a pollutant, is not a serious consideration (Table 4-3-2). In the No. 1 zone the sulphur

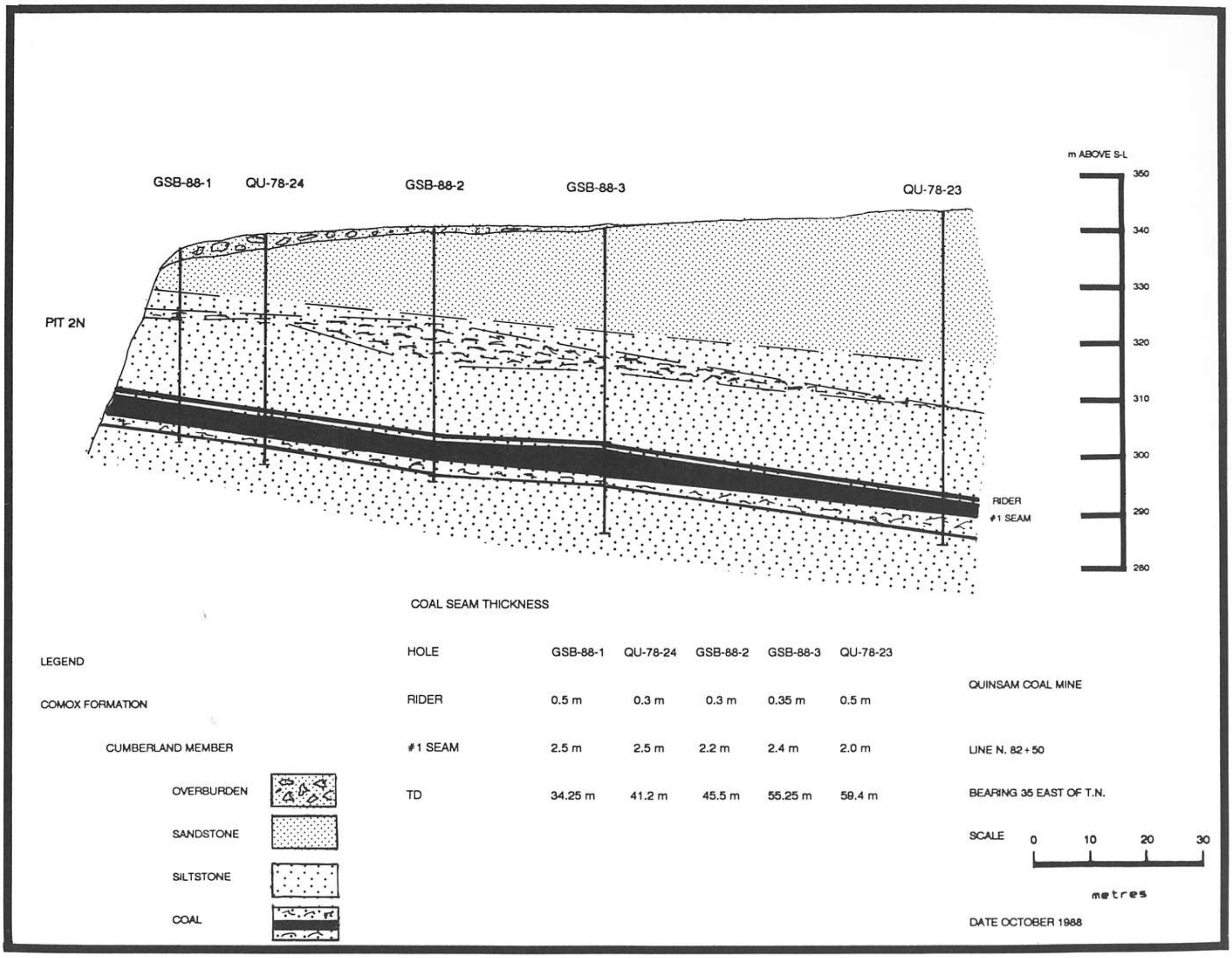

Figure 4-3-2. Type section, Comox Formation coal measures.

(pyritic and total) decreases with depth, with the highest concentrations in the rider; this is typical of a marine incursion subsequent to deposition of the coal. The sulphate content is low (0.01 per cent), unlike that of highly weathered or oxidized coals. In general the total sulphur is well within the acceptable limits for power generation and cement manufacture (Table 4-3-3).

The major components of the ash (Table 4-3-4) are SiO_2 at 26.63 per cent, CaO at 28.99 per cent, Al_2O_3 at 21.34 per cent and Fe_2O_3 at 7.93 per cent for the No. 1 seam. Lime acts as a mild flux and in this case it is noticeably high. The P_2O_5 content is very low at 0.39 per cent, well below the maximum of 1 per cent allowable in clinker for cement manufacture.

Chlorine at 0.017 per cent, is well below the maximum of 0.1 per cent above which it would cause ash fouling in boilers (Table 4-3-5). Fluorine is very low, however in cement manufacture a certain amount is beneficial as it acts as a flux in the burning of clinker. Mercury appears to be higher in parts of No. 1 rider than the main seam.

The ash fusion temperatures fall in the medium range (Table 4-3-6), between 1350 and 1300°C.

The reflectance values place the coals in the high-volatile bituminous category ($R_o$max between 0.50 and 1.12 per cent); the No. 1 rider is generally lower in reflectance values than both the No. 1 and No. 2 seams (Table 4-3-7).

CONCLUSIONS

The subsurface coal sampling program started in 1988 as a pilot project with a modest budget. Three diamond-drill holes were put down using an x-ray drill, one hole recovering EX core and the other two an IAX core. A short fourth hole was drilled using a packsack drill. A total of 138.3 metres was drilled, from which 115 coal samples were taken from an aggregate coal thickness of 11.5 metres. The core recovery was excellent due mainly to the driller's technique, and the character of the coal and the sediments.

As a result of the 1988 project, the Geological Survey of Canada participated in the 1989 field project, doubling the drilling budget and assuming responsibility for all analysis except the reflectance and low-temperature ashing which will be done in-house. The raw analytical data will be published at a later date as an Open File.

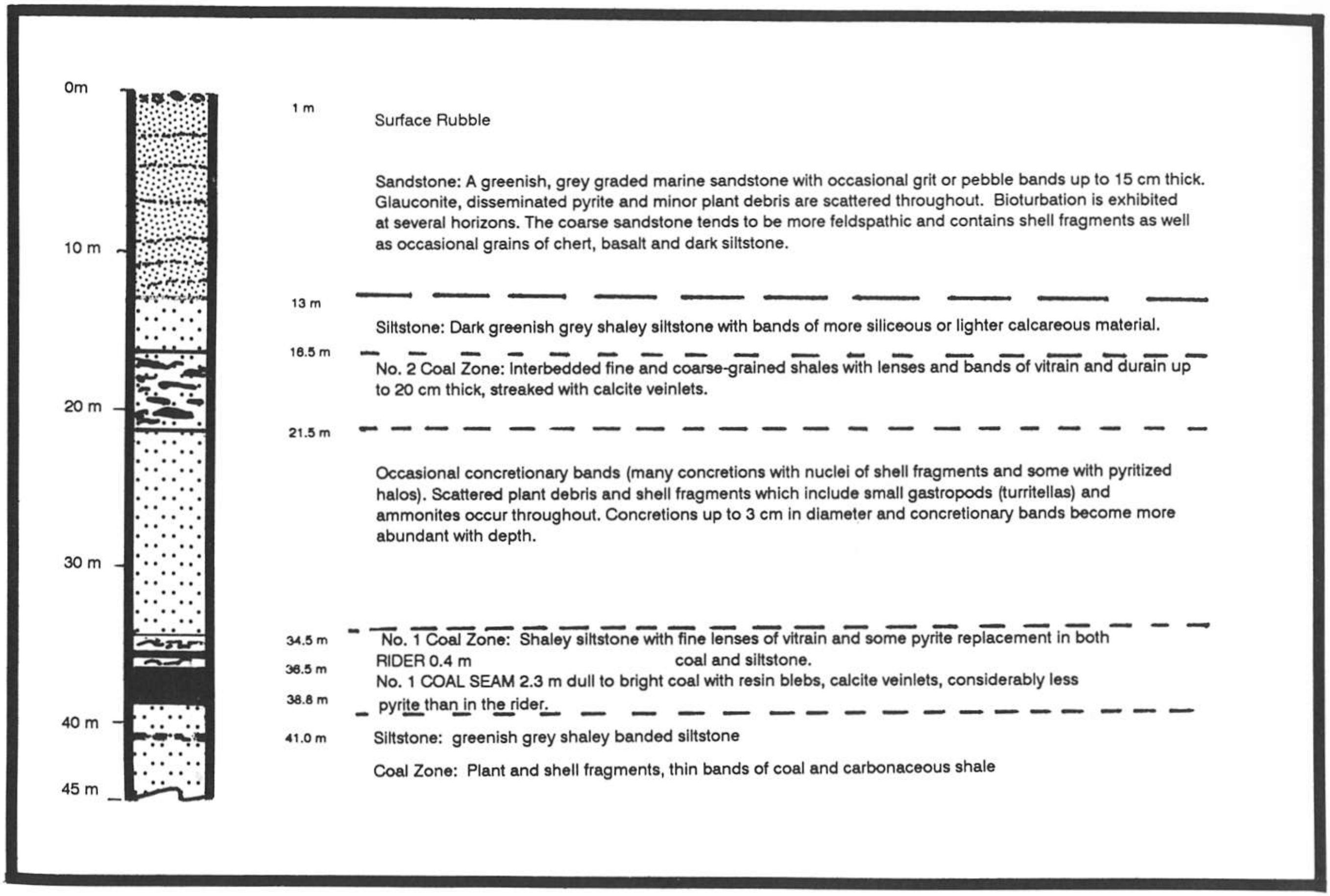

Figure 4-3-3. Cross section, Quinsam mine.

TABLE 4-3-5
AVERAGE VALUES OF CHLORINE, FLUORINE AND MERCURY IN QUINSAM DRILL-CORE SAMPLES

SEAM	Cl %	F ppm	Hg ppb
#2	<0.010	<5	225
#1 RIDER	0.017	<5	263
#1	0.017	<5	175

TABLE 4-3-6
AVERAGE VALUES OF ASH FUSION TEMPERATURE IN REDUCING ATMOSPHERE FOR QUINSAM DRILL CORES

Seam	Initial Deformation °C	Spherical °C	Hemispherical °C	Fluid °C
#2	1380	1410	1430	1450+
#1 RIDER	1248	1278	1300	1313
#1	1318	1338	1348	1363

TABLE 4-3-7
AVERAGE AND RANGE OF VITRINITE REFLECTANCE OF QUINSAM DRILL-CORE SAMPLES

Coal Seam	Number of Samples	Mean Maximum Average	Mean Maximum Range	Mean Random Average	Mean Random Range
#2	8	.68	.58–.78	.65	.56–.73
#1 RIDER	14	.54	.47–.69	.52	.45–.67
#1	89	.72	.60–.85	.68	.60–.82

ACKNOWLEDGMENTS

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NOTES

SUBSURFACE COAL SAMPLING SURVEY, TELKWA AREA, CENTRAL BRITISH COLUMBIA (93L/11)

By Alex Matheson and Brad Van Den Bussche

KEYWORDS: Coal geology, Telkwa, Lower Cretaceous, coal measure stratigraphy, prospector drill, core recovery.

INTRODUCTION

The 1989 coal sampling survey at Telkwa followed the pilot project at Quinsam mine the preceding year (Matheson, 1990, this volume). The budget for the drilling was augmented by funds from the Institute of Sedimentary and Petroleum Geology, enabling a total of 280 metres to be drilled, double that of the previous year. All coal exposures near the drill sites were sampled in addition to the coal recovered as drill core. The drilling was performed under contract by Neill's Mining Company which had replaced the X-ray drill with the Prospector 89, a new drill manufactured by Hydrocore Drills Limited.

LOCATION OF STUDY AREA

The drilling took place in the Telkwa coalfield located to the southwest of the village of Telkwa on the Yellowhead Highway, 18 kilometres southeast of Smithers in west-central British Columbia (Figure 4-4-1). The Canadian National Railway passes through the village and connects it to the port of Prince Rupert 370 kilometres to the west.

THE PROSPECTOR 89 HYDRAULIC DRILL

The Prospector 89 drill (Plate 4-4-1) is light and portable with a total weight of about 200 kilograms (excluding the rods). The weight of the heaviest component, the engine, is 45 kilograms. The unit, which is mounted, is powered by a 16-horsepower air-cooled Briggs and Stratton engine and can drill to a vertical depth of 150 metres. A double swivel core barrel with an internal diameter of 35.0 millimetres (IAX) was used. A total of 280 metres was drilled in nine holes; core recovery varied from 90 to 98 per cent.

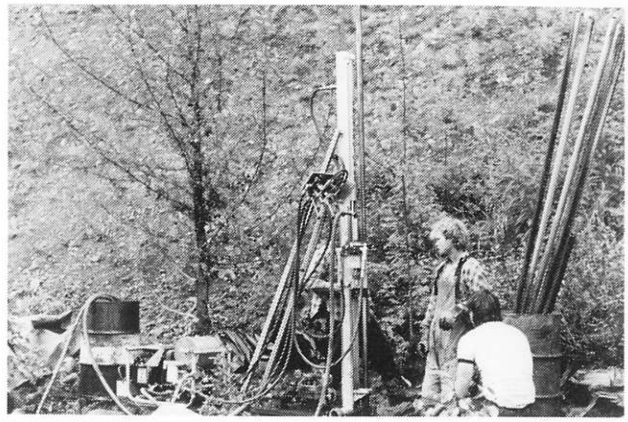

Plate 4-4-1. The Prospector 89 drill.

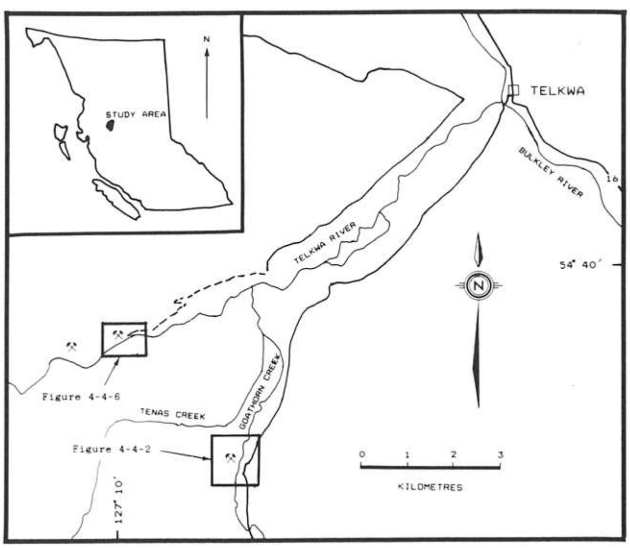

Figure 4-4-1. Location map, Telkwa coalfield, showing location of drilling areas.

GEOLOGICAL SETTING

Outcrop in the area is sparse and it is only in some of the valleys that the coal measures have been exposed by river erosion. The Lower Cretaceous Telkwa coal measures of the Skeena Group consist of interbedded marine and nonmarine sediments divided into three units (Koo, 1984). This sequence unconformably overlies volcanic rocks of the Jurassic Hazelton Group. The area was later intruded by Tertiary dikes and sills. Widespread block faulting forming horsts and grabens has been postulated from Crows Nest Resources Limited drill-hole data. The sediments in the vicinity of the drillsites strike 340° to 350° and dip 10° to 30° east.

DESCRIPTION OF THE COAL MEASURES

The Telkwa coal measures are about 400 metres thick with ten major correlatable coal seams recorded, amounting to an aggregate thickness of up to 24 metres of coal.

The lower unit comprises siltstones, sandstones and grits overlying a discontinuous basal conglomerate (Plate 4-4-2). Distribution of the conglomerate is controlled by paleotopography. Some thin coal seams and the No. 1 coal seam occur in this unit which attains a thickness of up to 120 metres in places. The No. 1 seam, near the top of the unit, ranges up to 3.5 metres in thickness. Paleosols are common throughout. Near the base the sandstones become a reddish purple in colour indicating the proximity of the underlying volcanic basement.

The middle unit of medium to fine-grained sandstones, siltstones and mudstones ranges up to 140 metres thick and is devoid of any carbonaceous material.

The upper unit comprises up to 300 metres of mudstones, siltstones and sandstones, and is characterized by an absence of coarse-grained material. The coal occurs in the lower 180 metres. There are nine correlatable coal seams varying from 0.5 to 6.0 metres thick, with an aggregate thickness varying from 13 metres to 21 metres. These seams contain the indicated reserves. Paleosols occur throughout the lower half of the unit.

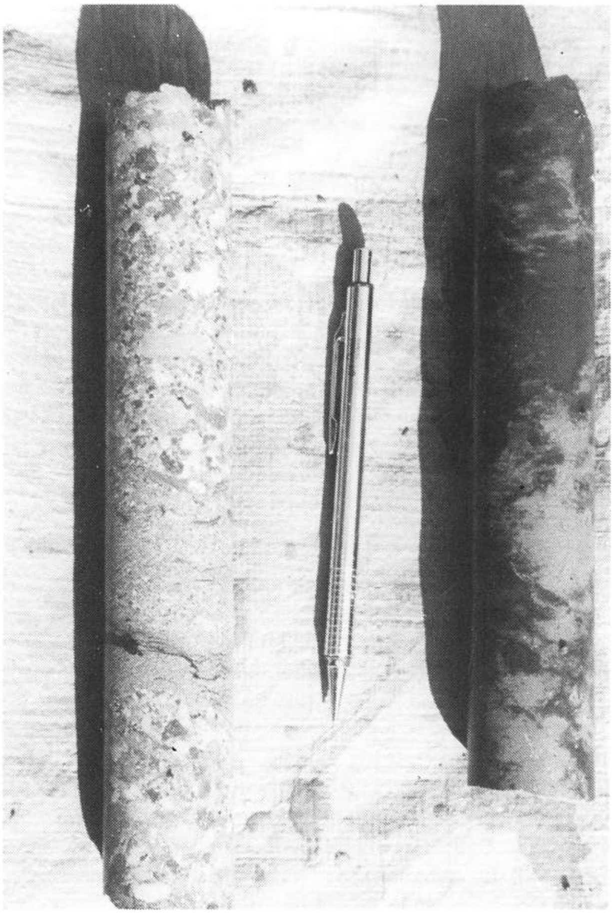

Plate 4-4-2. The basal conglomerate of the lower unit of the Telkwa coal measures.

SAMPLING AND ANALYSIS

A total of 226 samples were taken in the Telkwa coalfield during the 1989 field season; 197 from drill core and 29 from outcrop. Of the drill-hole samples, sixty were from GSB89-01 and 02, thirteen were from GSB89-03 and 04, four from GSB89-05, three from GSB89-07, and 117 were from GSB89-08 and 09. The 29 outcrop samples represent eight seams at seven sample sites.

Outcropping coal seams were described prior to sampling. After cleaning a face across the seam, channel samples were taken perpendicular to bedding. An attempt was made to collect at least three samples at each site in order to represent the upper, middle and lower parts of the seam. Samples ended after 50 centimetres or where a parting occurs, whichever came first.

The coal from the drill core was generally sampled in 20-centimetre increments or shorter intervals as dictated by partings. All samples comprise the entire section of core and were crushed to -20 mesh. Petrographic rank determinations will be done in-house by the vitrinite reflectance method. Analyses will also be made using x-ray defraction on low-temperature ash samples. The following analyses will be carried out by a private laboratory under the auspices of the Institute of Sedimentary and Petroleum Geology: proximate, ultimate, sulphur forms, calorific value, free swelling index, ash analysis, chlorine, fluorine, mercury contents and ash fusion. At the request of Dr. Fari Goodarzi the remainder of the core, after the coal had been removed, was sent to the Institute of Sedimentary and Petroleum Geology in Calgary, primarily for petrographic examination of the maceral composition of the carbonaceous material in the mudstones, siltstones and shales.

DRILLING

Drill holes GSB89-01 and 02 (Figure 4-4-2), 28.5 and 25 metres deep respectively, were each drilled 7 metres from the west bank of Goathorn Creek and 70 metres apart. The northern hole is about 100 metres south of the Bulkley Valley mine site. The coal exposed by the river was sampled

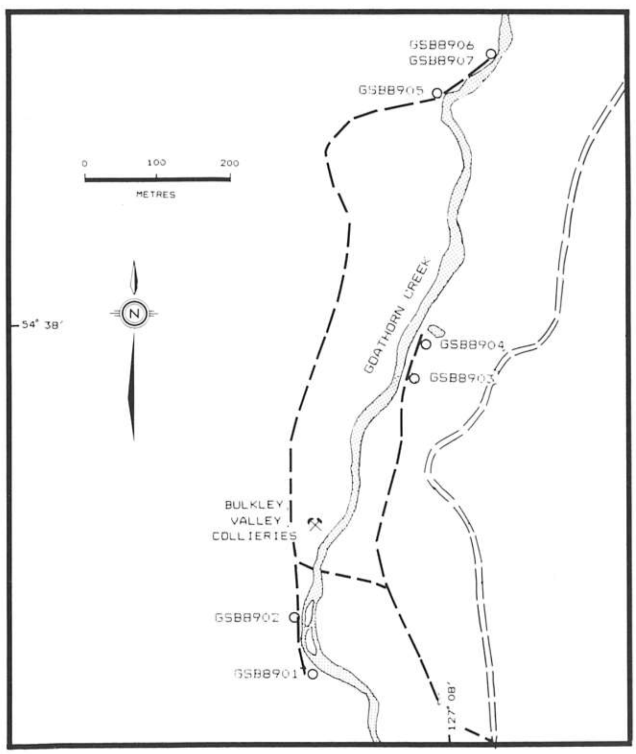

Figure 4-4-2. Location of drill holes along Goathorn Creek near the Bulkley Valley mine (*see* also Figure 4-4-1).

adjacent to each drill site. Both holes were spudded fairly low in the upper unit, along strike from each other. It is assumed that Nos. 2, 3, 4 and 5 seams were intersected (Figure 4-4-3). There is evidence of seams, splitting, pinching and swelling within the short distance separating the holes.

Drill holes GSB89-03 and 04 (Figure 4-4-2), 52.0 and 25 metres deep respectively, are located on the east bank of the Goathorn Creek about 250 metres north of the Bulkley Valley mine site. They are 55 metres apart and about 15 metres from outcrops in the river. They were spudded in the lower unit of the Telkwa coal measures and each intersected three coal zones, in all probability below No. 1 seam (Figure 4-4-4). The holes were stopped before reaching the basement.

Hole GSB89-05 is located 350 metres north of GSB89-04 on the west side of Goathorn Creek, 22 metres from a cliff face (Figure 4-4-2) and was drilled to a depth of 45.6 metres. It cut rocks of the lower unit which are generally coarser grained than the upper unit. Four minor coal zones were intersected.

Holes GSB89-06 and 07 were drilled 100 metres to the northeast (Figure 4-4-2). Hole GSB89-06 (Figure 4-4-5) was abandoned at 19.2 metres because the coal seam exposed in the river was not intersected and appeared to have been eroded. This was proven to be correct when Hole GSB89-07 was drilled halfway to the outcrop, 12 metres from the westbank of the Goathorn Creek. It intersected 0.7 metre of coal as expected and was stopped at 10.3 metres.

Holes GSB89-08 and 09 (Figure 4-4-6) are located south of the Avelling mine and north of the Telkwa River. GSB89-08 is 20 metres from the river and has a depth of 33.4 metres (Figure 4-4-7). It intersected 11 metres of coal in five seams. GSB89-09 is 40 metres from river, roughly along strike. It was drilled to 43.3 metres and intersected five coal seams with an aggregate thickness of 9.6 metres. Both holes are in the upper unit of the Telkwa coal measures and possibly

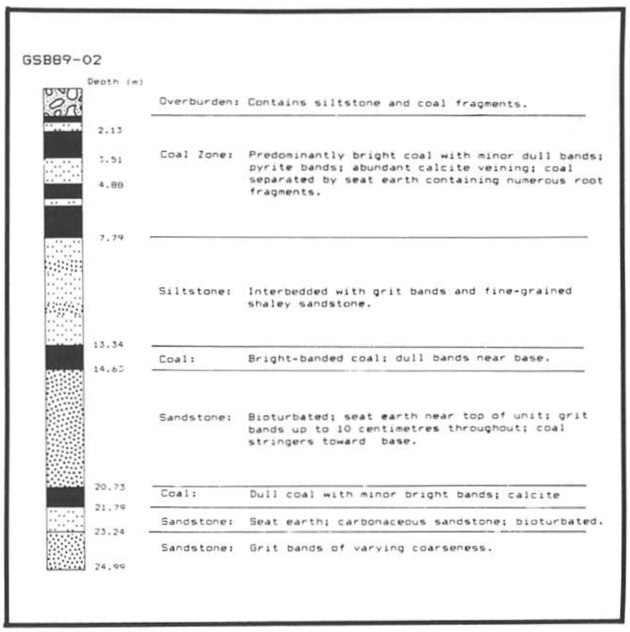

Figure 4-4-3. Simplified stratagraphic log of Hole GSB89-02 (for location *see* Figure 4-4-2).

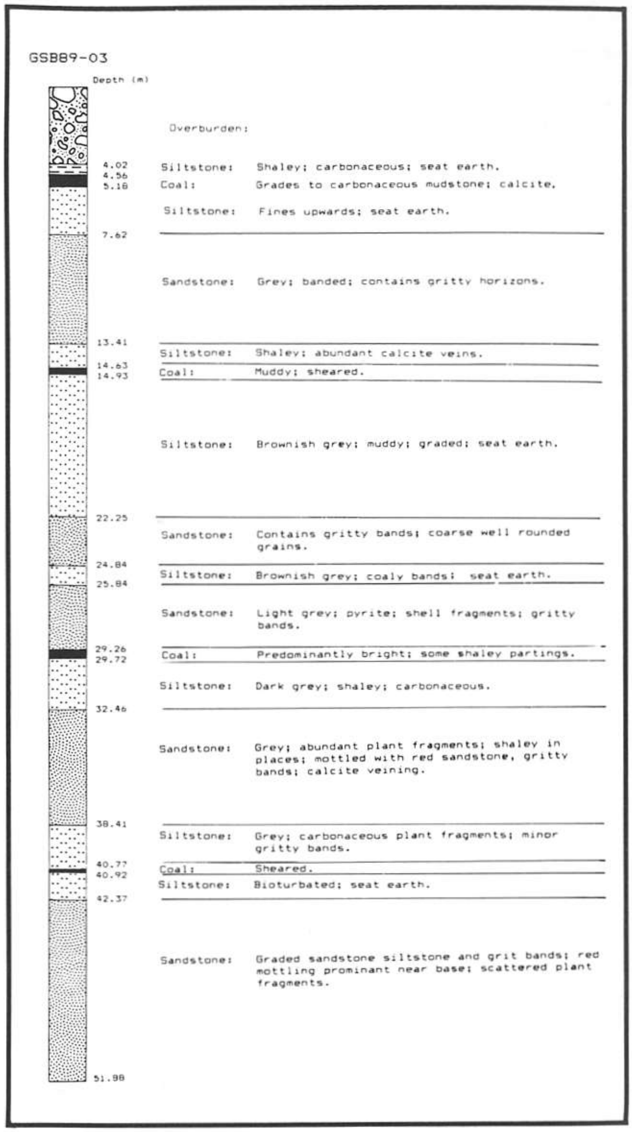

Figure 4-4-4. Simplified stratagraphic log of Hole GSB89-03 (for location *see* Figure 4-4-2).

low in the sequence as indicated by the grouping of the coal seams. Here again there is evidence of seams splitting, thinning and swelling over a very short distance.

CONCLUSIONS

The drill achieved an average core recovery of 95 per cent, which is excellent, and is well suited to drilling in sedimentary rocks. Core recovery in the coal was particularly good, about 97 to 98 per cent. This, however, may be partially attributable to the character of the coal. In the streams the coal is more resistant to erosion than the surrounding sediments.

The drill is easily portable, in sections, along a prepared foot trail. The success of the project has resulted in plans by two major coal mining companies to use the drill for exploration work and for coal-quality control in pit areas.

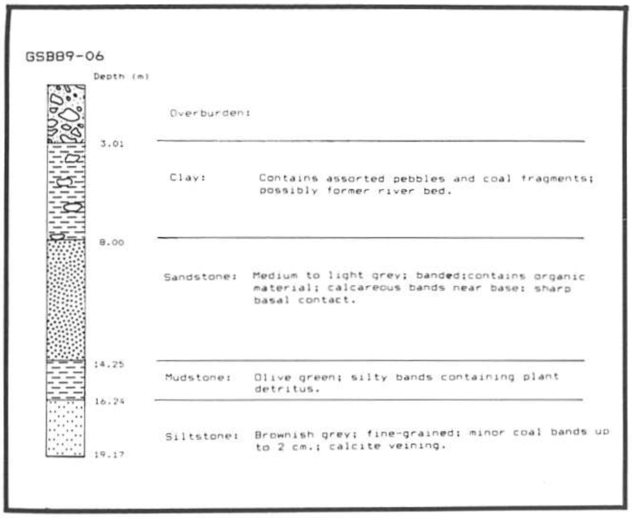

Figure 4-4-5. Simplified stratagraphic log of Hole GSB89-06 (for location *see* Figure 4-4-2).

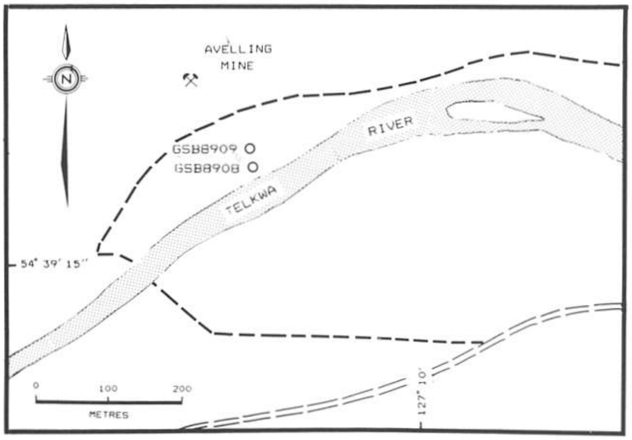

Figure 4-4-6. Location of drill holes along the Telkwa River near Avelling mine (*see* also Figure 4-4-1).

ACKNOWLEDGMENTS

The authors would like to extend their appreciation to Brian McKinstry of Crows Nest Resources Limited for his very kind assistance, to Fari Goodarzi for his keen interest and participation in the program, to their colleagues of the Coal Subsection, to Kenny Neill for his assistance in the field and crushing room, and to Rebecca Arnet for her stenographic support.

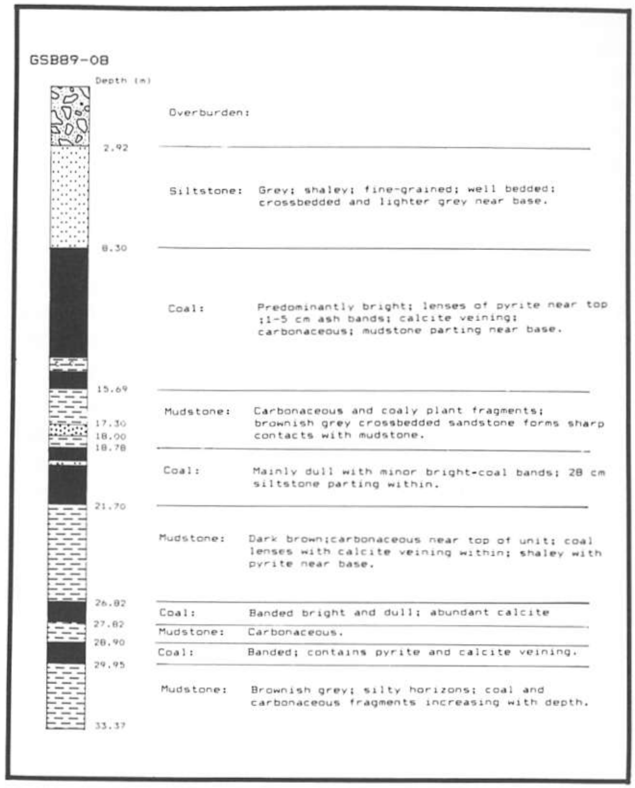

Figure 4-4-7. Simplified stratagraphic log of Hole GSB89-08 (for location *see* Figure 4-4-6).

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STRATIGRAPHY AND SEDIMENTOLOGY OF THE LOWER SKEENA GROUP, TELKWA COALFIELD, CENTRAL BRITISH COLUMBIA (93L/11)

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KEYWORDS: Coal geology, sedimentology, Telkwa coalfield, Skeena Group, Red Rose Formation.

INTRODUCTION

Significant coal resources occur in the Lower Cretaceous Skeena Group near the town of Telkwa in central British Columbia. The coal deposits have been mined on a small scale since the early 1900s and a major exploration project designed to delineate further exploitable coal reserves is currently in progress (Handy and Cameron, 1983). Early attempts to resolve the lithostratigraphy and subsequently interpret the depositional history of the coal measures have been hampered by the lack of outcrops and subsurface information. As a result of recent exploration a large quantity of drill cores, geophysical logs and data are now available, facilitating a detailed sedimentological and coal-quality study of the coal measures in the Telkwa area.

The objectives of this study are: to describe and define the stratigraphy and lateral facies changes of the Skeena Group; to interpret the depositional history of coal measures in the Telkwa area; and to relate the quality and mineability of the coals to their depositional environments. The results of this study will contribute to understanding of the regional geology of Cretaceous basins and evolution of the Skeena arch in the Intermontane Belt in central British Columbia. In addition, the results will aid in the development of depositional models to predict variations in coal quality in other basins.

This report summarizes the preliminary results of fieldwork completed during the summer of 1989. It describes the stratigraphy and major facies changes, and interprets the depositional history of the eastern half of the Telkwa coalfield.

LOCATION AND REGIONAL GEOLOGY

The Telkwa coalfield is in the Intermontane Belt of west-central British Columbia, approximately 18 kilometres south of Smithers, near the town of Telkwa (Figure 4-5-1). The Telkwa River and Goathorn Creek run through the basin, dissecting it into three segments, locally known as the North Telkwa, West Goathorn and East Goathorn areas. The East Goathorn area is the subject of this paper.

The coalfield is located on the northern flank of the Skeena arch, near the southern limit of Jurassic sediments in the Bowser Basin. The coal measures, which are included in the Lower Cretaceous Skeena Group, are an erosional remnant of a larger Cretaceous basin which Hunt and Bustin (in press) refer to as the "Nazko basin". Local tectonism was a significant factor during the deposition of the coal measures.

The Skeena Group comprises a Lower Cretaceous sedimentary unit and an Upper Cretaceous volcanic unit. Sutherland Brown (1960) applied the names Red Rose Formation and Brian Boru Formation to the lower sedimentary and upper volcanic units respectively. Tipper and Richards (1976) follow Sutherland Brown's nomenclature, but include the Lower Cretaceous Kitsun Creek Formation, the Rocky Ridge volcanics and other unnamed sediments at the base of the Skeena Group. Tipper and Richards assigned the coal measures in the Telkwa area to undifferentiated Lower Cretaceous Skeena Group. MacIntyre *et al.* (1989) include them in the Red Rose Formation. In this discussion, the sediments of the Telkwa coalfield are referred to as Lower Skeena Group. The name Red Rose Formation will not be applied to the Telkwa coal measures; Tipper and Richards describe the Red Rose Formation as containing chert-pebble conglomerates, whereas in the Telkwa area, conglomerates consist entirely of volcanic clasts.

Deposition of the Skeena Group sediments began during the Early Cretaceous, following regional uplift and erosion of

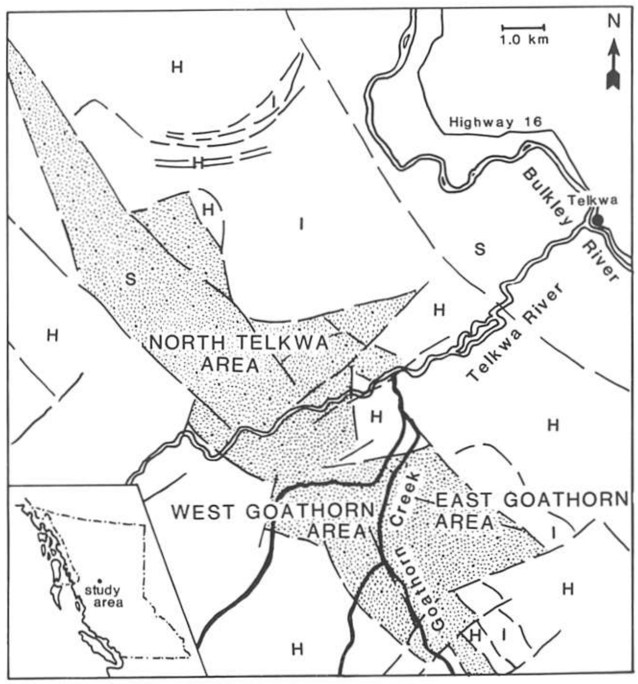

Figure 4-5-1. Location map of study area. S = Skeena Group, H = Hazelton Group, I = Tertiary intrusions. Telkwa coalfield is shown stippled. (Modified after MacIntyre *et al.*, 1989).

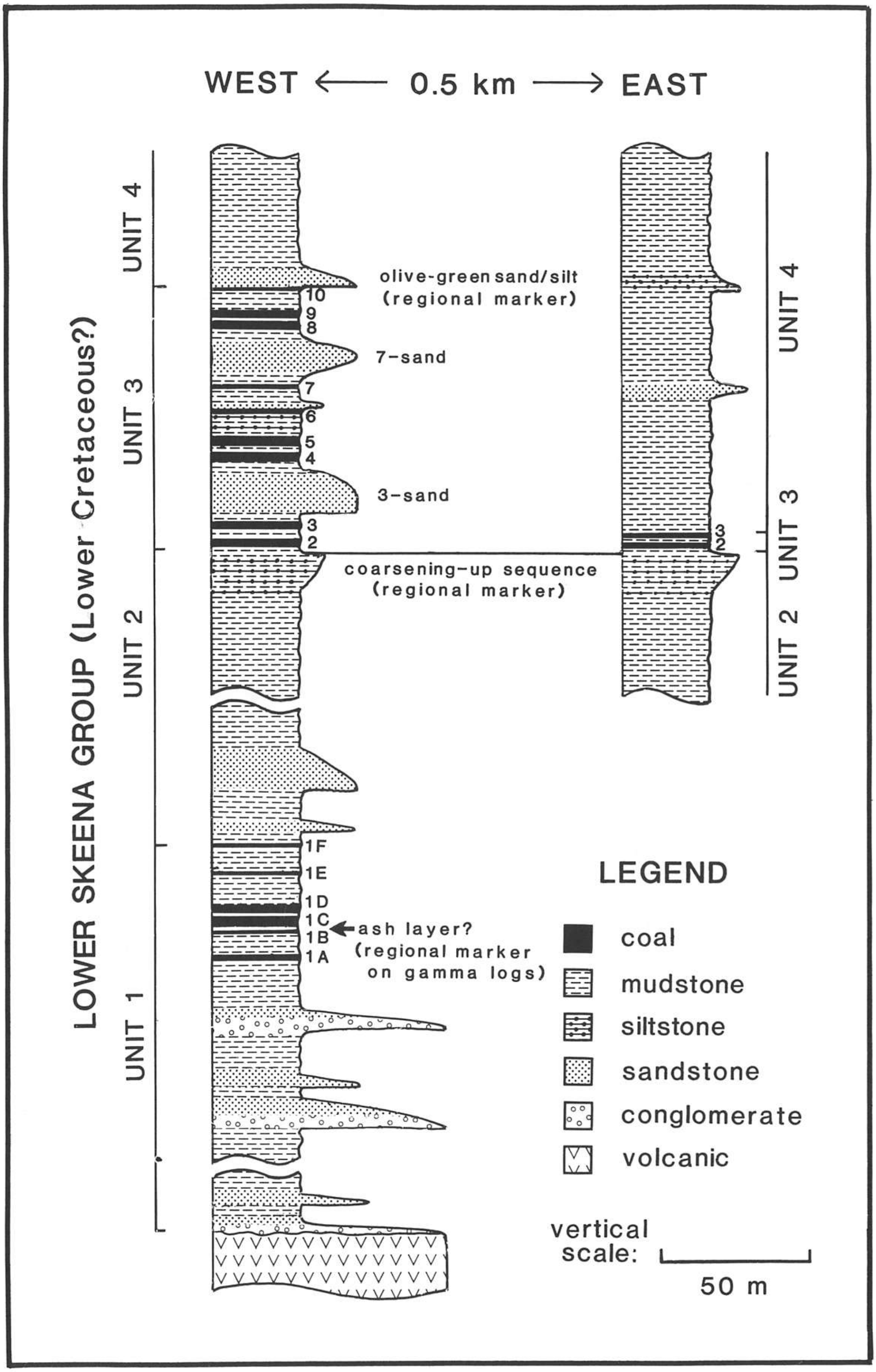

Figure 4-5-2. Representative stratigraphic columns of the Telkwa coalfield.

the Skeena arch. The sediments of the Skeena Group were transported southwest across the arch from the Pinchi belt and Columbian orogen (Tipper and Richards, 1976). The lower sedimentary unit is comprised of marine or continental and coal-bearing beds, which typically contain an abundance of fine-grained detrital muscovite (Leach, 1910; Sutherland Brown, 1960; Tipper and Richards, 1976).

LOCAL STRATIGRAPHY AND STRUCTURE

The Telkwa coalfield is fault bounded on all sides by the Lower Jurassic Telkwa Formation of the Hazelton Group (MacIntyre *et al.*, 1989) (Figure 4-5-1). The displacement on the faults is primarily vertical and estimated to be between 50 and 300 metres (Koo, 1983). Several high-angle faults strike north and west across the basin, with normal or reverse displacement of up to 30 metres (Koo, 1983). In addition, thrust faults strike northwest across the southeastern corner of the coalfield causing repetition of strata in some drill holes. Porphyritic Tertiary dikes and sills intrude the sediments throughout the coalfield, and a large granodiorite and quartz monzonite intrusion is present in North Telkwa (Koo, 1983).

More than 500 metres of Lower Skeena Group unconformably overly Hazelton Group volcanics. Plant fossils found in Skeena sediments in the Telkwa coalfield were dated as Early Cretaceous by Hacquebard *et al.* (1967). Palynological data from the base of the stratigraphic succession indicate that the oldest strata are Neocomian (Handy and Cameron, 1983).

Over most of the study area, the stratigraphic succession is divisible into four informal units based npon gross lithology (Figure 4-5-2). Coal occurs in ten "zones" (Handy and Cameron, 1983); each zone consists of one or more seams which are split by a thickness of organic-rich mudstone that is less than or equal to the thickness of the coal seams themselves. The one exception is Coal Zone 1, in which all the laterally correlatable seams are grouped together as one zone, despite the fact that the seams may be separated by as much as 8 metres of carbonaceous mudstone.

The four lithological units defined in this study correspond to those defined by Handy and Cameron (1983), and the three lowest units correspond roughly to those recognized by Koo (1983). The lowest unit (Unit I) unconformably overlies the volcanic basement and consists of fine to coarse-grained crossbedded lithic sandstone, volcanic-derived conglomerate, siltstone, mudstone and coal (Coal Zone 1). Coal Zone 1 contains a highly radioactive interval which is a regional marker on gamma ray logs. Unit II overlies Coal Zone 1 and consists primarily of monotonous silty mudstone, with occasional siltstone and sandstone beds. The top of Unit II is defined by a 10 to 15-metre coarsening-upward sequence that serves as a regional marker on geophysical logs. Unit III, which overlies the coarsening-upward sequence, comprises micaceous sandstone, siltstone and mudstone, and up to nine coal zones (Coal Zones 2 through 10). The uppermost unit, Unit IV, overlies Coal Zone 10 and is characterized by monotonous silty mudstone similar that in Unit II. A 2 to 4 metre thick bright green sandstone or siltstone marks the base of Unit IV and serves as a regional marker in the drill core.

The stratigraphic succession changes markedly from west to east. In the eastern margin of the study area, although the same stratigraphy is recognized, Unit III is considerably thinner, lacks sandstone, and eontains only the two lowest coal zones (Coal Zones 2 and 3).

UNIT I:

Unit I comprises conglomerate, sandstone, siltstone, mudstone and coal between the unconformable contact with volcanic basement and the top of Coal Zone 1 (Figure 4-5-2). It ranges in thickness from 50 metres to more than 100 metres, reflecting paleotopographic relief on the underlying basement.

The contact of the Skeena Group with volcanic basement was not observed in drill holes from the East Telkwa area. In the North Telkwa area, volcanics are bleached from green to brown or maroon near the contact. The basal strata of the Skeena Group consist of horizontally bedded or crossbedded, poorly sorted granule or pebble conglomerate, or coarse sandstone. The clasts are subangular and derived from Hazelton Group volcanic rocks. Pyritized coal spar and mudstone clasts are common.

In general, the sediments of Unit I consist of fining-upward sequences, 2 to 10 metres thick, comprising green and maroon carbonaceous conglomerate and sandstone, separated by organic-rich mudstone and thin coal seams. The fining-upward sequences have sharp, erosional bases. Sorting is poor at the base and moderate at the top. The top of Unit I contains Coal Zone 1, which consists, on average, of six seams (locally termed seams 1A, 1B, 1C, 1D, 1E and 1F) interbedded with carbonaceous mudstone in a 30-metre interval. Coal seams 1A, 1D, 1E and 1F are usually less than 1 metre thick, whereas coal seams 1B and 1C are 1 to 3 metres thick. All six seams are not found throughout the area; often they are thin, laterally discontinuous and impossible to correlate.

UNIT II:

Unit II is between Coal Zones 1 and 2 and is comprised primarily of monotonous grey mudstone or silty mudstone, with occasional light grey beds of bioturbated sandstone (Figure 4-5-2). Unit II has a minimum thickness of approximately 100 metres and is thought to exceed 140 metres (Koo, 1983). No complete section of Unit II was penetrated in a single drill hole.

There is an overall decrease in sandstone relative to mudstone from the base to the top of the unit. The top 50 metres consists almost entirely of silty mudstone. Silty mudstone throughout the unit has occasional centimetre-scale, very bioturbated (mottled) siltstone laminae, rare gastropods and inoceramid-like fragments (P. Smith, personal communication, 1989), and little organic matter other than rare pyritized carbonaceous fragments. The sandstone beds have gradational lower and upper contacts and range in thickness from less than 1 metre to 30 metres. Often the sandstone is thoroughly bioturbated, resulting in a mottled texture. A variety of trace fossils occur, including *Teichichnus–Skolithos–Paleophycus* assemblages and monospecific *Macaronichnus* and *Helminthopsis* assemblages. In rare undisturbed intervals, pelecypod shell hashes, crossbedding, and isolated granules and pebbles occur.

The top of Unit II is marked by a 10 to 15-metre coarsening-upward sequence, abruptly overlain by organic-rich mudstone followed by Coal Zone 2. The coarsening-upward sequence can be correlated throughout the study area. It is characterized by an upward-increasing abundance of irregular, planar, lenticular and wavy bedded siltstone or fine-grained sandstone laminae and beds.

Unit III:

Unit III averages about 90 metres thick and includes the sediments between the base of Coal Zone 2 and the top of Coal Zone 10 (Figure 4-5-2). It comprises a diverse sequence of mudstone, siltstone, sandstone and nine distinct coal zones.

Coal Zone 2 averages 2 to 4 metres in thickness and consists of one or two 0.5 to 3-metre seams. Coal Zone 3 is 2 to 4 metres thick and comprises three seams 0.5 to 2 metres thick. Where split, Coal Zones 2 and 3 are separated by up to 4 metres of organic-rich mudstone and irregular millimetre-scale siltstone or fine sandstone beds. Coal Zone 3 is gradationally or sharply overlain by light grey, fine-grained bioturbated sandstone (herein referred to as "three-sand") up to 18 metres thick (Figure 4-5-2). Three-sand has abundant thin mudstone laminae. Its upper contact with organic-rich mudstone and Coal Zone 4 is usually sharp. Occasionally, three-sand is rippled or crossbedded with scour surfaces and shell hashes, but bioturbation is the predominant structure and the trace fossils *Skolithos, Zoophycus* and *Planolites* are present. Three-sand also contains a diverse assemblage of marine pelecypods (P. Smith, personal communication, 1989).

Three-sand thins and disappears from west to east across the coalfield (Figure 4-5-3). Its disappearance coincides with the loss of Coal Zones 4 through 10 in the eastern half of the study area (Figure 4-5-2). Thus, in the eastern half of the study area, Coal Zone 3 is directly overlain by Unit IV; whereas elsewhere Unit IV overlies Coal Zone 10.

On average, Coal Zone 4 is 3.5 metres thick and consists of one or two seams 0.5 to 3 metres thick. Coal Zone 5 is 2 to 3 metres thick and comprises one or two seams 0.5 to 4 metres thick. Coal Zones 4 and 5 are separated by up to 3.5 metres of organic-rich mudstone. Above Coal Zone 5, strikingly regular, thinly laminated (1 millimetre) lenticular siltstone and carbonaceous mudstone are present. Well-preserved plant fossils and syneresis cracks are common. Coal Zone 6 is 2.5 to 4 metres thick and consists of one or two 0.5 to 3-metre seams. It is generally overlain by carbonaceous rooted mudstone, but occasionally by a thin, fine-grained, 1 to 2-metre sandstone resting on an erosional contact. The sandstone has limited lateral extent and fines upward into organic-rich mudstone followed by Coal Zone 7.

Coal Zone 7 comprises a single 1-metre seam. It is overlain by mudstone with moderately abundant carbonaceous fragments and occasional silt laminations and rooted intervals. The mudstone grades upward into a 3 to 11 metre thick, fine to medium-grained grey-green sandstone referred to as "seven-sand" (Figure 4-5-2). It is horizontally bedded or massive, with occasional crossbeds or scour surfaces. Thin mudstone beds with the trace fossil *Planolites* are common within seven-sand. Organic matter is abundant in roots and thin carbonaceous laminations or scattered throughout. Seven-sand coarsens and then fines upward, grading into mudstone and Coal Zones 8 and 9.

Coal Zones 8 and 9 each contain a single seam 1 to 2 metres thick. The seams are separated by less than 2 metres of carbonaceous mudstone, and in places Coal Zones 8 and 9 are considered one zone. Coal Zone 9 is overlain by organic-rich mudstone with siltstone interbeds. In some areas, a horizontally bedded fine-grained, green-grey sandstone, up to 5 metres thick with frequent mudstone laminations and moderately abundant organic matter, overlies Coal Zone 9. Coal Zone 10 is less than a metre thick, where present.

Unit IV:

Unit IV overlies Coal Zone 10 in the western part of the study area and Coal Zone 3 in the eastern part, and is at least 40 metres thick. It is unconformably overlain by Quaternary sediments. Most of the unit resembles the upper part of Unit II, consisting primarily of silty mudstone with the trace fossil *Helminthopsis*, thin mottled siltstone laminations, bioturbated sandstone beds, and very rare carbonaceous matter.

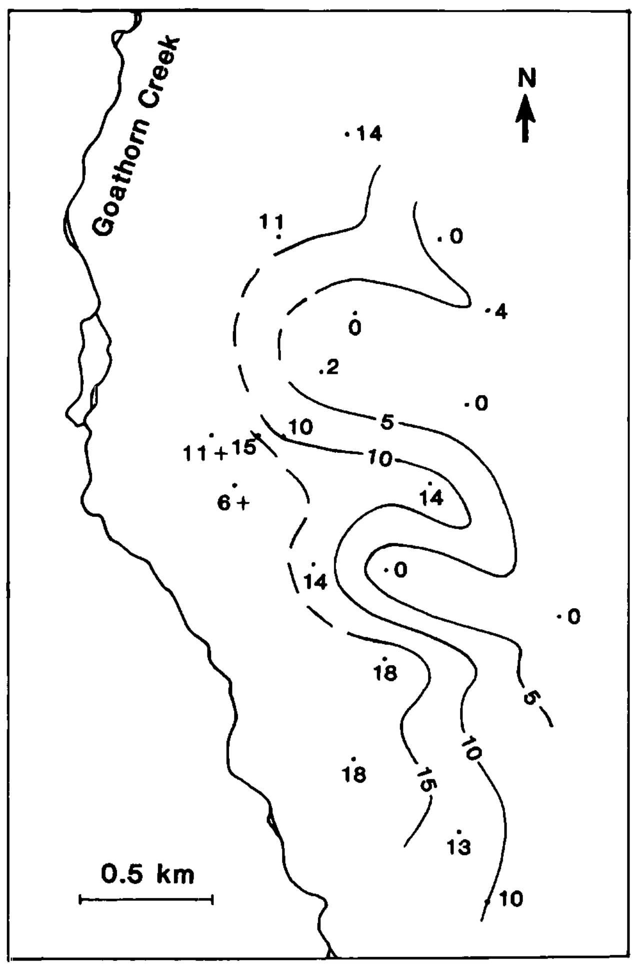

Figure 4-5-3. Isopach map (in metres) of three-sand, Telkwa coalfield.

In the western half of the study area, the base of Unit IV is marked by a thick, chloritic, bright green, massive, very fine grained sandstone or siltstone 2 to 4 metres thick. It sharply overlies Coal Zone 9 or 10. It fines upward very gradationally into the silty mudstone described above. In the eastern half of the study area, a similar green siltstone occurs at the same stratigraphic horizon (Figure 4-5-1), but it no longer marks the base of Unit IV.

DEPOSITIONAL ENVIRONMENTS

In previous studies, the Telkwa coal measures have been described as lacustrine (Hacquebard *et al.*, 1967), fluvial and shallow marine (Koo, 1983), and fluvial, deltaic and marine (Handy and Cameron, 1983). Detailed core logging during the 1989 field season, together with trace and macrofossil identification has led to further refinements in the interpretation of the depositional environments of the strata.

The Lower Skeena Group in the Telkwa area was deposited in alternating nonmarine, transitional, and open-marine environments. Unit I was deposited in a fluvial environment. Palaeotopographic lows were filled with coarse-grained sediments derived from the eroded Hazelton volcanic basement. Eventually a peat-forming floodplain was established, resulting in the formation of Coal Zone 1. Peat formation was periodically interrupted by migrating channels, resulting in laterally discontinuous, thin seams.

Unit II was deposited in a marine lower-shoreface environment. This is suggested by the presence of inoceramids, *Helminthopsis, Teichichnus, Skolithos* and *Paleophycus* in the mudstone and siltstone (Ekdale *et al.*, 1984). The paucity of organic matter (restricted to rare occurrences of coal spar and pyritized plant matter) is consistent with a lower-shoreface marine environment. The burrowed sandstone of Unit II, with relatively more organic matter and rare shells and granules, may indicate transitions to upper-shoreface conditions. The presence of *Macaronichnus* near the top of some of the sandstones is indicative of relatively higher energy conditions (Ekdale *et al.*, 1984), suggesting that some of the sandstone may have been deposited in a shoaling-upward environment.

The coarsening-upward sequence near the top of Unit II is a regressive sequence marking the transition between marine-shoreface sediments and nonmarine organic-rich rooted mudstone and Coal Zone 2. The absence of sediments characteristic of foreshore or beach environments suggests that the marine sediments were prograded over by a lower energy nearshore environment, such as an interdistributary bay.

Much of Unit III was deposited in a terrestrial environment as evident from the abundance of coal zones, but occasionally, marine conditions existed. Marine pelecypods are present in three-sand. Extensive bioturbation of three-sand and the presence of *Skolithos* and *Zoophycus* indicate that the sandstone was deposited in a shoreface environment (Ekdale *et al.*, 1984). Occasional ripples, crossbeds, shell hashes, and scours filled with coarse sand or granules suggest wave action, so it is probable that three-sand was deposited in an upper-shoreface (transitional with foreshore) environment.

The sandstones commonly present above Coal Zones 6 and 9 were deposited in a fluvial environment. This is suggested by their fining-upward character, their limited lateral extent, and their lateral facies association with carbonaceous mudstone and coal. They may represent thin channel deposits. Seven-sand was also deposited in a subaerial environment, evident from the abundance of rooted beds.

Unit IV marks a return to marine or transitional depositional environments. The presence of *Helminthopsis* is indicative of a lower-shoreface or offshore environment (Ekdale *et al.*, 1984).

DISCUSSION

The sediments of the Lower Skeena Group in the East Goathorn area of the Telkwa coalfield record a complex depositional history. Unit I was deposited in a fluvial environment on eroded volcanic basement. Peat growth took place in fluvial floodplains frequently interrupted by flooding and sediment influx. Deposition of Unit I was terminated by a marine transgression. Unit II was deposited in a marine shoreface environment. A subsequent regression occurred, resulting in deposition of the continental coal-bearing sediments of Unit III. Coal formed in peat marshes close to a shoreline, and later in fluvial floodplains. Unit IV was deposited in a shoreface environment during a second marine transgression.

The major facies change in the upper part of Unit III, from nonmarine coal-bearing strata in the west to transitional marine or marine strata of Unit IV in the east (Figure 4-5-2) indicates that the sea transgressed east to west across the area. Marine water inundated the eastern margin of the coalfield after the deposition of Coal Zone 3, and reached the western part of the coalfield after the deposition of Coal Zone 10. If this interpretation is valid, the source of the Lower Skeena Group in the study area must have been to the west. Elsewhere, it is generally thought that the Skeena Group has an eastern provenance (Tipper and Richards, 1976).

FUTURE WORK

In the future, the West Goathorn and North Telkwa areas will be examined in greater detail, to supplement the data from the East Goathorn area. Laboratory studies in progress, which include palynology, paleontology and petrology, will further refine the litho and biostratigraphy of the Lower Skeena Group in the study area. Coal-quality parameters, including sulphur content, ash content, and seam thickness and lateral distribution, will be interpreted in the context of the sedimentology of the coal measures.

ACKNOWLEDGMENTS

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TUMBLER RIDGE, NORTHEAST BRITISH COLUMBIA (93P/2, 3, 4; 93I/14, 15)

By W.E. Kilby and D.J. Hunter

KEYWORDS: Coal geology, Tumbler Ridge, stratigraphy, structure, automated data analysis.

INTRODUCTION

This study completes the geological mapping project done in the Bullmoose and Kinuseo areas. Work during this segment of the program concentrated on corrections to previous mapping, and additional mapping in a portion of the 93P/2 map sheet. In addition to completion of surface mapping, work began on integration of subsurface information and application of computer-based analysis procedures.

The study area contains the two producing coal mines which comprise the northeast British Columbia coal development. About one half the value of British Columbia's coal production comes from these two mines. This is approximately one eighth of the total value of mineral production in the province. The Sukunka, Bullmoose, Wolverine, Grizzly North and Grizzly South producing gasfields and the shut-in Murray field are also located in the study area. Interest in coalbed methane as an economical energy resource adds a new resource value to this area as well as all coal-bearing areas of the provinces. Exploration programs have outlined significant additional coal resources which only require improved markets to reach production. Significant new natural gas discoveries have encouraged major exploration activity for this commodity in the study area. The Tumbler Ridge Project of which this season's work was a part, will document the structure and stratigraphy of the area, evaluate coal maturation, estimate coal resources and coalbed methane potential and provide all raw data in computer processable form. This status report briefly describes the stratigraphy of the area, the data distribution and illustrates the penetrative fold-axis orientation of the area.

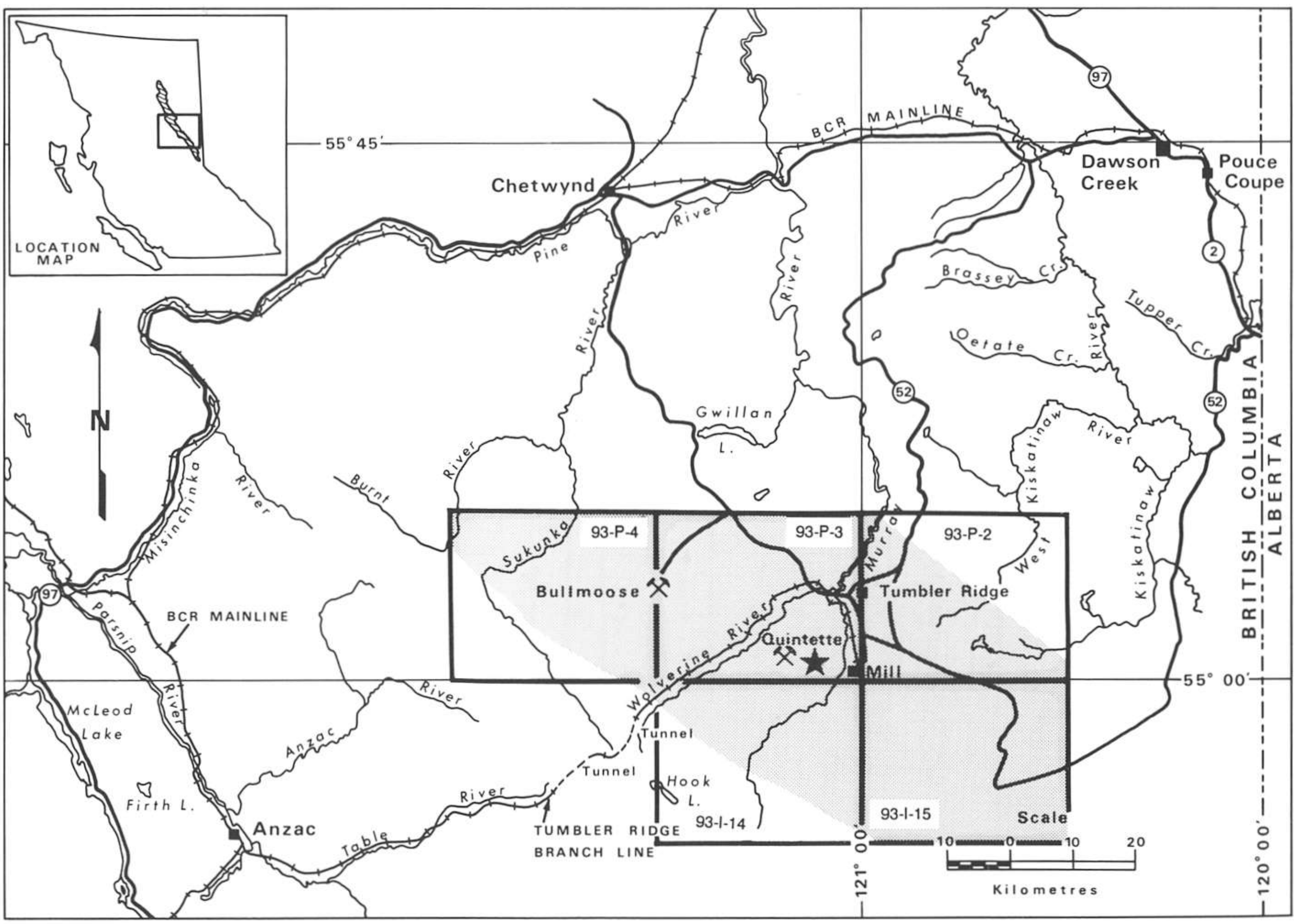

Figure 4-6-1. Location map of the study area. Stippled area illustrates the zone in which mapping was concentrated.

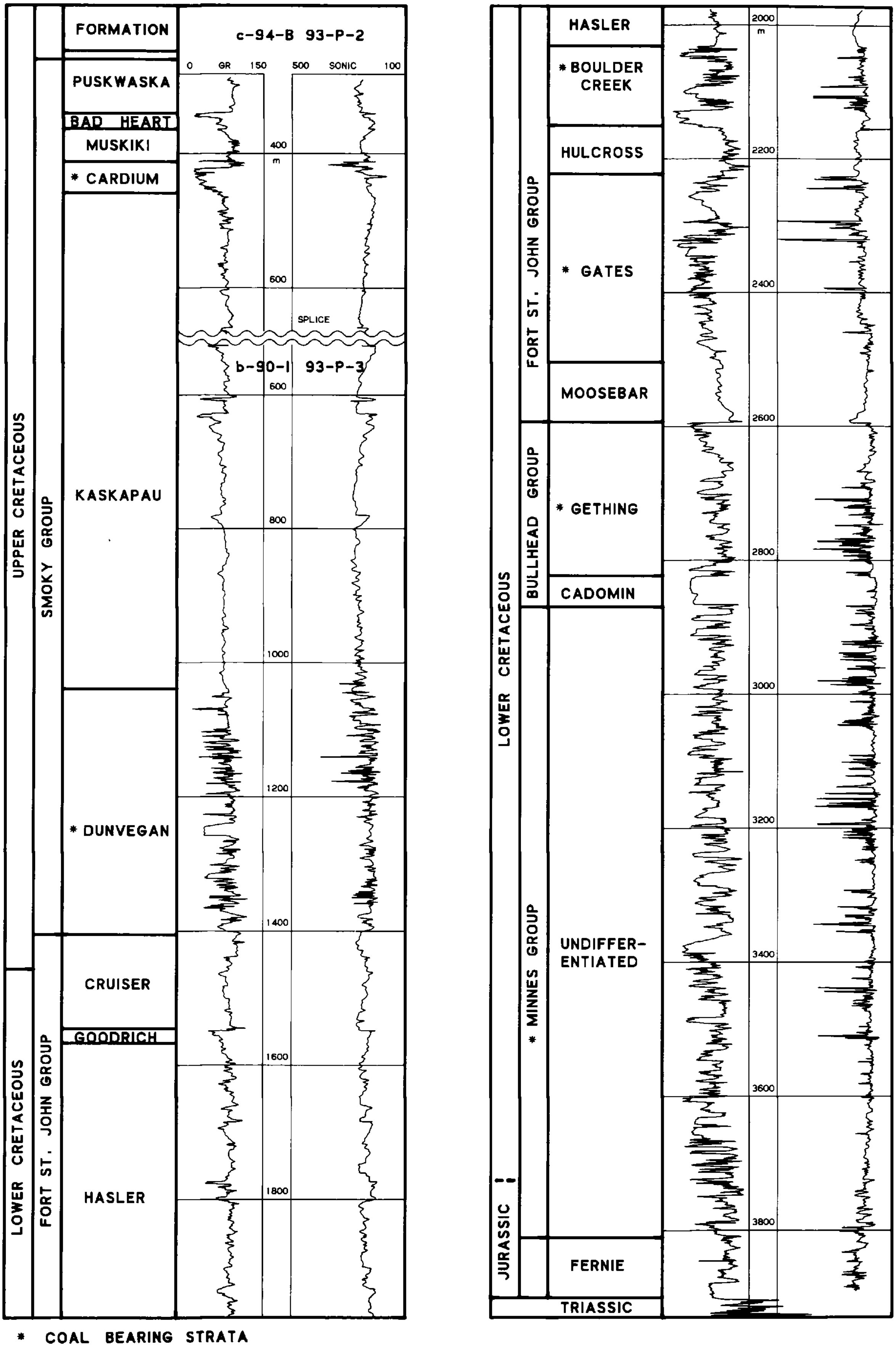

Figure 4-6-2. Table of formations with their geophysical log responses. Coal-bearing formations are noted with a star. No suitable log was available for the Wapiti Formation.

LOCATION

The map area is located in northeastern British Columbia and is centred at approximately latitude 55°N and longitude 121°E (Figure 4-6-1). Mapping covered parts of five 1:50 000 map sheets and was bounded on the west by the Rocky Mountain Front Ranges and on the east by a convenient outer foothills structure.

The town of Tumbler Ridge is situated near the centre of the map area and provided an excellent base for fieldwork. Three paved and all-weather roads connect Tumbler Ridge with other communities such as Chetwynd and Dawson Creek. Within the area there is generally good access to most of the coal-bearing strata by means of petroleum and coal exploration roads, forestry and railroad access roads. The drainage and physiography of the area have been described by Kilby and Wrightson (1987a), and Kilby and Johnston, (1988a).

PREVIOUS WORK

This project draws extensively on the efforts of past workers as it is essentially a compilation and re-interpretation with relatively minor infill mapping. Petroleum exploration in the area has provided excellent data sources including Jones (1960) and Hughes (1956). The federal government effort in the area has been led by Stott, (1967, 1968, 1973 and 1982). University studies include the work of Carmichael (1983) and Leckie and Walker, (1982). Previous work under this project has been reported by Kilby and Wrightson (1987a, b, c) and Kilby and Johnston (1988a, b, c).

1989 FIELD ACTIVITIES

Mapping during the 1989 field season concentrated on the re-examination of previously mapped areas. Some of the early maps were based extensively on compiled data and interpretation. Several problem areas were identified by industry users of the maps, as well as by re-interpretation using the expanded database. Those areas that warranted review were the priority for this year's work. In general the coal-bearing areas are very well mapped and have required virtually no revisions of the existing coal company interpretations. However, in areas which were mapped on a regional scale more than 15 years ago it was found that significant changes are required in the positioning of formation contacts. Much of this re-interpretation is now possible due to new information from recent exploration activities not available to previous workers.

FUTURE ACTIVITY

Production of 1:50 000-scale geology maps of the study area will be completed in early 1990. An automated structural analysis will be performed in order to construct a regional digital model of the map area. Both in-house software and commercially available programs will be used. A publication will be produced to describe the structure and stratigraphy of the area as well as demonstrate the numerical analysis techniques used during the study. All data compiled or generated during the project will be made available in digital form as Open File releases. For example, files of the outcrop and drill data will be available, as will CAD files of the geology maps and the constructed digital deposit models along with analysis programs.

STRATIGRAPHY

Formations mapped during this project ranged from the Jurassic Fernie Formation to the Upper Cretaceous Wapiti Formation. A representative section of this interval is about 4500 metres thick (Figure 4-6-2). Coal occurs in the Minnes Group and the Gething, Gates, Boulder Creek, Dunvegan, Cardium and Wapiti formations. The major coal-bearing formations are the Gething, Gates and Wapiti with the only current production coming from the Gates Formation. When evaluating the coalbed methane potential of a section, seams less than a metre thick may prove economic. Under this scenario all coal-bearing formations in this section are of interest.

The stratigraphy is essentially an alternating sequence of marine shales and marine or nonmarine clastics, resulting from marine transgressive and regressive cycles. Examination of the natural gamma trace in Figure 4-6-2 illustrates this feature. This geophysical method measures the natural radioactivity of rock and as such is excellent for distinguishing the shaliness of the strata. A marine shale will generally have a high gamma count while a clean sandstone will have a low count. Using this relationship, several coarsening-up sequences indicative of marine regressions can be identified. For example, in the upper Gething, Moosebar to Gates transition and the Goodrich, Cardium and Bad Heart formations. The sonic log, which measures interval transit times indicates coal horizons together with other features. Formation identification based solely on single outcrops is dubious; in areas of isolated sandstone and shale outcrop, interpretation relies heavily on subsurface data. Where exposure is good there is little difficulty in correctly identifying stratigraphic position.

DATA

Large amounts of geological information have been collected during this project, necessitating the use of computer techniques for data handling and compilation. The data are stored in three major files.

Outcrop information was collected from the best geology maps available. The Canada–B.C. Coal Information Collection Project facilitated the capture of the majority of the outcrop data used in this study. Orientation, stratigraphic position, geographic position, structure type and data source are recorded for each outcrop location providing about 22 000 data points (Figure 4-6-3).

Subsurface information from coal company boreholes was obtained from work carried out during the Canada–B.C. Coal Information Collection Project and log analysis was performed as part of the Tumbler Ridge Project. Information from petroleum and natural gas wells was obtained by log analysis performed during the Tumbler Ridge Project. For ease of storage and manipulation these two datasets are stored in a manner similar to outcrop information. Each subsurface intersection that is recorded is treated as a single outcrop. There are 1205 coal exploration boreholes (Figure 4-6-4) which generally only penetrate the coal-bearing horizons and tend to be less than 500 metres in depth. Sixty-one petroleum and natural gas wells (Figure 4-6-5) penetrate much deeper and provide stratigraphic and in some cases structural data across many formations.

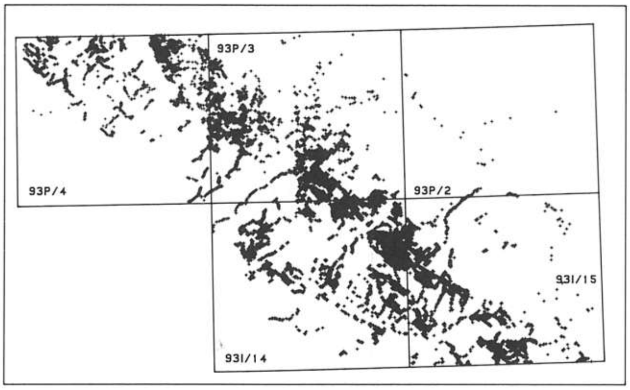

Figure 4-6-3. Distribution of outcrop data within the study area.

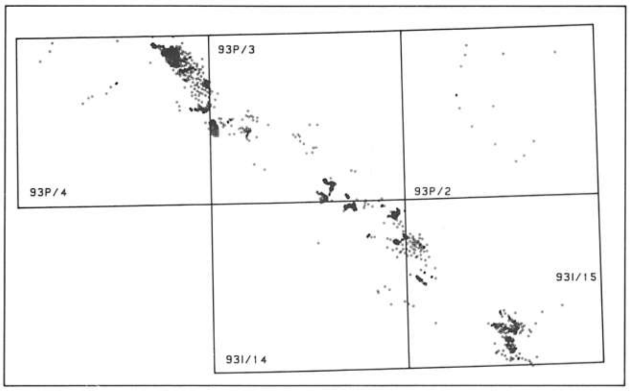

Figure 4-6-4. Distribution of the coal exploration boreholes within the study area.

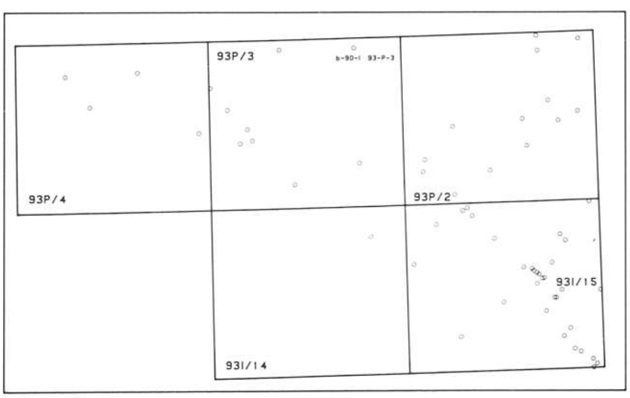

Figure 4-6-5. Distribution of the petroleum and natural gas exploration and production wells in the study area.

STRUCTURE

In the study area the structural style varies eastward across the regional trend of the Rocky Mountain Foothills, depending upon lithology and distance from the Front Ranges. Folding in the Minnes Group is extremely complex, usually consisting of short-wavelength inclined chevron folds which are not pervasive. Large rounded folds are present only when thick competent units of strata are involved. In the middle of the sequence under study, the Gates Formation is dominated by major long-wavelength folds, often forming large box anticlines. The uppermost section which is located along the eastern edge of the study area, shows strata simply deformed into gentle warps associated with the Alberta syncline. Smallscale faults are numerous, occurring in nearly every outcrop, but very few major regional faults are present, exceptions being the Bullmoose and Gwillan Lake faults. All faults identified in the area are thrusts. Figure 4-6-6 is a pi-diagram of all the bedding orientation data from the study area (approximately 21 000 points). The plot forms a girdle of poles to bedding indicative of cylindrically folded strata. Following Charlesworth *et al.* (1976), a numerical procedure was used to obtain a more precise calculation of the regional fold axis. The eigenvalue/eigenvector technique calculated a regional fold axis orientation of 309°/1°. The major and intermediate eigenvalues are approximately equal and the minimum eigenvalue is more than an order of magnitude less than the other two axes. This result is characteristic of a good cylindrical distribution of bedding poles. The cylindrical nature of deformation in the area suggests the use of the down-plunge projection technique will be successful in examining the structure.

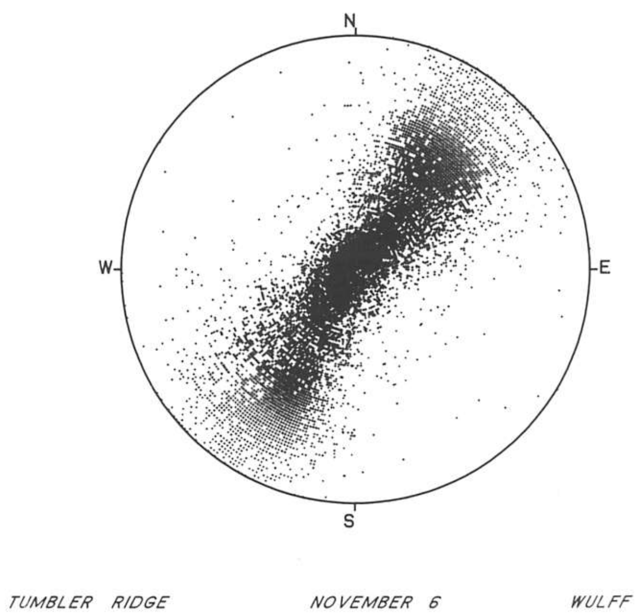

Figure 4-6-6. Pi-diagram of the bedding orientation data for the whole study area, about 22 000 outcrops.

ACKNOWLEDGMENTS

The authors would like to thank members of the coal and petroleum industry for their constructive comments and discussions about the study area. In particular D. Johnson (Quintette Coal Limited), M. Cooper (B.P. Resources Canada Limited), C. Bickford (consultant) and B. Wrightson (consultant). Quintette Coal Limited's Housing Division has supplied members of this project with excellent accommodation for the last two field seasons.

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NOTES

LITHOTYPE CHARACTERISTICS AND VARIATION IN SELECTED COAL SEAMS OF THE GATES FORMATION, NORTHEASTERN BRITISH COLUMBIA (93P/3)

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W. Kalkreuth and K.C. Pratt
Geological Survey of Canada

KEYWORDS: Coal geology, Bullmoose mine, Gates Formation, coal lithotype, maceral, image analysis.

INTRODUCTION AND OBJECTIVES

A combined field and laboratory investigation of selected coals within the Gates Formation in the Rocky Mountain foothills, northeastern British Columbia, was undertaken during the summer of 1988, and continued in 1989, with the purpose of gaining a better understanding of the sedimentological factors controlling variations in coal composition. This project involves examination of both the organic and inorganic strata. The initial results of the organic aspect of the study are reported here.

Previous investigations of Gates Formation coals have been regional in scope (Kalkreuth *et al.*, 1989; Kalkreuth and Leckie, 1989), and have involved definition of bulk compositional characteristics. This investigation focuses on in-seam variation as well as between-seam variation. The primary objectives of the research are threefold: to determine the petrographic composition of coal lithotypes within the Gates Formation; to document lateral and stratigraphic variation in lithotypes and maceral composition; and, to interpret the sedimentological factors controlling lithotype and maceral distribution (coal sedimentology). It is anticipated that the results of this study will be used to develop a methodology for predicting variations in coal quality, and to gain a better understanding of the characteristics of the Lower Cretaceous wetland environments.

REGIONAL AND LOCAL GEOLOGIC SETTING OF STUDY AREA

The Lower Cretaceous (Albian) Gates Formation contains the thickest, most economically viable coal seams in the study area. The coals outcrop in the Rocky Mountain foothills in the vicinity of Tumbler Ridge (Figure 4-7-1). The regional geology was initially described by Stott (1968, 1982). Detailed sedimentological studies of the Gates Formation (Figure 4-7-2) were done by Leckie (1983) and Carmichael (1983, 1988).

The Lower Cretaceous strata consist of a series of transgressive-regressive clastic wedges deposited in response to periodic uplift of the Canadian Cordillera (Smith *et al.*, 1984). The Moosebar (marine) and Gates formations and their subsurface equivalents, the Wilrich, Falher and Notikewin members of the Spirit River Formation (Figure 4-7-2) form one of these major sedimentary packages. Within both formations there are coarsening-upward sequences corresponding to progradational cycles of an active delta complex. Leckie (1986a) documented seven cycles within the Moosebar and Gates formations. The coastline is interpreted to have consisted of a series of high-energy, wave-dominated arcuate or cuspate deltas, with a major depocentre existing in the Bullmoose Mountain – Mount Spieker area (Leckie, 1986a). There is extensive intertonguing of marine and nonmarine strata in the study area. The northern limit of economic coal deposits in the Gates Formation is in the vicinity of the Bullmoose mine leases; north of the Bullmoose properties, the formation consists primarily of marine-shelf sediments (Stott, 1982; Leckie and Walker, 1982).

The depositional environment of Gates Formation coals is discussed in Kalkreuth and Leckie (1989) and Kalkreuth *et al.* (1989). Kalkreuth and Leckie's research was regional in

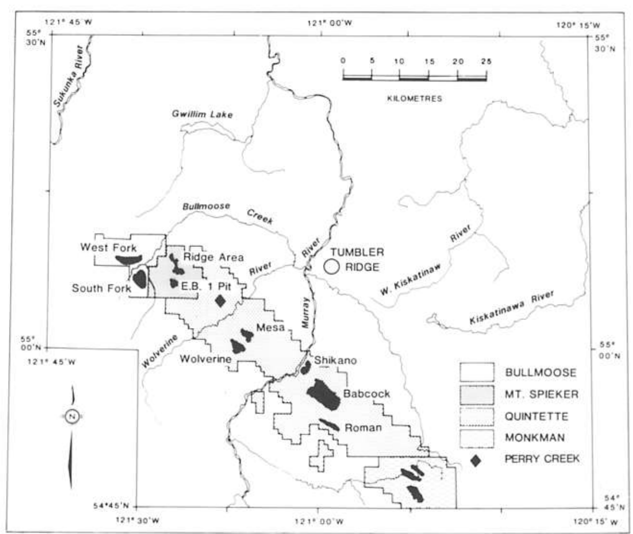

Figure 4-7-1. Location map of study area. Shaded areas indicate major coal leases. Diamond indicates location of outcrop sample of J seam on Perry Creek Road. Named areas are current production sites or areas of proposed development. Modified from Matheson (1986).

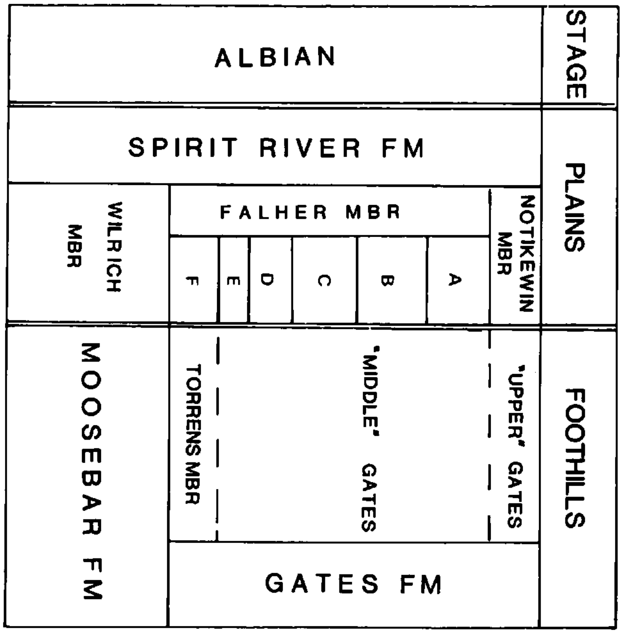

Figure 4-7-2. Stratigraphic chart of a portion of the Lower Cretaceous formations in northeastern British Columbia. Modified from Leckie (1986) and Carmichael (1988).

scope; whole-seam channel samples were collected from active mines throughout the Foothills trend, from Bullmoose southeastward to Mountain Park, Alberta. They interpreted the coals as having formed in protected environments shoreward of a high-energy, wave-dominated coastline. Coals formed in this environment are petrographically distinct, being composed of relatively low vitrinite (average 57 per cent), relatively high inertinite (average 42 per cent) and negligible liptinite concentrations. Semifusinite and inertodetrinite are the primary inertinite macerals. Petrographic (tissue preservation and gelification) indices indicate a predominantly forest-moor type wetland environment that allowed for extensive surface degradation of organic material.

Fieldwork in 1989 was centred in the Bullmoose mine area (Figure 4-7-1). Six seams of economic thickness are present at Bullmoose (Drozd, 1985), designated, from oldest to youngest, A1, A2, B, C, D and E. Rapid facies changes occur in the Bullmoose area. The Falher D, a thick, coarsening-upward marine unit occurring stratigraphically between the B and C seams at Bullmoose Mountain, is replaced by lagoonal (?) carbonaceous shale and siltstone 4 kilometres to the south, in the South Fork pit area (Leckie, 1986b). The paleoshoreline existed in the area just north of Bullmoose mine and was oriented roughly west-northwest. Predominantly nonmarine conditions appear to have prevailed in the Bullmoose Mine area.

METHODS

Research discussed in this report has both field and laboratory segments. Fieldwork involved section description, correlation and sample collection. Laboratory research focused on delineating the compositional characteristics of the individual coal lithotypes.

FIELD: SECTION CORRELATION AND SAMPLE COLLECTION

Fieldwork in 1989 was concentrated in the South Fork pit area of Bullmoose mine (Figure 4-7-1). Five Gates Formation coal seams, A1, B, C, D and E are exposed. Lithotypes of the seams were described according to a modified Australian classification scheme as bright, banded bright, banded coal, banded dull, dull, fibrous and sheared (Diessel, 1965; Marchioni, 1980; Lamberson and Bustin, 1989; Lamberson *et al.*, 1989; Lamberson *et al.*, 1989). A minimum thickness of 1 centimetre was used to delineate a lithotype. At least three sections of each seam, at different locations within the mine, were described. Wherever possible, representative samples of lithotypes were collected from each site.

Section profiles of each seam were drawn and an attempt was made to correlate zones using a minimum thickness of 1 centimetre. Lithotypes were subsequently regrouped and sections redrawn using a minimum thickness of 5 centimetres. Exceptions were made for the occurrence of fibrous coal and mudstone; the unique environmental significance of each of these lithologies is lost if combined with another lithotype. Two examples of section correlations are shown in Figures 4-7-3 and 4-7-4.

LABORATORY: MACERAL POINT-COUNT ANALYSIS

Lithotype samples collected from the Bullmoose and Quintette areas during the 1987, 1988 and 1989 field seasons were processed by two types of petrographic analysis: standard point-count and image analysis. Representative lithotype samples were crushed to -20 mesh, split and mixed with a polyester resin to make standard 2.54-centimetre petrographic pellets. The pellets were polished to a 0.05 micrometre aluminum oxide. Three hundred points were counted (mineral matter free) using the established maceral classification scheme (Bustin *et al.*, 1985). Mineral matter was counted separately. In total, 83 crushed particle pellets were examined. Percentage composition on a mineral matter free basis was calculated and averaged by lithotype (Table 4-7-1).

LABORATORY: IMAGE ANALYSIS

Image analysis was performed using a Zeiss IBAS 2 system following a procedure modified from Pratt (1989a, b). The primary objective of this analysis was to determine bulk changes in maceral composition with respect to stratigraphic position. Oriented bench samples (stratigraphic direction preserved) representing the entire B seam were prepared for image analysis. The oriented blocks were sawn in half. Half of the block was embedded in a polyester resin and subsequently polished to a 0.05 micrometre aluminum oxide. In total 100 blocks were prepared. A line of traverse was determined which provided a profile of the block perpendicular to bedding, following the path illustrated in Figure 4-7-5. Four fields were evaluated along a horizontal plane. The stage moved one field vertically and then proceeded in

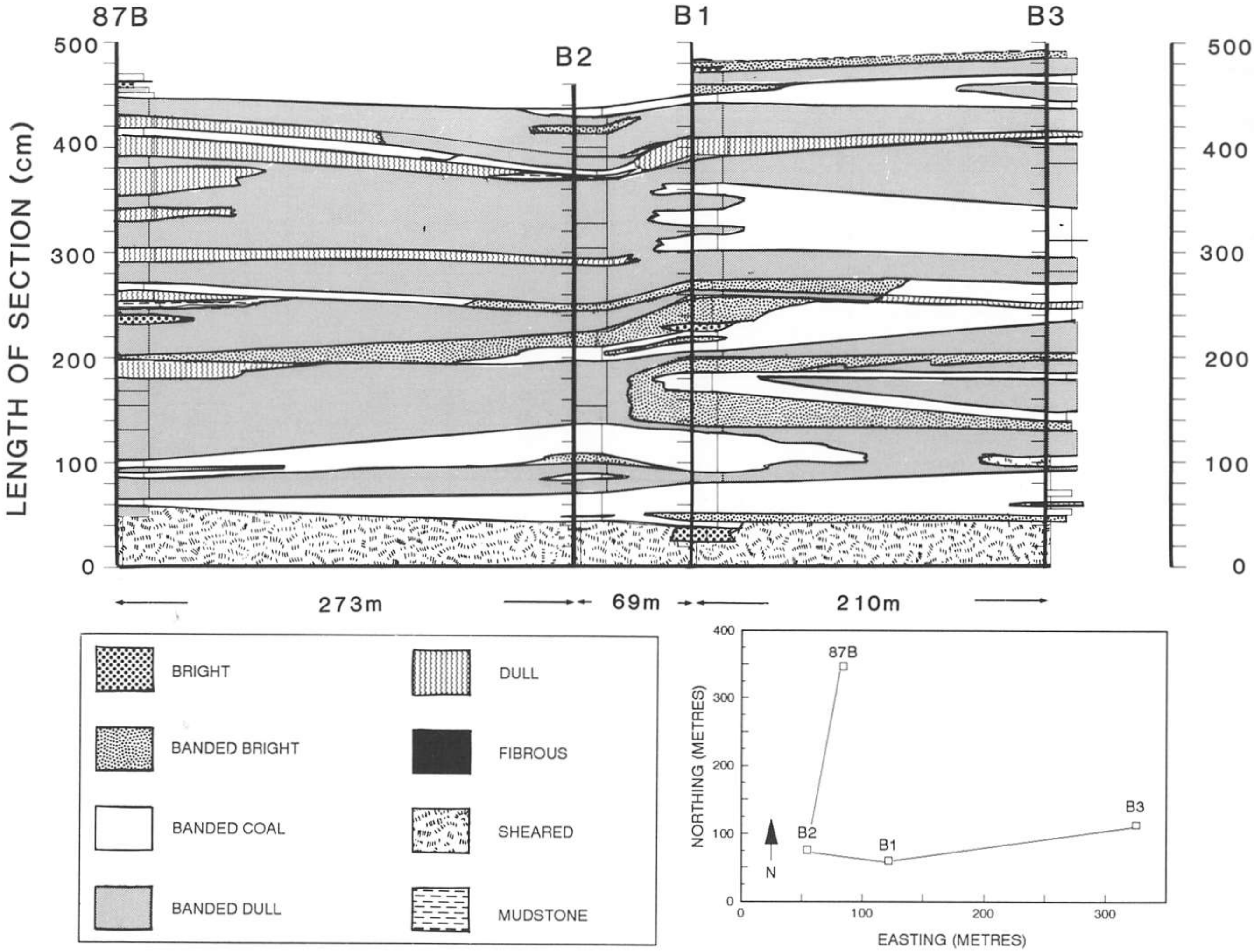

Figure 4-7-3. Section correlation for Bullmoose B seam. Datum for correlation is the floor of the seam. Inset map shows relative locations of section sites.

the opposite direction for four fields. This process was repeated across the vertical distance of the block. The distance between fields is 0.5 millimetre and each field represents an area of 0.19 square millimetres.

Prior to doing the analyses, a number of individual fields were analyzed in order to determine the grey level thresholds between maceral groups. An example histogram is illustrated in Figure 4-7-6. Three threshold values (between liptinite, vitrinite, low-reflecting inertinite and high-reflecting inertinite) were identified at grey-levels corresponding to reflectances (per cent in oil) of 0.88, 1.24 and 2.00. During routine analysis, the frequency of occurrence of each maceral group for each field was determined from the reflectance histogram (Figure 4-7-6). The relative percentage of each group was then calculated. The percentage values were then averaged with the three other fields on the horizontal part of the traverse to yield an approximation of the average maceral composition of each 0.5-millimetre stratigraphic thickness of the block. The end result of the traverse is a maceral compositional profile of the block, displayed in a stacked bar fashion, as illustrated in Figure 4-7-7. The plotting program eliminated areas of the block with cracks or holes.

RESULTS AND PRELIMINARY CONCLUSIONS

FIELD OBSERVATIONS

In previous studies (Lamberson and Bustin, 1989; Lamberson *et al.*, 1989), some general characteristics of the lithotypes were described which may be expanded upon with the additional data gathered during the 1989 field season. The banded lithotypes predominate in all of the seams studied. Banded dull and banded coal are the most common in A1, B and C seams. Banded bright and banded coal are more common in D and E seams. Mudstone partings are rare, thin and lenticular when present in A1, B and E seams, whereas they tend to be thick and laterally extensive in C and D seams. Two types of banded dull and dull coal exist: a mineral-rich and a mineral-poor variety.

Fibrous coal occurs in accumulations thick enough to constitute a lithotype, however, it is quite rare. The thickest lens of fibrous coal found is 2.5 centimetres thick. Bright coal and fibrous coal are lenticular, which may reflect their origin from individual logs and stems. Banded bright coal

TABLE 4-7-1
MACERAL COMPOSITION

		BRIGHT (15)	BANDED BRIGHT (13)	BANDED COAL (15)	BANDED DULL (16)	DULL (7)	SHEARED (8)	ROCK (9)
TELOCOLLINITE	AVERAGE	6	11	9	8	5	10	16
	STANDARD DEV	6	8	8	8	6	5	15
TELINITE	AVERAGE	19	16	13	11	7	11	6
	STANDARD DEV	16	10	4	5	2	9	5
PSEUDOVITRINITE	AVERAGE	45	24	15	5	1	21	14
	STANDARD DEV	23	12	9	5	1	11	20
DESMOCOLLINITE	AVERAGE	17	21	28	22	16	28	4
	STANDARD DEV	17	10	11	12	7	15	4
VITRODETRINITE	AVERAGE	1	7	8	14	12	21	28
	STANDARD DEV	1	13	11	18	22	14	21
TOTAL VITRINITE	AVERAGE	89	80	72	61	42	91	68
	STANDARD DEV	9	10	14	19	25	4	35
SEMIFUSINITE	AVERAGE	4	9	12	17	28	2	5
	STANDARD DEV	5	7	7	10	13	2	5
FUSINITE	AVERAGE	3	4	7	6	10	3	5
	STANDARD DEV	3	3	7	4	13	2	8
INERTODETRINITE	AVERAGE	3	6	8	15	18	3	22
	STANDARD DEV	3	4	5	9	12	1	26
TOTAL INERTINITE	AVERAGE	10	19	27	39	57	8	32
	STANDARD DEV	9	10	14	19	24	4	35
SPORINITE	AVERAGE	0	0	0	1	1	0	0
	STANDARD DEV	1	0	0	1	1	0	0
TOTAL LIPTINITE	AVERAGE	1	0	1	1	1	1	0
	STANDARD DEV	1	0	1	0	1	1	0
TOTAL AVERAGES		100	100	100	100	100	100	100

NUMBER OF LITHOTYPE SAMPLES INDICATED IN ( )

often contains abundant thin lenses and laminae of fibrous coal.

Maceral Point Count Analyses

The averages and standard deviations of the maceral point count analyses by lithotype are reported in Table 4-7-1. Figure 4-7-8 graphically illustrates the average compositions. There is a marked decrease in vitrinite, with a concomitant increase in inertinite, from bright to progressively duller lithotypes. This variation appears in large part to reflect a decrease in the pseudovitrinite population. Among the vitrinite maceral group, there is an increase in the percentage of the more degraded vitrinite varieties, desmocollinite and vitrodetrinite, from the brighter to duller coals. In all lithotypes except sheared coal, semifusinite is the most abundant inertinite maceral. Micrinite and macrinite are nowhere abundant. In agreement with Kalkreuth *et al.* (1989) and Kalkreuth and Leckie (1989), there is very little liptinite in any of the lithotypes (1 per cent or less). The paucity of liptinite in these coals may be due to difficulties associated with its recognition at elevated coal ranks (medium volatile), or its near absence in the original wetland. Low liptinite concentrations have been noted in Gates Formation coals in the Cadomin-Luscar, Mountain Park area of southwestern Alberta (Kalkreuth and Leckie, 1989).

The composition of carbonaceous mudstones in the coal appears to be variable, but most closely resembles banded dull coal. Inertodetrimite is the most abundant inertinite maceral in these rocks. Sheared coals are consistent in composition, and most commonly resemble the bright coal lithotype.

The origin of psuedovitrinite has most commonly been attributed to oxidation (Kaegi, 1985). Its occurrence in the brighter, less degraded lithotypes raises questions as to the origin of the maceral in these coals.

Image Analysis

Preliminary image analysis results indicate a good correlation between the bulk maceral composition as determined by image analysis and by standard point-count analysis for each lithotype. Table 4-7-2 compares results obtained for maceral composition by image analysis and the bulk maceral composition averages obtained by point-counting. The most significant problem with the technique is in determining the percentage of liptinite. In image analysis, edge effects around cracks, holes, scratches, mineral matter etc., produce a gradation in values that are lower than the adjacent maceral. For this reason, liptinite values obtained by image analysis are higher than those obtained by point-counting and often reflect mineral matter concentration (Pratt, 1989a, 1989b).

Section Correlations

Section correlations for B seam and D seams are illustrated in Figures 4-7-3 and 4-7-4. The two seams differ in lithotype composition and stratigraphy, reflecting what are interpreted to be differences in depositional conditions. Throughout the time the peats were being deposited, the paleoshoreline is

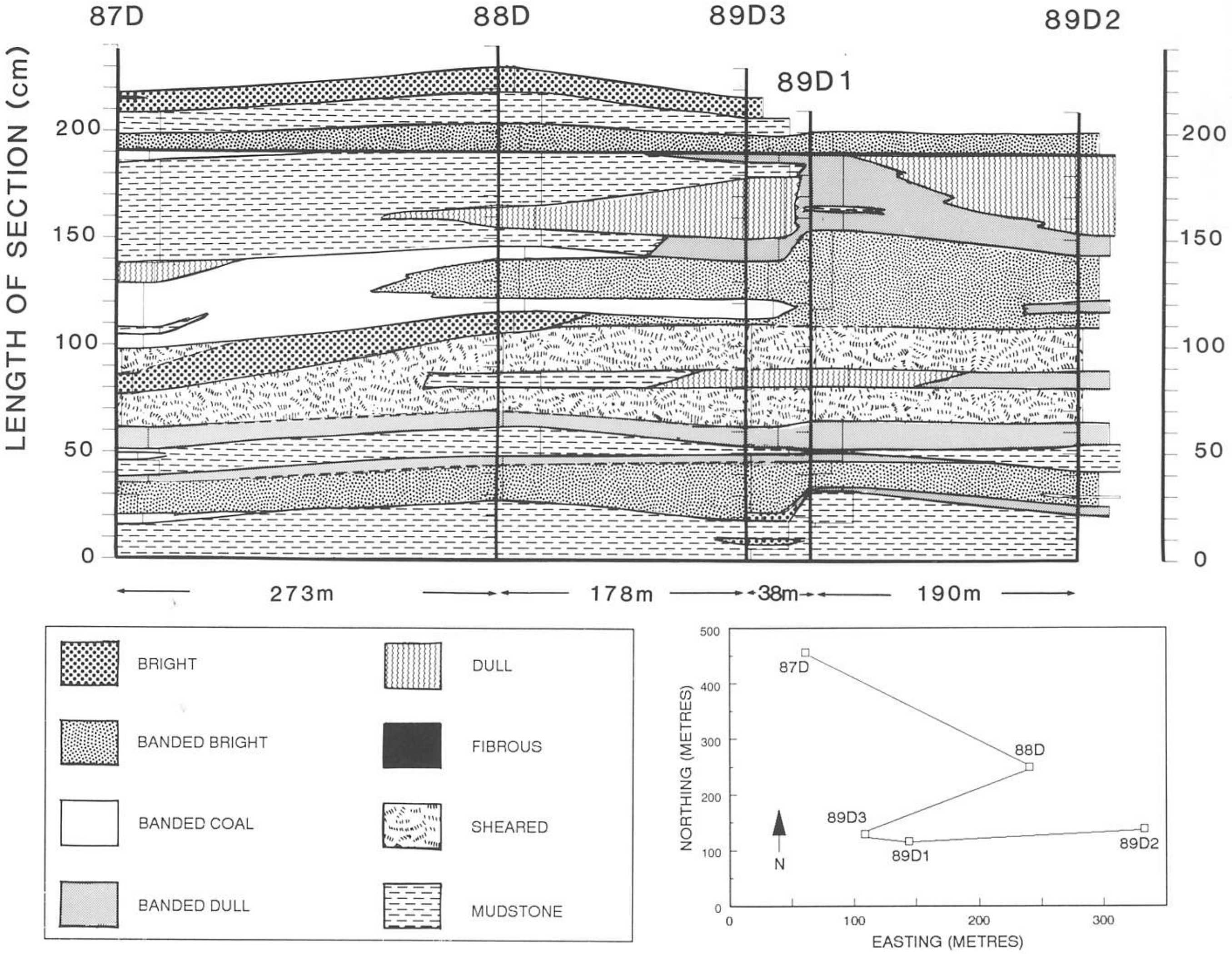

Figure 4-7-4. Section correlation of Bullmoose D seam. Location for seam base problematic in some locations. Datum for correlation is the base of a banded bright layer occurring across the section line, near the top of the seam. Inset map shows relative section locations.

interpreted to have been located immediately north of Bullmoose Mountain; a major fluvial channel was located in the Mount Spieker area (Leckie and Walker, 1982; Leckie, 1986a, b). The proximity of these two features may have influenced coal stratigraphy.

Within the B seam (Figure 4-7-3), banded dull coal is the only lithotype which can be consistently correlated through the section; the other lithotypes are more restricted in their areal distribution. In all of the sections examined the basal part of B seam is sheared, although the thickness of the sheared interval varies. In sections 87B, 89B2 and 89B1 there is a general increase in the percentage of the duller lithotypes from the base to the top of the seam. The trend is not as well developed in 89B3. There is also a decrease in the duller coal lithotypes from north to south. No significant mudstone interbeds are present.

The B seam appears to have developed in an area protected from clastic input. However, degradation levels appear to have been quite high (as evidenced by the predominance of the duller lithotypes). Depositional conditions were variable from south to north and east to west, as well as from the onset to cessation of the peat-forming episode.

D seam (Figure 4-7-4) is dominated by banded bright coal, sheared coal and mudstone interbeds. Unlike the B seam, the brighter lithotypes and mudstone layers are correlative across the section. The sheared zone averages about 50 centimetres thick, and is found approximately 0.5 metre above the base of the seam.

Within the unsheared zones of D seam there is a cyclic repetition of lithotypes. Banded bright coal is stratigraphically succeeded by mudstone and dull coal, followed by a return to banded bright or bright coal. This cyclic repetition is interpreted to represent fluctuations in wetland type due to repeated influx of clastic material from adjacent fluvial channels.

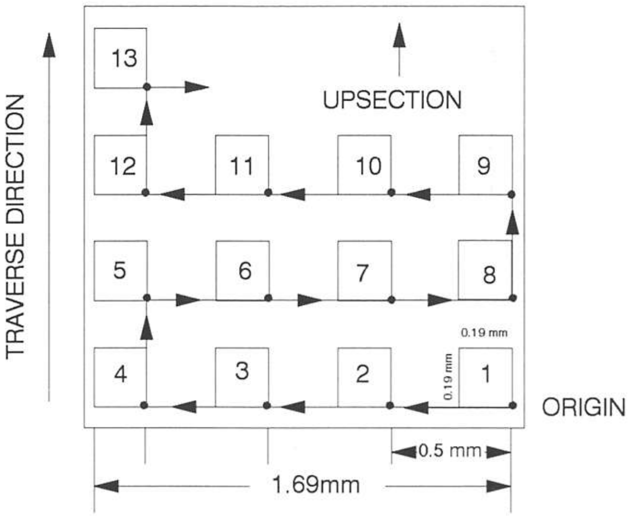

Figure 4-7-5. Traverse path of microscope on coal block during image analysis. Not to scale.

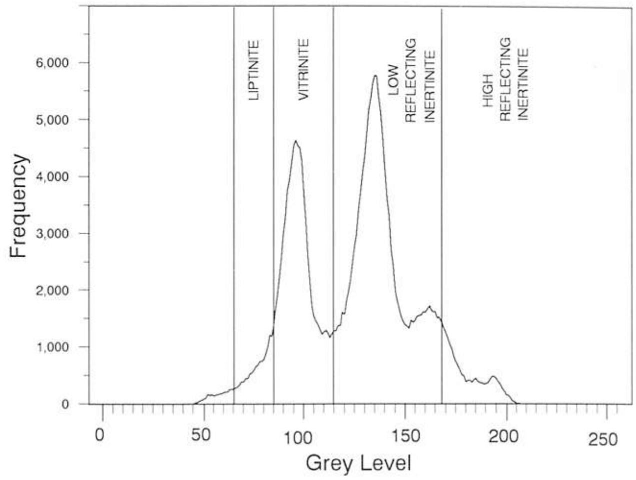

Figure 4-7-6. Grey-level histogram of a field enriched in low-reflecting inertinite showing location of compositional thresholds.

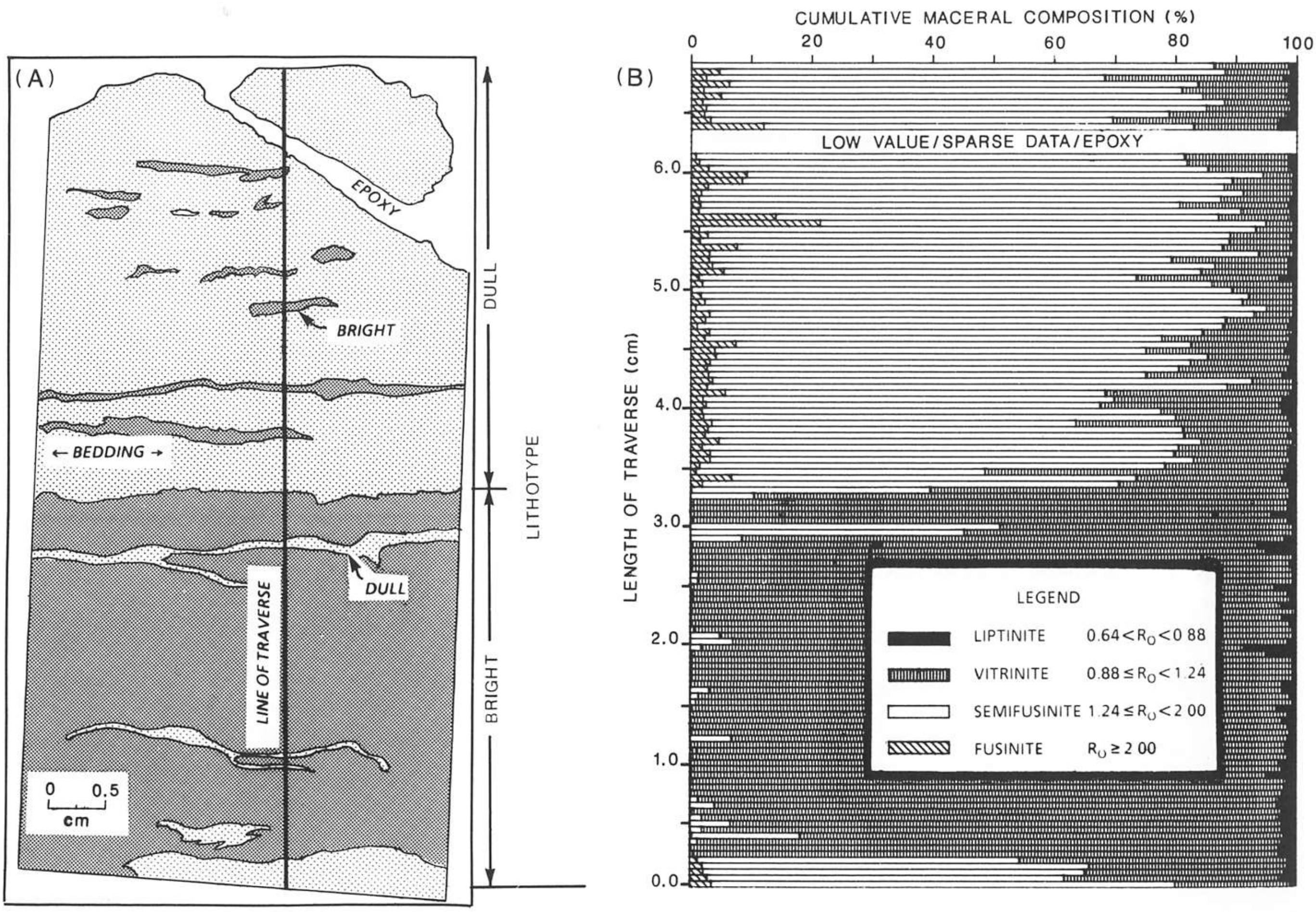

Figure 4-7-7. (A) Schematic drawing of coal block B66a showing location of bright and dull layers. (B) Image analysis plot of block B66a showing compositional changes along line of traverse.

TABLE 4-7-2
COMPARISON OF MACERAL COMPOSITION

	DATA SOURCE	BLOCK NUMBER	VITRINITE	LOW REFLECTING INERTINITE	HIGH REFLECTING INERTINITE	LIPTINITE
BRIGHT	PC	N/A	89	4	6	1
BANDED BRIGHT	PC	N/A	80	9	10	0
BANDED COAL	PC	N/A	72	12	15	1
BANDED DULL	PC	N/A	61	17	21	1
DULL	PC	N/A	42	28	28	1
SHEARED	PC	N/A	91	2	6	1
BANDED COAL	IA	B21	81	27	2	2
BANDED DULL	IA	B21	69	27	2	2
BRIGHT	IA	B43a	82	8	1	10
DULL	IA	B43a	61	26	10	3
FUSAIN	IA	B43a	23	29	42	6
SHEARED	IA	B14	97	1	1	3
BANDED DULL	IA	B14	67	29	4	1
FUSAIN	IA	B14	33	36	27	5
BANDED COAL	IA	B14	56	39	3	2
BRIGHT	IA	B66a	92	7	1	1
DULL	IA	B66a	15	79	4	2

PC=POINT COUNT; IA=IMAGE ANALYSIS

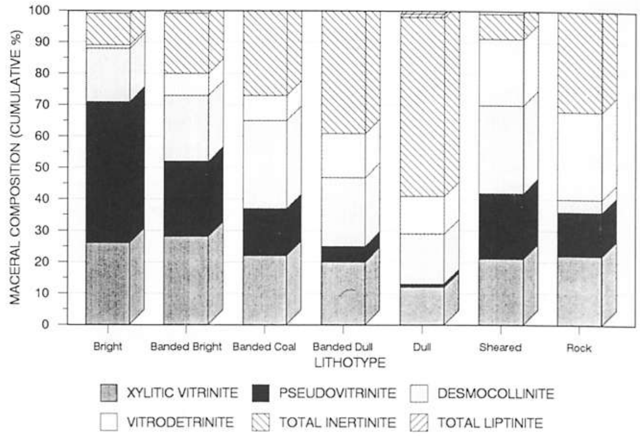

Figure 4-7-8. Average maceral composition by lithotype (volume per cent), mineral matter free basis, as determined by traditional point counting methods. Xylitic vitrinite = (telinite + telocollinite).

SUMMARY

Mapping lithotype variations within the coal seams at the Bullmoose mine provides a framework for interpreting the characteristics of the original wetland environments. Coal seams in the mine area differ from one another in terms of their seam stratigraphy. In addition, variations are present within each seam, implying that depositional conditions were not uniform throughout the original wetland environment. Distance from the paleoshoreline or active fluvial channels is believed to have been a major factor controlling seam compositional variations.

Lithotypes appear to be compositionally distinct, with a decrease in vitrinite content from the brighter to duller lithotypes. The results of this study suggest that a variety of wetland environments were present; each lithotype may reflect one, or several different environments. Two wetland types, one open to clastic input and the other protected from clastic input, are illustrated by the D and B seams, respectively.

FUTURE RESEARCH

Future research will concentrate on analyzing more lithotype samples in order to more precisely characterize their maceral composition. In addition, an investigation into the occurrence and characteristics of pseudovitrinite, which is inordinately abundant in the bright and banded bright lithotypes, will be undertaken. Because the coals are composed of banded coal and banded dull lithotypes, the pseudovitrinite is not abundant when considered on a whole-seam basis. The seam stratigraphy, as determined by the lithotype correlations, will be further interpreted in terms of the enclosing clastic strata.

ACKNOWLEDGMENTS

The authors thank Mr. David Malcom and the staff at Bullmoose mine for their cooperation and logistical support. Ms. Doni Jacklin's assistance in the field and in sample preparation is gratefully acknowledged. Support for this project was provided by the British Columbia Ministry of Energy, Mines and Petroleum Resources and the Geological Survey of Canada.

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PRELIMINARY SURVEY OF THE COAL RESOURCES OF UPPER CRETACEOUS ROCKS, NORTHEASTERN BRITISH COLUMBIA (93, 94)

By Barry Ryan

KEYWORDS: Coal geology, Dunvegan Formation, Wapiti Formation.

INTRODUCTION

The area covered by this study area forms a triangle in the northeast corner of British Columbia. The east side of the triangle extends 600 kilometres south from the British Columbia, Alberta, Northwest Territories common boundary point and the north side extends 250 kilometres west from the same point.

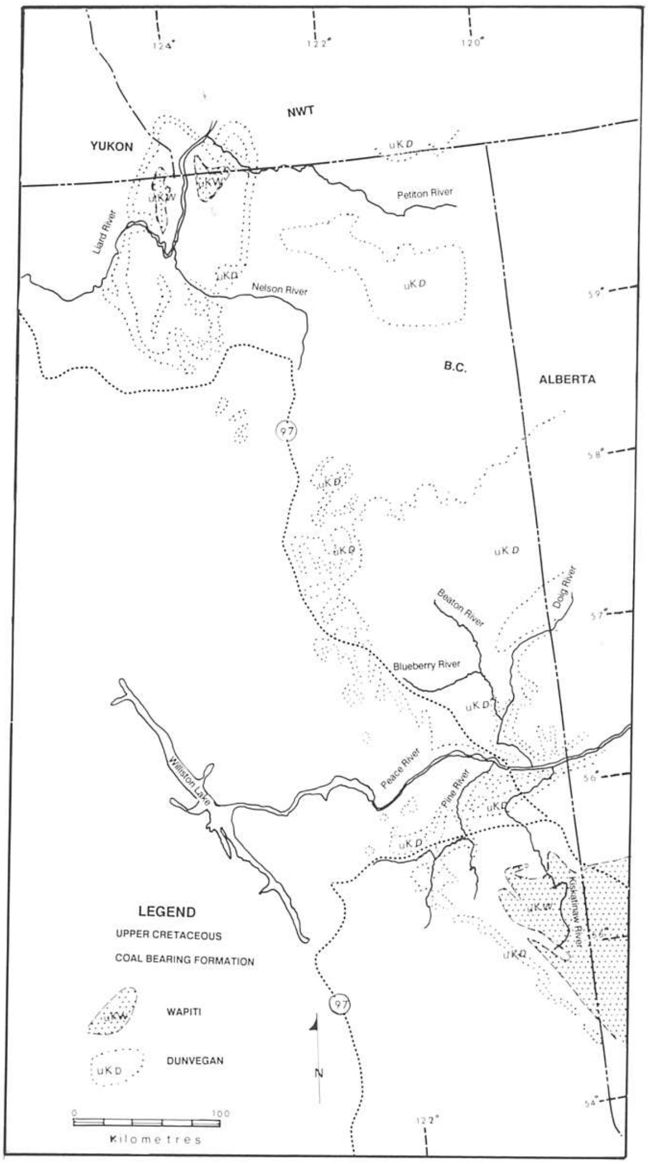

Figure 4-8-1. Location of Dunvegan and Wapiti formation outcrops in Northwestern British Columbia.

There are two major Upper Cretaceous coal-bearing formations in this area separated by marine shales: the Dunvegan Formation of Cenomanian age and the Wapiti Formation of Late Cretaceous to Early Paleocene age. Figure 4-8-1 illustrates the general extent of outcrops of Dunvegan and Wapiti formations in Northeastern British Columbia. Coal also occurs in at least one small Tertiary basin.

The lower part of the Dunvegan Formation includes a 30-metre section containing up to four coal seams, all generally less than 50 centimetres thick. Lateral extent of the coal seams within the Dunvegan is not known, but they are absent in some outcrop locations. Drilling in the Wapiti Formation south of Dawson Creek indicates that a single seam, generally less than 2 metres thick, is nearly always present at the base of the formation. A Tertiary basin containing about 5 metres of lignite extending over an area of approximately 5 by 15 kilometres is located on the Alaska Highway at Mile 533.

In general the open-pit coal resource potential of the Upper Cretaceous appears at this time to be marginal even for local markets. It is possible that the Tertiary coal basins may provide more resource potential.

1989 FIELD PROGRAM

The 1989 field program consisted of two weeks fieldwork in September. The late start did not permit a longer or more detailed field study. In preparation for the fieldwork all geological literature on the area was checked for references to coal occurrences, and a number of people were contacted who might have detailed knowledge of coal in the area. Generally the area is rolling farmland or wooded and swampy. Rivers have incised canyons into the landscape to produce good outcrops. It was found that the low-rank coals survive as boulders in the rivers and that boulder prospecting would probably be effective for a limited distance from source.

Roads in the vicinity of Chetwynd, Dawson Creek and Fort St. John were driven extensively. North of Fort St. John the Alaska Highway and the Liard Highway (No. 77) were checked to Coal River and the Northwest Territories border, respectively.

An attempt was made to examine as many known coal occurrences as possible, as well as to prospect in obvious places adjacent to roads. Fifteen coal samples were collected; in most cases they are extensively weathered and will probably only be useful for total ash and vitrinite reflectance rank determinations. All coal occurrences identified in the literature or found during the season are tabulated.

TABLE 4-8-1
LOCATION OF UPPER CRETACEOUS COAL OUTCROPS

Occurrence	Formation	Coal Description	Location Description	Reference
Lone Prairie	Dunvegan	One to four coal seams each less than 30 cm thick; very weathered; flat dipping.	15 km southeast of Chetwynd; 2.2 km east of the Lone Prairie/Tumbler Ridge road intersection on the Lone Prairie road.	Present study
Moberly Lake	Dunvegan	Single seam less than 30 cm thick; very weathered; flat dipping.	9 km north of Chetwynd on Highway 29; east side of road.	Present study
Island Lake road	Wapiti	Single seam; 15 cm thick; near base of Wapiti.	Off Highway 2, 30 km south of Pouce Coupé on Island Lake road 1.27 km north of Kiskatinaw River.	Present study
Kiskatinaw bridge, Alaska Highway	Dunvegan	Three coal seams from top to bottom, 50, 70 and 20 cm thick in 30 cm of sand and mudstone stratigraphy.	Between Dawson Creek and Fort St. John under bridge over Kiskatinaw River; north side.	Present study
Peace River	Dunvegan	Three thin coal seams from top to bottom, 10, 35 and 10 cm thick.	5.5 km south of bridge over Peace River and 100 m east of Alaska Highway.	Present study
Doig River	Dunvegan	Boulders up to 20 cm in length in river bed.	35 km northeast of Fort St. John and 20 km above confluence with Beaton River a secondary road crosses Doig River.	Present study
Lower Cache Creek road	Dunvegan	One 90-cm seam including 55 cm of clean coal.	35 km north of Fort St. John on Highway 97; 7.2 km on Lower Cache Creek road to southeast.	Present study
Blueberry River	Dunvegan	A few small boulders of coal in river bed.	55 km north of Fort St. John and 20 km east of Buik.	Present study
Coal River	Tertiary	Lignite boulders in river bed; outcrop approximately 10 km upriver. Basin about 5 km by 15 km.	Mile 533. (Kilometre 858) Alaska Highway.	Present study McConnell (1891), McLearn and Kindle (1950), Williams (1944)
Wapiti	Wapiti	Single seam at base of formation; up to 2 m thick.	South of Dawson Creek.	Gulf (1981)
Coldstream Creek	Dunvegan	Coal occurrences, no thickness specified.	Southeast of East Pine on Highway 97 between Dawson Creek and Chetwynd.	Selwyn (1877)
South of Pine River valley	Dunvegan	Thin coal beds.	South of Smokey River.	Spieker (1921)
Pine River canyon	Dunvegan	Four coal seams top to bottom, 15, 20, 61 and 70 cm thick.	East of Chetwynd near Wartenbe Mtn.	Selwyn (1877)
Pine River	Dunvegan	61 cm coal seam.	Near East Pine on Highway 97.	Williams (1934)
Kiskatinaw River	Dunvegan	51 cm seam.	5 km above mouth of Kiskatinaw River.	Williams (1934)
Doig River	Dunvegan	76 cm seam and 30 cm seam.	16 km above mouth of Doig River.	Williams (1934)
Alaska Highway Mile 66	Dunvegan	Thin coal seams in quarry.	Mile 66 (Kilometre 106), Alaska Highway, quarry.	Hage (1944)
Table Mtn.	Dunvegan	30 cm coal seam near base of Dunvegan.	Mile 354 (Kilometre 370), Alaska Highway.	Williams (1944)
Liard River Pretty Hill	Wapiti	50 cm seam.	15 km upriver from Fort Liard.	Hage (1945)
Liard River	Tertiary	Two small Tertiary coal basins.	20 km south of Watson Lake on Alaska Highway.	Dowling (1915)
Petitot River	Dunvegan	Coal in mudstone, no thickness given.	Near Highway 77 and B.C. border.	Stott (1982)
Kotaneelee River	Wapiti	38 cm coal seam, poor quality.	20 km up the Liard River from Fort Liard, 3 km above mouth of Kotaneelee River.	Hage (1945)

STRATIGRAPHY

Dunvegan Formation

The Dunvegan Formation is of Cenomanian age; it contains mainly nonmarine sands and conglomerates, but marine shales are also present. The thickness ranges from 150 to 200 metres. Generally the Dunvegan outcrops as light buff-coloured, massive, coarsely crossbedded sandstones with some mudstone or silty zones; channels are often visible. The base of the Dunvegan is usually underlain by dark marine shales of the Shaftesbury or Sully formations; the top is overlain by the marine shales of the Kaskapau Formation.

Thin coal seams occur in the lower part of the formation. No reports of seams thicker than 1 metre have been found.

Wapiti Formation

In British Columbia south of Dawson Creek, the Wapiti Formation consists of nonmarine clastic sediments, mainly sandstones and siltstones. It has an estimated thickness of 400 metres and ranges in age from Late Cretaceous to Early Paleocene. The base is underlain by the Chungo sandstone of the Puskwaskau Formation. A thin coal seam marks the base of the Wapiti. The coal-bearing sections found in Alberta in the upper part of the Wapiti appear to have been eroded in British Columbia.

MINING HISTORY

There are no reports of commercial mining of Upper Cretaceous coals in Northeastern British Columbia. Minor amounts of coal have been taken from seams in the Dunvegan Formation in Beatton and Kiskatinaw rivers in the past. During construction of the Alaska Highway, lignite at Coal River was used for heating camps.

UPPER CRETACEOUS AND YOUNGER COAL OUTCROPS

Table 4-8-1 locates and describes all coal occurrences found during this study or mentioned in the literature. Generally references to coal occurrences in the literature are lacking in detail, some exceptions are as follows: Gulf Resources Canada Ltd. (1981) provides a detailed evaluation of the resource potential of the Wapiti Formation south of Dawson Creek; McLearn and Kindle (1950) and Williams (1944) provide some details on the Coal River lignite basin.

ACKNOWLEDGMENTS

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NOTES

STRATIGRAPHY OF COAL OCCURRENCES IN THE BOWSER BASIN

By H.O. Cookenboo and R.M. Bustin,
The University of British Columbia

KEYWORDS: Coal geology, Groundhog, Klappan, Bowser basin, stratigraphy, Currier Formation, McEvoy Formation, Sustut Group.

INTRODUCTION

Coal has been known in the Groundhog coalfield in the northern Bowser basin since the turn of the century (Buckham and Latour, 1950), but because of the remoteness of the area and the complexity of the geology the resource potential remains poorly understood. In order to better define the stratigraphy, sedimentology and coal resource potential, a field study of the Groundhog coalfield and surrounding areas has been undertaken. The purpose of this paper is to report the results of fieldwork conducted in the northern Bowser basin during the 1989 field season.

TRADITIONAL BOUNDARY OF THE GROUNDHOG COALFIELD

The boundaries of the Groundhog coalfield (Buckham and Latour, 1950) have never been formally defined, but traditionally have included coal-bearing strata from the Groundhog Range to north of Mount Klappan. Exploration was active in the Groundhog coalfield for a 10-year period ending in 1915. More recently, exploration activity has been centred near Mount Klappan, in the northern part of the coalfield, in an area now known as the Klappan coalfield (Koo, 1986). The approximate limits of the Groundhog and Klappan coalfields are shown in Figure 4-9-1.

STRATIGRAPHY

Stratigraphy proposed for the Groundhog and Klappan coalfields is summarized in Figure 4-9-2 (after Cookenboo

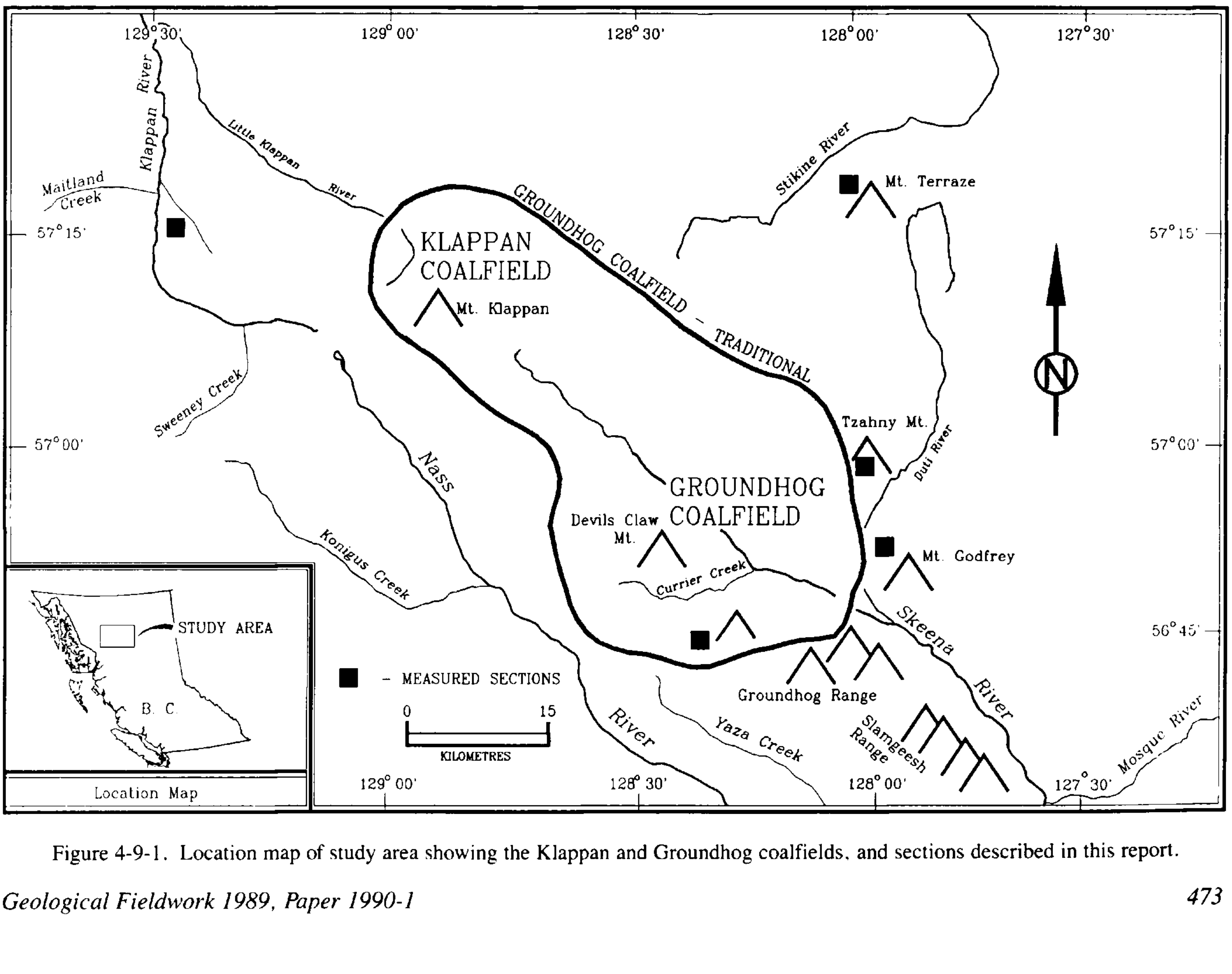

Figure 4-9-1. Location map of study area showing the Klappan and Groundhog coalfields, and sections described in this report.

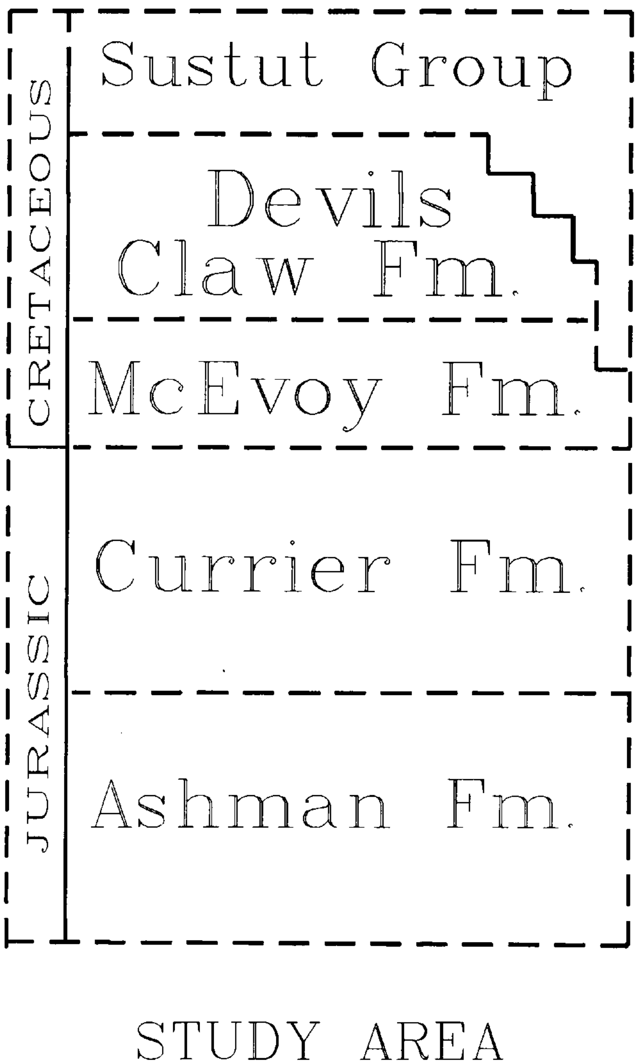

Figure 4-9-2. Stratigraphy of the northern Bowser basin.

and Bustin, 1989). From oldest to youngest, the stratigraphic units recognized in the area of the coalfields are the Ashman, Currier, McEvoy and Devils Claw formations. Sustut Group rocks occur to the east of the coalfields and include at least one coal-bearing exposure. Previous study (*c.f.* Bustin and Moffat, 1983) has documented a substantial coal resource in the Currier Formation throughout the traditional Groundhog coalfield area. Exposures examined in the 1989 field season have allowed extension of the lower contact of the Currier Formation to the top of the Ashman Formation by including the transitional facies of the informally named upper Jackson unit (Bustin and Moffat, 1983) in the Currier Formation (Cookenboo and Bustin, in press).

Three stratigraphic intervals of good coal development have been identified in the study area: the oldest occurs in the Currier Formation, the second is assigned to the McEvoy Formation and the youngest is here considered to be in strata correlative with the Sustut Group. Differences in number, thickness and rank of coal seams that have been observed in preliminary study of each interval may prove to be of great significance to exploitation of the coal resource. The best development in terms of thickness and number of seams is anthracite and meta-anthracite coals in coal measures of the Currier Formation. Thinner semi-anthracite coals typify the McEvoy Formation. High volatile bituminous coals were encountered in thinner seams in the Sustut Group.

The stratigraphic positions of the three coal-bearing zones are shown in a cross-section through the study area (Figure 4-9-3). From the cross-section, it is apparent that coal measures comprise only a fraction of the total stratigraphic thickness. Much of the remaining strata consist of lithologically monotonous, marginal marine deposits. Because of the structural complexity of the area, correlations in these rocks are difficult and the thickness assigned to each interval should be regarded as preliminary.

CURRIER FORMATION COAL ZONE

The primary coal zone of the Klappan and Groundhog coalfields occurs in the lower Currier Formation, in strata of Middle to Late Jurassic age (Cookenboo and Bustin, in press). Good coal development occurs in this zone in sections measured across a broad area of the northern Bowser basin. Currier Formation coal seams are anthracite and meta-anthracite in rank, and are the thickest in the study area, reaching 8 and 10 metres in thickness in two seams east of the Klappan River. Seams 3 to 5 metres thick have been measured in several areas, including Mount Klappan, Tzahny Mountain and Currier Creek (Bustin and Moffat, 1983).

Three sections, one 5 kilometres northwest of Panorama Mountain and two others east of Nass Lake, were measured within the Groundhog and Klappan coalfields in the 1989 field season, and two other sections of good Currier coal development were found east of the Klappan River and on Tzahny Mountain, outside the area traditionally considered part of the Groundhog coalfield.

PANORAMA MOUNTAIN

Five coal seams 1 to 3 metres thick occur in a 125-metre section northwest of Panorama Mountain. This section is well exposed on the east side of a north-trending ridge, 4 kilometres south of Currier Creek. Facies are consistent with deltaic deposition, as interpreted for the Currier Formation (Bustin and Moffat, 1983). This section is considered roughly correlative with a section previously measured on the south flank of Devils Claw Mountain, north of Currier Creek, which has been dated as late Callovian in age, based on palynoflora (G. Rouse, personal communication, 1989).

EAST OF THE KLAPPAN RIVER

Thick coals occur on the northeast side of a northwest-trending ridge east of the Klappan River, near its confluence with Maitland Creek. Two poorly exposed coal seams totalling 18 metres in thickness occur over a 55-metre interval near the base of nearly 600 metres of strata. These two thick seams are mostly covered by talus and vegetation and the top of each is marked by springs. At least five other seams between 0.5 and 2 metres thick occur higher in the section.

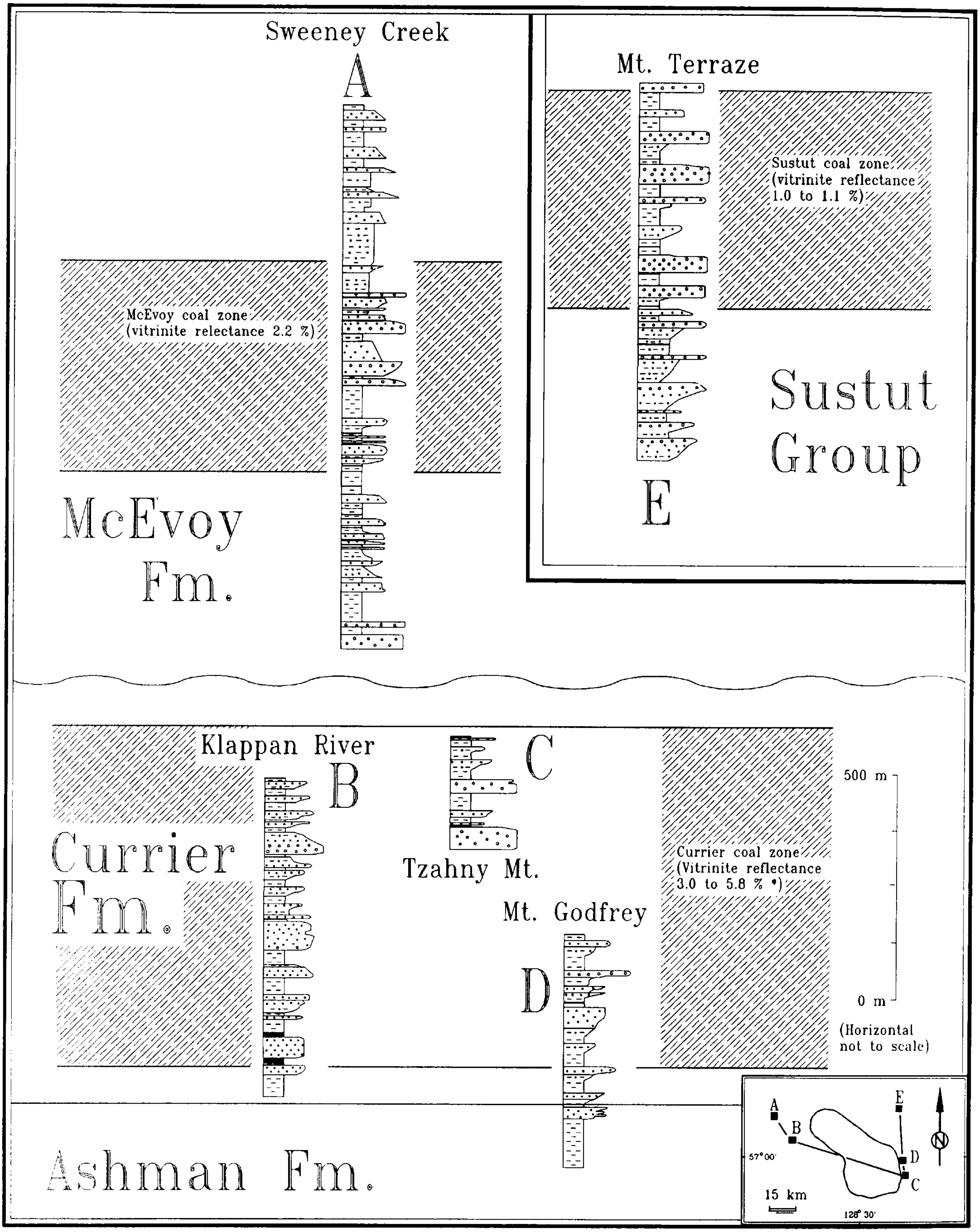

Figure 4-9-3. Stratigraphic cross-section across the north-central Bowser basin, showing the three zones of good coal development. Vitrinite reflectance values are from this report, except for Currier Formation (Bustin, 1984).

The thick coals found east of the Klappan River occur within transitional marine facies characterised by coarsening-upward sequences of mudstones, sandstones and conglomerates, with abundant biogenic structures and rare shells, in addition to plant remains. This section is correlated with the coal-bearing zone of the lower Currier Formation that is exposed within the Klappan and Groundhog coalfields 20 kilometres to the southwest. Measured vitrinite reflectance values from the two thick seams near the base of this section were 3.0 and 3.7 per cent, indicating that the coal is anthracite rank. These values are within the range previously reported from the Currier Formation, supporting the correlation with the Klappan and Groundhog coalfields. This correlation extends the reported range of the Currier Formation beyond the coalfields for the first time.

TZAHNY MOUNTAIN

Coal seams exceeding 2 metres in thickness occur in coarse-grained facies tentatively assigned to the Currier Formation on Tzahny Mountain. The structure in the area is complex, including faults and tight folds, but an apparently continuous section 205 metres thick was measured including two coal seams more than 2 metres in thickness. The facies is coarser than the Currier Formation in the Groundhog and Klappan coalfields, with conglomerates to 32 metres thick accounting for nearly 30 per cent of the section. However, the coarse nature of the facies is probably misleading, and is at least partly a result of preferential exposure of the more resistant conglomerates.

MCEVOY FORMATION COAL ZONE

Good coal development was identified in sections measured south of Sweeney Creek and tentatively assigned to the McEvoy Formation.

SWEENEY CREEK

Six seams of 1 to 3 metres thickness are exposed within a 300-metre-thick succession near Sweeney Creek. Two exposures on ridges south of Sweeney Creek were measured, with the thickest and most complete section exceeding 1000 metres in thickness. The coal occurs in a facies of interbedded mudstones and sandstones that is interpreted as deltaic in origin. The strata at Sweeney Creek are a finer grained facies of the McEvoy Formation than that previously described in the type area of the Klappan and Groundhog coalfields east of the Nass River (Cookenboo and Bustin, 1989). Coals are thicker in the Sweeney Creek area than in the McEvoy Formation within the coalfields, where seams are typically less than 0.5 metre thick.

Coal in the Sweeney Creek area is semianthracite rank, with a measured vitrinite reflectance value of 2.2 per cent. This coal is lower rank than most of the coal previously reported from the Currier Formation (Bustin and Moffat, 1989), consistent with the stratigraphically higher position and younger age of the McEvoy Formation. The McEvoy Formation is dated as mid-Cretaceous (upper Barremian or Aptian to middle or late Albian) east of the Nass River (Cookenboo and Bustin, 1989).

SUSTUT GROUP COAL ZONE

A zone of good coal development is exposed on Mount Terraze, 15 kilometres east of the Groundhog coalfield, in strata tentatively assigned to the Sustut Group (Cookenboo and Bustin, in press).

MOUNT TERRAZE

Four coal seams between 0.5 and 2 metres thick were measured in the upper 200 metres of a more than 600-metre-thick exposure on the northeast flank of Mount Terraze. The coal seams are poorly exposed on talus slopes between resistant conglomerate cliffs. Because of the poor exposure, the measured seams probably represent a minimum value for the total thickness of coal in the section. Additional coal potential is suggested by the local structure. The top of the section is a dip-slope at the the top of Mount Terraze, dipping down towards Ella Creek to the east, suggesting the possibility that related coal-bearing strata may exist in the Ella Creek valley.

Vitrinite reflectance values of 1.0 and 1.1 per cent have been measured from these coals. Such values are equivalent to high volatile bituminous rank. These coals are significantly lower rank than the semianthracite to meta-anthracite of Currier and McEvoy Formation coals. Coals of this rank have previously only been reported in northern British Columbia from the Sustut Group (Bustin, 1984; Smith, 1989). This relatively low rank supports the assignment of the these coals to a third and probably younger coal zone, and is consistent with assigning the strata to the Sustut Group (Cookenboo and Bustin, in press).

ACKNOWLEDGMENTS

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NOTES

Industrial Minerals Studies

PERLITE AND VERMICULITE OCCURRENCES IN BRITISH COLUMBIA

By G.V. White

KEYWORDS: Industrial minerals, perlite, vermiculite, Frenier, Francois Lake, Nechako River, Port Clements, Joseph Lake, Sowchea Creek.

INTRODUCTION

The term "perlite" applies to hydrated volcanic glass of rhyolitic composition which, when heated to temperatures as low as 510°C, will expand to form a white, porous, lightweight material. In its expanded form, perlite is used primarily as an insulating aggregate in plaster and concrete, as a loose-fill insulation, in horticultural applications and as a filtering agent. Generally perlite is thought to form by the secondary hydration of obsidian. Combined water in perlite exists in two forms, molecular and as hydroxyl ion, with the ratio of one to the other varying with location. Deposits are restricted to volcanic belts ranging in age from Tertiary to Quaternary. Commercial deposits of perlite have a variety of textures grading from pumiceous to obsidian, which appear to be related to depth of burial.

"Vermiculite" is a micaceous mineral that rapidly expands on heating, to produce a low density material. Like perlite, vermiculite swells when heated, with individual flakes expanding up to 30 times – a process called exfoliation. If heated in an oxidizing atmosphere vermiculite turns dull grey or tan but if heated in a reducing atmosphere the product is bronze or gold. These light-weight commercially valuable products are commonly used in construction, agriculture or

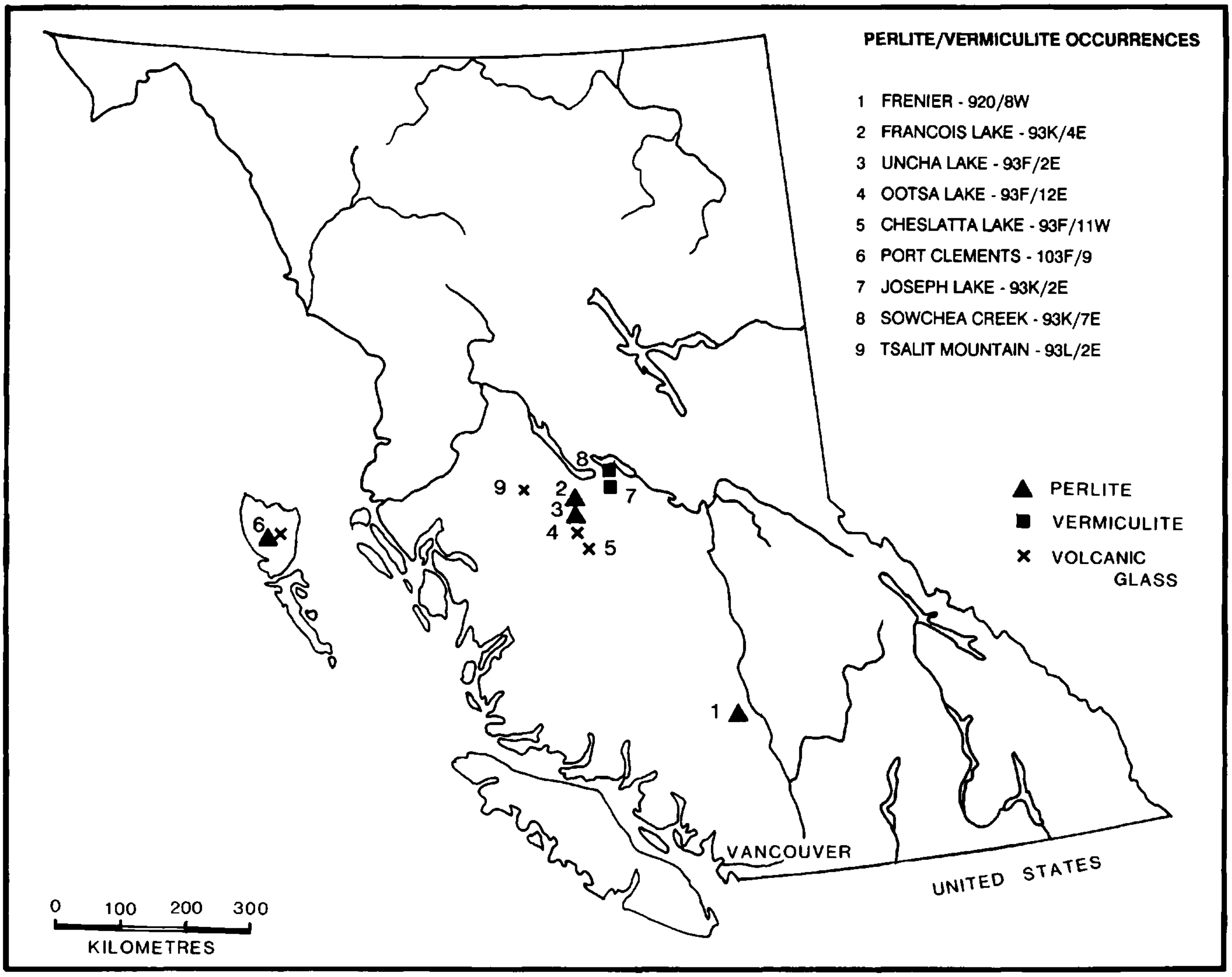
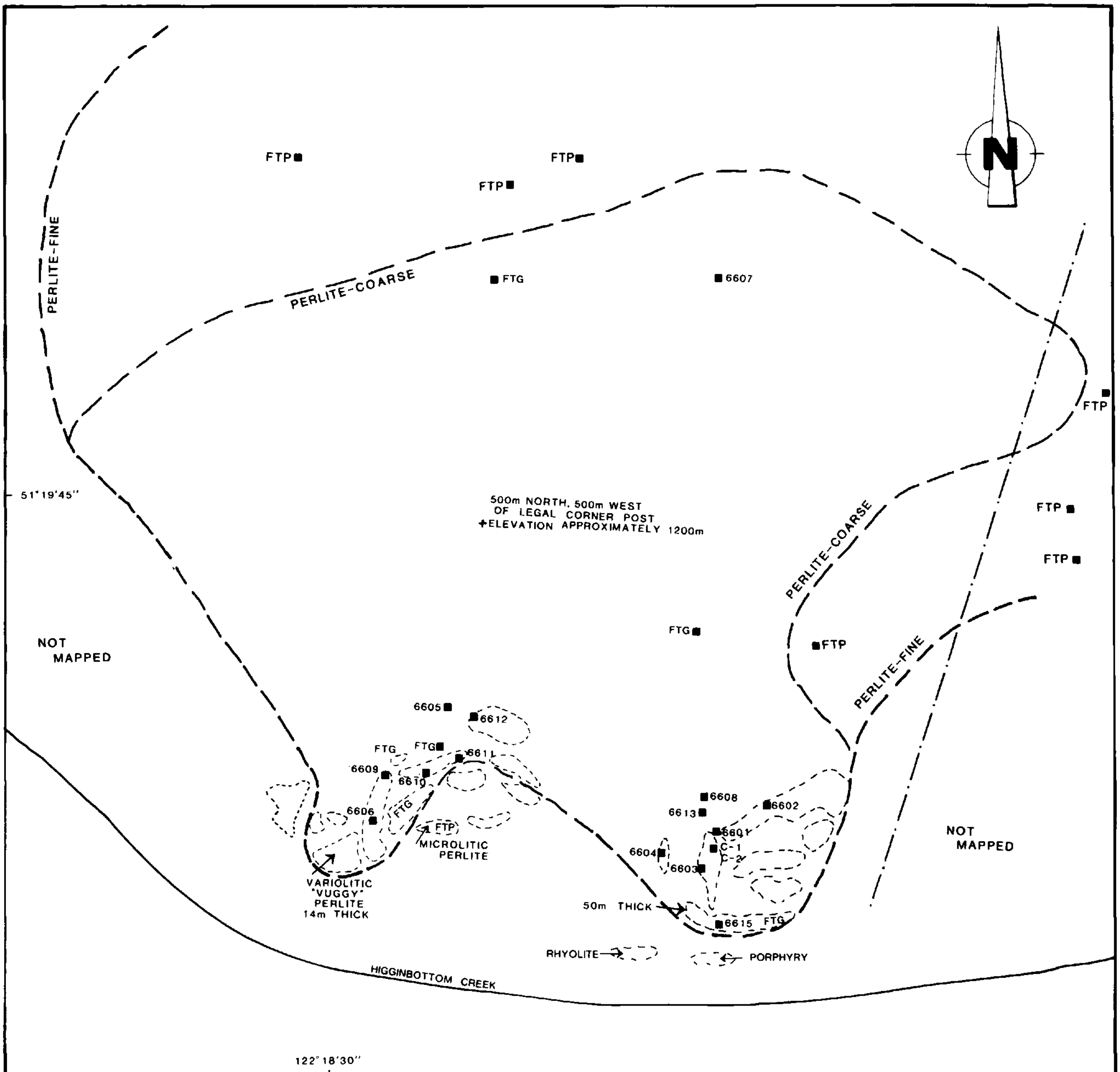


Figure 5-1-1. Locations of perlite/vermiculite occurrences in British Columbia.



SAMPLE No.	CRUDE BULK DENSITY Gms/cc	PRODUCT BULK DENSITY Gms/cc	EXPANSION FACTOR	PERCENTAGE RETAINED
6601	1.21	0.12	10.08	87.0
6602	1.13	0.17	6.64	78.4
6603	1.26	0.09	14.00	73.4
6604	1.20	0.22	5.45	73.5
6605	1.17	0.25	4.68	79.2
6606	1.22	0.13	9.38	70.03
6607	1.24	0.12	10.33	85.68
6608	1.23	0.12	10.25	76.80
6609	1.21	0.21	5.76	79.97
6610	1.18	0.21	5.62	78.60
6611	1.19	0.31	3.84	68.20
6612	1.17	0.22	5.32	76.78
6613	1.24	0.20	6.20	81.25
6615	1.19	0.23	5.17	
C-1	1.25	0.12	10.42	
C-2	1.22	0.13	9.32	

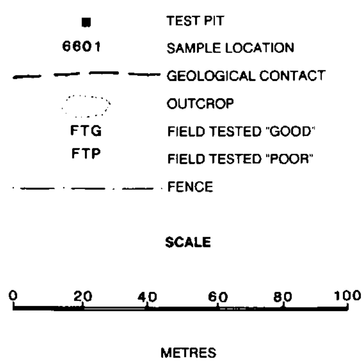


Figure 5-1-2. Frenier perlite deposit located 60 kilometres northwest of Clinton, B.C. (92O/08W); adapted from Horne, 1983.

for horticultural applications. To be commercially viable, vermiculite deposits must contain flake material larger than 65 mesh with the recovery greater than 20 per cent by volume. Vermiculite is usually mined from altered basic intrusive rocks or ultramafic layered metamorphic rocks. Biotite is the most common parent mineral.

At present, all perlite and vermiculite products used in British Columbia are imported from the United States. As a first step to assess the potential to develop local production, a study of known occurrences was undertaken. Of ten known perlite occurrences in the province, only three had partial geological descriptions. Three vermiculite occurrences with a sizeable tonnage potential were reported in 1987 and 1988 but only one has a report in the Ministry's files. A systematic, "grassroots" evaluation of six volcanic glass (perlite) deposits and two vermiculite occurrences was carried out during 1989 (Figure 5-1-1). Sites were selected on the basis of their accessibility to established transportation networks, production record (if any) and geological setting. In most instances the geological, physical and chemical characteristics of each site were unknown.

This report describes these occurrences and their physical characteristics and, where possible, evaluates their potential as a commercial source of perlite or vermiculite. All sites are accessible by road. Representative samples collected at each site were heated using a hand-held propane torch. This simple test is effective in determining whether the material is expandable.

Preliminary results indicate five perlite occurrences contain expandable material (Frenier, Francois Lake, Uncha Lake, Blackwater Creek, Gold Creek) and material from both vermiculite showings swells when heated (Fraser Lake, Fort St. James). Laboratory and process testing of bulk samples are required to assess whether the occurrences contain material which meets industry specifications, and to compare the quality of perlite from individual locations.

PERLITE OCCURRENCES

FRENIER DEPOSIT (MINFILE 92O 072)

The Empire Valley perlite deposit was discovered by Lawrence Frenier in 1949. It is located approximately 60 kilometres northwest of Clinton at 1200 metres elevation on the eastern slope of Blackdome Mountain. The property, developed by Aurun Mines Limited and named after the prospector who discovered it, saw production from 1983 through 1985 producing 1000, 2000 and 3000 tonnes respectively. Crude perlite was shipped by truck, initially to the company's pilot-plant in Aldergrove, B.C., and later to its processing facility in Surrey where the expanded product was marketed under the tradename Aurolite. The mine has been inactive since 1986 because of transportation difficulties resulting from an old, low-capacity bridge across the Fraser River. Various industry sources indicate that the quality of Frenier perlite is far superior to any imported rock.

GEOLOGY

Volcanic rocks at the site are assigned to the Eocene Kamloops Group and consist of devitrified rhyolite tuff, vesicular rhyolite flows, rhyolite crystal tuff, perlite and volcanic breccia with clasts of varied composition (Green and Trupia, 1989). In outcrop the perlite is a homogeneous, light grey, glassy rock, crosscut by veins of opalline silica and pitchstone. When heated using a hand-held propane torch, crushed perlite expands rapidly to many times its original size, similar to heated corn kernels.

The deposit consists of a flat-lying flow of volcanic glass with occasional shards of glass welded together to form tuff. Flow direction has not been established but the deposit is massive, appears domed, and exhibits perlitic (onion skin) textures.

RESERVES

The shape of the orebody is illustrated in Figure 5-1-2. The deposit has been divided into "coarse" and "fine" perlite with inferred reserves calculated by Aurun Mines Limited of 3.8 million tonnes, using an average thickness of 30 metres and a specific gravity of 2.3. There is possiblity that the estimated resource reported by the company can be increased.

FRANCOIS LAKE PROSPECT (MINFILE 93K 001)

First reported by G.M. Dawson in 1876 and staked in 1948 by N.B. Davis of Ottawa, Ontario, the Francois Lake perlite showing was sold to Western Gypsum Products Limited of Winnipeg a year later. In 1953 this company produced 1100 tonnes of perlite from a quarry located on the north shore of Francois Lake approximately 22 kilometres south of the town of Burns Lake. The mineral was processed at the company plant in Calgary, but its eventual use and value are not known. The quarry has not been worked since that time.

GEOLOGY

The Francois Lake perlite occurs in a package of rocks considered to be Eocene to Oligocene and possibly Paleocene in age, consisting of rhyolite, dacite and associated tuffs and breccias minor andesite, basalt and conglomerate (Tipper, 1963).

Volcanic glass crops out at four separate sites (Figure 5-1-3) and is medium grey on weathered surfaces but dark grey to black on freshly exposed outcrops. Commonly, exposed perlite crumbles into marble-sized angular pellets. When heated with a hand-held propane torch, perlite from Sites A and B expands a similar amount to that tested at the Frenier deposit. At Sites C and D, however, tested volcanic glass did not expand.

Perlite beds at Site A strike northeast and dip 15° to 35° northwest. The rock exhibits typical onion-skin texture with radiating fractures perpendicular to strike. In places it is brecciated and siliceous with pronounced flow banding and, in all locations, perlite beds sit in sharp contact with cherty rhyolite both above and below exposed outcrop. At the lakeshore, perlite is exposed in a 2-metre bed over 15 metres. Twenty metres away from the shore much of the overburden was removed during quarry development and a small stockpile remains, but fresh outcrop is not exposed.

At Site B, 300 metres north of the lake, perlite is exposed intermittently for 110 metres along an access road. At the north end of the roadcut, fresh perlite is exposed continuously for 50 metres. The 15-metre-thick bed strikes northeast dipping 30° northwest and lies on coarse, grey tuff and under 10 metres of unconsolidated silty overburden.

Volcanic glass crops out at two sites 300 metres north of Site B (Site C). It is medium grey in colour, but unlike rock at Sites A and B, contains hard, dense, spherulitic aggregates up to 2 centimetres in diameter. No other rock types are exposed at this site but small outcrops of rhyolite occur to the west.

At Site D, 100 metres northeast of Site C, a trench 24 metres long and 1.5 to 2 metres deep exposes volcanic glass striking northwest and dipping 20° southwest. Grey to brown cherty rhyolite crops out immediately east of the exposed perlite and white rhyolite crops out to the west.

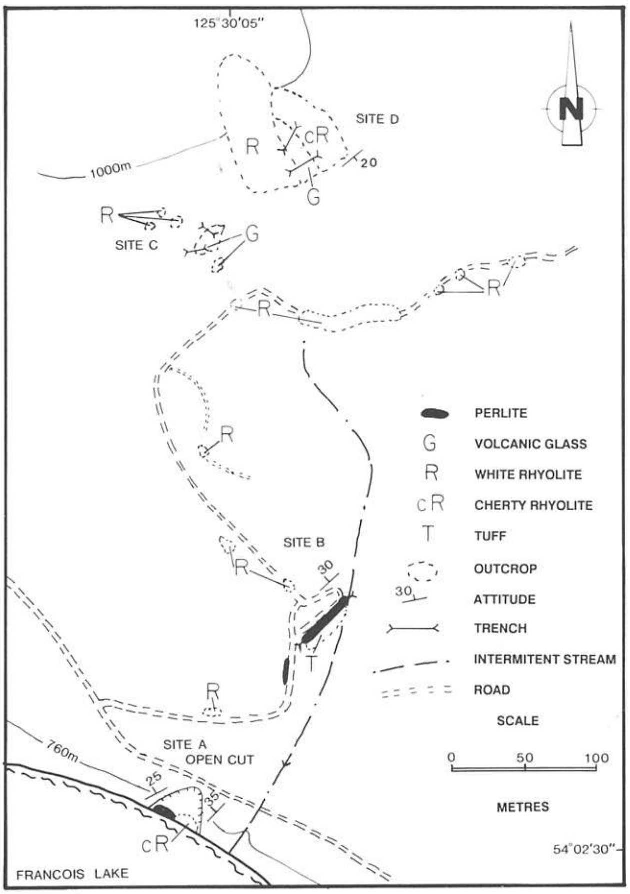


Figure 5-1-3. Francois Lake perlite prospect 22 kilometres south of Burns Lake (93K/04E); adapted in part after McCammon (1949).

Nechako River Area Perlite/ Volcanic Glass Occurrences

The Uncha Lake (perlite), Cheslatta and Ootsa Lakes (volcanic glass) occurrences are in a volcanic package mapped by Tipper (1963), as Paleocene (?), Eocene and Oligocene Ootsa Lake Group, consisting of rhyolite, dacite and associated tuffs and breccias, minor andesite, basalt and conglomerate. A brief description of each occurrence follows:

UNCHA LAKE PROSPECT (MINFILE 93F 026)

Originally staked in 1953 by C.S. Powney and J. Rasmussen of Fort St. James and their associates, the Uncha Lake perlite prospect has been explored by trenching and limited laboratory processing tests. The property is located 40 kilometres south of the town of Burns Lake on the northwest slope of Dayeezcha Mountain between 975 and 1125 metres elevation. British Columbia Minister of Mines reports indicate that in 1955 Technical Mines Consultants Limited exposed six mineable perlite layers along a zone 850 metres long and 500 metres wide. The company reported the layers are "irregular in width and attitude, lying interbedded in a folded series of rhyolites striking generally northeast and dipping about 70° to the southeast". James (1955) reports the maximum exposed width of at least two layers exceeds 45 metres, and that in some places interbedded rhyolite is sufficiently narrow to permit practical open-pit mining of two or more layers from one pit. Currently the property is inactive and the old trenches are partially filled. Past company records are not available so the following description is based on field observations only.

GEOLOGY

Figure 5-1-4 shows the rock types mapped at the perlite prospect and their distribution. A description of each follows:

Perlite is intercalated with light to dark grey porphyritic and sometimes cherty rhyolites and ranges in colour from brown to medium grey to black to pale green. It often has a good pearly lustre but when exposed for periods of time tends to break down into 2 to 3-centimetre subangular fragments. Uncha Lake perlite expands when heated with a hand-held propane torch although not as rapidly as samples from the Frenier deposit.

Perlite is exposed in trenches south of the access road but not enough bedrock is exposed to determine whether these occurrences represent a single unit. Significantly, fresh, medium grey perlite is exposed along a ridge west of the trenched area. Structural information is limited but exposures in trenches indicate the host rhyolite strikes northeast and dips steeply southwest.

Rhyolite, in sharp contact with perlite, ranges from white to dark grey in colour. Both white and grey varieties contain 1 to 7-centimetre bands of darker "cherty" quartz (chalcedony?) or patches, up to 3 centimetres across, of light green silica possibly indicative of hydrothermal alteration. Rhyolite is occasionally porphyritic with 1 to 5-centimetre rectangular phenocrysts of potassium feldspar in a fine-grained matrix. Near the southern end of the access road siliceous angular fragments, 5 to 7 centimetres across, are observed in rhyolite.

OOTSA LAKE SHOWING (MINFILE 93F 028)

Three kilometres southeast of False Hill and 800 metres north of Intata Reach, at 975 metres elevation, a 2 to 3-metre bed of volcanic glass/chert is exposed along the side of a hill

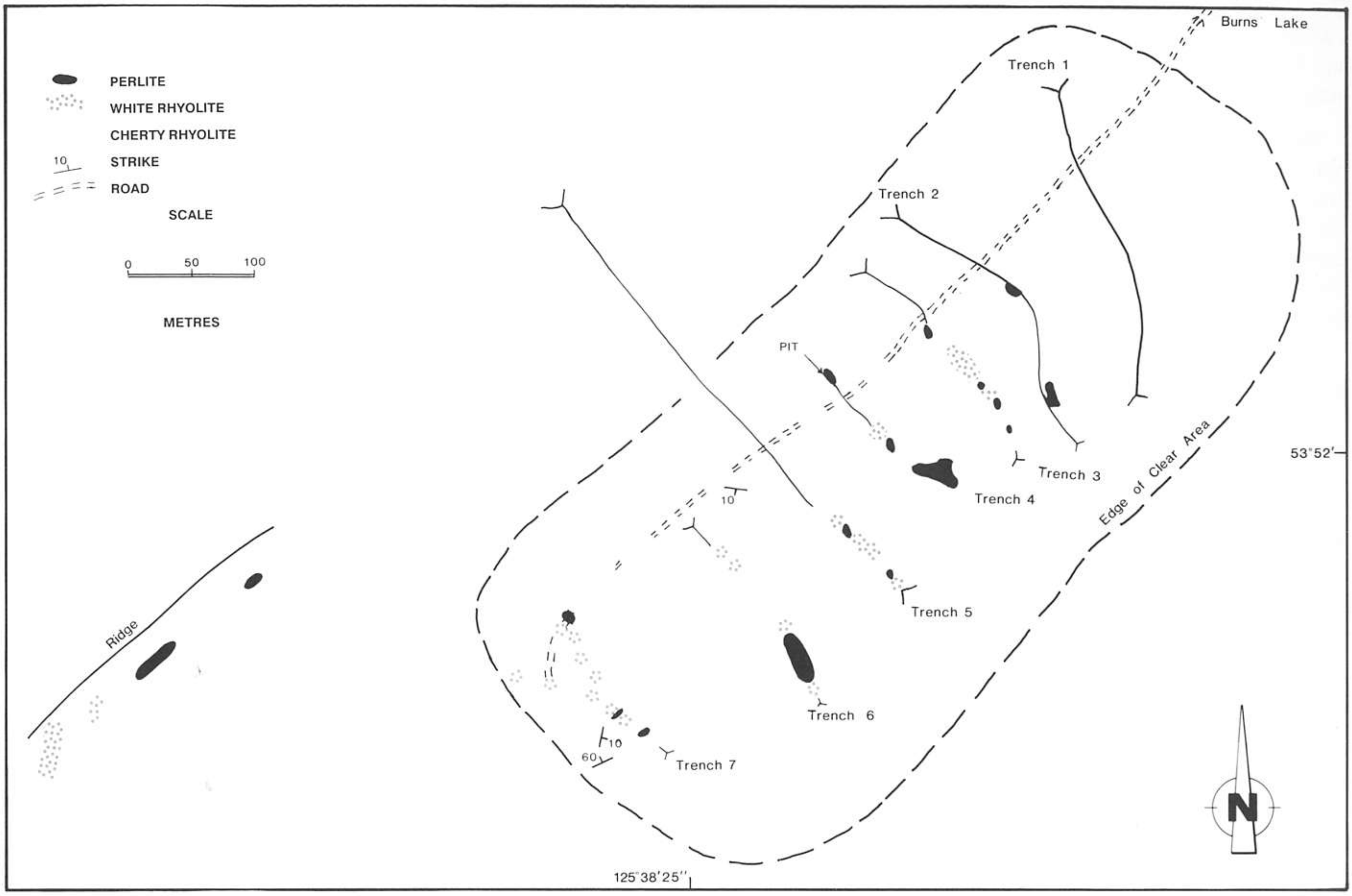


Figure 5-1-4. Uncha Lake perlite prospect located 40 kilometres south of Burns Lake (93F/13E).

for 250 metres (Figure 5-1-5). The bed strikes northeast, dipping variably 35° to 70° northwest, in pale to medium green cherty rhyolite. Commonly, 4 to 6-centimetre beds of pale green chert alternate with beds of dark grey to black volcanic glass. The bed splits into two separate seams approximately one-third of the way along its length. When heated with a hand-held propane torch the rock "explodes" without apparent swelling.

CHESLATTA LAKE OCCURRENCE (MINFILE 93F 027)

It required extensive searching in heavily forested and till-covered terrain to locate a 5 to 7-metre outcrop of weathered,

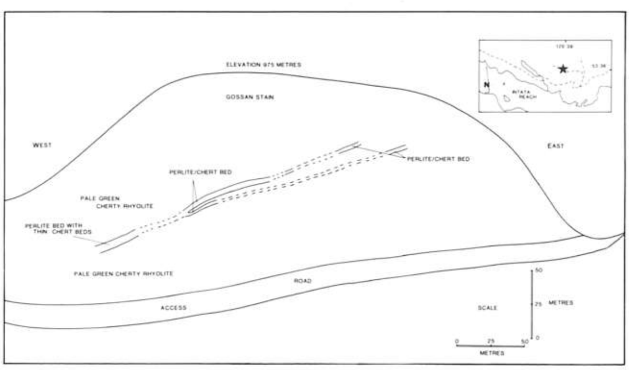


Figure 5-1-5. Sketch section of the Ootsa Lake perlite showing located 66 kilometres south of Burns Lake (93F/12).

medium grey to pearly black volcanic glass. The showing was located 1.5 kilometres south of Cheslatta Lake at latitude 53°43′15″N and longitude 125°27′W. The volcanic glass occurs as lenses in dirty white to light grey rhyolite. Samples heated using a propane torch glowed red but did not expand.

TSALIT MOUNTAIN VOLCANIC GLASS (MINFILE 93L 258)

Volcanic breccia, glassy rhyolite lava, tuff and dikes were reported by Church (1972) near Fenton Creek, west of Tsalit Mountain. This sequence, which is the uppermost part of the Francois Lake Group, has been dated as Eocene. The reported chemical analysis of the glass indicates a significant percentage of water but the glass does not expand when heated.

PORT CLEMENTS AREA

BLACKWATER CREEK (MINFILE 103F 022)

Occurrences of volcanic glass and perlite are documented by Sutherland Brown (1968, page 175) in dikes and flow-like masses in rhyolitic units of the Masset Formation, Queen Charlotte Islands. Proximity to tidewater, with the possibility of using inexpensive water transportation to reach major market areas along the coast, makes the Graham Island occurrences particularly interesting. Two of these sites (perlite) and three sites recently mapped by Cathy Hixon of the

Geological Survey of Canada (personal communication, 1989) were tested using a propane torch. Material at Sites 2, 4 and 5 expanded when heated while rhyolitic units at the two other sites did not (Figure 5-1-6).

Site 2: Medium grey to black (fresh surfaces) perlite crops out for 85 metres along a roadcut immediately northeast of bridge Q9 on Blackwater creek. The bed strikes north and dips 65° east. Till and forest cover away from the road prevent examination of the unit but tested samples expanded to several times their volume. They did not however, expand as quickly or as much as the Frenier perlite.

Site 4: Black, medium-grained glassy dacite containing 2 to 4-millimetre phenocrysts of potassium feldspar violently popped when heated, not swelling gradually as other tested perlite. This expandable rock crops out for 300 metres along a roadcut above Florence Creek.

Site 5: Several large boulders of medium grey perlite (3 to 5 metres across) were found along the road just north of Gold Creek. Samples expanded when heated but did not compare in volume to Frenier perlite. Traverses above the site failed to locate perlite in place.

Sampling and testing of volcanic glass and perlite occurrences in this study was restricted to the Port Clements area for logistical reasons. Other felsic and perlitic units in the Masset Formation may contain expandable rock and should be the focus of exploration programs in the Queen Charlotte Islands as the accessibility of other sites improves.

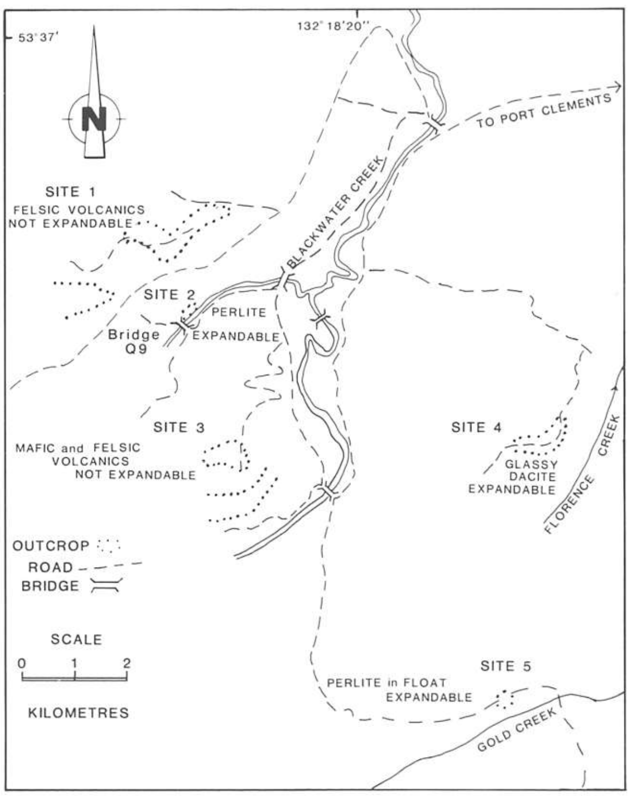


Figure 5-1-6. Perlite prospects, Port Clements area, Queen Charlotte Islands.

VERMICULITE

JOSEPH LAKE OCCURRENCE (MINFILE 93K 077)

The Joseph Lake vermiculite prospect, 14 kilometres southeast of Fraser Lake, was staked in 1987 by J. Steiner of Fraser Lake. The area is underlain by Jurassic granite, granodiorite and quartz diorite (Tipper *et al.*, 1979). Granite at the site is medium grained (2 to 3 millimetres), light grey in colour (with a pink tone), and contains rectangular plagioclase phenocrysts up to 7 centimetres long. An access road crossing the prospect exposes a weathered zone approximately 75 metres long (Figure 5-1-7). Weathered rock has a distinct reddish orange colour and contains mica flakes which swell when heated with a propane torch. Immediately northeast of this zone fresh granite crops out along a prominent ridge. In many places mica flakes from fresh-looking rock also expand on heating.

The nature and percentage of expandable mica concentrated in the weathered zone is not yet known and analytical, petrographic and processing tests are required to evaluate the potential of this prospect.

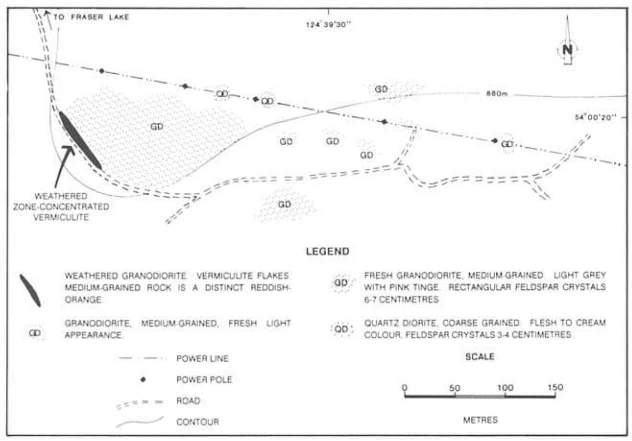


Figure 5-1-7. Joseph Lake vermiculite prospect 14 kilometres southeast of Fraser Lake (93K/02E).

SOWCHEA CREEK SHOWING (MINFILE 93K 083)

The Sowchea Creek vermiculite prospect, staked by A. Almond in 1987, is located 17 kilometres southwest of Fort St. James. Most of the area is covered by glacial till but Jurassic medium-grained hornblende diorite crops out north of Sowchea Creek. Here, vermiculite derived from weathered diorite is concentrated along a 150-metre zone (Figure 5-1-8). Laboratory and petrographic analyses of samples collected from the zone are required to establish the percentage of vermiculite and its extent.

Both the Joseph Lake and Sowchea Creek expandable mica prospects occur in both fresh and weathered zones of Jurassic intrusions. Other similar intrusions are mapped in the region (Tipper *et al.*, 1979), and should be the focus of prospecting or exploration programs.

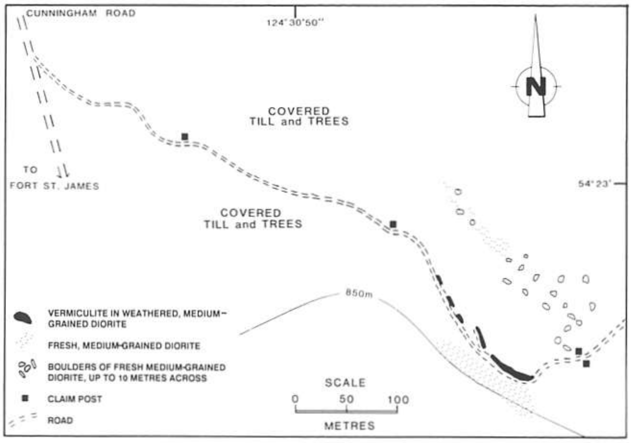


Figure 5-1-8. Sawchea Creek vermiculite prospect located 17 kilometres southwest of Fort St. James (93K/07E).

SUMMARY

Of the six perlite and two vermiculite sites described, one, the Frenier deposit, located west of Clinton, has reserves outlined. Two other sites, Francois Lake and Uncha Lake contain significant amounts of expandable perlite and are potential commercial sources. The extent of expandable perlite at Blackwater Creek is not known. Material tested at Ootsa Lake and Sites 4 and 5, south of Port Clements, did not swell when heated but rather "exploded", unlike volcanic glass from Cheslatta Lake and Tsalit Mountain, which glowed red without expanding.

Additional exploration is necessary to fully assess the potential of each site and processing tests on bulk samples are required to document each deposit's potential to produce perlite/vermiculite which will meet industry specifications.

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NOTES

SEDIMENTARY PHOSPHATES IN THE FERNIE BASIN: DEVELOPMENT OF NEW TECHNOLOGY FOR DIRECT APPLICATION TO SOILS (82G AND 82J)

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KEYWORDS: Economic geology, sedimentary phosphates, Fernie basin, zeolites, Princeton basin, agrogeology.

INTRODUCTION

The Canadian fertilizer and agriculture industries are entirely dependent on a reliable and economic source of phosphate rock. Canada currently imports all of its phosphate rock requirements from Florida and the western United States. Increased demand and expected depletion of the high-grade Florida deposits, however, suggest that Canada may have to seek a new source of supply, face considerably higher import costs, or develop its own resources in the future.

The most recent evaluation of phosphate potential in British Columbia has focused on the sedimentary phosphates of the Fernie basin in the southeastern part of the province. A province-wide inventory of phosphate resources conducted by Butrenchuk (1987), reported that the best potential for development occurs at the base of the Jurassic Fernie Formation in the Fernie basin. At present, this phosphorite, which averages 18 to 20 per cent P_2O_5, cannot compete with the Florida deposits.

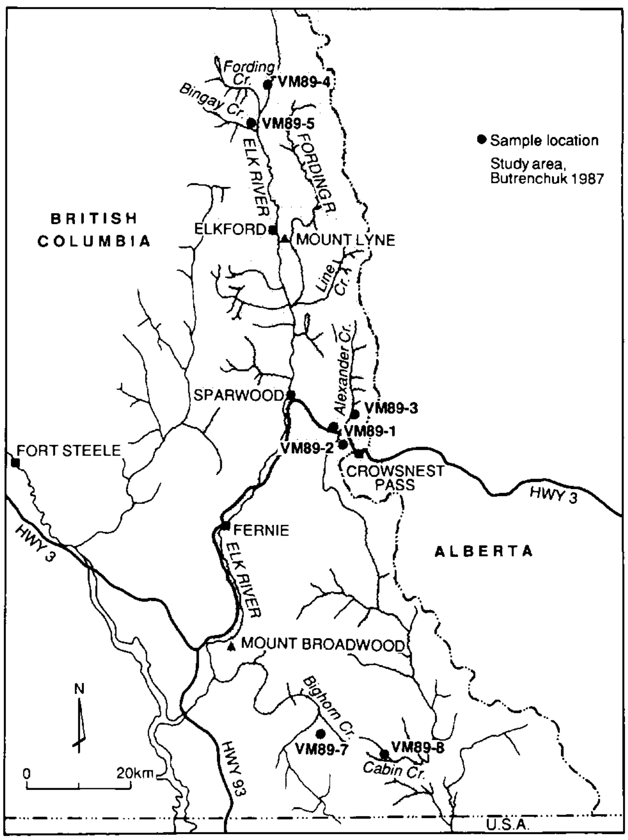

Figure 5-2-1. Simplified geological map of the Fernie basin showing sample locations. Modified from Butrenchuk (1987).

The objective of this study is twofold: to evaluate the suitability of ground phosphate rock from the Fernie basin for direct application as a fertilizer, and to develop new technology for the utilization of British Columbia phosphate rock and zeolites in agriculture. This would provide an alternative to existing chemical fertilizers and increase the economic potential of the British Columbia deposits. This study is an extension of previous work on industrial zeolites in the Princeton basin (Marcille, 1989).

Using Butrenchuk's inventory as a guide, samples of the basal Fernie phosphorite were collected from seven exposed sections, VM89-1 to VM89-8 (Figure 5-2-1). These correspond to sample locations 3 (Highway 3), 31 (Alexander), 4 (Crow), 13 (Abby), 14 (Bingay), 42 (Bighorn), and 38 (Cabin) in the earlier study (Butrenchuk, 1987). This report summarizes the results of the laboratory characterization and initial dissolution experiments. Future work will include further laboratory and greenhouse studies.

GENERAL GEOLOGY

The Fernie basin is a broad, canoe-shaped synclinorium lying within the thrust and fold belt of the Rocky Mountains, and covering roughly 2000 square kilometres of southeastern British Columbia. In the basin, the phosphatic shales, pelletal phosphorite and lesser limestone, siltstone and sandstone of the Jurassic Fernie Formation unconformably overlie the fine clastic sediments and carbonates of the Triassic Sulphur Mountain and Whitehorse formations. The Fernie Formation shows a general thickening westward, reaching thicknesses of 70 to 365 metres (Butrenchuk, 1987) and is overlain by nonmarine, coal-bearing Cretaceous strata.

A pelletal phosphorite bed, 1 to 2 metres thick and averaging 20 per cent P_2O_5 occurs at the base of the Fernie Formation. Deposited as a single bed or as two beds interlayered with phosphatic shale, the phosphorite reflects a period of rapid marine transgression and slow clastic accumulation during Sinemurian time. The phosphate occurs as dark brown pellets of apatite and organic matter in a finer grained matrix. Comprising 50 to 85 per cent by volume of the phosphorite, the pellets are well sorted, subrounded, essentially structureless and 0.1 to 0.3 millimetre in diameter. The basal phosphorite is overlain by phosphatic shales of variable thickness and a yellow, calcareous marker bed a few centimetres thick.

In addition to the synclinal fold of the Fernie basin, certain structural features are significant in the evaluation of the phosphate resources. Extensive structural deformation has resulted in repetition of beds and local thickening of the basal phosphorite. Areas of thickened phosphorite beds are of economic significance, as are overturned beds; the more competent Triassic strata form a more desirable back for underground mining (Butrenchuk, 1987).

MINERALOGY

The mineralogy of the phosphorite samples was determined using a Rigaku Geigerflex x-ray diffractometer under the following operating conditions: copper k-alpha radiation, step-scan mode, 4 second count time and 0.05° step width.

Apatite is present in all samples as are quartz and minor amounts of feldspar. Calcite is a constituent of VM89-1 to VM89-4, and dolomite occurs in trace amounts in VM89-1 and VM89-2.

Using a revision for microcomputers by Paul Benoit, University of Alberta, the least-squares powder diffraction program by Appleman and Evans (1975) was used to calculate the a-cell parameters for the apatites from their respective d-spacings. Values for VM89-2 to VM89-8 fit the range defined by McClennan and Gremillion (1980) for francolite (0.9322 to 0.9376 nanometres). The a-cell value of 0.9377 nanometres for VM89-1, however, suggests that the apatite in this phosphorite is fluorohydroxyapatite.

In addition to lowering the ore grade of the phosphate rock, the presence of certain accessory minerals has deleterious effects on commercial processing. Quartz and feldspar can interfere with grinding and beneficiation, increasing wear on plant equipment. Futhermore, the presence of acid-soluble silicates results in the release of impurities such as potassium, iron, aluminum and magnesium during acidulation. Carbonates, which generally cannot be removed by beneficiation, present problems in chemical processing. During acidulation, carbonates cause excessive foaming, consume acid and contribute impurities, particularly magnesium from dissolved dolomite. It is, however, possible to separate carbonate minerals from the phosphate rock by calcining, but this process, which converts carbonates to oxides, followed by slaking and separation, is costly (McClennan and Gremillion, 1980).

GEOCHEMISTRY

Three samples, VM89-3, VM89-4 and VM89-8, were selected for further study and were analysed using a Philips XRF spectrophotometer at the Department of Geology, McMaster University. The results are presented in Table 5-2-1.

The P_2O_5 contents range from approximately 30 to 34 per cent. While their relative order is consistent, the actual values are notably higher than those of 22 to 25 per cent cited by Butrenchuk (1987). Similarly, the silica contents are considerably lower than those determined in the earlier study. These differences may be due to variability within the phosphorite bed.

CaO/P_2O_5 and R_2O_3/P_2O_5 ratios were calculated and are presented in Table 5-2-1 ($R_2O_3 = Al_2O_3 + Fe_2O_3 + MgO$). The following ranges are those acceptable to commercial processing plants (Butrenchuk, 1987):

- P_2O_5 : 27 to 42 per cent
- CaO/P_2O_5 : 1.32 to 1.60
- R_2O_3/P_2O_5 : less than 0.1
- MgO : less than 1.0 per cent

All three samples fall within the acceptable ranges with the exception of the R_2O_3/P_2O_5 value of 0.12 for VM89-8. In the remaining categories, however, VM89-8 has the most desirable specifications.

TABLE 5-2-1
X-RAY FLUORESCENCE DATE FOR VM89-3, VM89-4 AND VM89-8

	VM89-3	VM89-4	VM89-8
SiO_2 (%)	5.34	7.56	10.97
Al_2O_3 (%)	1.11	1.03	1.71
Fe_2O_3 (%)	0.53	0.51	1.94
MgO (%)	0.40	0.40	0.35
CaO (%)	47.72	46.99	45.43
Na_2O (%)	0.34	0.30	0.26
K_2O (%)	0.15	0.15	0.31
TiO_2 (%)	0.07	0.07	0.16
MnO (%)	0.01	0.01	0.02
P_2O_5 (%)	29.81	31.15	33.96
S (%)	1.45	1.06	0.44
Total C (%)	3.75	2.97	2.04
Organic C (%)	1.94	1.69	1.89
Carbonate C (%)	1.81	1.28	0.15
LOI (%)	13.05	10.82	5.35
CaO/P_2O_5	1.60	1.51	1.34
R_2O_3/P_2O_5	0.07	0.06	0.12

DIRECT AVAILABLE PHOSPHORUS

When phosphate rock is to be used for direct application, the reactivity of the rock is a major concern. Although this is only one of several factors affecting its agronomic efficiency, the solubility of the rock, as determined by various extracting solutions, shows a strong correlation with both yield and plant uptake of phosphorus (IFDC, 1979). Leon *et al.* (1986) found that the relationship between reactivity and plant response was strongest when neutral ammonium citrate was the extractant: a modified version of their methodology was used in this study.

The values for direct-available P_2O_5 (expressed as per cent of rock) are 0.50, 0.63, 0.63, 1.10, 1.27, 1.19 and 1.17 respectively for VM89-1 to VM89-8. A comparison of these values with other phosphate rocks used for direct application indicates that these samples are moderately reactive (Leon *et al.*, 1986).

DIRECT APPLICATION OF PHOSPHATE ROCK

In 1979, the direct use of finely ground phosphate rock accounted for 4 per cent of world phosphate fertilizer consumption, with the U.S.S.R. accounting for over 70 per cent of the reported use (IFDC, 1979). This may, however, be an underestimate of actual use due to unreported consumption and inclusion with chemical fertilizer use.

The effectiveness of direct application is limited in many soils by the relatively low solubility of phosphate rock. A simplified dissolution reaction for hydroxyapatite is as follows (Chesworth *et al.*, 1987):

$$Ca_5(PO_4)_3OH + 7H^+ = 5Ca^{2+} + 3H_2PO_4^- + H_2O$$

By the law of mass action, an accumulation of calcium ions inhibits the breakdown of phosphate rock and the subsequent release of phosphorus to plants. A possible solution to this problem is the addition of zeolites, naturally-occurring aluminosilicate minerals which have a characteristically high cation exchange capacity and an affinity for certain cations. Acting as a sink for Ca^{2+}, zeolites can enhance the dissolution of phosphate rock (Lai and Eberl, 1986).

Natural zeolites suitable for this system are indigenous to British Columbia. Based on previous findings, a clinoptilolite from the Princeton basin was chosen for the current study. This zeolite, Tailings Ash (VM88-9), has an average cation exchange capacity of 147 milli-equivalents per 100 grams, and an average calculated zeolite content of 67 per cent (Marcille, 1989).

Given the negative effect of calcium ions on the dissolution of apatite, it is apparent that phosphate rocks with little or no calcite and dolomite are desirable for direct application. Of the sampled deposits, carbonates are absent in VM89-5, VM89-6 and VM89-8. A comparison of the CaO/P_2O_5 ratios in Table 5-2-1 suggests that VM89-8 is better suited for direct application than VM89-3 and VM89-4.

DISSOLUTION STUDIES

In an effort to better understand the mechanisms involved in the phosphate rock–zeolite system and to maximize its effectiveness, controlled dissolution experiments were conducted. In the first study, the effect of both untreated versus ammonium-exchanged zeolite of various grain sizes on the dissolution of VM89-8 phosphate rock was studied. In brief, the procedure involved shaking 200 milligrams of powdered phosphate rock with 1.00 gram of zeolite in 30 millilitres of distilled water for four days at 25°C. The solutions were then centrifuged and the supernatants were analysed colorimetrically for water-soluble phosphorus using a Technicon II Auto Analyser.

For the grain size fractions studied (powdered, 0.053 to 0.25 millimetre, 0.25 to 0.5 millimetre, and 0.5 to 1 millimetre), phosphorus released was proportional to grain size. Furthermore, the ammonium-exchanged zeolite was more effective than the untreated zeolite in increasing the solubility of the phosphate rock: The amount of phosphorus released from the exchanged samples was approximately twice that from the untreated samples. The increased effectiveness of ammonium-saturated zeolite is probably due to the fact that roughly 30 per cent of the exchange sites in the untreated zeolite are initially occupied by Ca^{2+}. Saturating the zeolite with ammonium ions removes this native Ca^{2+}, thereby increasing its capacity to adsorb calcium from the system.

In the second experiment, the relationship between the grain size of the zeolite and the rate of phosphate rock dissolution was studied further in order to determine the most effective size. The procedure was similar to the previous experiment except that only ammonium-saturated zeolites were used, and two additional size fractions (1 to 2 millimetres, 2 to 4 millimetres) were included. As expected, the amount of phosphorus released increased with grain size, ranging from 2.82 ppm to 6.45 ppm. For grain sizes greater than 0.5 millimetre, however, this increase became negligible.

The third experiment compared the solubility of three phosphate rocks, VM89-3, VM89-4 and VM89-8, in the presence of zeolite over time. The procedure was similar to the previous study except that only one size fraction (0.5 to 1.0 millimetre) of zeolite was used and samples were shaken for time intervals ranging from 1 to 20 days. Control samples of VM89-8 without zeolite were shaken for each of the time periods.

The results of the third experiment are illustrated in Figure 5-2-2. After a slight decrease between Day 1 and Day 2, the amount of phosphorus released increased with time for each of the phosphate rocks, as expected. This rate of increase was most linear for VM89-8. After 20 days, the amount of phosphorus released was greatest for VM89-4 followed by VM89-3, VM89-8, and finally, VM89-8 without zeolite. A comparison of VM89-8 with and without ammonium-saturated zeolite reveals the positive effect the presence of zeolite has on phosphate rock solubility.

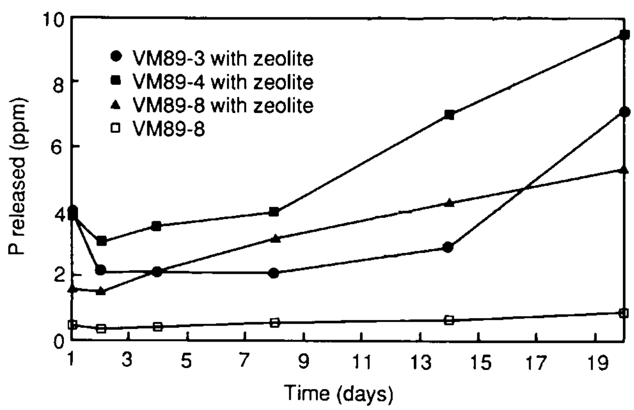

Figure 5-2-2. Plot of phosphorus released versus time for VM89-3, VM89-4 and VM89-8 with zeolite and VM89-8 without zeolite, as determined in experiment 3.

ECONOMIC CONSIDERATIONS

The basal phosphorite of the Fernie Formation is potentially significant to the Canadian fertilizer and agricultural industries. It has an estimated resource potential of 400 million tonnes of phosphate rock with an average P_2O_5 content of 20 per cent (Butrenchuk, 1987). Of this amount, the Cabin Creek deposit contains approximately 34 million tonnes and the Crow deposit has greater than 2 million tonnes.

Western Canada imports 2 million tonnes of phosphate rock annually from the United States at a cost of 70 to 85 Canadian dollars per tonne, including transportation costs

(MacDonald, 1988). Although phosphate rock is currently in oversupply, the availability of rock for export from the United States is expected to decrease by the end of the century. This prediction is based on a projected increase in demand and a depletion of the higher grade Florida deposits. In the near future, Canada will likely have to seek alternative sources for its phosphate rock. A Canadian source would not only lead toward national self-sufficiency, but may also result in lower fertilizer prices due to reduced transportation costs for raw materials.

The use of finely ground phosphate rock as a direct source of phosphorus for crops presents another potential market for the British Columbia deposits. As processing requirements are minimal, a wider range of P_2O_5 contents are acceptable. The viability of this market is increased in western Canada by its close proximity to suitable zeolite deposits, and the growing popularity of alternative agricultural practices (*i.e.* low-input and organic farming).

CONCLUSIONS

Direct application of ground phosphate rock to soils may provide an alternative to chemical fertilizers and a market for the basal Fernie phosphorite. While each of the deposits studied may be suitable for direct application, the Cabin Creek deposit (VM89-8) appears to have the greatest potential: calcite and dolomite are absent in the x-ray diffraction analysis, the ratio of CaO/P_2O_5 of 1.34 is relatively low, and its reactivity is relatively high (1.17 per cent direct available P_2O_5). Furthermore, Cabin Creek is a deposit of substantial size with an estimated 34 million tonnes of phosphate rock averaging 20 per cent P_2O_5. Based on the results of the third dissolution experiment, the Crow deposit (VM89-3) may also be well suited for direct application.

The potential of the Fernie phosphorite for direct application is increased by the occurrence of zeolites in the nearby Princeton basin. Initial dissolution experiments indicate that the addition of a Princeton zeolite to a system containing phosphate rock from the Fernie basin increased the solubility of the phosphate rock. This effect was maximized by using a size fraction of 0.5 to 1.0 millimetre for the zeolite, and by presaturating the exchange sites with ammonium ion.

Future research will involve additional controlled laboratory studies to better understand the phosphate rock–zeolite system and to maximize its performance. Greenhouse experiments will be conducted to test its long and short-term effectiveness.

ACKNOWLEDGMENTS

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FOUR MARL DEPOSITS WITHIN THE SKEENA RIVER DRAINAGE (103I/16W, 103P/1W, 93M/5E)

By M.L. Malott

KEYWORDS: Industrial minerals, marl, bog lime, Skeena River, Buccaneer of the North, Gee Kid, Robinson Lake, Wilson Kettle.

INTRODUCTION

Marl deposits have been known from the Skeena drainage, in the Terrace-Hazelton region, since the early 1930s. Several deposits were identified at that time as possible sources of lime for neutralizing acidic soils. Although only minor amounts of marl have been mined, these occurrences have received more attention recently. Marl has a potential use in neutralizing acid mine drainage and in the acidic cycles of mill processes.

The four marl deposits discussed in this article, Buccaneer of the North, Gee Kid/Lime Lake, Wilson Kettle and Robinson Lake are all within a few kilometres of the Skeena River between Terrace and Old Hazelton (Figure 5-3-1). The term marl is used in this report to indicate a friable mixture of greater than 40 per cent calcium carbonate together with insoluble detritus and noncarbonate plant material. The colour is usually white or buff, but grey to brown or black shades occur as the organic content increases. Bog lime is an alternate term for marl.

REGIONAL GEOLOGIC SETTING

The Terrace to Hazelton area lies within the Stikine Terrane, a component of the Intermontane Belt. Lower Jurassic volcanics outcrop over a large area to the south of the marl deposits. In the Bowser basin to the north, Jurassic and Lower Cretaceous siltstones, argillites and greywackes underlie a veneer of Quaternary Fraser glacial land forms which host the marl deposits discussed in this article. The Fraser glaciation, beginning 25 000 to 30 000 years ago, was the most recent glacial advance (Clague, 1984). During this time, coalescent piedmont glaciers moved southwestward through the study region. Commencing approximately 15 000 years ago, as the glacial age waned, the coalescent ice mass downwasted forming several separate glaciers within the Skeena Valley. These isolated glaciers gradually lost mobility and stagnated, leaving behind fluvioglacial deposits and glacial landforms such as kettle depressions and kame terraces as they melted.

BUCCANEER OF THE NORTH MARL DEPOSIT (MINFILE 103I 001)

A marl deposit known as Buccaneer of the North (or Bockner of the North), occurs 46 kilometres north-northeast of Terrace and approximately 1 kilometre west of the Canadian National Railway Ritchie siding on the west side of the Skeena River (Figure 5-3-1). Access is by crossing the river from a sand bench beyond the north edge of the Kwa-tsa-lix Indian Reserve 4 at Klootch Canyon. A trail on the west bank of the Skeena leads 1 kilometre to the deposit. Alternatively, an old road can be followed southwestward from Ritchie siding.

The property was first staked in 1931 and discussed in the British Columbia Minister of Mines Annual Reports for 1931 (p. A72), 1932 (p. A90) and 1935 (p. C34). Duffel and Souther (1964) report that several railcar loads of marl were shipped to Terrace in 1935 and used for soil dressing.

TOPOGRAPHY AND SURFICIAL GEOLOGY

The marl deposit lies at an elevation of 130 metres in an ephemeral kettle lake within an abandoned channel of the Skeena River (Figure 5-3-2). To the east there is a bench 100 metres wide which gives way to kames that rise gradually 50 metres, separating the lake from the Skeena River channel (Figure 5-3- 3).

The glaciofluvial sedimentary cover which blankets the area consists of gravel and sand, thicker than 10 metres, deposited during the waning stages of the Fraser glaciation (Clague, 1984). Postglacial accumulations of alluvium and colluvium flank the glacial deposits on the north, east and south.

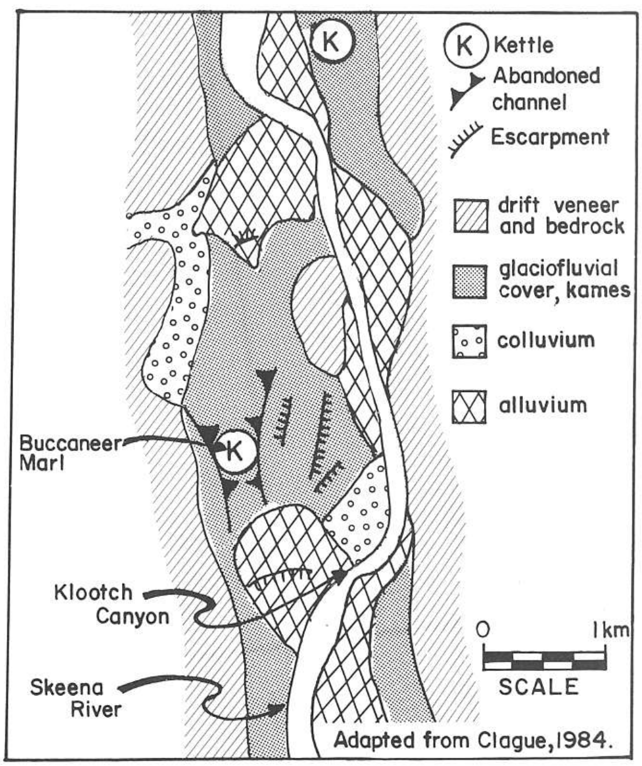

Figure 5-3-1. Geology of the Terrace–Smithers region.

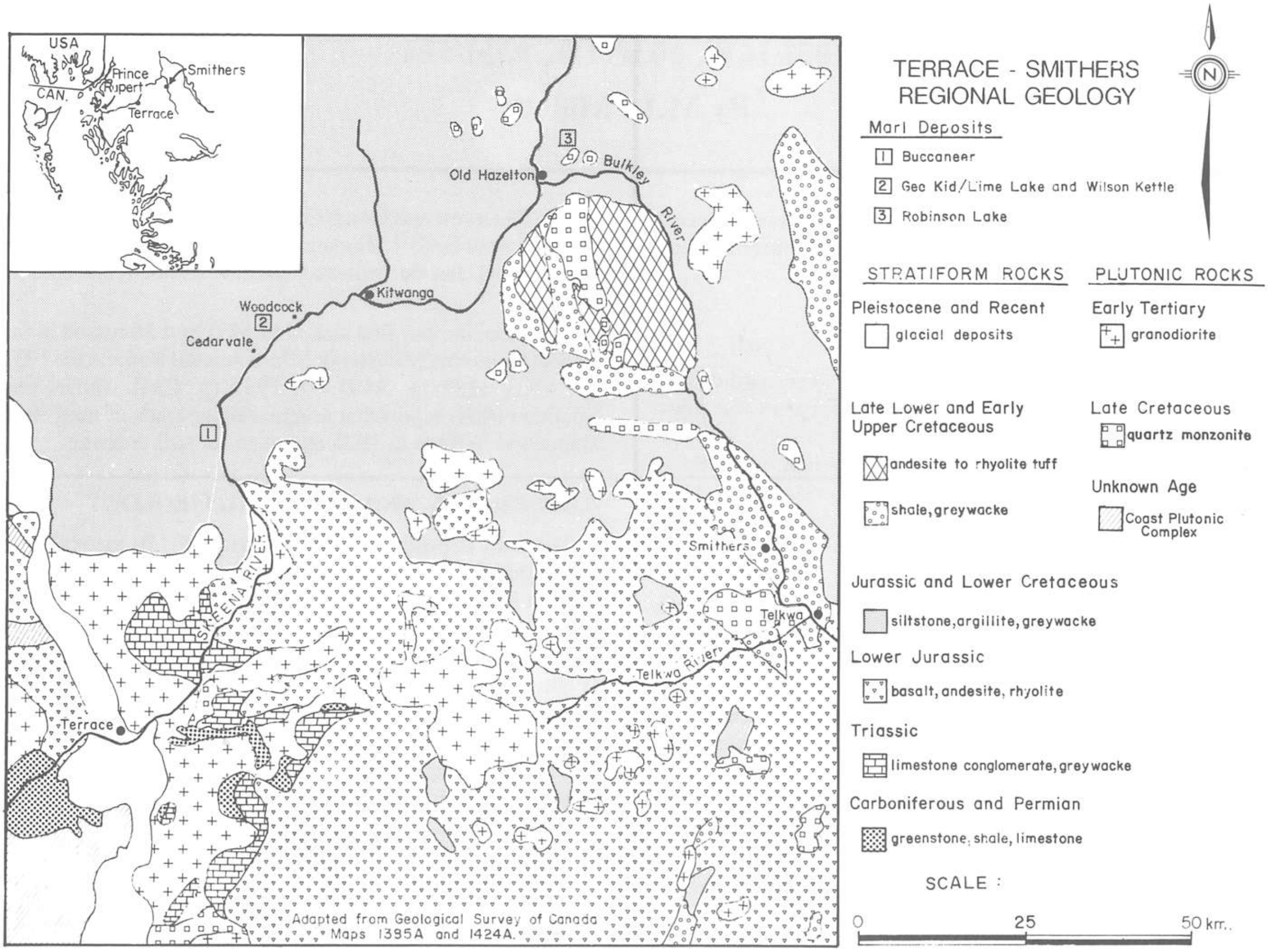

Figure 5-3-2. Surficial geology in the Buccaneer of the North area.

VEGETATION

Aspen and birch fringe the lake amidst a mature stand of lodgepole pine 20 to 30 metres tall. Marsh grasses and horsetails, *Equisetum* sp., fill the intermittent lake site. Nucules from the algal family *Characeae* have been identified (J. White, GSC, personal communication) from the marl and the algae are assumed to be present in the lake.

HYDROLOGY

The deposit lies within a depression that collects runoff locally, and from an intermittent stream (dry at the time of the property visit) which enters in the southwest corner (Figure 5-3-2). An earlier report (B.C. Minister of Mines Annual Report, 1935) indicates that a stream cutting through calcareous argillites to the west of the marl disappears into the glacial debris flanking the meadow. This stream may be the dry creek mentioned above, or may be an additional source of water for the catchment basin which has no known outlet. The water supply appears to have a yearly cycle. In the fall, winter and spring, water collects and then completely evaporates and percolates away through the late spring to early fall.

SIZE

The deposit lies within a depression approximately 110 metres wide by 115 metres long. The marl underlies the depression and extends at least 70 metres beyond it to the south-southeast, beneath a gently dipping bench. The depth of the marl on the bench, in the vicinity of an old dragline cut (Figure 5-3-3), is known to be 1.5 metres for a length of 70 metres. In the 1935 report the marl is described as being 10 metres deep at the eastern end of the bench. The present owner reports that hand drilling has encountered marl to a depth of 9 metres within the depression.

Systematic sampling to define the extent of the marl within the depression and bench has not been attempted. A rough estimate in the 1935 report suggests approximately 65 000 tonnes of wet crude marl could be present in the 90 by 90 by 9 metre bench area based on the marl occupying 0.5 cubic metre per tonne. Using a linear relationship (Figure 5-3-4) between the moisture content of crude marl and tonnes of dry marl per cubic metre, (Macdonald, 1982) an estimate can be made of the dry marl under the bench and depression. A volume of 63 250 cubic metres of marl may exist within the depression assuming dimensions are 110 by 115 by 5 metres.

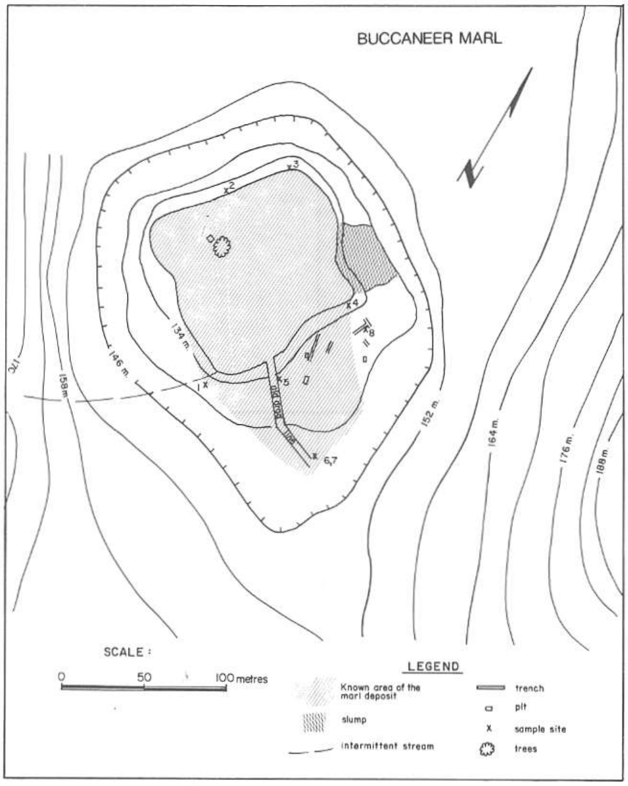

Figure 5-3-3. Topography and deposit configuration of Buccaneer of the North.

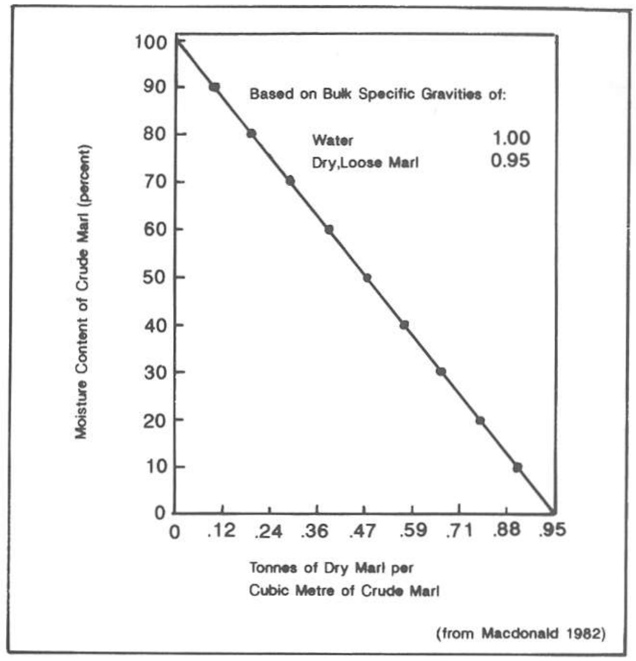

Figure 5-3-4. Estimation of dry marl available in the wet crude.

Assuming a moisture content varying between 30 and 50 per cent then the depression may contain between 30 000 and 41 000 tonnes of dry marl. Again assuming the same variable moisture content, 34 600 to 47 400 tonnes of dry marl may exist on the bench. The deposit may therefore contain between 64 600 and 88 400 tonnes of dry marl in total.

ANALYSIS AND COMPOSITION

Oxide analyses revealed that the samples are all marls with the total CaO content ranging from 39 to 51 per cent and averaging 45 per cent (Table 5-3-1). The $CaCO_3$ equivalence of these values ranges between 70 and 92 per cent with an average of 82 per cent. Contaminants such as silica, aluminum and iron are present in quantities averaging less than 8, 2 and 0.7 per cent, respectively. Other elements, as determined by spectrographic analysis, are present in amounts ranging from 0.5 per cent to traces. (Table 5-3-2).

Dispersal of the marl in a detergent solution indicated the presence of numerous organics. Root fragments and small wood fragments, some charred, are abundant. Fragments of aquatic mosses as well as pelecypod and gastropod shells are more abundant than *Characeae* nucules or the more rare calcareous charophyte axes (J. White, GSC, personal communication, 1989).

GEE KID/LIME LAKE AND WILSON KETTLE MARLS

Several marl localities are situated about 1.5 kilometres northwest of the Skeena River and approximately 6 kilometres northeast of Cedarvale (Figure 5-3-1). One is known in the literature as the Gee Kid, or locally as Lime Lake. The second is a dry lake bed 750 metres to the southwest of Gee Kid and here named the Wilson Kettle marl. Both are accessible by gravel road from Kitwanga. On the north side of Wilson Creek gravel pit travel west 1 kilometre on a logging road. From here Wilson Kettle is approximately 100 metres to the west, in a depression. Continue on the winding logging road approximately 750 metres north and east to the end of the road; Gee Kid is 200 metres farther to the northeast. The Gee Kid marl was staked in 1936 and discussed by Kindle (1937). The presence of marl in Wilson Kettle has not been reported previously.

TOPOGRAPHY AND SURFICIAL GEOLOGY

Gee Kid lies at 340 metres elevation with the Nass Range mountains rising to the north and west. The marl occurrence is situated 1.5 kilometres north of the Skeena River, on a 1000 by 250 metre bench blanketed with glacial drift more than a metre in thickness. Wilson Kettle is at about 325 metres elevation. Measuring approximately 500 by 300 metres, it is a depression, probably a kettle, in the glacial drift covering the area.

VEGETATION

The forest in the vicinity of the Gee Kid occurrence has been clear cut. Mature 30-metre lodgepole pine and spruce with some aspen surround Wilson Kettle which is covered with marsh grasses and has a number of dead spruce near the periphery.

HYDROLOGY AND SIZE

The Gee Kid occurrence collects local runoff in a lake measuring about 230 by 110 metres which drains into the

TABLE 5-3-1
CHEMICAL ANALYSES OF GRAB SAMPLES FROM THREE MARLS*

Field Number	SiO_2	TiO_2	Al_2O_3	Fe_2O_3 (T)	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5	BaO	LOI**	SUMS	$CaCO_3$#	CO_2	S	FeO
	Buccaneer																
1	3.33	0.03	0.61	0.31	<0.01	0.55	49.10	0.33	0.22	0.01	0.02	43.16	97.69	87.64	41.0	0.03	0.44
2	14.50	0.18	3.29	1.27	0.01	0.68	39.09	0.77	0.40	0.06	0.04	37.69	97.96	69.78	34.8	0.03	1.32
3	12.50	0.09	2.67	0.74	0.02	0.78	44.61	0.42	0.18	0.06	—	38.04	100.11	79.61	38.5	0.04	1.61
4	13.57	0.18	3.32	1.49	0.02	0.75	39.92	0.78	1.17	0.27	0.04	37.27	98.78	71.25	35.2	0.04	1.24
5	7.61	0.09	1.83	0.78	0.01	0.40	45.54	0.50	0.30	0.04	0.03	41.09	98.22	81.28	37.7	0.03	1.10
6	3.71	0.03	0.78	0.14	0.02	0.97	51.39	0.10	0.04	0.03	—	42.62	99.83	91.73	41.7	0.03	0.73
7	2.70	0.02	0.46	0.28	<0.01	0.50	49.92	0.32	0.17	0.01	0.02	42.13	96.54	89.11	40.1	0.02	0.28
8	5.60	0.07	1.33	0.61	0.01	0.40	47.47	0.43	0.25	0.02	0.03	41.58	97.79	84.73	38.8	0.02	1.06
Average	7.94	0.09	1.79	0.70	0.01	0.63	45.88	0.55	0.34	0.06	0.03	40.45	—	81.89	38.48	0.03	0.97
	Wilson Kettle																
9	8.12	0.05	1.80	0.48	0.03	0.86	42.68	0.32	0.11	0.05	—	45.34	99.84	76.18	40.6	0.38	6.05
	Robinson Lake																
10	63.95	0.85	16.40	6.98	0.06	1.71	0.99	2.22	1.41	0.02	—	5.28	99.87	1.77	1.96	0.06	4.06
11	46.37	0.63	13.79	4.71	0.07	0.99	2.34	1.52	0.90	0.18	—	28.45	99.95	4.18	29.0	0.38	—

Note: Samples 1-9 are an unconsolidated, nongritty, crumbly paste containing organics (detrital moss, root and algal fibre). Field descriptions are:
Sample 1 – light grey, slightly moist, 10% organics
Sample 2 – mottled light and medium green-grey, slightly moist, 10% organics
Sample 3 – mottled light and medium green-grey, slightly moist, 10% organics
Sample 4 – mottled light and medium green-grey, very moist, 10% organics
Sample 5 – light grey, very slightly moist, 5% organics
Sample 6 – white with light grey layers, dry, 15% organics
Sample 7 – light grey, dry, 10% organics
Sample 8 – light grey, dry, 15% organics
Sample 9 – light grey-brown, moist paste, 20% organics
Sample 10 – dark grey, clay rich, gritty, very moist paste
Sample 11 – black, clay-rich, gritty, very moist paste
* all values are percentages
(T) Total Iron
** LOI values ($\pm$2 per cent absolute) are elevated due to the high calcium carbonate content and predominantly represent volatiles such as CO_2, H_2O, Fl and Cl.
# Calcium Carbonate Equivalence

TABLE 5-3-2
SPECTROGRAPHIC ANALYSES OF GRAB SAMPLES FROM THREE MARLS*

Sample	Si	Al	Mg	Ca	Fe	Pb	Cu	Zn	Mn	Ag	V	Ti	Ni	Co	Na	K	W	Sr	Ba	Cr	Ga	Zr
Buccaneer																						
3	7.0	5.5	0.7	>10	1.0	—	Tr	Tr	0.1	—	—	0.06	—	—	0.5	<0.3	—	Tr	Tr	Tr	—	—
6	1.5	2.0	0.4	>10	0.4	—	Tr	Tr	0.06	—	—	0.03	—	—	0.3	<0.3	—	Tr	Tr	—	—	—
Wilson Kettle																						
9	>10	2.0	0.3	>10	0.6	—	Tr	—	0.09	Tr	—	Tr	—	—	0.3	<0.3	—	Tr	Tr	—	—	—
Robinson Lake																						
10	>10	4.0	0.6	0.25	2.7	—	Tr	—	0.1	Tr	—	0.06	—	—	1.0	<0.3	—	Tr	Tr	—	—	—
11	>10	>10	1.6	1.2	7.0	—	Tr	—	0.12	Tr	.02	0.3	Tr	Tr	>2	>2	—	Tr	0.1	—	Tr	Tr

* all values are percentages

Skeena River through an outlet on the southeast side. A beaver dam crossing the northeast section of the lake has raised the water level.

Wilson Kettle is oval in shape, measures approximately 170 by 100 metres and is bisected by a small stream entering on the north and draining to the south toward the Skeena River. It is not known if the lake bed is now permanently dry or whether it floods periodically.

In the time intervening since the field visit, local residents reported that a small beaver dam at the southern end of the area gave way and exposed approximately half a metre of white to light grey marl. This indicates that the marl deposit may extend the 170-metre length of the kettle, but its depth is unknown.

Sampling and Analysis

With the extensive flooding caused by the beaver dam across Lime Lake, the water level has risen to such an extent that sampling can only be done by boat. A boat was not available at the time of the field visit and repeated attempts to sample the present lake shore encountered only gravel.

However, the presence of a high calcium carbonate content in the water is suggested by the abundance of several species of freshwater snails along the present shoreline.

Wilson Kettle was sampled with a shovel in the middle of the northeast end. At this location approximately 50 centimetres of dark brown to black, wet peat overlies a wet light grey to brown marl with abundant organics. The oxide analyses of this sample indicate a 43 per cent CaO content (or 76 per cent $CaCO_3$ equivalence) together with 8 per cent SiO_2 and 2 per cent Al_2O_3 contaminants (Table 5-3-1).

ROBINSON LAKE (MINFILE 93M 103)

Ten kilometres northeast of Old Hazelton a marl deposit is reported in Robinson Lake (Kindle, 1954; Figure 5-3-1). Access is by the Silver Standard/Nine Mile Mountain gravel road which begins 3.2 kilometres north of the Hagwilget bridge. At a point 8.5 kilometres along the Silver Standard road an old logging road provides access on foot to the northeast corner of the lake, 200 metres to the north.

TOPOGRAPHY AND SURFICIAL GEOLOGY

The lake is at 470 metres elevation on the eastern margin of the Skeena River valley. Till greater than 1 metre thick blankets the area and the lake occupies a depression 630 metres long, with an average width of 120 metres.

VEGETATION

Mature 20 to 30-metre spruce surround the lake, except for the northwest section which has been clearcut. Small patches of birch and aspen together with a stand of 30 to 40-metre hemlock and cedar border the northeast shore. Marsh grasses are abundant along the edge of the lake. As well as the grasses, *Sphagnum* sp. moss, Labrador tea, and especially horsetails, *Equisetum* sp., are abundant in the wet northern portion.

HYDROLOGY

Robinson Lake collects local runoff and is drained to the southwest by Two Mile Creek from which the town of Hazelton draws its water. Beaver dams across the northeast and the southwest ends of the lake have considerably raised the water level and created an extensive marshy area to the northeast.

SIZE

Kindle (1954) reports that the shallower parts of the lake are bottomed by marl, particularly in the south where it is at least 4 metres deep. At present the beaver dams have raised the water level to such an extent that the dimensions of the deposit can only be determined by sampling from a boat.

SAMPLING AND ANALYSIS

Attempts to sample the marl along the southeast shoreline encountered a dark grey, clay-rich, gritty paste. Analysis shows about 1 per cent CaO with SiO_2, Al_2O_3, and Fe_2O_3 being the principal components (Table 5-3-1, Sample 10). Another sample was taken in the northeast, above the beaver dam, where a number of mounds of light grey sediment dot the lake bottom. These mounds appear to be deeper sediment, possibly marl, brought to the surface by gases, noticeably rich in hydrogen sulphide. The gas is probably produced from decaying organics deeper in the sediments. Penetrating only 35 centimetres with a shovel, not enough to reach the light grey material, obtained a sample of black, slightly gritty, clay-rich paste. (Table 5-3-1, Sample 11). Although CaO is present, SiO_2, Al_2O_3 and Fe_2O_3 are the major constituents (Table 5-3-2).

DISCUSSION

MARL FORMATION

Marl deposits usually have four features in common: they are associated with present or past groundwater discharge areas; they are located in topographically low, poorly drained sites; they have highly permeable recharge areas nearby and they are fed by groundwater with high Ca^{++} and $(HCO_3)^-$ion concentrations (Macdonald, 1982). A study by Thiel (1930) revealed that high, irregular morainal topography favours marl accumulation. He found the greatest number of large marl deposits located in coarse outwash sands and gravels.

The marl deposits in the Skeena drainage are situated in depressions on permeable glaciofluvial landforms which facilitate groundwater discharge. Adjacent areas of high relief allow rapid recharge within the water cycle and contain permeable sediments rich in calcium carbonate.

A number of theories have been put forward to explain the formation of marl in lakes. Generally they fall into two broad categories: physiochemical processes such as carbon dioxide degassing and thermal stratification, or biologically linked processes, such as calcium carbonate precipitation by the algal family *Characeae* and by blue-green algae, or through accumulation of invertebrate remains. Duston *et al.* (1986) concluded that calcium carbonate precipitation is probably the result of a complex interaction between these physiochemical and biological processes. Physiochemical factors certainly play a role in the formation of the deposits discussed here and biological processes also appear to be active. *Characeae* are known to be present and gastropods have been found associated with at least one deposit.

DEPOSIT CLASSIFICATION

Marl deposits can develop through a series of stages, as depicted in Figure 5-3-5. The Buccaneer deposit typifies the gradual in-filling of a kettle depression since the last glacial retreat approximately 10 000 years ago.

A seepage-ponded classification (MacDonald, 1982) best characterizes the Buccaneer deposit. The kettle depression collects local runoff, with the water level fluctuating according to annual precipitation and evaporation cycles. The precipitation of calcium carbonate is facilitated by the hummocky terrain with, presumably, short groundwater-flow systems. Groundwater flows through permeable and reportedly calcium carbonate rich sedimentary rocks and permeable surficial sediments. In conjunction with these physical features, the presence of the algae *Characeae* and the phys-

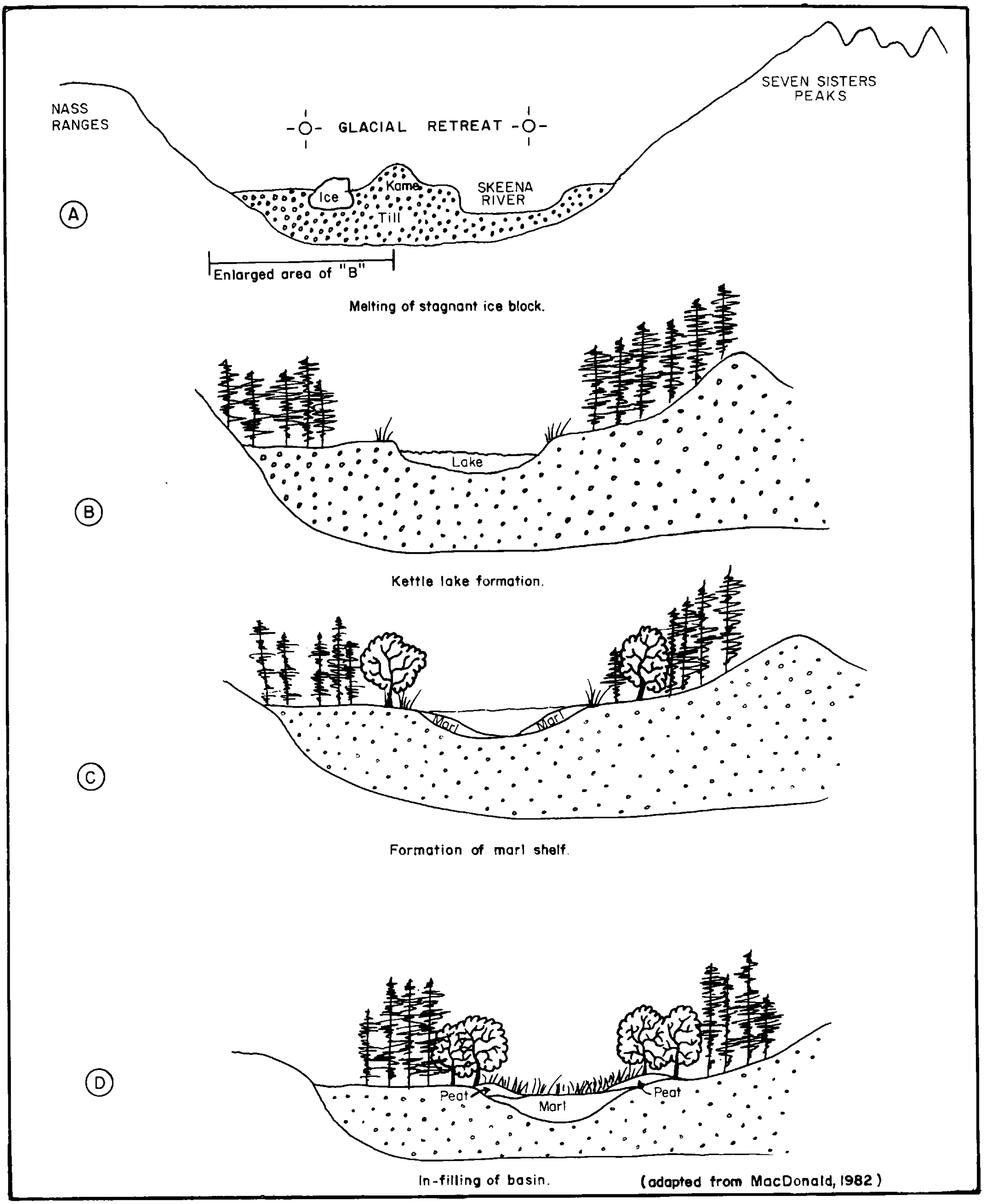

Figure 5-3-5. Development stages of a marl lake.

iochemical factors of thermal stratification and carbon dioxide degassing probably cause the precipitation of calcium carbonate. Grasses cover the ephemeral lake and the locality is presumably in the late stages of the development of a marl deposit.

Gee Kid and Wilson Kettle marls are also best described as seepage-ponded deposits. Both sites appear to be kettle depressions which collect water and favour the deposition of marl. Precipitation of marl may be on-going in Lime Lake due to the raised water level. Wilson Kettle appears to be only intermittently wet, if flooded at all, and is covered by grasses. It is probably in the later stages of marl development (Figure 5-3-5).

The Robinson Lake marl may be a shoreline-fringe deposit, as classified by Macdonald (1982). Adjacent to a high-relief area to the north and east, the lake collects local groundwater flow and calcium carbonate is precipitated in shallow water by thermal stratification, probably aided by the biological carbonate-fixing ability of *Characeae*.

SUMMARY

Of the four marl deposits studied, two, Buccaneer of the North and Wilson Kettle, contain good quality marl. The other two, Gee Kid and Robinson Lake reportedly contain marl, but are presently flooded and need further study.

The Buccaneer is estimated to contain as much as 88 400 tonnes of dry marl averaging 82 per cent calcium carbonate with a low percentage of contaminants. After removal of the thin vegetative cover, the marl could readily be scooped out and trucked to the railway only a kilometre away.

The Wilson Kettle deposit contains marl of good quality with only a small percentage of contaminants, bnt further sampling is needed to determine its dimensions and overall quality. A maintained all-weather road passes within a kilometre of the deposit.

The extent and quality of the Gee Kid and Robinson Lake deposits was not ascertained because of extensive flooding, but exploitation of the Robinson Lake marl would be difficult as the town of Hazelton draws its water from the lake.

In summary, the marl in the Terrace-Hazelton area is of good quality, adjacent to transportation, readily removable, and at least one site has significant estimated reserves.

ACKNOWLEDGMENTS

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NOTES

Applied Geochemistry

1989 REGIONAL GEOCHEMICAL SURVEY, SOUTHERN VANCOUVER ISLAND AND LOWER MAINLAND* (92B, 92C, 92F & 92G)

By John L. Gravel, Wayne Jackaman and Paul F. Matysek

KEYWORDS: Regional Geochemical Survey, reconnaissance stream sampling, moss-mat sediment, stream sediment, stream water, Victoria, Cape Flattery, Alberni, Vancouver.

INTRODUCTION

Three Regional Geochemical Surveys covering southern Vancouver Island and Lower Mainland were conducted by the Geological Survey Branch in 1989 (Figure 6-1-1) Map sheets covered are: Victoria (92B), Cape Flattery (92C), Alberni (92F) and Vancouver (92G). Approximately 2430 paired sediment and water samples were collected over 25 000 square kilometres at an average density of one site every 10.5 square kilometres. As in 1988, moss-mat sediment samples were obtained on Vancouver Island (1397 sites), stream sediments were collected on the mainland (1033). The following report highlights mineral potential and survey parameters of the 1989 RGS program and a brief description of orientation studies conducted in preparation for the 1990 program.

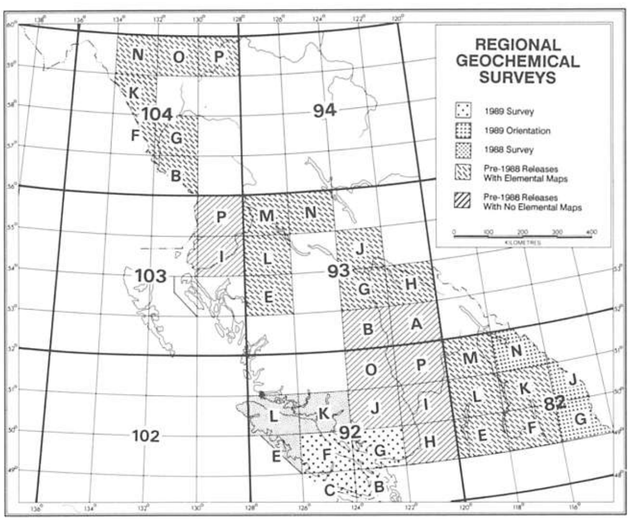

Figure 6-1-1. Current status of British Columbia Regional Geochemical Survey program.

HIGHLIGHTS

Some of the oldest mining camps in British Columbia are located on southern Vancouver Island and the Lower Mainland, for example, the Brittania and Tyee-Lenora massive sulphide deposits. The high number of past and present producers, volume of exploration activity and discovery of numerous new prospects attest to the high potential of the area which, until this year, lacked systematic geochemical coverage. Innovations introduced in 1988 (Gravel and Matysek, 1989) proved highly successful and were incorporated in the 1989 program. Potentially, moss-mat sediment sampling on southern Vancouver Island may permit merging of 1988 and 1989 surveys resulting in a contiguous 2800-site database covering all the major lithologies. New developments in 1989 include:

- Establishing a three-year analytical contract ensuring greater continuity in sample results;
- Preparation of three stream-sediment standards for monitoring quality and continuity between surveys; and
- Lowering the detection limit for silver to 0.01 ppm, thereby improving its usefulness as a pathfinder.

SURVEY AREA FEATURES

GEOLOGY AND PHYSIOGRAPHY

Understanding local geology and physiography is essential for informed interpretation of RGS results. A geological underlay (Roddick *et al.*, 1976) on element plots and surficial geology maps (Fulton *et al.*, 1982) are provided within RGS packages for this purpose. An abridged description of geology and physiography based on studies by Holland (1964), Muller (1977) and Massey and Friday (1988, 1989) is given in Table 6-1-1 and Figure 6-1-2.

Southern Vancouver Island is underlain by Insular Belt rocks comprising mid-Paleozoic to mid-Tertiary volcanics and sediments representing island arc and back-arc basin sequences. Accreted onto the southwestern margin is a sliver of trench and slope volcanics and sediments belonging to the Pacific Belt. Intermediate to felsic plutons intruded Insular rocks during the Jurassic and Tertiary periods. Tertiary intrusions in the Pacific Belt are of mafic composition. The physiography of southern Vancouver Island is dominated by the Vancouver Island Ranges with the Nanaimo Lowlands bordering the southern and eastern coast.

Granitic to dioritic intrusions of Jurassic to Eocene age containing elongated pendants of Triassic to Cretaceous metavolcanics and metasediments constitute the Coast plutonic complex underlying most of the Mainland. The Coast complex is represented physiographically by the Coast Mountains and the Georgia Lowlands. Tertiary sediments form the Fraser Lowlands.

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

TABLE 6-1-1 Abridged Description of Physiography and Geology of 1989 RGS Project Area

PHYSIOGRAPHY

AREA	UNIT	SUBUNIT	LOCATION	QUATERNARY GEOLOGY	DESCRIPTION
VANCOUVER ISLAND	Coastal - Trough	Nanaimo Lowland	East coastal margin and southern tip	Thin till and colluvium on low hills, thick glaciofluvial deposits along coastal margin	Rolling hills below 600 m a.s.l., ridges separated by narrow valleys, box-like canyons in incised overburden along coast
		Vancouver Island Ranges	NW trending range forming core of island	Colluvium on steep slopes and summits, till on lower slopes (glacio-)fluvial deposits in valleys	Very rugged, U-Shaped valleys, 2200 m peaks near Buttle Lake, summit elevations decrease to SE, dissected Tertiary surface
		Estevan Coastal Plain	3 km wide strip along west coast	Mantle of bedrock derived colluvium, (glacio-)fluvial sediments, till, marine deposits along coast	Flat and featureless to low rolling hills, rock cliffs and platforms, pocket beaches
	Insular Mountains	Fiordland	Peninsulas and islands along west coast	Colluvium on steep slopes and summits, till on lower slopes	Land rises abruptly to 600 - 900 m a.s.l., rounded and timbered hilltops
		Alberni Basin	8 x 40 km basin around Port Alberni	Moderate to thick till and (glacio-)fluvial deposits, some marine sediments	low lying (below 300 m a.s.l.) and low relief, fault bounded on east side by Beaufort Range
MAINLAND	Coastal Trough	Georgia Lowland	3 - 15 km strip along west coast	Thin colluvium and till on summits and slopes, broad U-shaped valleys with thick fluvial deposits	Tertiary erosional surface rising gently to the east with increasing dissection, 600 m contour marks divide with Coast Mountains
		Fraser Lowland	South of Coast Mountains, forms Fraser valley	Thick glacio-fluvial, fluvial and marine deposits	Depositional environment since mid-Tertiary (4500 m thick), wide flat valleys separate low (<300 m) hills cored by overburden or bedrock
	Coast Mountains	Pacific Ranges	Adjacent to Georgia Lowland 125-160 km wide belt	Barren summits, colluvium and till slopes, thick (glacio-) fluvial deposits in valleys	Incision of Tertiary surface increases eastward until completely eroded leaving jagged peaks and U-shaped valleys

GEOLOGY

VANCOUVER ISLAND

Stratified rocks

Name	Age	Description
Metchosin Group	Eocene	Pillow basalt
Nanaimo Group	Upper Cretaceous	Epiclastic sandstone, shale, conglomerate
Leech River Group	Jurassic and Cretaceous	Metasediments and metavolcanics, greywacke, phyllite, schist
Westcoast Gneiss Complex	Jurassic	Gneiss, metaquartzite, marble, agmatite, amphibolite
Bonanza Group	Lower Jurassic	Andesitic to rhyolitic lavas, tuff breccia
Vancouver Group		
Parsons Bay Formation	Upper Triassic	Calcareous siltstone, shale, limestone, greywacke, conglomerate, breccia
Quatsino Formation	Upper Triassic	Limestone, marble
Karmutsen Formation	Upper Triassic	Basaltic pillow lava, breccia, minor limestone
Sicker Group (including) (Buttle Lake) (Group)	Devonian to Pennsylvanian	Lower (Sicker) contains basaltic to rhyolitic volcanics. Upper (Buttle) composed of clastic and carbonate sediments

VANCOUVER ISLAND

Plutonic Rocks

Name	Age	Description
Sooke plutonic suite	Tertiary	Augite gabbro
Tertiary intrusions	Early to mid-Tertiary	Quartz diorite to quartz monzonite, includes Catface plutonic suite
Island plutonic suite	Middle Jurassic	Granite, granodiorite, quartz monzonite, quartz diorite, gneissic diorite

MAINLAND

Stratified Rocks

Name	Age	Description
Gambier Group	Upper Cretaceous	Greenstone, volcanic breccia argillite, minor conglomerate, limestone and schist; includes Fire Lake Group
Vancouver Group		
Karmutsen Formation	Upper Triassic	Basaltic pillow lava, breccia and minor limestone
Metamorphics	Paleozoic to Triassic	Amphibolite, schist, quartzite, minor crystalline limestone, greenstone

Plutonic Rocks

Name	Age	Description
West Coast Complex	Cretaceous to early Tertiary	Quartz monzonite, granodiorite, quartz diorite, diorite, gabbro

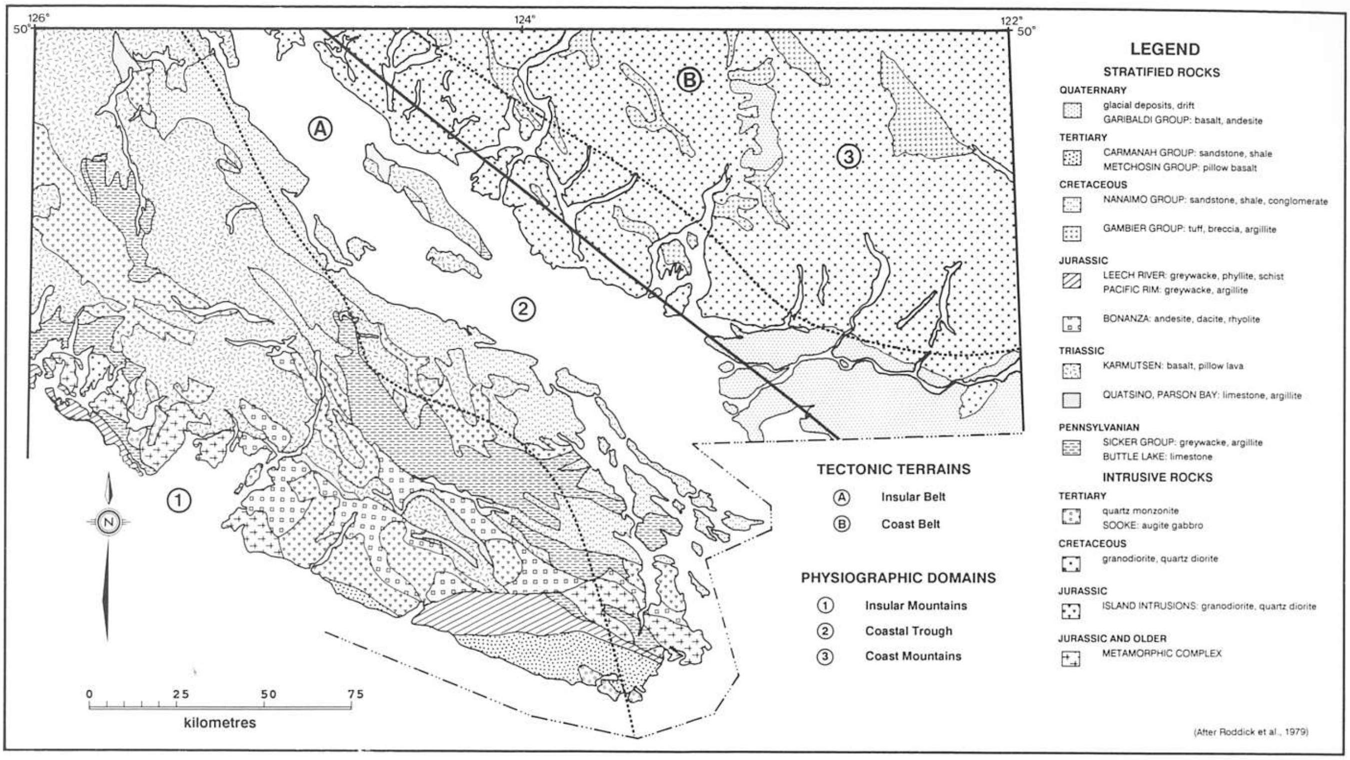

Figure 6-1-2. Geology and physiography of 1989 Regional Geochemical Survey area.

MINERALIZATION AND EXPLORATION POTENTIAL

The style of mineralization found in the 1990 RGS- release area can be divided into four broad categories: massive sulphides, porphyries, quartz veins and skarns. Table 6-1-2, based on a compilation of work by Muller and Carson (1969) and numerous property studies by ministry geologists, subcategorizes these deposits giving type examples, commodities, hostrock and related intrusions.

Mineral potential is very high as attested by 21 past producers including Brittania, a world class massive sulphide deposit of 47 million tonnes grading 1.1 per cent copper, 0.3 per cent zinc, 0.03 per cent lead, 3.4 grams silver and 0.3 gram gold per tonne. A study of local mining history from the turn of the century to the present reveals two periods of activity separated by the Great Depression in the 1930s. Spurred by World War II and the rebuilding of industrial nations during the 1950s and 1960s, local mining reached its peak. At times upwards of eight mines were in operation. Mining activity has dropped sharply since 1975 where today, only the massive sulphide deposit at Buttle Lake is in operation. There are however, 17 potential producers located in the program area. Notable deposits in this class include Catface and Gambier Island, both copper ± molybdenum porphyries in the 200 million tonne range and Lara, a copper-lead-zinc massive sulphide prospect with approximately 500 000 tonnes of indicated reserves.

Exploration trends, based on assessment reports filed with the Ministry of Energy, Mines and Petroleum Resources, show a marked increase in activity over the 5-year period examined (Table 6-1-3). From 1984 to 1988 the number of reports increased by 75 per cent from 96 to 168 while the nature of the work matured. Eighty per cent of work completed in 1984 was grassroots exploration compared to 40 per cent in 1988. The prime mineral target also changed during this period, from near equal concentration on massive sulphides (40 per cent) and precious metal quartz-vein systems (50 per cent) in 1984, to predominantly the latter (20 per cent *versus* 65 per cent) in 1988. The pattern of exploration however, remained fairly constant as Sicker Group rocks are favourable hosts for either deposit type. Frequently the

TABLE 6-1-2 Common Deposit Types and Associated Geological Environment

DEPOSIT CLASS	DEPOSIT SUBCLASS	TYPE DEPOSIT	COMMODITIES	HOST ROCK	INTRUSIONS
Massive Sulphide	Volcanogenic Massive Sulphide	Buttle Lake, Brittania	Pb, Zn, Cu (Au)	Felsic flows in upper Sicker Group, Gambier and Fire Lake Group pendants in Coast Complex, Harrison Lake Formation	None
	Magmatic Massive Sulphide	Tofino Nickel	Ni, Fe, Cu (Pt, Pd)	Karmutsen Formation	Feeder zone for Karmutsen volcanics
Porphyry	Cu±Mo Porphyry	Gambier Is., Catface	Cu, Mo		Tertiary "Catface" quartz diorite
	Cu±Mo±Au Stockwork	Mount Washington	Cu, Au, Mo	Karmutsen Formation, Sicker Group,	Early tertiary quartz diorite
Quartz Veins	Epi & meso-thermal Au-qtz veins	Mount Washington, Ashloo	Au, Ag, Cu, As, Sb, Au, Ag	Bonanza Group, Sicker Group Karmutsen Formation, Intrusives	Early Tertiary quartz diorite
	Gold in qtz shears	Debbie	Au, Cu, Ag	Sicker Group, Bonanza Group Leech River Group; along Tertiary shear zones	
	Gold in sulphide veins	Thistle	Au, Cu	Upper Duck Lake Formation in Sicker Group; along Tertiary shear zones	
Skarns	Iron-gold skarns	Yellow Kid	Fe (Cu, Au)	Karmutsen Formation, Quatsino Formation	Jurassic Island plutonic suite
	Copper skarns	Blue Grouse	Cu (Ag, Au)	Bonanza Group	Island Intrusions Flds-por dikes
	Basemetal skarns	Cambrian-Chieftain	Cu, Zn, Ag (Au, Mo)	Limestone member Gambier Group	Quartz diorite of Coast plutonic complex

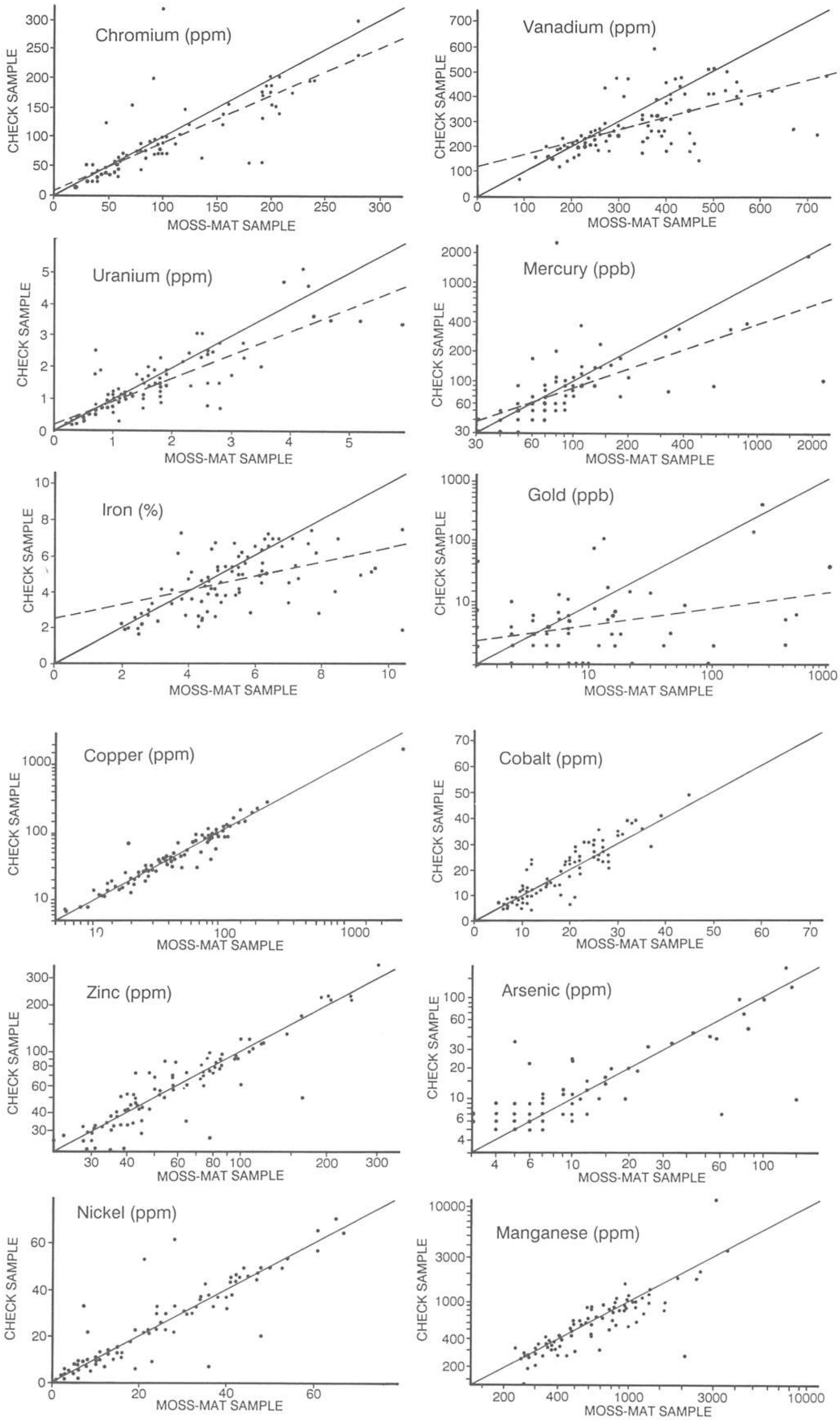

Figure 6-1-3. Scatter diagrams comparing concentrations of elements dispersed hydromorphically (copper, zinc, nickel, cobalt, arsenic and manganese) and elements dispersed as heavy minerals (chromium, uranium, iron, vanadium, mercury and gold) in 96 paired stream-sediment check samples and moss-mat sediment samples. Solid sloped lines are the unity lines, points plotting along them are sample pairs having equal elemental concentrations. Sloped dashed lines are least squares regression lines (After Matysek *et al.*, 1989.)

TABLE 6-1-3
Five-Year Compilation of Exploration Work in 1989 RGS Area Based on Assessment Reports

NTS MAP SHEET	1984 TYPE	a	b	c	d	1985 TYPE	a	b	c	d	1986 TYPE	a	b	c	d	1987 TYPE	a	b	c	d	1988 TYPE	a	b	c	d
92B	11MS	7	3	1		8MS	3	3	2		15MS	6	4	5		10MS	5	3	2		11MS	2	5	3	1
	3QV	2		1		6QV	5	1			5QV	3	2			19QV	14	2	3		15QV	9	2	3	1
						1SK		1			1SK		1												
92C	2MS	2				5MS	1	3	1		3MS	3				3MS	3				8MS	2	6		
	2PO	2				2PO	1	1			1PO			1		1PO	1				13QV	7	3	3	
	10QV	9	1			11QV	9	1	1		12QV	5	6	1		10QV	6		4		4SK		1	3	
	2SK	2				1SK		1			7SK	6	1			2SK	2								
92F	15MS	11	1	3		19MS	10	6	3		14MS	10	4			11MS	7	3	1		9MS	4	2	3	
	26QV	22	3	1		1PO	1				44QV	33	6	5		54QV	30	10	14		70QV	20	23	27	
	8SK	7	1			32QV	19	6	6	1	14SK	7	2	5		14SK	10	3	1		17SK	7	7	3	
						10SK	9	1																	
92G	9MS	9				5MS	4		1		8MS	3	3	2		6MS	3	2	1		6MS	2	2	2	
	8QV	7		1		4QV	2		2		2PO	1		1		6QV	3	1	2		12QV	7	3	2	
						1SK		1			9QV	6	1	2							3SK	3			
TOTAL	96	80	9	7		106	64	25	16	1	135	83	30	22		136	84	24	28		168	63	54	49	2

LEGEND

Deposit Type (TYPE)

MS = Massive Sulphide
PO = Cu-Mo Porphyry,
QV = Au-Quartz Veins
SK = Au-Fe-Cu-Pb-Zn Skarn

Exploration Activity

a = Grassroots: prospecting, minor (<$10K) ground surveys
b = Intermediate: mapping, moderate (>$10K) ground surveys
c = Advanced: drilling, trenching, extensive (>$50K) surveys
d = Development: Underground development, surface stripping for open pits, ore definition drilling, etc.

Example:

NTS MAP SHEET	1984 TYPE	a	b	c	d
92B	11MS	7	3	1	

Interpretation:
In 1984 on map sheet 92B, eleven assessment reports were filed for properties with massive sulphide potential, of which seven were at the grassroots stage, three were at the intermediate stage and one was at the advanced stage.

emphasis on deposit type was seen to shift on the same property. Several factors can be sited as having influenced exploration patterns, overriding most of these was flow-through funding of junior resource companies with their main emphasis on gold-bearing deposits. Events over the past 2 years in mineral markets (lower gold prices and higher base metal prices) and federal government policy changes (structure of flow-through funding) could erode the number of claims worked in the near future, with emphasis returning to polymetallic deposits.

Subsequent to the 1990 RGS release, primary exploration targets will undoubtedly be polymetallic massive sulphide deposits and precious metal quartz veins with activity centred upon Sicker Group uplifts and Coast Complex pendants. Secondary targets will be epithermal vein systems related to Tertiary intrusions and various gold-bearing skarns associated with Vancouver Island and Bonanza Group rocks. The following deposit types have received less general attention but are well worth noting and may form the basis of exploration programs extending from searches of the RGS database:

- Tofino Nickel:
  - A magmatic massive sulphide deposit containing copper, nickel and platinum group metals.
  - Believed to be hosted by an ultramafic intrusive phase of the Karmutsen volcanics.
  - May represent feeder zones for flood basalts similar to Russian deposits at Noril'sk.
- Ashlu:
  - Mesothermal precious metal quartz vein systems with tellurides and PGM values.
  - Hosted by granodiorite phase of Coast plutons.

1989 SAMPLING PROGRAM

BACKGROUND

Use of moss mats as a sampling medium in RGS programs was initiated in 1988, solely on northern Vancouver Island. Findings from orientation studies (Matysek and Day, 1988) carried out the previous year showed the bedrock, climate and physiography of Vancouver Island restricted accumulation of fine-grained sediment (−80 mesh or −177 micron fraction) within stream channels. Moss mats were selected as a sampling medium due to their abundance in Vancouver Island stream beds and high concentration of fine-grained sediment.

Results from the 1988 RGS program were highly encouraging. In order to demonstrate the benefits of moss-mat sediment samples, check samples consisting of routine stream sediment were taken at one in every twenty moss-mat sample sites. A total of 96 check samples were collected. In general, moss-mat samples provided several advantages:

- On average, moss mats contain five times the amount of fine-grained sediment found in conventional samples at the same site, reducing the number of samples rejected due to insufficient sediment.
- Samplers find moss mats easier to locate and collect.
- Hydromorphically dispersed elements (copper, nickel, zinc, etc.) display near identical concentrations for the two media types (Figure 6-1-3).
- Elements which commonly form heavy detrital minerals (chromium, gold, etc.) are concentrated in moss mats, with a suggestion that the degree of enrichment is correlative to specific gravity (Figure 6-1-3).
- Reproducibility for elements prone to "nugget" effect, appears to increase for moss-mat samples, however, further testing is required.
- Moss mats suffer less from seasonal variability as they trap sediment only during floods when the mat is inundated, at this stage gold is liberated and mobilized from the various placer traps within the stream bed (W.K. Fletcher, personal communication, 1989).

Speculation on the nature of sediment trapping by moss mats, based on the above results, favours two controlling mechanisms:

(1) Filter trapping whereby fine sediment is caught in the dense growth of plant fibres as water passes through the mat; and

(2) Density sorting, a cyclical process whereby sediment deposited during a waning flood stage is selectively removed during the next waxing stage. The susceptibility of a mineral grain to removal is directly related to its specific gravity.

Given the above conditions, a fine-grained sediment will accumulate in which heavy minerals are concentrated preferentially to light minerals. Experiments are in progress to test this hypothesis.

SAMPLE COLLECTION

MPH Consulting Ltd. was contracted to carry out the 1989 RGS sampling program. The field crew consisted of five samplers and a party chief stationed in a mobile field camp. Ministry representation was maintained throughout the program, providing advice, crew training, site inspections and quality control. Sampling began on July 15 and was completed by September 20.

On Vancouver Island, moss-mat sediments and stream waters were sampled at 1397 sites while 1033 sites were sampled for stream sediment and water on the Lower Mainland (Table 6-1-4). For comparison, stream-sediment check samples (90 sites) were collected on Vancouver Island while moss-mat check samples (40 sites) were collected on the Mainland. Sites were restricted to primary and secondary drainages having catchment basins of less than 10 square kilometres. Streams in provincial parks and large municipalities such as Vancouver, Victoria and Nanaimo were avoided. In total, 2430 sites were sampled in 250 person-days, giving a sampling rate of 9.8 sites per sampler per day. Approximately 26 000 square kilometres were covered at an average density of 1 site every 10.7 square kilometres. Sixty per cent of sites were accessible by truck, trail-bike or boat, the remaining sites were reached by helicopter.

Sites were selected in pre-season by the authors, however, samplers were given flexibility to modify site locations based on accessibility and availability of samplable material.

TABLE 6-1-4 Sample Distribution in 1989 RGS Program Area

92B VICTORIA					
MAP SHEET	MOSS MATS	STREAM SEDS	TOTAL SITES	AREA Km^2	DENSITY $SITES/Km^2$
05	42	0	42	577	13.7
12	69	0	69	880	12.8
13	42	0	42	750	17.9
14	2	0	2	175	87.5
TOTAL	155	0	155	2382	15.4

92C CAPE FLATTERY					
MAP SHEET	MOSS MATS	STREAM SEDS	TOTAL SITES	AREA Km^2	DENSITY $SITES/Km^2$
08	18	0	18	215	11.9
09	113	0	113	976	8.6
10	52	0	52	392	7.5
11	4	0	4	25	6.3
13	4	0	4	43	10.8
14	29	0	29	300	10.3
15	119	0	119	968	8.1
16	105	0	105	963	9.2
TOTAL	444	0	444	3882	8.7

92F NANAIMO					
MAP SHEET	MOSS MATS	STREAM SEDS	TOTAL SITES	AREA Km^2	DENSITY $Km^2/SITE$
01	98	0	98	1003	10.2
02	99	0	99	972	9.8
03	111	0	111	953	8.6
04	50	0	50	523	10.5
05	82	0	82	636	7.8
06	95	0	95	713	7.5
07	42	0	42	808	19.2
08	2	3	5	285	57.0
09	0	18	18	478	26.5
10	8	5	13	383	29.5
11	88	0	88	690	7.8
12	24	0	24	210	8.7
13	49	0	49	452	9.2
14	39	0	39	792	20.3
15		9	9	223	24.8
16		87	87	808	9.3
TOTAL	787	122	909	9929	10.9

92G VANCOUVER					
MAP SHEET	MOSS MATS	STREAM SEDS	TOTAL SITES	AREA Km^2	DENSITY $SITES/Km^2$
01	0	34	34	290	8.5
02	0	1	1	10	10.0
04	11	0	11	388	35.3
05	0	13	13	226	17.4
06	0	52	52	508	9.8
07	0	79	79	903	11.4
08	0	84	84	948	11.3
09	0	90	90	987	11.0
10	0	101	101	1008	10.0
11	0	102	102	891	8.7
12	0	79	79	905	11.5
13	0	108	108	918	8.5
14	0	85	85	1003	11.8
15	0	6	6	31	5.2
16	0	77	77	898	11.7
TOTAL	11	911	922	9914	10.8

	MOSS MATS	STREAM SEDS	TOTAL SITES	AREA Km^2	DENSITY $Km^2/SITE$
GRAND TOTALS	1397	1033	2430	26,107	10.7

Preparation of Samples

All samples were returned to a central depot within three days following collection, to avoid rotting. A drying and processing facility was built in Port Alberni for samples gathered on Vancouver Island. Mainland samples were shipped to Rossbacher Analytical Laboratory in Burnaby for processing.

Sediment sample drying was improved over last year by spreading each sample on large paper-lined trays which were stacked on mesh racks inside a drying shed. Average drying time was 2 to 3 days compared to 2 to 3 weeks needed last year when drying samples within their bags. Following drying, each sample was processed and inspected following methods outlined by Gravel and Matysek (1989).

Subsequent to field processing and inspection, samples were sent to Rossbacher Analytical Laboratory in Burnaby, British Columbia for final processing. A sediment sample processing routine was set-up for inspection, weighing, sieving and packaging of samples in preparation for analysis.

Blind duplicate and reference standard materials were inserted by Rossbacher in each sequence of twenty samples to monitor precision and accuracy of analytical results. In combination with field duplicates, it is possible to assess sample concentration variability due to geology (including mineralization), choice of sample site, subsampling, analytical digestion and determination methods, allowing us to distinguish between truly background and anomalous values.

Sample Analysis

Barringer Magenta Ltd. in Calgary, Alberta was chosen to analyze both water and sediment samples. Table 6-1-5 outlines analytical procedures for the various elements. These methods were employed in last year's survey with success. One minor modification has been introduced, a palladium inquart in the fire assay will permit the analysis of silver at a detection limit of 0.02 ppm or 20 ppb, 5 times lower than past surveys.

By maintaining consistent sampling, processing and analytical methods for the 1988 and 1989 surveys, sample results should be highly comparable. Recognizing continuity of most major lithological packages across the survey areas, merging databases will improve statistical inferences such as calculation of background and anomalous concentrations for various rock types.

Table 6-1-5. Methods and Specifications for Sample Analysis (after Matysek et al., 1989)

Element	Units	Detection Limits	Sample Weight	Digestion Technique	Determination Method	
Gold Silver	ppb ppm	1 ppb 10 ppb	10 g	Fire assay fusion - Palladium inquarting agent	FA-AA	Atomic adsorption spectrophotometry after digestion of doré bead by aqua regia
Cadmium Cobalt Copper Iron Lead Manganese Nickel Zinc	ppm ppm ppm % ppm ppm ppm ppm	0.2 ppm 2 ppm 2 ppm 0.02 % 2 ppm 5 ppm 2 ppm 2 ppm	1 g	3 ml HNO_3 let sit overnight, add 1 ml HCl in 90°C water bath for 2 hrs cool add 2 ml H_2O wait 2	AAS	Atomic absorption spectrophotometer using air-acetylene burner and standard solutions for calibration, background corrections made for Pb, Ni, Co, Ag, Cd.
Molybdenum	ppm	1 ppm	0.5 g	Al added to above solution		
Barium Vanadium Chromium	ppm ppm ppm	10 ppm 5 ppm 5 ppm	1 g	HNO_3-HCl-HF taken to dryness, hot HCl added to leach residue		
Bismuth Antimony	ppm ppm	0.2 ppm 0.2 ppm	2 g	HCl - $KClO_2$ digestion, KI added to reduce Fe, MIBK and TOPO for extraction	AAS	Organic layer analyzed by atomic absorption spectrophotometry with background correction
Tin	ppm	1 ppm	1 g	Sintered with NH_4I, HCl & ascorbic acid leach	AAS	Atomic absorption spectrophotometry
Arsenic	ppm	1 ppm	0.5 g	Add 2 ml KI & dil. HCl to 0.8M HNO_3 - .2M HCl	AAS-H	2 ml borohydride solution added to produce AsH_3 gas which is passed through heated quartz tube in the light path of atomic absorption spectrophotometer
Mercury	ppb	10 ppb	0.5 g	20 ml HNO_3 & 1 ml HCl	AAS-F	10% stannous sulphate added to evolve mercury vapour, Atomic Absorption Spect. determination
Tungsten	ppm	1 ppm	0.5 g	K_2SO_4 fusion HCl leach	COLOR	colorimetric: reduced tungsten complexed with toluene 3,4 dithiol
Fluorine	ppm	40 ppm	0.25 g	$NaCO_3$-KNO_3 fusion H_2O leach	ION	Citric acid added and diluted with water, Fluorine determined with specific ion electrode
Uranium	ppm	0.5 ppm	1 g	nil	NADNC	Neutron activation with delayed neutron counting
LOI	%	0.1 %	0.5 g	Ash sample at 500°C	GRAV	Weight difference
pH - water	pH unit	0.1	25 ml	nil	GCE	Glass-calomel electrode system
U - water	ppb	0.05 ppb	5 ml	Add 0.5 ml fluran solution	LIF	Place in Scintrex UA-3
F - water	ppb	20 ppb	25 ml	nil	ION	Fluorine ion specific electrode

RELEASE INFORMATION

Results of the 1989 program will be available to the public in three release packages; RGS-24 Victoria-Cape Flattery (NTS 92B-C), RGS-25 Alberni (92F) and RGS-26 Vancouver (92G).

- Hard copy release packages contain sample location (1:100 000 and 1:250 000 scale) and geochemical maps (1:250 000 scale for each element) together with a data booklet giving field and analytical results, summary statistics, anomaly ratings and discussion of methods and specifications.
- Digital format packages will consist of MS-DOS formatted, 5¼″ floppy diskettes containing listings of field and analytical information, and discussion of methods and specifications. Sample location maps will be included.

Release of RGS-24, 25 and 26 is tentatively scheduled for late June or early July, with release centres in Nanaimo and Vancouver.

1989 RGS ORIENTATION PROGRAM

Orientation surveys were completed in the eastern Rocky Mountains in preparation for next year's RGS program; map sheets covered were: 82G Fernie, 82J Kananaskis Lakes and 82N Golden. Detailed stream sampling orientation programs provide data on:

- Basic reconnaissance to determine regional climatic and physiographic environments as controls on mineral dispersion;
- General accessibility of sites;
- Applicability of various sampling media in terms of availability and geochemical response in mineralized and unmineralized drainages; and
- Characteristics of type deposits and associated dispersion patterns in local drainages.

SAMPLING DESIGN

In total, 337 samples were collected from twelve streams draining known showings and nine background creeks (Table 6-1-6). Selection of mineralized stream drainages for detailed study was based on several criteria:

- Known mineral occurrence at or near the headwaters;
- Each deposit type represented by at least one detailed study;
- Stream length before entering a major valley must be at least 4 kilometres to demonstrate down-stream dispersion pattern; and
- Good access to permit sampling along the entire length of stream.

Optimum sampling and analysis parameters will be determined from these drainages by comparative testing of media types and size fractions (Table 6-1-7).

Background creeks were chosen where geology, size of catchment basin, physiography and climate matched mineralized creeks, thereby providing a framework for comparison. Within these drainage basins, duplicate moss-mat and stream-sediment samples, a bulk stream sediment and a heavy mineral concentrate sample were collected at a single site, typically 4 to 6 kilometres down-stream of the headwaters.

TABLE 6-1-6 1989 RGS Orientation Survey - Map Sheets 82G, 82J & 82N

STREAM NAME	NTS	Deposit Type	Commodity Elements	Pathfinder Elements	Samples Collected MM	SS	BS	HMC
Roche	82G	Stratabound	Cu	Au, Ag, Mo, U	17	17	5	2
Phillips	82G	Stratabound	Cu	Zn, Br	17	17	7	2
Maus	82G	Vein	Pb, Zn	Ag, Au	16	16	7	2
Rabbit Foot	82G	Stratabound	Pb, Zn	B	11	11	3	2
Boulder	82G	Vein	Au, Ag	Pb, Zn, Cu	12	12	4	1
Madias	82J	Vein	Zn, Pb, Ag, Au	As, Sb, Ba	3	4	1	3
Rock Canyon	82J	Carbonatite	F, REE	Same	18	18	8	2
McMurdo	82N	Vein	Pb, Zn, Au, Ag	Same	13	14	6	2
Albert Canyon	82N	Vein	W, Zn	Cu, Mo, B	3	3	2	1
Woolsey	82N	Vein	Pb, Zn, Ag	Au, Sn	11	12	4	1
Clabon	82N	Vein	Sn, Pb, Zn	Ag, Be	10	11	4	2
				Totals	131	135	51	20

MM = Moss Mat Sediment SS = Routine Stream Sediment BS = Unsieved Bulk Sediment
HMC = Sieved Bulk Sediment

Table 6-1-7 Orientation '89 Sample Pattern and Size Fractions for Detailed Study of Mineralized Creeks

MEDIA TYPE COLLECTED	STATION SPACINGS	SIZE FRACTIONS
Moss-mat & routine stream sediment	500 metres	-40 to +80 mesh and -80 mesh
Duplicate moss-mat and routine stream sediments	Uppermost and lowermost sites	-40 to +80 mesh and -80 mesh
Bulk Samples	1000 metres	-40 to +80 mesh, -80 to +150 mesh and -150 mesh
Heavy mineral concentrates	lowermost sites	-80 mesh non-magnetic

Samples were shipped to Acme Analytical in Vancouver for preparation and analysis by 30 element ICP emission spectroscopy. A split of each sample will also be analysed by neutron activation at Becquerel Lab in Ontario. The combined results will provide approximately 50 elements having acceptable detection limits and, information on speciation for those elements overlapping the two methods.

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PRELIMINARY REPORT ON THE DISTRIBUTION AND DISPERSION OF PLATINUM IN THE SOILS OF THE TULAMEEN ULTRAMAFIC COMPLEX, SOUTHERN BRITISH COLUMBIA*
(92H/10)

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KEYWORDS: Applied geochemistry, Tulameen complex, platinum, chromite, dunite, soil, till, colluvium.

INTRODUCTION

Exploration for platinum in British Columbia is hampered by a lack of data on its distribution and behavior in the surficial environment. To obtain some of the information required and suggest practical guidelines for exploration, this study, involving systematic sampling and analysis of soils and other media in the vicinity of a platinum deposit in the Tulameen district of southwestern British Columbia, was initiated in 1988 (Fletcher, 1989). Preliminary results of work in progress are described.

DESCRIPTION OF THE STUDY AREA

LOCATION AND ACCESS

The field area, on Grasshopper Mountain north of the Tulameen River, approximately 25 kilometres west of Princeton, is divided into the main study area, on the southern slope of the mountain, and a second, smaller area near the summit (Figure 6-2-1). The lower part is accessible by pack trail from the Tulameen River road. Several kilometres of recently constructed drill road on the north and west sides of Grasshopper Mountain connect with the Lawless Creek forestry road and provide access from both the village of Tulameen and the Coquihalla Highway.

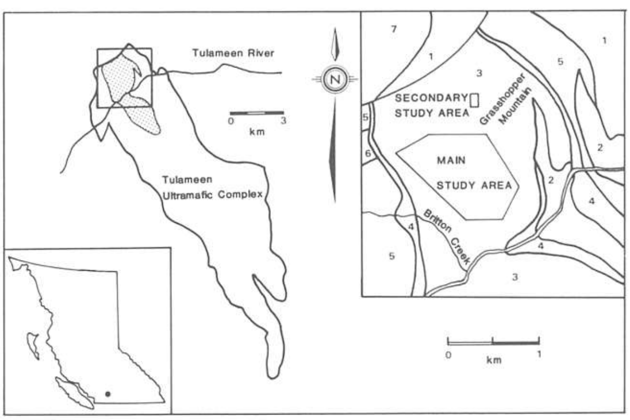

Figure 6-2-1. Location of the study areas and generalized geology, Tulameen ultramafic complex, B.C. (1) Nicola Group, (2) syenogabbro, (3) dunite, (4) olivine clinopyroxenite, (5) hornblende clinopyroxenite, (6) mylonitic rocks, (7) Eagle granodiorite (modified after Nixon and Rublee, 1988).

BEDROCK GEOLOGY

Grasshopper Mountain is the northern segment of the dissected dunite core of the Tulameen ultramafic complex, a Late Triassic zoned Alaskan-type ultramafic-gabbroic intrusion within metasedimentary and metavolcanic rocks of the Upper Triassic Nicola Group. Geology of the complex, described by Findlay (1969) and Nixon and Rublee (1988), comprises a dunite core surrounded by crudely concentric shells of olivine clinopyroxenite, hornblende clinopyroxenite and gabbroic rocks (Figure 6-2-1).

The focus of the current study is within the dunite core. The dunite is typically fine grained, extensively serpentinized, and weathers buff-brown on exposed surfaces. Platinum mineralization is restricted to chromite-rich dunite (Nixon *et al.*, 1989; St. Louis *et al.*, 1986). Platinic chromite is best exposed in the summit region and in the subvertical Cliff Zone area (Bohme, 1987) of Grasshopper Mountain, as randomly distributed massive to discontinuous pods, segregations, schlieren and disseminated grains. Schlieren are randomly distributed throughout the dunite and are thought to represent foliation-concordant remnants of former cumulate layers (Nixon and Rublee, 1988). Several zones of mineralization have been defined within the well-exposed central region of the core (Bohme, 1987), but till-covered areas are largely untested.

Platinum-group minerals (PGMs) include several platinum-iron alloys and platinum antimonides occurring as euhedral to subhedral inclusions within chromite grains, as anhedral platinum arsenide (sperrylite) interstitial to chromite grains, and platinum as a solid solution in native copper (St. Louis *et al.*, 1986; Nixon *et al.*, 1989). Nickel sulphides include pentlandite, violarite, and possibly millerite. PGMs are typically less than 30 microns in size, although grains up to 120 microns have been observed.

PHYSIOGRAPHY AND SURFICIAL GEOLOGY

Grasshopper Mountain lies on the western margin of the Thompson Plateau in its transitional zone with the Cascade Mountains. The flatter topography of the summit region, with a maximum elevation of 1500 metres, is deeply dissected by the steep, forested slopes of the Tulameen River and its tributaries. Prominent cliffs and sparsely forested scree

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

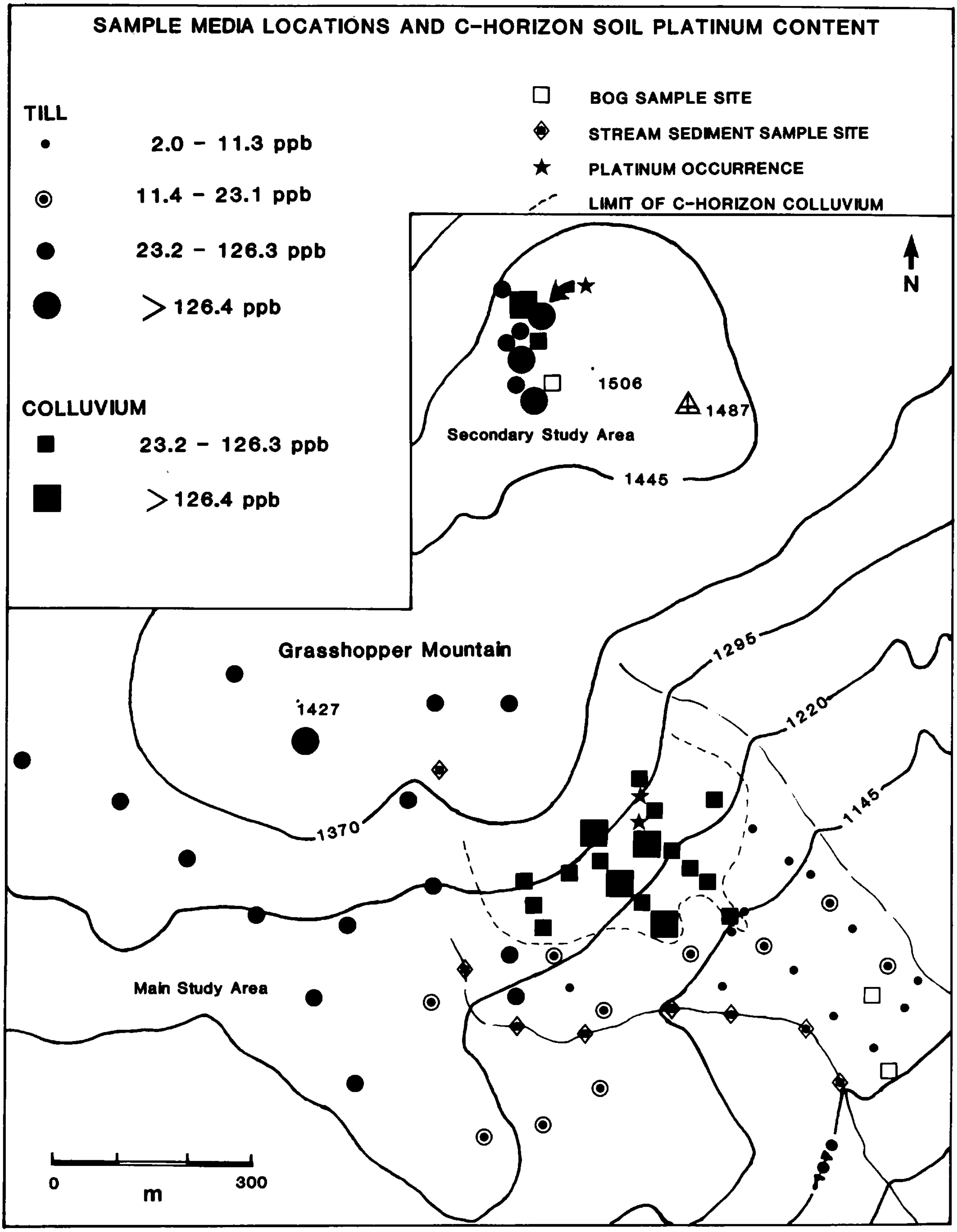

Figure 6-2-2. Sample locations and -70 mesh C-horizon-soil platinum content, Grasshopper Mountain, Tulameen, B.C.

slopes occur on the southeast and southwest faces of the mountain above the Tulameen River and Britton Creek. The mountain, with stands of mature Douglas fir, is largely unlogged near the study areas.

The region was ice-covered during the Pleistocene glaciation (Rice, 1947) and the forested slopes of Grasshopper Mountain are now mantled with glacial till. Near the summit this is thin and discontinuous, and disintegrating bedrock locally forms near residual soils. In this vicinity glacial striae indicate a south-southwesterly direction of ice movement. There is a thick postglacial apron of active scree below the cliffs (Cliff Zone) on the southeast face of the mountain and on the western slopes till is buried by stabilized dunitic colluvium. Fluvial channel deposits are exposed in roadcuts at two localities. At one site three soil parent materials – till, fluvioglacial outwash and colluvium – were observed. At lower elevations, fluvioglacial sediments are exposed in the valleys of the Tulameen River and its tributaries.

Small seepage-zone bogs occur in the lower part of the main study area and there are perched bogs on the summit. A small intermittent stream, Grasshopper Creek, flows through the area into the Tulameen River.

SOILS

Soil development, limited by high relief and active colluvial processes, is generally juvenile. The thickness of surficial LFH horizons seldom exceeds a few centimetres.

The main study area has three principal physiographic units (unstable colluvium, steep forested slopes, and base-of-slope areas) each with distinctive soil types. Unstable colluvium is characterized by orthic regosols. Genetic horizons are absent, although at some sites downslope of serpentinite outcrops the fines content increases with depth and there is an orange-brown surface coloration. Steep forested slopes exhibit a range of soil types with humo-ferric podzols and minor orthic regosols predominating in the east and eutric brunisols in the west. Development of near-residual brunisolic soils is most pronounced on gentler slopes and ridges and on the summit area where colluvial activity is minor. Relatively flat base-of-slope areas are characterized by seepage zones, small bogs and gleysolic soils.

SAMPLE COLLECTION, PREPARATION, AND ANALYSIS

SAMPLE COLLECTION

Soils, stream sediments and associated banks, bogs and waters were sampled (Tables 6-2-1 and 2). Soils were profiled and sampled in 76 pits dug in the two separate study areas during 1988 and 1989 (Figure 6-2-2). Duplicate samples of 10 to 15 kilograms each were typically collected from mineral horizons. Soil pedons were classified according to the Canadian System of Soil Classification (Agriculture Canada Expert Committee on Soil Survey, 1987).

Stream sediments were collected at approximately 100-metre intervals from seven sites along Grasshopper Creek. Moss-mat samples were also collected, if present, and bank samples were taken at five of the sites. Bog-centre and marginal samples were taken from three bogs. Following the spring snowmelt in 1989, waters were sampled and pH measured at stream sediment sites, bogs and some soil pits. Waters were filtered to <0.45 micron and acidified with hydrochloric acid in the field.

TABLE 6-2-1
DISTRIBUTION OF SOIL PROFILES ACCORDING TO STUDY AREA AND PARENT MATERIAL, GRASSHOPPER MOUNTAIN, B.C.

	Main Study Area	Secondary Study Area	Total
Till	38	11	49
Colluvium	25	2	27
Total Soil Profiles	63	13	76

TABLE 6-2-2
DISTRIBUTION OF SAMPLE MEDIA ACCORDING TO STUDY AREA, GRASSHOPPER MOUNTAIN, B.C.

	Main Study Area	Secondary Study Area	Total
Mineral horizons	149	31	180
LFH horizons	40	7	47
Stream sediments	8	—	8
Mossmats	5	—	5
Bank samples	10	—	10
Bog samples	4	2	6
Water samples	17	—	17

SAMPLE PREPARATION

Preparation of samples involved wet sieving a representative split of all C-horizon soil, stream sediment, moss-mat and bank samples to obtain a −70 mesh (<212 microns) fraction. Two-hundred-gram splits were then taken with a Jones riffle and ground to approximately −200 mesh in a tungsten carbide ring mill. Ground LFH and dried bog samples were ashed in a muffle furnace at 700°C.

ANALYTICAL TECHNIQUES

Samples were analyzed for three groups of elements and major element oxides at a commercial laboratory.

(1) Pt-Pd-Au-Rh by lead-fire assay on 10.0 gram subsamples using an ICP-MS finish in 1988 and ICP-AES in 1989. For platinum concentrations greater than 10 ppb analytical precision, based on duplicate analyses, is better than ±20 per cent at the 95 per cent confidence level.

(2) As-Sb-Bi-Ge-Se-Te by hydride generation. A 0.5 gram sample was digested with 3 millilitres 3:1:2 HCl-HNO_3-H_2O at 95°C for one hour and diluted to 10 millilitres with water. Hydrides are then determined by ICP-AES.

(3) Whole-rock analysis. A 0.2 gram sample was fused with $LiBO_2$, dissolved in 100 millilitres nitric acid, and analyzed for Si, Al, Fe, Mg, Ca, Na, K, Mn, Ti, P, Cr, and Ba by ICP-AES. Loss on ignition was also determined.

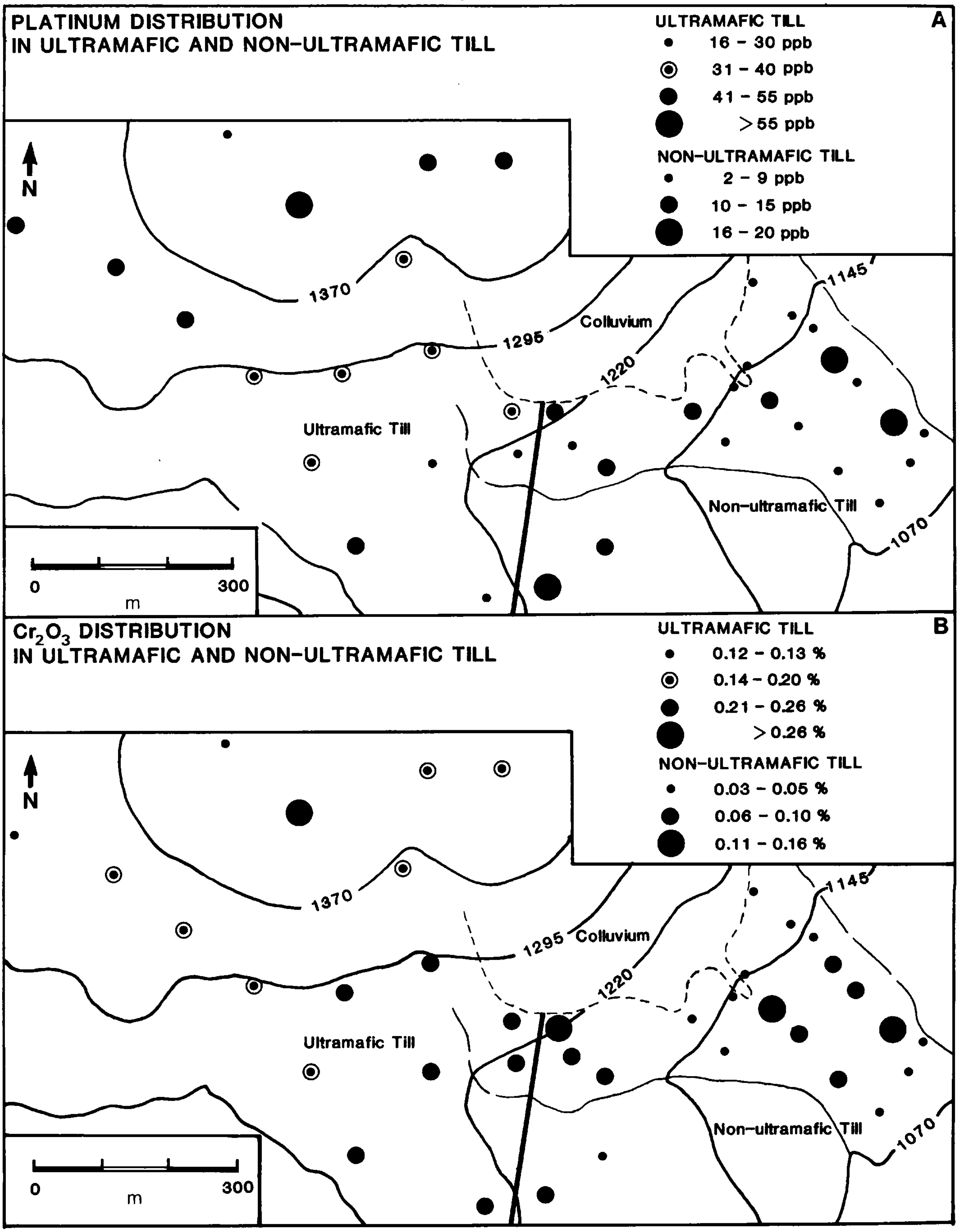

Figure 6-2-3. Distribution of Platinum (A) and Cr_2O_3 (B) in ultramafic and non-ultramafic till, main study area, Grasshopper Mountain, Tulameen, B.C.

Ashed LFH and organic samples were analyzed for Pt-Pd-Au-Rh as above. Platinum content of waters was determined at the Geological Survey of Canada, Ottawa, Ontario.

RESULTS

C HORIZON SOILS

Preliminary analytical results of the platinum content of soils on till and colluvium parent materials are shown in Figure 6-2-2. Mean and median concentrations of platinum and selected elements are given in Table 6-2-3.

TILL COMPOSITION

Two distinct non-overlapping till populations of MgO concentrations occur in the main study area below and adjacent to the colluvium. The south-southwest-trending boundary between the two is shown in Figure 6-2-3. The population in the western half of the area has a mean MgO content of 16.51 per cent and is associated with generally higher values of platinum and chromium. The second population, in the eastern half of the area, has a mean MgO content of 5.60 per cent. These are considered to be predominantly ultramafic and a non-ultramafic exotic till, respectively.

PLATINUM

Contour values for platinum were determined using log probability plots to separate lower (2.0-11.3 ppb) and upper (23.2-126.3 ppb) platinum populations in soils on tills (Figures 6-2-2 and 3). These populations generally correspond to soils derived from the ultramafic and non-ultramafic tills, which average 52 and 9 ppb platinum, respectively, in the main study area (Table 6-2-3). The mean platinum content of colluvium (120 ppb) is considerably greater. Median values, unaffected by outliers, give a better approximation of typical concentrations of platinum – 7.5 ppb in non-ultramafic till, 36 ppb in ultramafic till, and 88 ppb in colluvium. The maximum concentration (885 ppb) is in colluvium below one of the Cliff Zone platinum occurrences (Figure 6-2-4).

CHROMIUM

Distribution of chromium is generally similar to that of platinum. Thus, colluvium has the greatest mean Cr_2O_3 content (0.33 per cent) whereas soils on ultramafic and exotic tills have mean contents of 0.20 per cent and 0.07 per cent, respectively (Figures 6-2-3 and 4; Table 6-2-3). However, on a more detailed scale, there are differences. For example, within the area of ultramafic till a zone of higher chromium values lies downslope and to the southeast of an area of relatively high platinum values. Similarly, although an anomalous chromium value coincides with the maximum platinum concentration of 885 ppb in colluvium below the Cliff Zone (Figure 6-2-4), the chromium anomaly extends considerably farther, following a general trend for chromium content of colluvium to increase downslope.

OTHER ELEMENTS

Concentrations of palladium, gold, arsenic and antimony are summarized in Table 6-2-3. The most notable features are: (1) concentrations of palladium are generally less than 5 ppb except in soils derived from ultramafic till; (2) similar average arsenic concentrations in soils on all parent materials with 90 per cent of the values being greater than 10 ppm. High arsenic values, up to 43.5 ppm, are found downslope of the showings in the Cliff Zone; (3) the three to four times greater gold content of colluvial parent material from the

TABLE 6-2-3
MEAN, MEDIAN, AND RANGE OF PLATINUM AND OTHER SELECTED CONSTITUENTS OF C-HORIZON SOILS ACCORDING TO PARENT MATERIAL, GRASSHOPPER MOUNTAIN, B.C.

		Pt (ppb)	Pd (ppb)	Au (ppb)	As (ppm)	Sb (ppm)	MgO (%)	Cr_2O_3 (%)	CaO (%)	Na_2O (%)	K_2O (%)
NON-ULTRAMAFIC TILL	x	9.1	4.0	7.9	15.2	1.0	5.60	0.07	4.66	2.64	1.34
Main Study Area	M	7.5	3.0	6.0	14.3	0.9	5.44	0.05	4.75	2.58	1.34
(n = 21)	Min	2	2	2	8.1	0.5	3.86	0.03	3.13	2.34	1.09
	Max	20	15	34	23.2	1.8	8.23	0.16	5.51	3.20	1.64
ULTRAMAFIC TILL	x	52.8	8.9	8.4	15.8	0.3	16.51	0.20	3.71	1.27	0.83
Main Study Area	M	36.0	6.5	8.0	13.1	0.2	16.25	0.19	3.61	1.13	0.83
(n = 17)	Min	16	2	2	5.3	0.1	10.45	0.12	1.88	0.85	0.57
	Max	311	48	21	52.5	0.8	28.73	0.29	5.58	2.08	1.07
ULTRAMAFIC TILL	x	158.1	7.0	5.1	18.5	0.4	13.84	0.28	3.28	1.72	0.66
Secondary Study	M	89.0	2.5	4.0	15.6	0.4	13.21	0.26	3.46	1.83	0.73
Area (n = 8)	Min	42	2	1	7.4	0.1	10.64	0.12	1.24	1.06	0.35
	Max	455	36	10	30.5	0.8	19.06	0.51	4.48	2.29	0.79
COLLUVIUM	x	120.0	2.6	21.0	21.3	0.6	24.16	0.33	1.73	0.83	0.30
Main Study Area	M	88.0	2.0	18.5	15.8	0.5	23.29	0.32	1.90	0.85	0.22
(n = 25)	Min	24	2	2	7.5	0.2	14.29	0.20	0.35	0.08	0.05
	Max	885	5	56	56.3	1.2	32.78	0.50	3.29	1.44	0.82

x = mean
M = median
Min = minimum value
Max = maximum value

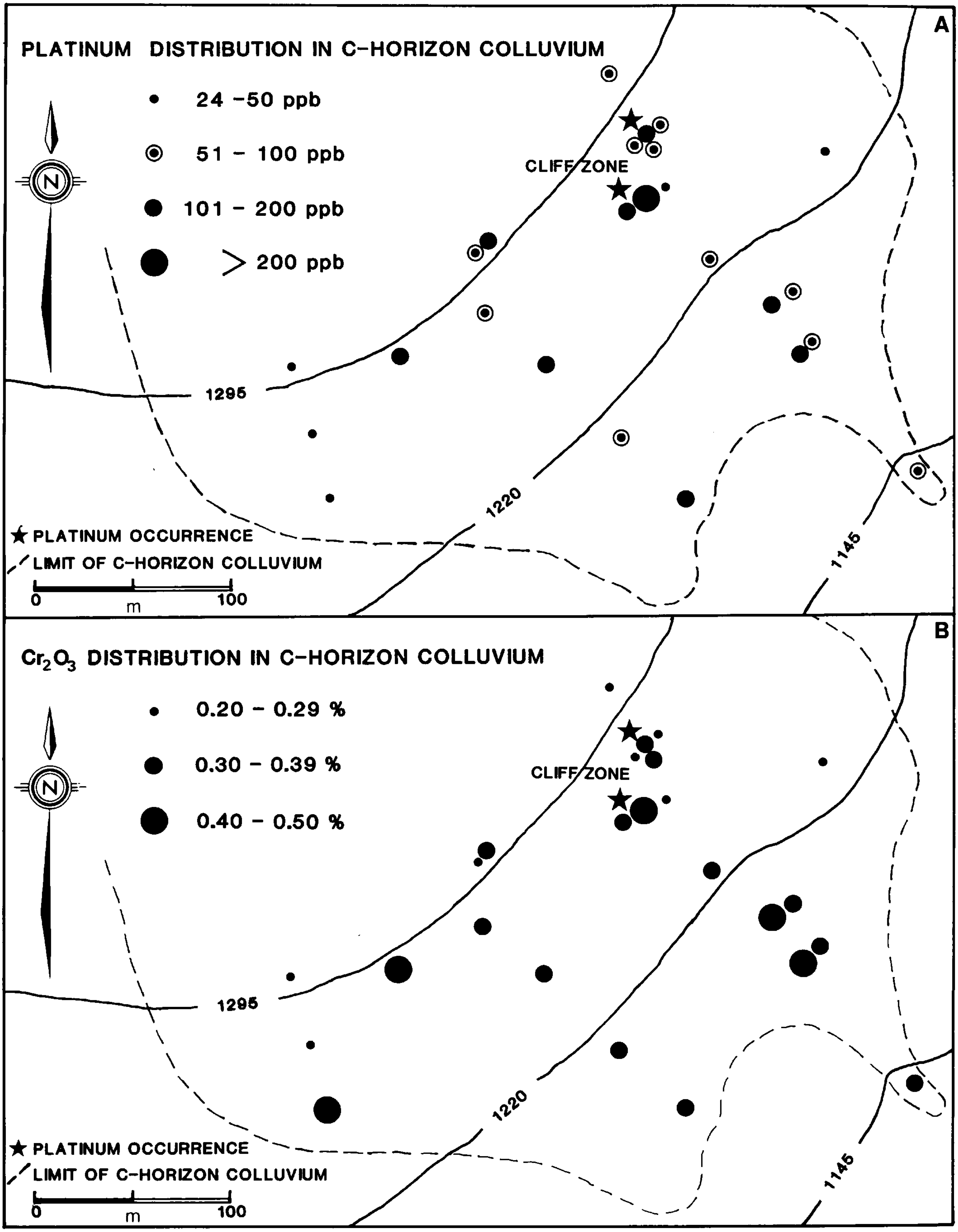

Figure 6-2-4. Distribution of Platinum (A) and Cr_2O_3 (B) in C-horizon colluvium, main study area, Grasshopper Mountain, Tulameen, B.C.

Cliff Zone (average 21.0 ppb with up to 54 ppb in serpentinized material) compared to soils on tills; and (4) antimony values (1.2 – 1.8 ppm) in the seepage zone at the base of the main study area that are three to six times greater than those upslope.

LFH Horizon Soils

Up to 167 ppb platinum has been obtained in LFH horizons from the secondary study area where a zone of high LFH platinum values appears to be indicative of the underlying mineralization. The few LFH horizons sampled from active colluvium contain up to 141 ppb platinum. Elsewhere platinum content of LFH horizons is low (typically less than 10 ppb) and erratic. The data have not yet been evaluated for contamination by mineral matter.

Stream Sediments and Waters

Platinum concentrations in stream sediments from Grasshopper Creek range from 8 to 91 ppb (Table 6-2-4). Associated moss-mats contain 8 to 47 ppb platinum. Most stream water and seep samples at lower elevations contain less than 1 part per trillion platinum. However, samples from the summit area contain 1.3 – 3.5 parts per trillion platinum.

TABLE 6-2-4
DISTRIBUTION OF PLATINUM IN STREAM SEDIMENTS AND MOSS-MATS, GRASSHOPPER CREEK, GRASSHOPPER MOUNTAIN, B.C.

Site	Pt in Sediment (ppb)	Pt in Moss-mat (ppb)	– 70# Sediment (%)	– 70# Moss-mat (%)	Stream Topography
1	18	17	10.7	40.7	L
2	78	8 (17)	9.5	36.3	B
3	11 (12)	—	25.2	—	M
4	8	11	20.2	25.5	M
5	53	47	18.1	41.8	G
6	20	—	26.7	—	G
7	91	—	33.6	—	B
8	32	23	51.6	75.6	B

L = Level to gentle
B = Break of slope
G = Gentle slope
M = Moderate slope

Site 8 is farthest upstream

– 70#(mesh) as Wt.% of – 10 mesh

Bogs

Two bogs in the lower till-covered part of the main study area contain up to 9 ppb platinum and a bog in the summit area contains up to 21 ppb. On the basis of very limited data, greater concentrations of platinum appear to be associated with bog-margin than bog-centre sites. Concentrations of antimony (0.8 to 4.1 ppm) in bogs are similar or slightly higher than those in base-of-slope soil samples.

DISCUSSION

The high chromium and MgO content of colluvium below the Cliff Zone clearly reflects its origin in mechanical weathering and mass wasting of the dunite cliffs. However, MgO concentrations (average 24.16 per cent) in the -70 mesh colluvium are appreciably lower than an average of 42.85 per cent for the dunite, based on analyses by St.Louis *et al.* (1986). It is not clear to what extent this is a result of weathering. In this context, platinum concentrations (average 120 ppb) in active colluvium are very similar to estimates of 48 to 180 ppb platinum in dunite, serpentinized dunite and serpentinite (St. Louis *et al.*, 1986). Chromitic dunite, with an average platinum content of 3410 ppb (St. Louis *et al.*, 1986) is the most likely source of the highest platinum value (885 ppb) in colluvium.

Based on a south-southwesterly direction for ice movement and composition of the dunite, the relatively high concentrations of MgO and chromium in tills from both the summit area and the western part of the main study area indicate that they contain relatively high proportions of dunitic material. Conversely, lower concentrations of both elements in tills in the eastern part of the area are probably indicative of their lower dunite content and greater compositional influence of rock units to the north or northeast. Determination of MgO and chromium content in soils may thus provide a useful method of delineating glacial dispersion, and mixing and dilution of material derived from the dunite core. In this case platinum concentrations in soils could be evaluated against soil chromium or MgO values (Figures 6-2-5 and 6). The presence of complex, composite soil profiles where shallow colluvium (upslope source) overlies till (up-ice source) suggests that without adequate care, routine soil sampling may result in erroneous interpretations of anomaly contrast and probable source.

The relationships between the remaining elements and platinum are not clear. However, arsenic and antimony concentrations might, in part, be related to the platinum arsenides and antimonides in the dunite and their redistribution during weathering. Relatively high values of antimony in base-of-slope and bog samples may indicate some hydromorphic dispersion for this element. Above average concentrations of gold in serpentinized colluvium are consistent with lithogeochemical results obtained by St.Louis *et al.* (1986).

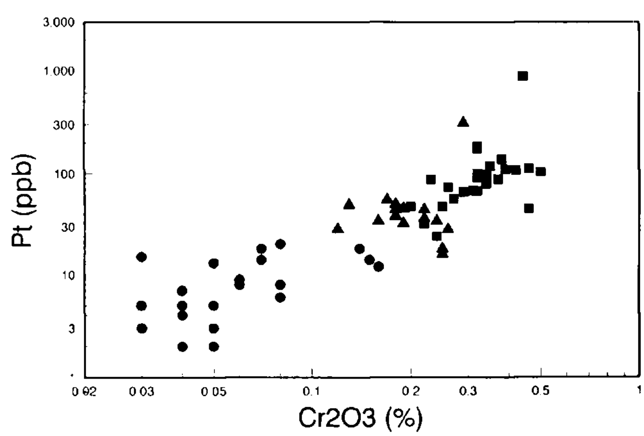

Figure 6-2-5. Pt (ppb) versus Cr_2O_3 (%) in colluvium (squares), ultramafic till (triangles) and non-ultramafic till (circles).

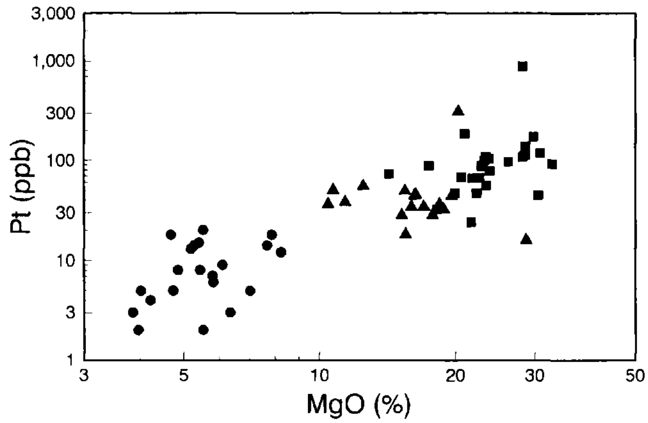

Figure 6-2-6. Pt (ppb) versus MgO (%) in colluvium (squares), ultramafic till (triangles) and non-ultramafic till (circles).

CONCLUSIONS

Glacial dispersion and post-glacial mass wasting are the dominant processes influencing distribution of platinum in the vicinity of platinum occurrences on Grasshopper Mountain. Platinum content of soils, ranging from 2 to 885 ppb, is strongly dependent on the amount of dunite in the soil parent material. This can be estimated roughly from soil MgO and chromium contents. The presence of composite till-colluvium soil profiles necessitates careful sampling for correct interpretation of geochemical patterns.

ACKNOWLEDGMENTS

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GEOLOGY OF PLACER DEPOSITS IN THE CARIBOO MINING DISTRICT, BRITISH COLUMBIA; IMPLICATIONS FOR EXPLORATION (93A, B, G, H)

By V.M. Levson, T.R. Giles, P.T. Bobrowsky and P.F. Matysek

KEYWORDS: Economic geology, Cariboo Mining District, placer gold, sedimentology, auriferous gravels, exploration.

INTRODUCTION

This report provides details on field activities in placer geology conducted by the Applied Geochemistry and Surficial Geology Section of the British Columbia Geological Survey Branch. In 1989, the provincial government expanded the area open to placer mining substantially, and the need for new exploration techniques in these regions was recognized. Studies of gold-bearing placer deposits were undertaken with the objective of developing a set of geological criteria useful for recognizing placer potential in undeveloped or poorly explored areas. The results of these studies will be used to develop regionally applicable exploration models. The complexity of the geology observed during this field season indicates that detailed sedimentological studies are required in order to understand both the distribution of placer deposits in a regional sense and the location and extent of pay streaks at the local level.

The Cariboo Mining District was selected for initial study because of its long history of placer gold production. Since 1860 the district has produced almost 100 000 kilograms (about 2.5 to 3 million ounces) of gold, more than any other placer area in British Columbia (B.C. Ministry of Energy, Mines and Petroleum Resources, 1963; Boyle, 1979). Several large (>200 cubic metres per day processed) placer mines are now active in the region, as well as over 200 small operations, including hand mining and exploration projects. Shallow placer gold deposits in the Cariboo have been largely depleted. They were primarily in present-day river valleys,

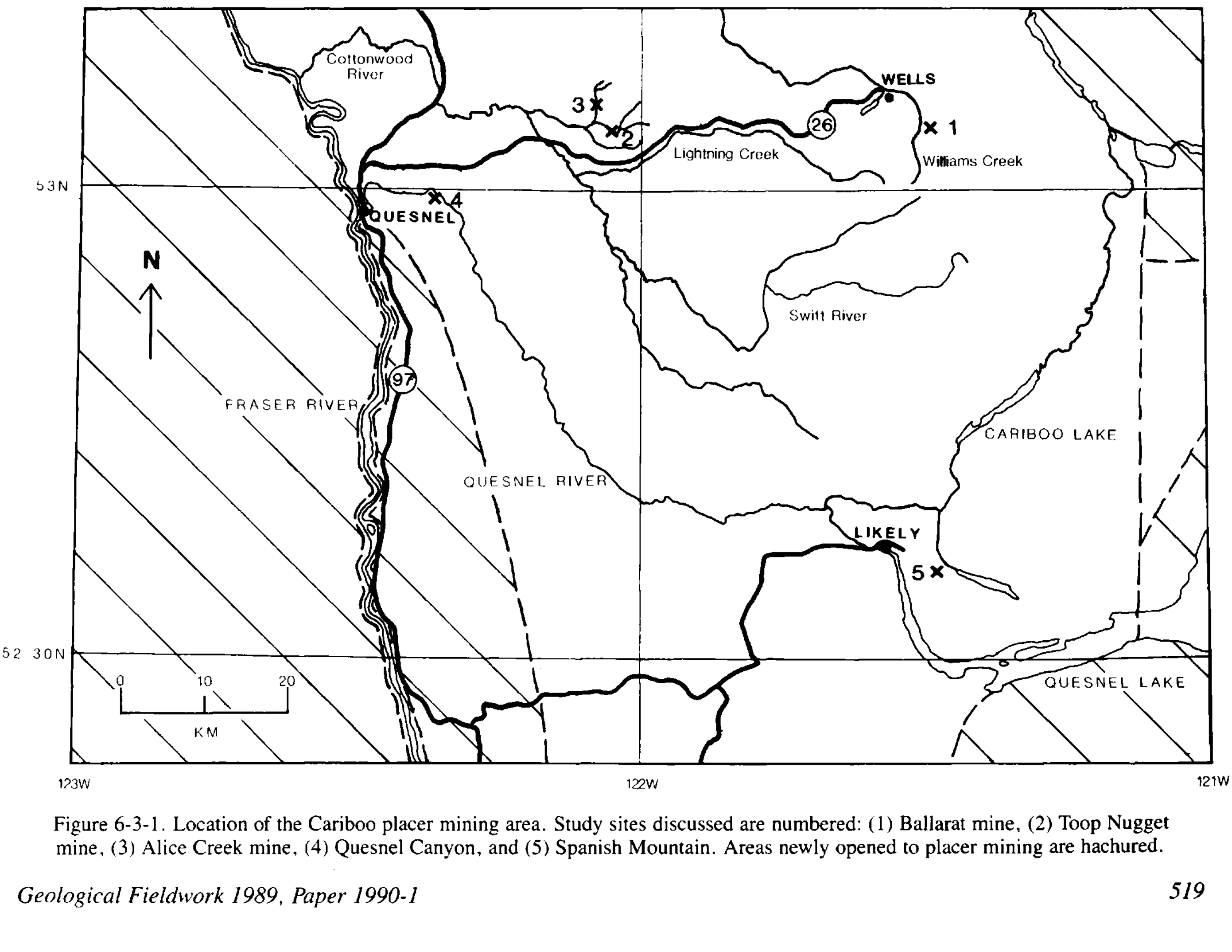

Figure 6-3-1. Location of the Cariboo placer mining area. Study sites discussed are numbered: (1) Ballarat mine, (2) Toop Nugget mine, (3) Alice Creek mine, (4) Quesnel Canyon, and (5) Spanish Mountain. Areas newly opened to placer mining are hachured.

where water for sluicing is readily available. Most large active operations (~70 per cent) are exploiting placers that are relatively deeply buried, or more distant from sources of water. Higher gold prices in the last 20 years have also allowed successful mining of previously uneconomic gravels, especially where large volumes of lower grade gravel are present. Although there are still many small operations in the Cariboo, the trend toward exploitation of deeper or larger volume placer deposits has resulted in the development of several large mines.

Detailed sedimentological evaluations of placer deposits have been made in unglaciated terrains such as on the White Channel gravels in the Klondike area of the Yukon Territory (Morison and Hein, 1987), however, the geology of placer deposits in glaciated areas is poorly understood. Regional studies in British Columbia have been done, but detailed sedimentological analyses are lacking. Sites selected for this study are producing mines offering good section exposure. Highwalls in the active mines were mapped and lithologic, pebble fabric and sedimentological studies were conducted. Samples were collected for textural, mineralogical and geochemical analysis. Gold production in each stratigraphic unit was determined, where possible, by discussions with miners. Detailed results of sample analyses are not presented in this report, rather a geological synopsis of several case studies is outlined. Additionally, a description of the placer geology at the Ballarat mine near Wells (Figure 6-3-1, Location 1) is given as an example of the on-going research.

PREVIOUS WORK

General descriptions of placer deposits in the Cariboo area first appeared in 1874 in British Columbia Minister of Mines Annual Reports. Johnson and Uglow (1926) completed descriptions of placer and lode gold deposits in the Wells-Barkerville area. Regional bedrock mapping was conducted by Tipper (1971) and more recently by Struik (1982). The Quaternary geology of the region was mapped by Tipper (1971) and recent investigations of the Quaternary and placer geology have been made by Clague (1987a, b and 1989). Depositional environments of Cariboo placer deposits have recently been discussed by Eyles and Kocsis (1988, 1989).

MAJOR PLACER DEPOSITS AND EXPLORATION IMPLICATIONS

Toop Nugget Mine

Discovered in 1972, the productive Toop Nugget mine (Figure 6-3-1, Location 2) lies in an area that has been frequently worked since the 1860s (*e.g.* shallow gravels at Mary, Alice and Norton creeks). Recent recognition of this deposit illustrates the potential for other new discoveries in equally well explored regions. At the Toop property, mining is mainly within a glacial meltwater channel that incised and removed much (at least 20 metres) of the overburden. Exploration for deeply buried auriferous gravels could benefit by focusing on meltwater channels where the overburden has been removed by natural processes.

Heavily oxidized and locally strongly cemented older (interglacial or preglacial) gravels (Plate 6-3-1) are overlain by two diamicton (till) units (1 and 3; Figure 6-3-2) which are separated in places by intertill sands and gravels (Unit 2). Early postglacial gravels (Plate 6-3-2) partially infill the meltwater channel and contain some gold, probably reconcentrated from underlying units. Gold recovered thus far comes mainly from the lower interglacial or preglacial gravels, but the intertill and early postglacial gravels have also produced gold. Some rich pay-zones were uncovered by the miners, with nuggets up to about 100 grams in size in the lower gravels (T. Toop, personal communication, 1989). The intertill and postglacial gravels also contain coarse gold nuggets. The hackly shape and presence of quartz in some nuggets suggests that the local bedrock may also be economically viable in this area.

The original exploration strategy of the miners assumed the older gravels represented a previous course of Lightning Creek. Similar buried valleys have been found locally (*e.g.* Alice Creek) and further potential is speculated, but as yet unrealized near Toop Nugget mine. In addition, the Toop

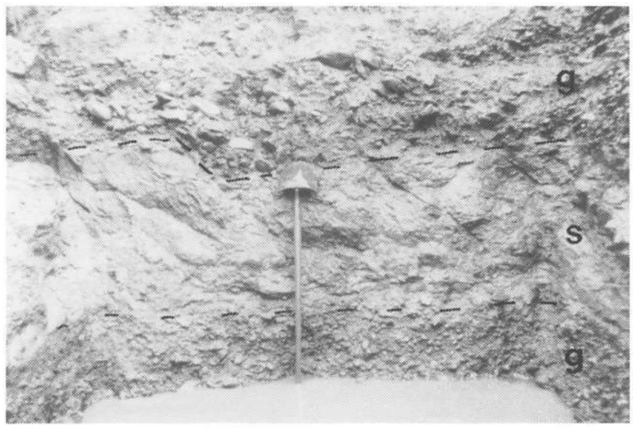

Plate 6-3-1. Oxidized and cemented auriferous sands (s) and gravels (g) in an active pit at the Toop Nugget mine. Poorly to moderately sorted, clast-supported, weakly imbricated pebble gravels are interbedded with faulted, pebbly sands and silts. The deposits are interglacial or preglacial in age.

Plate 6-3-2. Horizontally stratified, auriferous sands and gravels at the Toop Nugget mine interpreted as early postglacial deposits. The sediments overlie bedrock (r). The measuring rod is 4 metres long.

mine illustrates that exploration in meltwater-channel valleys elsewhere may provide a cost-effective means of locating productive preglacial and interglacial placer deposits.

Alice Creek

Gold-bearing gravels at this site, near the Toop mine (Figure 6-3-1, Location 3), are overlain by about 30 metres of interbedded diamicton, gravel, sand, silt and clay (Figure 6-3-3). Two diamicton units (Units 1 and 3) are interpreted as tills. Within Unit 1, intercalated silts and clays are laterally extensive and are interpreted as glaciolacustrine sediments, deposited during a temporary retreat of ice from the region during the late Pleistocene. Intertill sands and gravels (Unit 2) were probably deposited by a glaciofluvial stream.

The gold-bearing deposits (Figure 6-3-3; Unit 4) consist of interbedded gravels and sands interpreted as low-sinuosity braided-river sediments (Eyles and Kocsis, 1989). Unit 4 is 9 to 14 metres thick but only the upper 5 metres, dominated by horizontally stratified sands, is presently exposed. The gravels are poorly to moderately sorted, clast supported, discontinuously cemented and manganese and iron stained. Gold values increase toward the base of the gravels with the main pay zone in the lower 3 to 5 metres over bedrock. Paystreaks are sporadic, producing an average of 4 grams per cubic metre with a maximum return of about 9 grams (all gold concentration values reported here are based on approximate mining records). From 1986 to 1988, a total of 135 000 cubic metres of material was moved, of which only 11 000 cubic metres was washed with a resulting production of 43 kilograms (1375 ounces) of gold (Bob Patrick, personal communication, 1989).

This site provides an excellent example of the mining potential of deeply buried placer deposits. The cost of removing large volumes of overburden is offset by the potential richness of the deep gravels. In deposits of this type, detailed sedimentological and stratigraphic data are required to predict overburden depths and to identify the extent and volume of gold-bearing strata at present mines and in areas of active exploration.

Spanish Mountain

At the producing Spanish Mountain-McKeown mine (Figure 6-3-1, Location 5), gold is found in poorly sorted and crudely stratified coarse gravels interpreted as debris-flow deposits (Figure 6-3-4; Unit 2). Lenses of better sorted gravel, sand and silt occur throughout Unit 2 and are interpreted as fluvial channel deposits. The gold-bearing gravels are overlain by poorly exposed diamicton interpreted as till (Unit 1) suggesting that the placer deposits predate the last glaciation in the area. The gold-bearing sediments may be locally derived alluvium or subglacial deposits. Unit 2 appears to infill the upper part of an elevated erosional channel cut in bedrock. The orientation of the channel is not well defined but appears to be oblique to the regional northwesterly strike of bedrock and topography. The oblique orientation of the channel relative to ice flow would provide an ideal situation for the development of a subglacial cavity. Gold content is generally consistent throughout the mined sequence, averaging about 1 gram per cubic metre not including gold finer than 100 mesh (0.149 mm). The gold appears to originate locally and nuggets have been recovered (V. McKeown, personal communication, 1989). This illustrates that the upper part of buried channels can be productive even where sediments are relatively poorly sorted.

Drilling results indicate that the bedrock channel is up to 74 metres deep. The lower 50 metres of the channel is infilled with clean pebble and boulder gravels (P. McKeown, personal communication, 1989). The lower gravels have not been mined extensively but there is a high probability that they are gold bearing, particularly at their base. North of Spanish Mountain, along the Cariboo River, Clague (1987b) identified a buried channel similar to the Bullion mine near Likely (Plate 6-3-3) as a possible placer exploration target. The Bullion mine has produced over 3860 kilograms (120 000 ounces) of gold. There is a good probability that other rich buried-channel placer deposits occur in the Cariboo region and possibly other areas of British Columbia. The identification and testing of these, often obscure, channels should be a major focus for the placer exploration industry.

Plate 6-3-3. The Bullion hydraulic mine near Likely. Note the large volume of material removed and the typically stratified buried channel sediments.

Quesnel Canyon

At least three distinct gravel units (Figure 6-3-5) have been identified along this portion of the Quesnel River (Figure 6-3-1, Location 4). The lower gravels (Plate 6-3-4; Figure 6-3-5, Unit 4) consist of well-rounded quartzite, chert and volcanic clasts. At the base they are strongly cemented with iron and manganese oxides. These texturally mature gravels are believed to be Tertiary in age. Lithologic and paleocurrent data indicate a substantially different drainage pattern to that of the present river. The overlying gravel deposits (Unit 3) show greater lithologic diversity and may be interglacial in age. The uppermost gravels (Unit 2), which are overlain by glaciolacustrine deposits (Unit 1), are similar to other glaciofluvial gravels in the area.

Postglacial terrace gravels, unconformably overlying the older gravel units (Plate 6-3-5), are presently being mined in the area. Mining of the Tertiary gravels is uncommon due in part to their induration. Much of the placer gold in the region is believed to have been reconcentrated, from the cemented

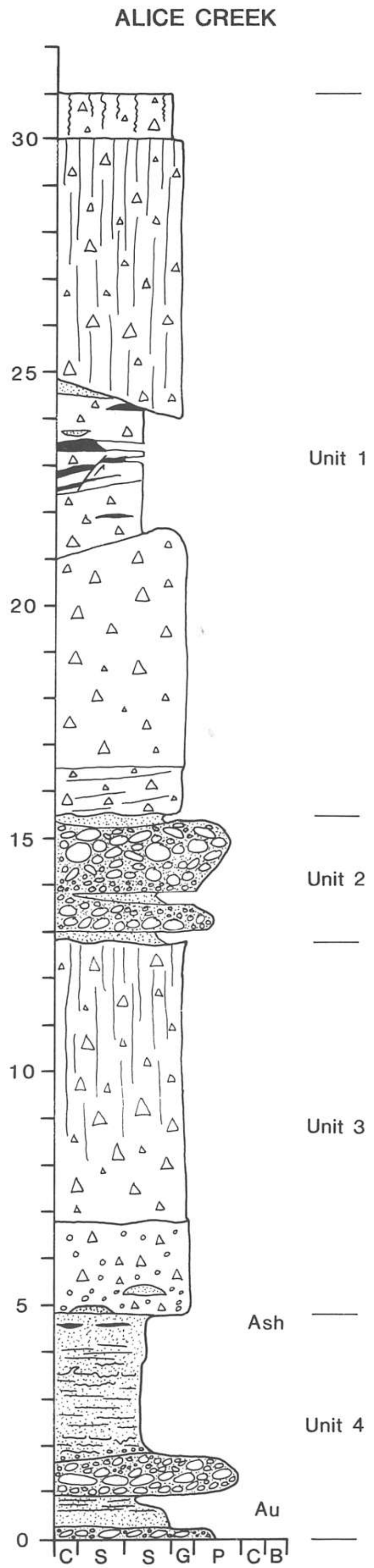

Figure 6-3-3. Composite stratigraphic column of old highwall exposures at Alice Creek mine.

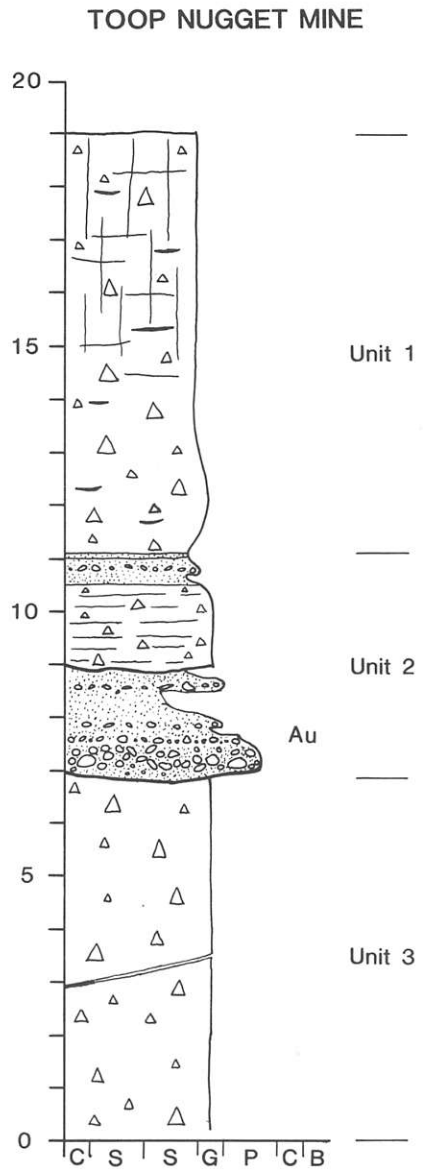

Figure 6-3-2. Stratigraphic column of old highwall exposure at Toop Nugget mine. See Figure 6-3-3 for legend. Horizontal scale: C–clay S–sand S–silt G–granule P–pebble C–cobble B–boulder

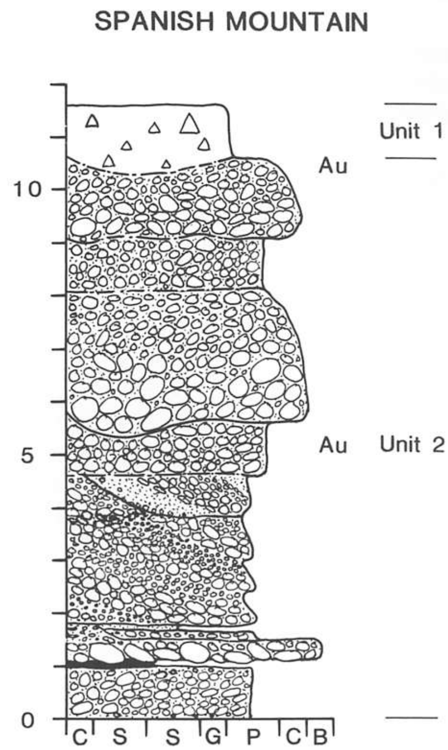

Figure 6-3-4. Stratigraphic column of highwall exposure at Spanish Mountain mine.

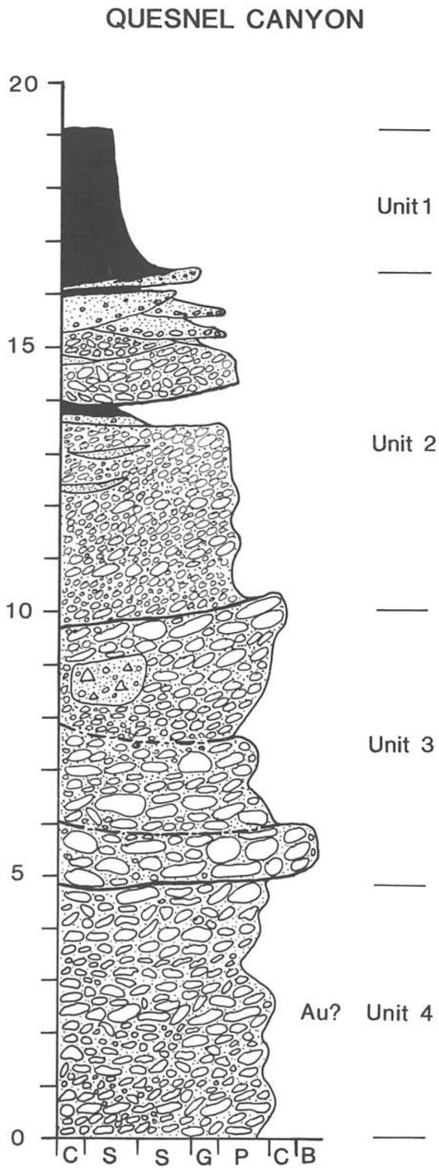

Figure 6-3-5. Stratigraphic column of exposure near Quesnel Canyon.

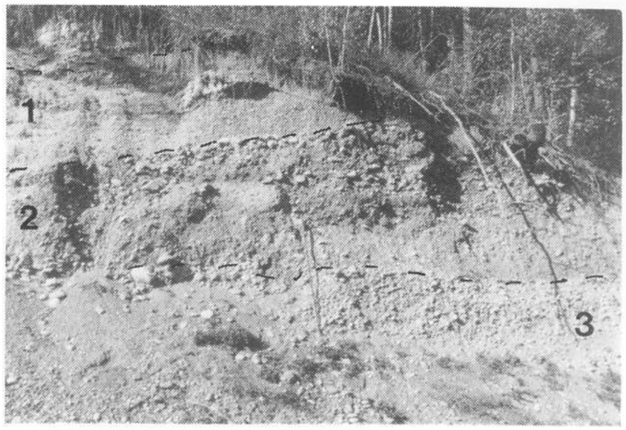

Plate 6-3-4. Silts, clays and glaciofluvial gravels of Unit 1(upper, grey units) unconformably overlying oxidized interglacial (Unit 2) and possibly Tertiary (Unit 3) gravels along the Quesnel River near Big Canyon.

Plate 6-3-5. Holocene terrace gravels (1) unconformably overlying older interglacial or preglacial gravels (2). Exposure is along an actively mined terrace. The measuring rod is 4 metres long.

Tertiary gravels, by younger river systems. Sedimentological and stratigraphic studies of the older river deposits are needed to understand the present distribution of placers in the Cariboo (Clague, 1989).

CASE STUDY – BALLARAT MINE

The Ballarat pit is about 2 kilometres north of Barkerville and is adjacent to Williams Creek, one of the richest gold-producing streams in British Columbia (Figure 6-3-1, Location 1). Williams Creek produced at least 1460 kilograms (47 000 ounces) of gold in the 10-year period from 1874 to 1885 (Holland, 1950). The gold-bearing strata underlie 2 to 8 metres of till deposits and are of alluvial, glaciofluvial and fluvial origin. The deposits infill an ancient bedrock channel and the lowermost and presumably richest gravels are as yet unmined. Gold from the mine is almost entirely finer than very coarse sand [99 per cent is smaller than 16 mesh (1.19 mm)]. This site illustrates the potential for the presence of other rich placer gold deposits in regions that historically have been heavily explored and mined.

Several highwalls at the Ballarat mine were mapped and described in detail during the course of this study. The major stratigraphic units exposed in the main highwall, oriented approximately north-south, are shown on Figure 6-3-6. The details of the Ballarat case study are provided to illustrate the type of geologic information that will aid in future development and exploration.

BEDROCK (UNIT 1)

Bedrock (Unit 1) is exposed at the northern and southern ends of the main highwall at the Ballarat mine (Figure 6-3-6). At the south end it consists of strongly altered muscovite-talc schist and phyllite with quartzite beds 1 to 10 centimetres thick. The schist is crosscut by several discordant quartz veins generally less than 1 centimetre wide. Bedrock exposures at the north end of the mine site are interbedded limestone, bedded quartzites, schist and phyllite.

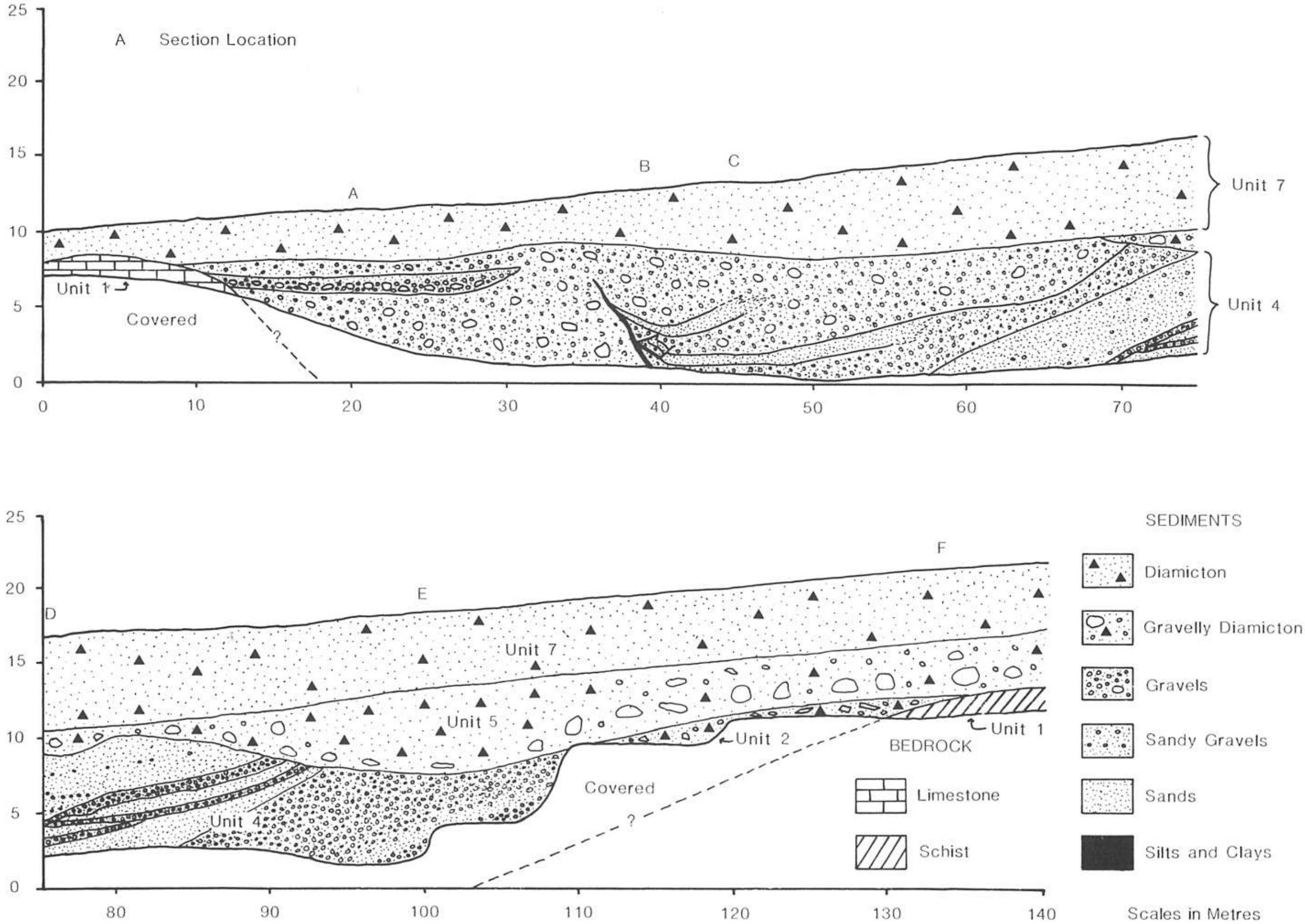

Figure 6-3-6. Major stratigraphic units exposed along the main highwall at the Ballarat mine.

MINEABLE GOLD-BEARING DEPOSITS

At least four stratigraphic units can be recognized in the deposits that are currently being mined. In general, gold increases with depth. The deepest and presumably the richest gravels were not exposed in October 1989. However, seismic results indicate that approximately 10 metres of gravel are present below the lowest exposed unit. Gold concentrations in the main pay gravels are up to approximately 2 grams per cubic metre (Figure 6-3-6; Units 3 and 4), 0.6 gram per cubic metre in gravelly diamictons (Units 2 and 5), and 0.1 to 0.2 gram per cubic metre in the upper diamicton (Bert Ball, personal commumication, 1989). The upper diamicton (Unit 7) is presently considered to be overburden but may be mined in the future with improved gold recovery techniques. Gold recovered is predominantly fine grained, with 99 per cent smaller than 16 mesh (1.19 millimetres). The gold has a fineness of about 840.

LOWER DIAMICTON (UNIT 2)

Discontinuous diamicton beds (Unit 2) overlie the bedrock highs along south and north sides of the Ballarat mine. These deposits are thin (<2 to 3 metres thick) and are restricted to areas adjacent to bedrock highs. The diamictons are matrix to clast supported with 50 to 80 per cent clasts. Locally derived angular schist, phyllite, quartzite and limestone dominate, but some subrounded to rounded pebbles are also present. The matrix is a sandy silt and contains abundant comminuted local bedrock material. The diamictons have a crude sub-horizontal stratification defined by horizontal tabular clasts, locally occurring in boudinaged and folded beds, and thin (<10 centimetres), tabular to trough-shaped lenses of fine sand. The sands are well sorted, have horizontal to wavy laminations and are locally interbedded with silts. Some beds directly overlying bedrock consist entirely of brecciated schist and, in places, vein quartz.

INTERPRETATION

The diamicton beds are interpreted to be debris-flow deposits derived from local bedrock highs. The clast lithology points to a local bedrock source that was probably originally disaggregated by *in situ* physical and chemical weathering. Previously rounded clasts were also incorporated into the flow deposits. Poorly sorted sediments with a fine matrix, disorganized fabric and gradational bed contacts are typical of modern debris-flow deposits (Bull, 1972;

Kochel and Johnson, 1984). Folded and boudinaged beds of local material indicate only minor transport. Stratified sands and silts were probably deposited by fluvial activity between debris-flow events.

Lower Gravels (Unit 3)

The lowest exposed gravels (Plate 6-3-6; Unit 3, not shown in Figure 6-3-6) in the Ballarat pit are moderately sorted, clast-supported, very compact pebbly gravels generally with a fine to medium sand matrix. Some beds exhibit planar and channel-fill crossbedding, and scoured lower contacts (<50 centimetres deep and 2 to 5 metres wide). These gravels generally dip 5° to the south and are unconformably overlain by large cobble to boulder gravels.

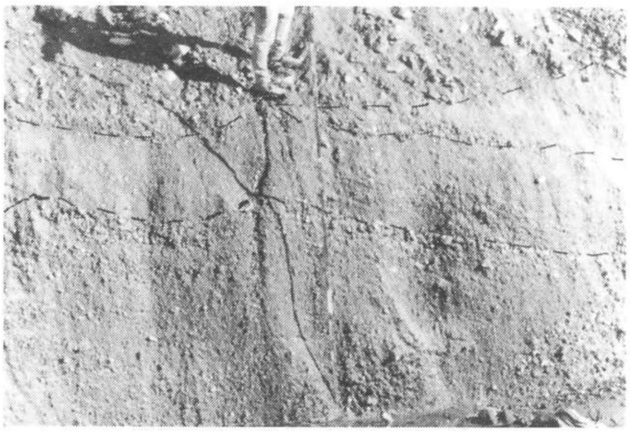

Plate 6-3-6. Lowest exposed gravels at the Ballarat mine (Unit 3). Note the stratification and scoured bed contacts (outlined).

INTERPRETATION

The degree of sorting, stratification and imbrication of the gravels suggests that they were fluvially deposited. Low-angle planar crossbeds probably represent deposition along the channel margins or as bar foreset beds. Scoured contacts indicate locally channelized flows, possibly in a braided-river system. The lithologic variability of these gravels (Figure 6-3-7) suggests that they were not derived solely from a local tributary but rather were deposited in the main valley system.

Middle Gravels and Sands (Unit 4)

A complex unit of sands and gravels (Unit 4), stratigraphically overlying Unit 3, is exposed along the main pit highwall. Lithologically, these gravels are composed almost entirely of quartzite, phyllite and schist.

At the base of the highwall (Figure 6-3-6, Section E), sand and gravel beds dip 25° north. Sandy beds are about 50 centimetres thick and consist of planar cross-stratified medium to coarse sand grading up into trough crossbedded medium sands and horizontally laminated silts. Convoluted bedding, normal faults, and load and flame structures are common. Gravel beds are up to 3 metres thick and are poorly

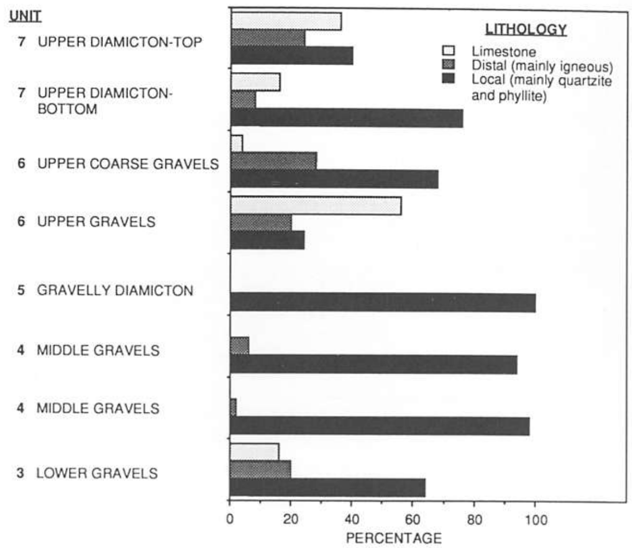

Figure 6-3-7. Lithology of main gravel and diamicton units at the Ballarat mine.

sorted and massive to crudely planar cross-stratified. Clasts are mainly small to large pebbles with few cobbles.

In the centre of the section (Figure 6-3-6, Section D; Plate 6-3-7) sandy units are 1 to 2 metres thick and dip consistently about 25° to the north. The dip of some beds decreases down-dip. Beds are 1 to 10 centimetres thick, laterally traceable for several metres, and mainly massive or normally graded. Thicker beds exhibit wavy parallel laminae with minor trough crosslaminae and low-angle climbing ripples. There are some small (10 cubic centimetres) angular inclusions of sandy gravel. Gravel beds are up to 30 centimetres thick, poorly sorted and massive. Bed contacts are sharp and conformable although lower contacts are locally scoured or marked by injection structures or sand intraclasts. Gravel beds generally pinch out down-dip or are eroded by more steeply dipping contacts of overlying units. Some gravel beds grade up-dip into the overlying gravel unit.

At Section C (Figure 6-3-6), moderately sorted, coarse sand and pebble-gravel beds occur in a deformed unit with a generalized trough shape. Poorly developed bedding in the sands is defined by textural changes between internally stratified sand beds up to 15 centimetres thick and sandy pebble beds up to 10 centimetres thick. Gravel beds are 5 to 20 centimetres thick and appear massive. Coarser beds are more poorly sorted, chaotic and in places they grade into pebbly sands. Beds either grade laterally into adjacent deposits or are eroded by overlying units. At the south end of this deformed zone, the sediments are generally better sorted, beds are thicker (up to 2 metres), and bed boundaries are sharper than to the north. The intensity of deformation increases to the north where the deformed zone is bounded by a steeply dipping lens of massive to poorly laminated silt and fine sand (Figure 6-3-6). Beds south of the silt lens are cut by reverse faults, folded and dip up to 35° to the south. The silt lens is folded, branched, locally boudinaged and characterized by soft-sediment deformation structures.

North of the silt lens (Section A, Figure 6-3-6), this unit consists mainly of a clast-supported, poorly sorted, sandy

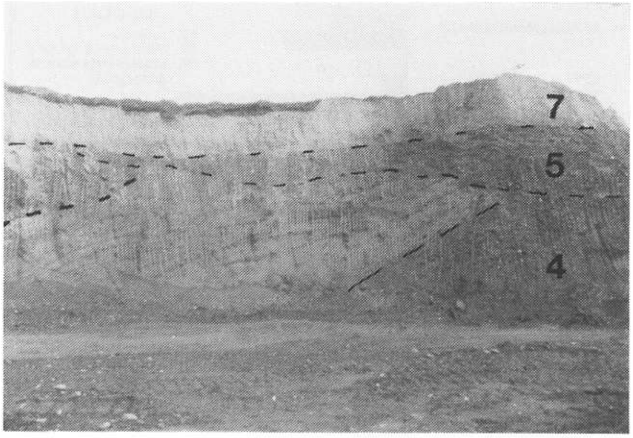

Plate 6-3-7. Steeply dipping sands and gravels in Unit 4 overlain by alluvium (Unit 5) and till (Unit 7). Section is about 15 metres high.

pebble-gravel with no apparent stratification. Cobble and boulder clusters up to 1 metre high and 3 metres wide are present. The clusters are clast supported and have a matrix that varies from moderately sorted sand to silty diamicton, locally with poorly developed convoluted laminae. Moderately sorted, locally openwork pebble-gravels occur adjacent to the clusters. In places the gravels grade into matrix-supported silty diamicton with clasts that are commonly angular, of local origin, and randomly oriented. The upper part of this unit contains a lens of poorly sorted gravel, 20 metres wide and up to 1.5 metres thick, that fines upward from cobbly gravel to medium pebble-gravel to coarse sand. The lens exhibits crude horizontal stratification and is trough shaped.

INTERPRETATION

Lithologic analysis of the gravels in this unit (Figure 6-3-7) and the general northerly dip of beds, indicate that they were derived almost entirely from a small tributary drainage to the south. The characteristically steep and consistent dip of beds and the topographic high to the south suggest a fan-delta environment. The lateral continuity of strata in this unit are suggestive of delta or fan-delta foreset beds. Massive, steeply dipping gravel beds with sharp planar contacts are typical of foreset beds. The foreset gravel beds pinch out or flatten in the down-dip direction and grade up-dip into overlying (topset) gravel beds. Massive and normally graded sands, horizontally laminated silts and fine sands, and climbing ripples are common in subaqueous environments. Load, flame and injection structures indicate rapid deposition onto saturated sediments. Sand intraclasts, local concave scouring, convoluted bedding, faults, and other deformation structures probably formed as a result of syndepositional slumping on over-steepened foresets. Massive to normally graded sands and gravels may reflect small sediment gravity-flow deposits which grade downslope into sand and silt beds. Angular gravel inclusions at the base of some sand beds are probably rip-up clasts.

Poorly sorted gravel and diamicton beds in this unit are interpreted as gravelly debris-flow deposits (*c.f.* Larsen and Steel, 1978; Burgisser, 1984). Disorganized to weakly imbricated, large clast clusters, such as those described above, form during the waning stage of high-discharge events (Brayshaw, 1984).

Deformation at the north end of the main highwall may be glaciotectonic in origin. High pore-fluid pressures in saturated sediments would result in deformation structures similar to those described. Compressive deformation would be expected along the margins of a glacier advancing from the east. This explanation is consistent with the observed decrease in deformation structures to the south.

UPPER GRAVELLY DIAMICTON (UNIT 5)

This unit is exposed only at the south end of the main highwall (Figure 6-3-6). The diamicton is clast supported and poorly sorted with a matrix of coarse sands and minor fines. Clasts are up to small cobble size and are weakly to moderately imbricated. The b-axes of tabular and disk-shaped clasts tend to dip to the southwest (Figure 6-3-8a). Some of the larger clasts have an irregularly laminated silt and clay armour up to 2 centimetres thick. Crude horizontal beds, up to 1 metre thick, exhibit normal grading with a poorly defined layer of cobbles to small boulders grading up into pebble gravels. Some beds display a thin, inversely graded zone at their base. Minor open work, moderately well sorted pebble beds occur. A large lens of diamicton occurs within the upper part of Unit 5. It is trough shaped and has gradational lower and upper boundaries.

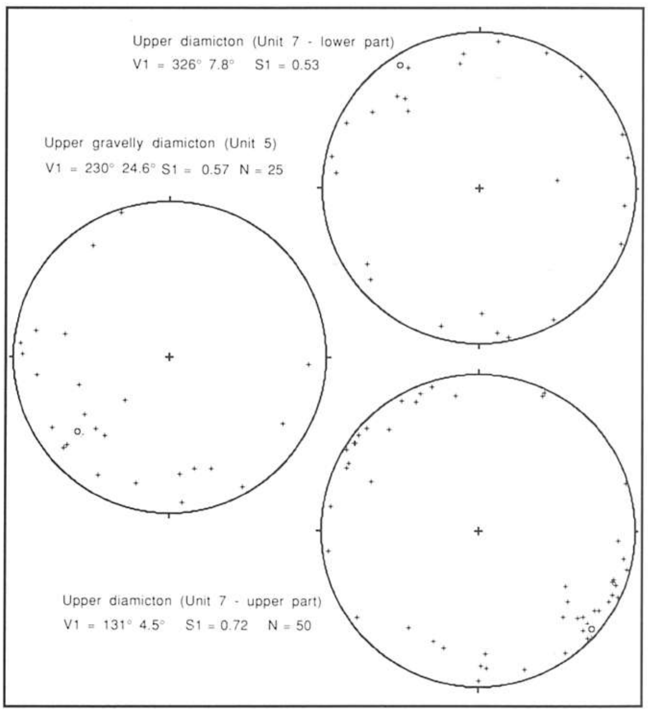

Figure 6-3-8. Stereo plots of the dip and dip direction of the intermediate axes of blade and disk-shaped clasts in gravelly diamicton (Unit 5), and the trend and plunge of the a–axes of elongated clasts in the upper and lower parts of Unit 7. V1 = orientation of maximum clustering (shown with a small circle) and S1 = normalized eigenvalue for V1.

INTERPRETATION

Diamicton beds in Unit 5 are interpreted as debris-flow deposits. Poor sorting and normal grading are typical and inverse grading is known to occur at the base of some debris-flow deposits (Burgisser, 1984). Indistinct stratification and weak imbrication have been observed in modern analogues (Kochel and Johnson, 1984). A preferred plunge of clast b-axes forming a girdled fabric (Figure 6-3-8) is also typical. Sorted beds indicate some fluvial activity, probably between debris-flow events. Lithologic and imbrication data (Figures 6-3-7 and 8) indicate that the flows originated from the southwest, and may therefore be associated with an alluvial fan.

UPPER GRAVELS (UNIT 6)

In the centre of the mine site the lower gravels (Unit 3) are unconformably overlain by chaotic to weakly imbricated, poorly sorted, large cobble to boulder gravels with a sand to pebble matrix (Unit 7, not shown on Figure 6-3-6). The coarse gravels are crudely stratified with beds up to 2 metres thick that dip 5° to 10° to the south. Bed boundaries are discontinuous and gradational. Clasts are mostly angular to subangular, especially near bedrock highs, but rounded cobbles and boulders dominate some beds. Stratification becomes more horizontal in the upper parts of the unit and the number of striated clasts increases toward the top.

INTERPRETATION

The coarse size of clasts, presence of striations, and high proportion of distally derived clasts (Figure 6-3-7) in these gravels suggests that they are glaciofluvial in origin. Deposits of this type are typical of proximal, braided gravel-bed rivers in glacial environments (Church and Gilbert, 1975). Gravelly braided streams are characterized by crude subhorizontal stratification (Hein and Walker, 1977). The increase in striated clasts towards the top of the unit may indicate the proximity of glacial ice, an interpretation which is further supported by the overlying diamicton, interpreted as till.

DIAMICTON (UNIT 7)

Unit 7 is generally 2 to 8 metres thick and consists primarily of massive, dense, matrix-supported diamicton. Clasts, often striated, are up to cobble size with rare boulders. They are mostly angular to subrounded with shattered and broken clasts common. The matrix is sandy silt. The lower contact of the diamicton is sharp and planar to gently undulatory.

The diamicton is highly variable in its sedimentary nature. At the south end of the main highwall (Figure 6-3-6) there are about 70 centimetres of interbedded silty diamicton and sandy silts at the base of the unit. Silt beds are horizontally laminated and laterally traceable for about 3 metres. They have few pebbles and some fine sandy laminae. Lineations of unknown origin on one bed trend at 160°. Irregular lenses and beds of poorly sorted coarse gravel up to 30 centimetres thick occur at the top of this unit and grade into the overlying massive diamicton.

In exposures west of the main highwall, Unit 7 consists of 4 metres of massive diamicton overlain by 2 metres of stratified diamicton and another 2 metres of massive diamicton. Diamicton beds within the stratified zone are 0.03 to 1 metre thick and vary in clast content and matrix texture. Contacts between beds are sharp and planar. Locally beds are separated by laminated clays, silts and fine sands with minor ripple-bedded sands. Some beds are deformed with clay beds containing rounded silt intraclasts and sand and silt beds with convoluted laminae. Beds are laterally traceable for two to several metres and have interbedded upper and lower contacts. Irregular gravelly lenses occur locally. Clast fabric data from the lower part of Unit 7 (Figure 6-3-8) indicate a strong preferred orientation of the long axis of clasts parallel to the main valley trend (northwest). Higher up in the diamicton clasts are more randomly oriented (Figure 6-3-8). The proportion of distal to local clasts decreases with depth (Figure 6-3-7).

INTERPRETATION

The massive diamicton of Unit 7 is interpreted as till deposited at the base of an over-riding glacier. Its massive, dense, matrix supported character is consistent with this interpretation. Diamictons with these characteristics, as well as with numerous striated and fractured clasts, a fine-grained matrix, a basal enrichment in local clasts, a strong a-axis fabric parallel to the main valley, and a sharp, planar lower contact, are typical of basal lodgement and meltout tills in mountain regions (Levson and Rutter, 1988 and references therein). Thin lenses of laminated clay, silt and sand probably are the result of pond sedimentation in small cavities at the glacier base. Diamictons interbedded with these sorted sediments were probably deposited as small debris flows within the cavities or by meltout of debris from the glacier base. Poorly defined gravel lenses may have been deposited in small cavities by subglacial streams. The random orientation of clasts in the upper part of the diamicton may be a result of resedimentation during postglacial times. Colluvial activity did not disrupt primary depositional characteristics in the lower part of the till.

GEOLOGIC SUMMARY – BALLARAT MINE

The sequence of deposits exposed at the Ballarat mine suggests that the following geologic events occurred during the late Quaternary. Prior to the last glaciation in the area, a bedrock-incised channel was occupied by a braided river that deposited the lower gravels (Unit 3). Coeval sedimentation along the margins of the channel was dominated by locally derived debris-flow deposits (Unit 2). Possibly as a result of the onset of glaciation, drainage in the vicinity of the mine was impeded, allowing development of a small fan-delta (Unit 4). Deposition of steeply dipping gravel and sand foreset beds initially dominated infilling of the channel. The upper gravelly diamicton (Unit 5) was deposited largely by a series of debris flows derived from the highlands to the south. The increase in locally derived debris-flow material, possibly due to the reduction in vegetative cover associated with glacial conditions, resulted in the progradation of alluvial fan sediments over the area. Coarse, glaciofluvial gravels (Unit 6) and basal lodgment and meltout tills (Unit 7) were deposited with the advance of glaciers. Some deformation of the underlying sediments occurred with the advance of ice.

Sorted sediments were deposited locally in subglacial cavities. Subsequent to deglaciation, resedimentation by colluvial processes was restricted to the upper part of Unit 7.

IMPLICATIONS OF BALLARAT CASE STUDY

Economic gold-bearing placer deposits at the Ballarat mine were deposited in a variety of sedimentary environments. The various units exposed are characterized by significant differences in grain size, sorting, clast roundness and stratification. Sedimentological interpretations indicate deposition in braided stream, deltaic and alluvial fan environments. Detailed sedimentological and stratigraphic analyses are required to understand the complex geologic origin of placer deposits of this nature. This information is necessary to determine the relationships of pay-streaks to sediment facies and to help project pay-zones into unmined areas.

Presently, economically viable gold occurs only in reworked sediments that were deposited prior to the last glaciation. The richest deposits are preglacial fluvial gravels with gold contents of about 2 grams per cubic metre. Mineable gold-bearing deposits of colluvial and alluvial origin produce about 0.6 gram of gold per cubic metre. Gold concentrations in till and enclosed sediments at the Ballarat site are low (0.1 to 0.2 gram per cubic metre), but may be worth processing with improved recovery systems.

Other bedrock channels along the margins of present valley bottoms may represent largely untapped sources of placer gold. Similar bedrock channels to that described at the Ballarat mine may occur throughout the Cariboo region and elsewhere in British Columbia. Exploration activities should focus on regional airphoto interpretation, detailed seismic cross-sections, large-diameter drilling, and other methods for the identification and evaluation of these hidden placer deposits.

CONCLUSIONS

Information gathered from geologic studies at active placer mines in the Cariboo region has several important implications for the placer industry. Exploration and production of deeply buried placers may be facilitated in some areas where natural processes have removed the overburden, such as along former glacial meltwater drainage courses. Preglacial and interglacial placers, deeply buried by till deposits, may be mineable in areas where detailed stratigraphic and sedimentological information is available to trace gold-bearing strata. In addition, some high-elevation buried-channel placer deposits, such as on Spanish Mountain, have a relatively thin glacial overburden and good gold concentrations in their upper part, and can therefore be productively mined. The potential for placer gold discovery in buried channels is good in the Cariboo and other traditional placer mining areas throughout the province. Stratigraphic and sedimentological studies of existing exposures are needed to help understand the paleodrainage patterns of preglacial and interglacial rivers and thereby identify probable gold-bearing buried channels.

Detailed geologic studies at the Ballarat mine site have identified several stratigraphic units with distinct sedimentologic characteristics. Gold occurs in economically viable qnantities in all of the sediments except the capping diamicton deposits. Comparisons of sediment facies at the Ballarat mine with the results of ongoing studies at other mine sites will lead to a better understanding of the relationships between gold concentrations and sedimentary environments. The analysis of sedimentary environments at the Ballarat mine illustrates the potentially complex geologic setting of placer gold deposits in glaciated areas. An understanding of the sedimentary origin of existing placers is needed to identify other sites where gold-bearing placers have been deposited and preserved from subsequent erosion. The Ballarat site also highlights the potential for locating new placer deposits in buried channels even in heavily exploited, traditional placer areas.

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NOTES

PALYNOLOGICAL DATING OF SEDIMENTS ASSOCIATED WITH PLACER GOLD DEPOSITS IN THE BARKERVILLE – QUESNEL – PRINCE GEORGE REGION, SOUTH-CENTRAL BRITISH COLUMBIA*
(93G)

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KEYWORDS: Economic geology, placer deposits, palynology, dating, correlations, Tertiary, Fraser Bend Formation.

INTRODUCTION

Placer deposits have been worked for many years in the Barkerville–Wells–Quesnel–Prince George region of British Columbia, with some yielding valuable amounts of gold. Throughout this long history, no direct evidence regarding the age of the placer deposits has been documented. It has been assumed by many geologists that the gold is associated with Pleistocene deposits, and a few have suggested that the placer gold was recycled from older Tertiary deposits, or that the deposits are quite variable in age, ranging from the late Tertiary through the Pleistocene (Clague, 1987; Eyles and Kocsis, 1988).

To attempt to resolve the uncertainties in dating, we proposed to apply palynological analyses to the placer deposits. The results reported here are the half-way record; it will take another year to zero in on the final picture. However, the preliminary results are interesting, and provide incentive to continue the search for the final answer.

RESULTS

Samples of the sediments were obtained from the localities numbered in Figure 6-4-1. Channel samples were taken from outcrop exposures to ensure a composite recovery of palynomorph facies. One of the first observations was that all of the outcrops examined contain volcanic clasts or are composed of volcanic ash, which was confirmed in the laboratory by identification of shards in microscopic preparations. From experience in this region, the occurrence of volcanics is limited to Tertiary successions. Palynologically, further support is given to a Tertiary age by the occurrence of *Cedrus perialata*, which is known to have become extinct in North America in the late Miocene (Rouse, 1977; Rouse and Mathews, 1979). In all of these sample analyses, care was taken to separate out any palynomorphs that may have been recycled from older deposits. Below is a list of palynomorphs recovered:

Conifer pollen:
- *Cedrus perialata*
- *Pinus haploxylon*–type
- *P. diploxylon*–type
- *P.*–3 (Piel)
- *P.*–4 (Piel)
- *Picea grandivescipites*
- *Podocarpus* **sp.**
- *Keteleeria* **sp.**
- *Pseudotsuga* **sp.**
- *Tsuga heterophyllites*
- *T. igniculus*
- *Juniperus* **sp.**

Angiosperm pollen:
- *Betula claripites*
- *Myricipites dubius*
- *Alnus vera*
- *cf. Quercus* **sp.**

Fungal spores:
- *Granatisporites* **sp.**
- *Dicellaesporites "taperensis" n.* **sp.**
- *Reduviasporonites anangus*
- *Multicellaesporites* **sp.**
- Tetracellate fungal spore
- *"Circulosporites"* **sp.**
- Fungal hyphae E (Norris)

Lycopods and Ferns:
- *Osmunda irregulites*
- *Lycopodium* **sp.**

Algal Cysts:
- *Ovoidites–1*
- *O.–2*
- *Cyclopsiella* **sp.**
- *Cf. Lejeunia* **sp.**

It is clear that the placer sediments sampled to date are all of late Tertiary age: they correspond to the Fraser Bend assemblages originally assigned to the late Middle Miocene, *circa* 17 to 13.5 Ma., Barstovian on the mammalian fossil printout (Rouse and Mathews, 1979; page 435). Equivalent outcrops are in the highest beds below the late glacial tills northwest of Quesnel. Hence, the general stratigraphic picture is late Tertiary sediments forming the upper stratigraphic component of the placer sequence.

During the next field season, sampling of outcrop exposures will be done on the Willow and Cottonwood rivers, because we feel that the best recovery of distinctive palynomorphs will be found there.

* This project is a contribution to the Canada/British Columbia Mineral Development Agreement.

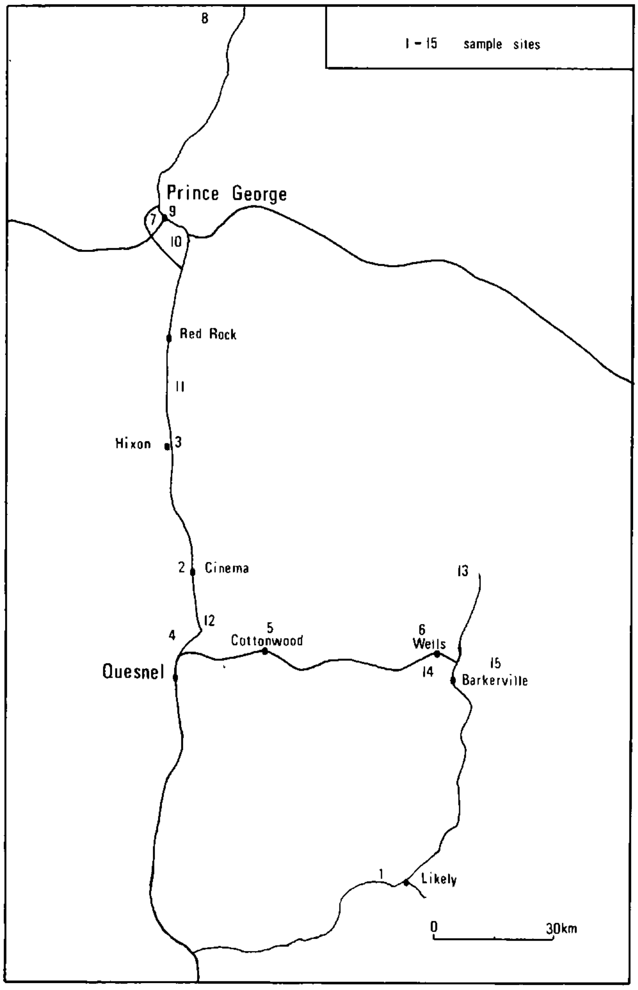

Figure 6-4-1. Distribution of collection sites in placer gold deposits.

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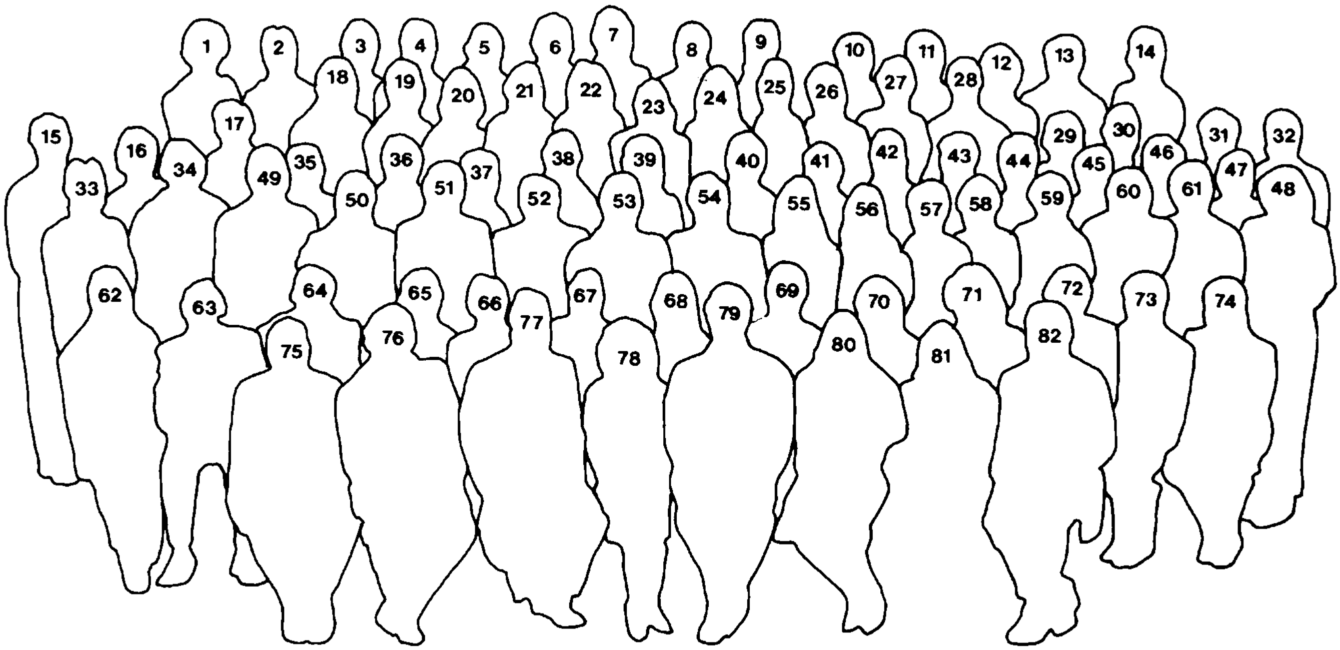

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