Field trip guidebook to the Upper Fir carbonatite-hosted Ta-Nb deposit, Blue River area, east-central British Columbia

A.S. Rukhlov, T.C. Chudy, H. Arnold, and D. Miller

British Columbia Geological Survey GeoFile 2018-6

Front cover: Sill-like body of calcite-carbonatite, Howard Creek. Photo by A.S. Rukhlov.

Back cover: Dolomite carbonatite with aligned fluorapatite megacrysts and dark-toned ferrikatophorite prisms set in a recrystallized ferroan dolomite matrix that readily oxidizes brown-red; BS-1 cut, Upper Fir. Photo by A.S. Rukhlov.

This publication is available, free of charge, from the British Columbia Geological Survey website: www.empr.gov.bc.ca/geology

This guidebook was initially prepared for a field trip sponsored by the Mineralogical Association of Canada as part of the Resources for Future Generations conference held in Vancouver, June 2018.
Field trip guidebook to the Upper Fir carbonatite-hosted Ta-Nb deposit, Blue River area, east-central British Columbia

A.S. Rukhlov
T.C. Chudy
H. Arnold
D. Miller

Ministry of Energy, Mines and Petroleum Resources
British Columbia Geological Survey
GeoFile 2018-6
Field trip guidebook to the Upper Fir carbonatite-hosted Ta-Nb deposit, Blue River area, east-central British Columbia

A.S. Rukhlov1a, T. C. Chudy2, H. Arnold1, and D. Miller1

2Department of Earth, Ocean and Atmospheric Sciences, The University of British Columbia, Vancouver, B.C., V6T 1Z4
aCorresponding author: Alexei.Rukhlov@gov.bc.ca


Abstract
Carbonatites are igneous rocks containing abundant primary carbonate minerals. These rare rocks generally form in intracratonic settings as part of crustal-scale dome and rift systems. Historically regarded as petrogenetic curiosities, recent interest in strategic metals has led to significant exploration for carbonatites. In the Canadian Cordillera, carbonatites were emplaced episodically, at ca. 810-700, 500, and 360-330 Ma, forming part of the British Columbia alkaline province, which defines a long (at least 1000 km), narrow (ca. 200 km) orogen-parallel belt. The ca. 810-700 Ma and 500 Ma carbonatites were injected during protracted breakup of the supercontinent Rodinia and passive margin development on the western flank of Laurentia. In contrast to these and to most carbonatites globally, the 360-330 Ma carbonatites, such as the Blue River area examples, are unusual. They were emplaced near the continental margin during subduction rather than in the cratonic interior during continent breakup. The Blue River carbonatites include at least 18 carbonatite and 2 alkaline, silica-undersaturated-rock occurrences. This field trip considers the characteristics, magmatic evolution, and mineralization of the Blue River carbonatites as represented by the Upper Fir complex, which hosts one of the largest and best studied Nb-Ta deposits in the Canadian Cordillera. Participants will examine outcrops and drill-core sections of amphibolite-grade metacarbonatites, related metasomatic rocks, syntectonic pegmatites, enclosing pelites and amphibolites of the Mica Creek assemblage (750-550 Ma), and Mesozoic-Cenozoic Cordilleran structures. We consider the tectonometamorphic overprinting of igneous features in the Upper Fir carbonatites, as recorded by paragenetic relationships, mineral chemistry, recrystallization, and retrograde deformation.

Exploration by Commerce Resources Corporation at the Upper Fir carbonatite complex established an NI 43-101-compliant resource of 48.4 million tonnes (Indicated) averaging 1,610 ppm Nb2O5 and 197 ppm Ta2O5 plus 5.4 million tonnes (Inferred) averaging 1,760 ppm Nb2O5 and 191 ppm Ta2O5. Sill-like carbonatite bodies up to 72 m thick are in isoclinal, recumbent, similar folds (in part rootless) that are refolded by southwest-vergent, open folds with parallel geometries. The deposit has been traced for about 1450 m north-south and about 800 m east-west and consists of mainly dolomite carbonatite and subordinate (<4%) medium- to coarse-grained calcite carbonatite. Calcite carbonatites form lenses up to a few m thick within dolomite carbonatite bodies. The carbonatites show contact-parallel foliation concordant with penetrative schistosity and compositional layering in the enclosing gneisses, schists, and amphibolites. Internal retrograde breccia and shear zones further obscure the original paragenetic record. The carbonatites are recrystallized, and display diverse fabrics. Granoblastic varieties, likely representing magmatic protoliths, occur mainly at the margins of the carbonatite bodies, whereas fine-grained, well-foliated and porphyroclastic varieties typically make up the central portions. This distribution correlates favourably with the composition of amphibole-group minerals and Nb-Ta oxide phases, suggesting that primary compositional zoning within individual dolomite carbonatite bodies controlled rheology and thus strain partitioning during deformation. The Nb-Ta mineralization consists of predominantly ferrocolumbite and pyrochlore-supergroup minerals.

Isotopic evidence and spatial and temporal associations with large igneous provinces containing ultramafic and alkaline silicate rocks indicate derivation of carbonatite magmas from sub-lithospheric mantle plumes. New isotopic and elemental compositions of minerals from Blue River carbonatites and related rocks are indistinguishable from worldwide carbonatites generated by such deep-mantle plumes. The 360-330 Ma Cordilleran examples formed along the western margin of Laurentia while subduction was taking place immediately to the west. Lithospheric extension related to this Late Paleozoic subduction is considered responsible for rifting the continent margin and initiating the Slide Mountain ocean as a back-arc basin. We suggest that this same back-arc extension triggered emplacement of the Blue River carbonatites, which were derived from a long-lived, deep-level mantle plume that was tapped episodically since the Neoproterozoic.

Keywords: carbonatite, British Columbia alkaline province, mantle plume, Blue River, Upper Fir, Ta-Nb deposit, pyrochlore supergroup, ferrocolumbite, molybdenite

British Columbia Geological Survey GeoFile 2018-6
1. Introduction

Carbonatites are rare magmatic rocks that contain abundant primary carbonate minerals, at least 30% (Mitchell, 2005) or 50% (Le Maitre, 2002). Given their extreme geochemistry, carbonatites are unlike other igneous rocks (Tuttle and Gittins, 1966; Heinrich, 1966; Le Bas, 1977; Woolley, 1987; Bell, 1989; Bell and Keller, 1995; Kogarko et al., 1995; Bell et al., 1998; Wall and Zaitsev, 2004). Carbonatites are generally found in intracratonic regions, such as crustal-scale domal swells and related rift systems (e.g., East African rift valley, Maimecha-Kotui and Kola alkaline provinces). Of more than 530 known occurrences, only two are from oceanic islands, the Cape Verdes and Canary Islands (Woolley and Kjarsgaard, 2008a), which emphasizes the important, yet still enigmatic, role of continental lithosphere in their genesis. Most carbonatites are spatially and temporally associated with much larger volumes of ultramafic and alkaline silicate rocks, suggesting genetic relationships between undersaturated silicate and carbonate magmas (Woolley, 2003; Mitchell, 2005; Woolley and Kjarsgaard, 2008b). Carbonatite magmas can form as low-degree partial melts of a carbonated mantle, as the products of fractional crystallization, or as immiscible liquids separated from carbonated silicate magmas (e.g., Wyllie, 1989). However, their ultimate mantle sources (lithospheric versus sub-lithospheric) remain controversial (e.g., Gerlach et al., 1988; Nelson et al., 1988; Gittins and Harmer, 2003; Bell and Rukhlov, 2004; Bell and Simonetti, 2010; Woolley and Bailey, 2012; Rukhlov et al., 2015; Hulett et al., 2016).

Isotopic evidence provides firm constraints on the origin of these unusual rocks. Found on all continents, carbonatites are well-suited for tracing secular mantle evolution. This is because their extreme enrichment in Sr and REEs, especially light REEs (LREE), buffers Sr and Nd isotopic compositions from changes due to contamination, and because they span ages from 3.0 Ga to the present (Bell et al., 1982; Bell and Blenkinsop, 1987a; Nelson et al., 1988; Bell and Rukhlov, 2004; Rukhlov et al., 2015). The Sr, Nd, Hf, and Pb isotopic signatures of carbonatites worldwide are similar to those of oceanic island basalts (OIBs). The mixing arrays defined by carbonatite data in multi-isotopic space rule out any oceanic island basalts (OIBs). The mixing arrays defined by carbonatite data in multi-isotopic space rule out any oceanic island basalts (OIBs). The mixing arrays defined by carbonatite data in multi-isotopic space rule out any oceanic island basalts (OIBs). The mixing arrays defined by carbonatite data in multi-isotopic space rule out any oceanic island basalts (OIBs). The mixing arrays defined by carbonatite data in multi-isotopic space rule out any oceanic island basalts (OIBs).
contemporaneous silicate rocks typically form regional swarms of individual occurrences across areas of 1000 km$^2$ (e.g., Blue River area; Pell, 1994; Mitchell et al., 2017). Individual occurrences such as the Upper Fir complex form highly strained and recrystallized tabular structures that may have originated as sills. These structures are a few dm to several tens of m thick and extend up to several hundred m along strike. They are typically composed of both calcite carbonatite and dolomite carbonatite end members in variable proportions, commonly with calcite-dolomite hybrids. Carbonatites rich in apatite and silicate and oxide minerals could be classified as phoscorites (Russell et al., 1954; Krasnova et al., 2004) or silicocarbonatites (>20 wt% SiO$_2$; Le Maitre, 2002).

This field trip considers the characteristics, magmatic evolution, and mineralization of the Blue River carbonatites as represented by the Upper Fir carbonatite complex (330 Ma), which hosts one of the largest and best studied Nb-Ta deposits in the Cordillera. Participants will examine drill-core sections (Appendix 1) and outcrops of metacarbonatites, related metasomatic rocks, syntectonic pegmatites, enclosing...
(semi)pelites and amphibolites of the Mica Creek assemblage (750-550 Ma) and Mesozoic-Cenozoic structures related to the Cordilleran orogeny. We will consider the tectono-metamorphic overprinting of igneous features in the Upper Fir carbonatites as recorded by paragenetic relationships, mineral chemistry, recrystallization, and retrograde deformation.

2. Field trip overview

The road log (Appendix 2) provides details of the stops and summarizes terranes along the route from Vancouver to Blue River. Throughout the text and in Appendix 2, the attitudes of planar fabrics are given as A/B°, where A is strike and B is dip in the direction indicated. A magnetic declination of 16.1° east is used.

The highway route from Vancouver to Blue River passes through the Coast and Cascade Mountains, the Interior Plateau, and the Columbia Mountains (Fig. 2), which are underlain by Proterozoic to Cenozoic rocks comprising several Cordilleran terranes. At stops along the way, participants will be able to observe outcrops of Late Paleozoic pillow basalts that represent remnants of the Slide Mountain ocean, which initiated by back-arc rifting during eastward subduction beneath the western margin of ancestral North America at the same time as carbonatite magmatism farther inland. We will also examine Pleistocene flood basalts.

The first part of this guidebook establishes the Upper Fir deposit in a global carbonatite context and summarizes the history of work in the Blue River area. We then briefly describe the geology and regional carbonatite occurrences of the Blue River area. The final part outlines the Upper Fir geology, mineralogy and geochemistry based on the findings of Commerce Resources Corp. (Kulla and Hardy, 2015), Chudy (2013) and ongoing studies of the Blue River carbonatites (Rukhlov et al., 2018).

We encourage discussion of mantle sources, evolution of parental carbonate magmas, and the unusual mineralization. In a wider context, the field trip considers the possibility that the back-arc extension responsible for initiating the Slide Mountain ocean also stretched the continental lithosphere farther eastward, triggering emplacement of the Blue River carbonatites by tapping a long-lived (Neoproterozoic to Cenozoic) mantle plume (Rukhlov et al., 2018).

3. Upper Fir deposit in a global context

Extensive exploration by Commerce Resources Corporation determined an NI 43-101-compliant resource of 48.4 million tonnes (Indicated) averaging 1,610 ppm Nb_2O_5 and 197 ppm Ta_2O_5 plus 5.4 million tonnes (Inferred) averaging 1,760 ppm Nb_2O_5 and 191 ppm Ta_2O_5 (Kulla and Hardy, 2015). The main ore minerals include the Ta- and Nb-rich pyrochlore supergroup and ferrocolumbite, with minor fersmite and nyrobeschynite (Chudy, 2013). Abundant ferrocolumbite and the unusual Ta-rich pyrochlore (1.05–40.56 wt% Ta_2O_5, average 20.27 wt% Ta_2O_5), which attains the microlite composition (i.e. Nb+Ta > 2Ti and Ta>Nb), distinguish the Nb-mineralization at Upper Fir from that of most carbonatite deposits (Van der Veen, 1963; Bakes et al., 1964; Heinrich, 1966; Mariano, 1982, 1989; Hogarth, 1989; Hogarth et al., 2000; Simandl et al., 2002; Chudy, 2013; Mackay and Simandl, 2014; Chakhmouradian et al., 2015; Mitchell, 2015).

Most primary carbonatite Nb deposits (e.g., Araxá, Brazil; St. Honoré and Oka, Quebec; Aley, British Columbia) have similar or higher Nb grades (0.12–1.6 wt% Nb_2O_5) as those at Upper Fir. Lateritic profiles above primary deposits host the world’s richest deposits (e.g., Araxá, Catalão-I and II, and Seis Lagos, Brazil), with extreme Nb enrichment of up to 6.7 wt% Nb_2O_5 (Mariano, 1989; Chakhmouradian, 2006; Mackay and Simandl, 2014; Chakhmouradian et al., 2015; Mitchell, 2015). However, the tantalum contents at Upper Fir are distinctly higher (up to 8,705 ppm Ta; Gorham et al., 2011b), resulting in anomalously low Nb/Ta ratios (0.76-556; average 8.9) relative to most carbonatites (average Nb/Ta = 35; Chakhmouradian, 2006). Examples of other carbonatite deposits with significant Ta enrichment include Crevier in Quebec, Mount Weld in Australia, and Belaya Zima in Russia (Mariano, 1989, Mackay and Simandl, 2014; Chakhmouradian et al., 2015).

Although carbonatitic pyrochlores with up to 26.9 wt% Ta_2O_5 have been reported (Požharitskaya and Samoylov, 1972), they tend to have <5 wt% Ta_2O_5, with global averages of 0.85 wt% Ta_2O_5 for ‘normal’ pyrochlores and 7.92 wt% Ta_2O_5 for U-rich pyrochlores (>2.0 wt% UO_2), resulting in much higher Nb/ Ta ratios (typically 200-500) compared with pyrochlores at Upper Fir (Nb/Ta = 0.45-56, average 3.2; Mariano, 1982, 1989; Hogarth, 1989; Simandl et al., 2002; Chudy, 2013; Mackay and Simandl, 2015; Mitchell, 2015). Pyrochlores can have considerable variations in Ta content within a single carbonatite complex, and primary pyrochlore crystals can exhibit both depletion and enrichment in Ta from core to rim (Hogarth, 1989 and references therein; Mariano, 1989; Simandl et al., 2002; Chudy, 2013; Chakhmouradian et al., 2015; Mackay and Simandl, 2015).

The Upper Fir has some similarities to pegmatite-hosted rare-metal deposits, in particular the low Nb/Ta ratios, the sill-like morphology, and the occurrence in clusters or swarms (e.g., Černý, 1992; Černý and Erict, 2005). The magmatic evolution of rare-metal pegmatites is characterized by progressive Ta-enrichment and roughly concentric zoning due mainly to in-situ differentiation.

Sporadic molybdenite (up to 1.7 cm and up to ~2 vol.%) in carbonatites and fenites (Fig. 3), with up to >2,000 ppm Mo across 2 m (Gorham et al., 2011b) is another peculiar feature of the Upper Fir (Appendix 3f), given that carbonatites are generally poor in Mo (average 12 ppm) and lack molybdenite (Woolley and Kempe, 1989). Although little work has focused on molybdenite from Cordilleran carbonatites (Rukhlov et al., 2018), it has been observed in other carbonatite and related-rock occurrences in British Columbia, including Perry River, Mount Grace, Wicheeda Lake, and the Mount Copeland past producer (Fig. 1; Currie, 1976; White, 1982; Höy, 1988; Trofanenko et al., 2016). With Mo production from a unique carbonatite-hosted...
Fig. 2. a) Morphogeological belts of the Canadian Cordillera (after Gabrielse et al., 1991; Hickin et al., 2017). b) Composite physiographic units (after Mathews, 1986; Hickin et al., 2017). c) Metamorphic and plutonic rocks (after Monger and Hutchison, 1971; Read et al., 1991; Monger, 2014).

British Columbia Geological Survey GeoFile 2018-6
Late Paleozoic subduction is considered responsible for rifting of the continental margin and initiating the Slide Mountain ocean as a back-arc basin (e.g., Nelson et al., 2013). We suggest that this same back-arc extension triggered emplacement of the Blue River carbonatites that were sourced from a long-lived, deep-level mantle plume that was tapped episodically since the Neoproterozoic.

4. History of work in the Blue River area


Although not recognized as carbonatites in early studies (McCammon, 1951, 1953, 1955; Rowe, 1958), carbonatites in the Blue River area (Fig. 4) have been intermittently explored since 1950 for vermiculite, uranium, niobium, tantalum and phosphate. Discovered by homesteader O.E. French in 1949, the Verity carbonatite occurrences near the Lempriere station, north of Blue River were initially evaluated for vermiculite by the Zonolite Corporation (McCammon, 1951). In 1951, the British Columbia Department of Mines identified uranpyrochlore in a sample of radioactive dolomite carbonatite from Verity, sent to them by O.E. French (McCammon, 1953). The whole-rock analysis returned 0.2 wt% $UO_3$ and 1.7 wt% $Nb_2O_5 + Ta_2O_5$ (Mariano, 1982). Rowe (1958) also reported uranium-bearing columbite identified in carbonate specimens sent by O.E. French to the Geological Survey of Canada. Further regional prospecting by the French family resulted in the discovery of the Paradise Lake and other carbonate occurrences east of the Verity showing (Fig. 4). Between 1952 and 1955, more carbonate outcrops were found in the area, including the Mill occurrence north of Verity, during uranium exploration by St. Eugene Mining Corporation Ltd. (McCammon, 1955; Rowe, 1958).

The first government reports on the Blue River carbonatites by J.W. McCammon (Mineralogical Branch of the British Columbia Department of Mines) and R.B. Rowe (Geological Survey of Canada) provide many field and mineralogical details, along with the results of vermiculite screen-expansion tests and chemical analyses of whole-rock, uranpyrochlore, and uranpyrochlore samples (McCammon, 1951, 1953, 1955; Rowe, 1958). At that time, the idea of magmatic carbonates (Högblom, 1895; Brogger, 1921; Soellner, 1927) seemed extreme for most to accept, and highly influential petrologists dismissed “these carbonate-bearing rocks as nonigneous” (e.g., Bowen, 1921; Daly, 1933; Johannsen, 1938; Shand, 1943) or considered them to be “limestones remobilized at crustal levels” (Brogger, 1921). Hence, it is not surprising
that the early studies described the Blue River carbonatites as “bands and lenses of mainly dolomitized limestone” of “a deceptively igneous appearance”, with variable amounts of
vernicular, biotite, white mica, apatite, magnetite, olivine, ilmenite, green amphibole, zircon, pyrrhotite, pyrite, uranian pyrochlore, and uranian columbite (McCammon, 1951, 1953, 1955; Rowe, 1958). Locally cut by pegmatities and coarse-grained hornblende-calcite veins, these carbonatite bodies are up to 46 m thick and extend for at least 300 m along strike. They contain accessory pyrochlore, as pale brown to reddish brown to almost black octahedral crystals typically forming penetration twins or irregular lumps up to 4.5 cm in diameter and disseminated small grains, and amphibole and mica grains lying within the foliation of the enclosing gneisses; an outcrop of sodalite syenite was also noted (McCammon, 1953, 1955; Rowe, 1958).

Several companies and individuals, including Southwest Potash Corporation, Kenco Explorations (Western) Ltd., Falconbridge Ltd., Vestor Exploration Ltd., A.N. Mariano, A. Rich, and J.A. Gower examined carbonatite occurrences at Verity, Paradise Lake, and in the surrounding area between 1955 and 1970 (Mariano, 1982; Kulla and Hardy, 2015). During that time, A.N. Mariano was engaged in carbonatite exploration worldwide on behalf of Kennecott Copper Corporation. He came across a report on the Blue River claims that were briefly evaluated by Kenco Explorations (Western) Ltd. in 1964. Although the Kennco assessment did not find the property of economic interest, A.N. Mariano recognized some of the unique aspects in the structure, geochemistry, and petrology of the carbonatites and initiated a research program to “further understand the nature and genesis of the Blue River occurrences and to reveal the presence of materials of possible economic importance” (Mariano, 1982).

Earl D. Dodson prospected the Howard Creek area to the west of the Paradise Lake occurrences on behalf of Falconbridge Ltd. sometime before 1967, when A. Rich, then a student at the University of Alberta, recognized carbonatites at Howard Creek. Rich carried out an exploration program including geological mapping at Paradise Lake, regional prospecting, and sampling (Rich and Gower, 1968). From 1967 to 1969, A.N. Mariano and J.A. Gower, manager of exploration for Kenco Explorations (Western) Ltd. and Professor of Mineralogy at the University of British Columbia examined the Blue River carbonatites (Mariano, 1979, 1982). They named mineralogically unusual, titanite-rich, calcite-clinoxyroxene-amphibole rock intimately associated with carbonatites, ijolites, and nepheline syenites at Howard Creek ‘lemprierite’ after the Mount Lempiere at the Howard Creek headwaters (Fig. 5). Mariano (1982) pointed out that the undersaturated alkaline silicate rock-carbonatite association was at odds with a sedimentary origin of the Blue River carbonatites. Rich and Gower (1968) further confirmed a magmatic origin based on the first isotopic analysis of C, O, and Sr in carbonate and whole-rock fractions from carbonatites and nepheline syenites at Verity, Paradise Lake, and Howard Creek. Their analyses revealed high Sr contents (1,100-5,650 ppm), which are at least ten times that of average sedimentary or metamorphic carbonate rocks, δ13C,ppm values of -3.7 to -5.6, and δ18O,ppm values of -19.9 to -25.5 (within the range of those from carbonatites elsewhere), and #Sr/86Sr values of 0.7035 (Verity carbonatites) and 0.7047 (Paradise Lake nepheline syenite), which are lower than those of crustal rocks (>0.706). These data indicate the sub-crustal origin of the Blue River carbonatites and the associated alkaline silicate rocks (Rich and Gower, 1968).

From 1976 to 1978, John Kruszewski explored for uranium and niobium in the Verity and Paradise Lake areas with ground magnetic and radiometric surveys and more stripping and trenching (Meyers, 1977). In 1979, A.N. Mariano brought the Blue River carbonatites to the attention of Anschutz (Canada) Mining Ltd. Between 1979 and 1982, the company carried out extensive exploration mainly for Ta and Nb in the Blue River area, with road construction, topographic surveys, and 1:4,000-scale mapping, airborne and ground geophysics, and diamond drilling (Ahroon, 1979, 1980; Aaquist, 1981, 1982a-c). This work defined a resource of about 2.13 million tons of 0.126 wt% Nb2O5 and 0.020 wt% Ta2O5 for the Verity deposit and in the discovery of the Bone Creek and Fir mineralized carbonatite occurrences (Ahroon, 1980; Aaquist, 1982a,c).

As part of his research on global carbonatites in the 1960s, A.N. Mariano carried out a detailed petrological, mineralogical, and geochemical study of the Blue River carbonatites that was summarized in confidential reports for Anschutz (Canada) Mining Ltd. (Mariano, 1979, 1982). Intended as a guide to aid Ta-exploration in the Blue River area, Mariano’s (1982) report provided detailed field and petrographic descriptions of the Blue River carbonatites and related rocks, along with a wealth of lithogeochemical and mineral-chemistry data that used spark source mass spectrometry and in situ methods such as cathodoluminescence microscopy, electron probe microanalysis (EPMA) and scanning electron microscopy with energy-dispersive X-ray spectrometry (SEM-EDX) on polished thin sections. These techniques proved very efficient in Blue River carbonatite exploration. Particularly efficient was the use of rapid, qualitative SEM-EDX to identify Ta=U-rich pyrochlore, and cathodoluminescence microscopy or a shortwave UV luminescence to identify fenitization (e.g., apatite), which may be difficult to recognize in the field. Mariano (1982) found anomalously high Ta contents (locally coupled with high U) in pyrochlores from the Blue River carbonatites, which result in unusually low Nb/Ta ratios of about 6. These ratios are unlike most carbonatite pyrochlores which have Nb/Ta ratios of 100-500 (Van der Veen, 1963; Bakes et al., 1964; Heinrich, 1966). Mariano (1982; 1989) also considered pyrochlore, ferrocolumbite, and fersmite to be primary phases in the Blue River carbonatites.

In 1978, J.T. Morton and A.E. Grant discovered the Mud Lake carbonatite occurrence (AEG and JTM claims) based on an anomalous (27 times background) scintillometer reading at a reddish-brown soil exposed in a roadcut along the newly built Mud River logging road about 10 km northeast of Blue Lake. Various exploration programs, including the AEG, JTM, and Kennecott programs, were undertaken to survey the Mud Lake carbonatite occurrence (AEG and JTM claims) based on an anomalous (27 times background) scintillometer reading at a reddish-brown soil exposed in a roadcut along the newly built Mud River logging road about 10 km northeast of Blue Lake.
Fig. 4. Carbonatites and related rocks of the Blue River area (after Pell, 1994; Rukhlov and Bell, 2010; Millonig et al., 2012, 2013; Millonig and Groat, 2013; Rukhlov et al., 2018). Geology and metamorphic isograds after Campbell (1968), Simony et al. (1980), Raeside and Simony (1983), Pell and Simony (1987), McDonough and Murphy (1990), McDonough et al. (1991a, b, 1992), Digel et al. (1998), and Murphy (2007).
River. The soil formed above a dolomite carbonatite body more than 4 m thick and extending for at least 150 m along strike, (Morton, 1979; White, 1980). Both forsterite and chondrodite were identified by X-ray diffraction of a soil sample containing visually identified phlogopite, pyroxene, and magnetite. Semi-quantitative spectrographic analysis of fresh carbonatite yielded 2.0 wt% P, 0.3-1.0 wt% Sr, 0.03-0.1 wt% Nb, 116 ppm U, 19 ppm Th, and detectable Y and REE (Morton, 1979).

In 1979, G.P.E. White, District Geologist of the Kamloops mining district examined the Mud Lake carbonatite (White, 1980). His report provided further geological details and assay data and quoted an Rb/Sr date of 370 Ma for the Mud Lake carbonatite, which was obtained by R.L. Armstrong and W.J. McMillan at the University of British Columbia and indicated that the carbonatite is significantly younger than the country schists. White (1980) also pointed out that the Mud Lake carbonatite (metabasaltite) has anomalously high Sr contents, similar to those of the Three Valley Gap carbonatite near Revelstoke and to South African dolomite carbonatites. Subsequent studies by White (1981, 1982, 1985) provided many field, mineralogical, and geochronological details for the Blue River carbonatites. Some of the important findings were two U-Pb dates of 325 Ma on zircons from the Verity carbonatites (donated by B. French and analyzed by R. Parrish at UBC), and K/Ar dates of 205±8 Ma on phlogopite from the Howard Creek carbonatite, and 92.5±3.2 Ma and 80.2±2.8 Ma on richterite from the Verity carbonatite obtained by J. Harakal at the University of British Columbia. The K/Ar dates indicate metamorphic overprinting of the Late Mississippian carbonatites during the Cordilleran orogeny, consistent with evidence of transposition of the bodies to a regional foliation, at least three phases of folding, and upper amphibolite-grade (kyanite to sillimanite zone) metamorphism (White, 1982, 1985). Regarding their economic potential, White (1982) noted columbite and pyrochlore in many of the Blue River carbonatites, and an occurrence of molybdenite in drill core from the Fir carbonatite. White’s (1985) report also includes a preliminary geological map of the Howard Creek carbonatite-nepheline syenite complex.

Detailed mapping at a scale of 1:24,000 by the Geological Survey of Canada in 1987 and 1988 located two new carbonatite occurrences in the Mount Cheadle area: 1) Serpentine Creek, a <100 m² body east of the Verity occurrence; and 2) Gum Creek, a ~10 m thick layer east of the Bone Creek and Fir carbonatites (Digel et al., 1989). Johnstone et al. (1990a, b) followed up exploration of the Mud Lake carbonatites and related rocks with prospecting, geochemical and geophysical surveys, and detailed petrographic descriptions.

Systematic studies of carbonatites and related rocks in British Columbia by J. Pell, initially at the University of British Columbia and later at the British Columbia Geological Survey, provided further geological maps and general descriptions.
Fig. 5. Examples of carbonatites and related rocks of the Blue River area (c, e, f, h, i; photomicrographs, plane-polarized light; field of view 5 mm): a) Sill-like body of calcite-carbonatite, Howard Creek. b) Cumulate texture in magnetite-rich phoscorite, Howard Creek. c) Poikilopropylpyroxenoblastic texture in calcite-titanite-amphibole clinopyroxenite ('lemprierite'), Howard Creek. d) and e) Granoblastic texture in foliated, olivine- and tetraferriphlogopite-bearing, dolomite-calcite carbonatite, Serpentinite Creek. f) Zirconolite in foliated, phlogopite- and olivine-rich carbonatite, Switch Creek. g) and h) Well-foliated, calcite-bearing nepheline-syenite, Paradise Lake. i) Granoblastic texture in titanite-rich, exoco¨ntact fenite, Upper Fir.

of the Blue River rocks and new geochemical analyses and geochronology (Pell, 1985, 1987, 1994; Pell and Höy, 1989). Based on available ages, Pell (1987) attempted to correlate pulses of alkaline and carbonatite magmatism in the Canadian Cordillera with the tectonic evolution of the Pacific and Atlantic margins of North America. Pell (1987) also suggested that variable prospectively of the carbonatites for Nb, REE, and fluorine may reflect their depth of emplacement and hence the degree of related metasomatism (fenitization) based on structural level within the host metamorphic core complex (Frenchman Cap dome) and the younger strata of the Omineca and Rocky Mountain fold-and-thrust belts.

In 2000, prices for Ta raw material spiked several times relative to those prevalent in the 1990s (e.g., Mackay and Simandl, 2014). The favourable Ta market prompted several companies to re-evaluate known carbonatites and to search for new Ta-Nb and REE targets in the Blue River area. In 2002, International Arimex Resources Inc. evaluated the Gum Creek carbonatite for its Ta-Nb-P-REE potential using limited rock and stream-sediment sampling. Reeder and Grywul (2002), Grywul and Reeder (2003), and Thom (2013) described further exploration of the Mud Lake carbonatites and associated rocks.

Extensive exploration undertaken by Commerce Resources Corporation in the Blue River area between 2000 and 2012 included regional prospecting, detailed mapping (1:2,500 scale), road construction, trenching, bulk sampling, drilling, subsurface 3D modelling, metallurgical testing, ground and airborne geophysics, regional drainage heavy-mineral (pan) concentrate- and soil-geochemical surveys, estimating mineral resources, and environmental and economic assessments (Dahrouge, 2001a, b, 2002; Dahrouge and Reeder, 2001; Reeder and Dahrouge, 2002; Smith and Dahrouge, 2002a, 2003; Dahrouge and Wolbaum, 2004; Davis, 2005; Rukhlov and Gorham, 2007; Gorham, 2008; Gorham et al., 2009, 2011a, b, 2013; Kulla and Hardy, 2015 and references therein). This work confirmed known Ta-Nb-P mineralization of the Verity and Fir carbonatites (Dahrouge, 2002; Smith and Dahrouge, 2003) and located many new carbonatites, such as Upper Fir, which was discovered in 2002 (Smith and Dahrouge, 2003). The Upper Fir deposit was delineated by extensive drilling between 2005 and 2011 and, thus far, represents the largest Ta-Nb deposit in the Blue River area (Kulla and Hardy, 2015). Other examples include the Felix, Little Chicago, Mona, Mud Creek, and RD carbonatites, and the Hodgie Zone REE showing (Gorham et al., 2009, 2011a, b). In addition, the extensive Mount Cheadle and Hellroar Creek Ta-Nb soil geochemical anomalies likely indicate near-surface carbonatite bodies (Gorham et al., 2009, 2011a, b). Commerce Resources Corp. (Zimtu Capital Corp.) co-funded several academic studies of the Blue River carbonatites and local geology (e.g., Millonig et al., 2012, 2013; Chudy, 2013; Gervais and Hynes, 2013; Millonig and Groat, 2013; Mitchell et al., 2017).

Further data on various aspects of the Blue River carbonatites were contributed by Hogarth (1989), Mariano (1989), Hogarth et al. (2000), Simandl et al. (2002, 2010), Rukhlov and Bell (2010), Millonig et al. (2012, 2013), Reguir et al. (2012), Chudy (2013), Millonig and Groat (2013), Mitchell (2015), Mackay and Simandl (2015), Fajber et al. (2015), Rukhlov et al. (2015, 2018), Chakhmouradian et al. (2016a, b, 2017), Mao et al. (2016), and Mitchell et al. (2017). Importantly, the compositional variations in pyroclones from the Verity carbonatite were interpreted to represent primary zoning that was preserved despite high-grade metamorphism (Hogarth, 1989; Hogarth et al., 2000). Chudy (2013) provided details on the mineralogy, geochemistry, and petrogenesis of the Upper Fir carbonatite complex. Mitchell et al. (2017) presented trace-element and Sr-Nd isotopic compositions of apatites from several Blue River carbonatites. These data rule out the possibility that these rocks evolved from a single parental magma and suggest a depleted mantle source similar to that of carbonatites worldwide. These findings were corroborated with further petrogenetic constraints in an ongoing mineralogical and O-C-S-Sr-Pb-Nd isotopic study of the Blue River carbonatites (Rukhlov et al., 2018).

5. Blue River area

5.1 General geology

The Blue River area is in the Omineca belt of the Canadian Cordillera, at the northeastern margin of the Shuswap metamorphic complex, in the Monashee Mountains (Fig. 2). The region saw multiple episodes of penetrative deformation from the Jurassic through to the Eocene (e.g., Monger and Price, 1979; Monger et al., 1982; Raeside and Simony, 1983; Scammell, 1993; Pell, 1994; Digel et al., 1998; Crowley et al., 2000; Evenchik et al., 2007; Simony and Carr, 2011). Barrovian metamorphism (garnet-staurolite-kyanite-sillimanite rocks) reached conditions for anatexis, which resulted in syntectonic plutonism (e.g., Ghent et al., 1980; Simony et al., 1980; Carr, 1992; Digel et al., 1988, 1998; Sevigny and Simony, 1989; Sevigny et al., 1989, 1990; Scammell, 1993; Ghent and Villeneuve, 2006; Gervais and Hynes, 2013). Consequently, the ca. 500 Ma and 360-330 Ma metacarbonatites and the enclosing (semi) pelites and amphibolites of the Mica Creek assemblage (750-550 Ma) record protracted deformation at mid-crustal and retrograde conditions (Ghent et al., 1977; Simony et al., 1980; Pell and Simony; 1981, 1987; Pell, 1994; Digel et al., 1988, 1998; Kraft, 2011).

The supracrustal rocks form part of a continuous belt containing the Horsethief Creek and the overlying Kazan groups (Neoproterozoic) and extending from the northern Selkirk Mountains in the southeast, through the Monashee Mountains, and into the Cariboo Mountains in the northeast (Campbell, 1968; Pell and Simony, 1981, 1987; Raeside and Simony, 1983; Digel et al., 1989, 1998; McDonough et al., 1991a, b, 1992; Murphy, 2007). These rocks are cut by steeply dipping, northwest-trending, west-side-down Eocene normal faults (North Thompson fault on the west; Southern Rocky Mountain Trench fault on the east) and are juxtaposed against imbricated North American basement (Malton gneiss complex) by the south-dipping Malton décollement, a mylonitic thrust zone to

British Columbia Geological Survey GeoFile 2018-6
the north (Fig. 4; Morrison, 1982; McDonough and Simony, 1988).

The lower pelite unit and the stratigraphically overlying semipelite-amphibolite unit (at least 1 km thick) of the Mica Creek assemblage, the lower division of the Horsethief Creek Group, underlie much of the Blue River area (Fig. 4; Simony et al., 1980; Raeside and Simony, 1983; Pell and Simony, 1987; McDonough and Murphy, 1990; McDonough et al., 1991a, b, 1992). These rocks represent turbidites and mafic rocks formed during breakup of the supercontinent Rodinia and subsequent development of the western margin of Laurentia from ca. 750 to 550 Ma (Bond and Kominz, 1984; Scammell, 1987; Scammell and Brown, 1990; Ross, 1991; Colpron et al., 2002). The ca. 500 Ma and the more widespread 360-330 Ma carbonatites and related rocks, including the Upper Fir complex (ca. 330 Ma) in the Blue River area were emplaced in a continent-margin setting (Pell, 1985, 1987, 1994; Pell and Höy, 1989; Rukhlov and Bell, 2010; Millonig et al., 2012, 2013; Millonig and Groat, 2013; Mitchell et al., 2017; Rukhlov et al., 2018).

Structural studies in the Blue River area recognized the Scrip nappe, a regional (>50 km long) southwest-vergent structure that has been attributed to Jurassic deformation and exhumation due to collision of the western margin of Ancestral North America with offshore Intermontane terranes (Raeside and Simony, 1983; Digel et al., 1989, 1998; Nelson et al., 2013; Monger, 2014; Sigloch and Mihalynuk, 2017). Subsequent protracted convergence of the Pacific and the North American plates resulted in crustal thickening and regional uplift in the mid-Cretaceous through to the Paleocene, refolding by at least two generations of isoclinal to open folds, a penetrative northeast to southeast dipping foliation, which developed at mid-crustal conditions, and retrograde reverse faults (e.g., Simony et al., 1980; Raeside and Simony, 1983; Digel et al., 1989, 1998; Kraft, 2011). In addition, 3D modelling of the Upper Fir carbonatite revealed late SE-NW contraction (Gorham et al., 2013; Kulla and Hardy, 2015). The latest brittle structures (with <1 m displacement) generally parallel the North Thompson valley and are probably related to Eocene extension and uplift, although some studies argue for continued contraction, until about 50 Ma (e.g., Parrish et al., 1988; Digel et al., 1998; Kraft, 2011). Syntectonic syenitic and granitic intrusions were emplaced throughout the Jurassic–Eocene orogenesis; at least some of these intrusions lack folds (Carr, 1992; Scammell, 1993; Digel et al., 1998; Gervais and Hynes, 2013).

Polyphase but localized metamorphism has been identified in the Blue River area. Metamorphism was between ca. 160 Ma and 56 Ma, with peak conditions of about 540-700°C at 6-8 kbar (e.g., Ghent et al., 1983; Digel et al., 1998; Scammell, 1993; Crowly et al., 2000; Tinkham and Ghent, 2005; Ghent and Villeneuve, 2006; Millonig et al., 2012, 2013; Gervais and Hynes, 2013). Deformation and metamorphic processes overprinted most of the primary igneous features of the carbonatites, which now comprise tightly folded layers up to several tens of m thick with contact-parallel foliation concordant with penetrative fabrics in the country gneisses, schists, and amphibolites (Rukhlov and Gorham, 2007; Gorham et al., 2013). Internal retrograde breccia and shear zones further obscure the original paragenetic record (Chudy, 2013).

5.2 Carbonatites and related rocks

The Blue River carbonatites include at least 18 carbonatite and 2 alkaline, silica-undersaturated-rock occurrences (Fig. 4; Mitchell et al., 2017). Table 1 provides details of some of the 360-330 Ma carbonatites, related rocks, and regional Neoproterozoic marbles. For additional details about these and other occurrences, the reader is referred to McCammon (1953, 1955), Rowe (1958), Rich and Gower (1968), Mariano (1979, 1982), White (1980, 1982, 1985), Aaquist (1982a), Johnston et al. (1990a, b), Pell (1994), Simandl et al. (2002), Rukhlov and Gorham (2007), Millonig et al. (2012, 2013), Chudy (2013), Millonig and Groat (2013), and Mitchell et al. (2017). Generally, they form folded, concordant bodies of dolomite and calcite carbonatites, mainly a few m thick and extending for up to a few 100 m along strike within the enclosing gneisses, schists and amphibolites of the Mica Creek assemblage (Figs. 5a, 6 and 7; Simony et al., 1980; Raeside and Simony, 1983; Pell and Simony, 1987; Digel et al., 1989, 1998; McDonough and Murphy, 1990; McDonough et al., 1991a, b, 1992). Alkaline
metasomatic rocks (fenites), including glimmerites and Na-rich amphibolites typically have sharp contacts with carbonatites and grade into unaltered country rocks across a few cm to a few m (Fig. 8). Clumps of magnetite±ilmenite±olivine±apatite several cm in diameter are common and represent fragmented cumulates (e.g., Fig. 5b; Mitchell et al., 2017). Coarse-grained, calc-silicate veins up to several dm wide (Fig. 9) commonly cut Fe-rich dolomite carbonatites, which weather to a reddish-brown saprolite with enrichment in apatite and silicate and oxide minerals. In contrast, Fe-poor calcite carbonatites typically weather to gravel (Fig. 10). Nepheline syenites associated with carbonatites and diverse ultramafic rocks are only at the Paradise Lake and Howard Creek occurrences (Figs. 5g, h and 11). Outcrops at most showings are discontinuous, due to extensive weathering, locally thick overburden and, at lower elevations, dense vegetation.

The coarse-grained, granoblastic or gneissic to strongly foliated porphyroclastic textures in the carbonatites and related alkaline silicate rocks record equilibrium metamorphic recrystallization and localized retrograde shearing (Chudy, 2013). Regardless of the metamorphic overprinting, the carbonatites retain primary paragenetic relationships, mineral chemistry, and isotopic compositions (Chudy, 2013; Mitchell et al., 2017; Rukhlov et al., 2018). The Upper Fir carbonatite is described in detail below.
<table>
<thead>
<tr>
<th>Rock type</th>
<th>Sample</th>
<th>Texture</th>
<th>Mineralogy</th>
<th>Notes</th>
<th>Equilibrium T°C (log(\text{O}_2))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fenite</td>
<td>16-ARU-126</td>
<td>Foliated, medium-grained, nematogranoblastic</td>
<td>Ret-Fktp, Aea, Cal, Fap</td>
<td>Ab, Kfs, Zrn</td>
<td>Al-Mg±Fe(^{2+})-rich Aea</td>
</tr>
<tr>
<td>Calcite-nepheline syenite</td>
<td>16-ARU-186</td>
<td>Foliated, medium-grained, porphyroclastic</td>
<td>Ab, Kfs, Ne, Sdl, Ann, Cal, Fap</td>
<td>Zrn, Ccn, Pcl, opq</td>
<td>Mg-Al-Ti-rich Ann; Hf-rich Zrn; red-brown Pcl</td>
</tr>
<tr>
<td><strong>Roadside 1</strong> (119°05’57.74” W, 52°23’32.96” N)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Calcite carbonatite</td>
<td>16-ARU-155</td>
<td>Foliated, coarse-grained, porphyroclastic</td>
<td>Cal, Fap, Ol, Di, Phil, Mgt</td>
<td>Zrn, Tphl, Ret, Py, Mgt, Pcl</td>
<td>668-731 725 (-13)</td>
</tr>
<tr>
<td><strong>Serpentine Creek 3</strong> (119°06’52.71” W, 52°22’55.33” N)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Calcite carbonatite</td>
<td>16-ARU-152</td>
<td>Foliated, porphyroclastic</td>
<td>Cal, Fap, Ol, Di, Phil, Mgt</td>
<td>Ed, Ilm, Po, Zrn, Srp</td>
<td>557 (-19)</td>
</tr>
<tr>
<td>Dolomite-calcite carbonatite</td>
<td>16-ARU-150</td>
<td>Foliated, coarse-grained, porphyroclastic</td>
<td>Cal&gt;Dol, Ol, Tphl, Ret-Kret, Mgt, Ilm</td>
<td>Fap, Po, Cep</td>
<td>569-714 538 (-19)</td>
</tr>
<tr>
<td><strong>Switch Creek</strong> (119°07’15.28” W, 52°24’40.72” N)&lt;sup&gt;1&lt;/sup&gt;</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Silicocarbonatite</td>
<td>16-ARU-093</td>
<td>Foliated, medium-grained, cumulate-porphyroclastic</td>
<td>Ol, Phil, Di, Cal, Zrt, Chl, Srp</td>
<td>Fap, Mgt, Ilm</td>
<td>511-558 (-23 to -21)</td>
</tr>
<tr>
<td>Rock type</td>
<td>Sample</td>
<td>Texture</td>
<td>Mineralogy</td>
<td>Notes</td>
<td>Equilibrium T° C (log $\log_{10}$ $\overline{\overline{f}}O_2$)</td>
</tr>
<tr>
<td>-----------</td>
<td>--------</td>
<td>---------</td>
<td>------------</td>
<td>-------</td>
<td>---------------------------------</td>
</tr>
<tr>
<td>Dolomite</td>
<td>16-ARU-06</td>
<td>Foliated</td>
<td>Dol (Fwnc, Po, Mgt), Cal, Ilm, Ab</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Calcite</td>
<td>16-ARU-02</td>
<td>Coarse-grained</td>
<td>Dol (Fwnc, Po, Mgt), Cal, Ilm, Ab</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Calcite</td>
<td>16-ARU-01</td>
<td>Coarse-grained</td>
<td>Dol (Fwnc, Po, Mgt), Cal, Ilm, Ab</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Calcite</td>
<td>16-ARU-02</td>
<td>Coarse-grained</td>
<td>Dol (Fwnc, Po, Mgt), Cal, Ilm, Ab</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fenite</td>
<td>16-ARU-02</td>
<td>Medium-grained</td>
<td>Dol (Fwnc, Po, Mgt), Cal, Ilm, Ab</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Calcite</td>
<td>16-ARU-175</td>
<td>Foliated, coarse-grained, porphyroelastic</td>
<td>Dol, Fap, Phl, Mgt, Cal</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Calcite</td>
<td>16-ARU-175</td>
<td>Foliated, coarse-grained, porphyroelastic</td>
<td>Dol, Fap, Phl, Mgt, Cal</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Windfall Creek</td>
<td>16-ARU-048</td>
<td>Foliated, medium-grained, granoblastic</td>
<td>Dol, Fap, Phl, Mgt, Cal</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ovancop Ridge</td>
<td>16-ARU-143B</td>
<td>Foliated, coarse-grained, porphyroelastic</td>
<td>Dol, Fap, Phl, Mgt, Cal</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ovancop Ridge</td>
<td>16-ARU-143B</td>
<td>Foliated, coarse-grained, porphyroelastic</td>
<td>Dol, Fap, Phl, Mgt, Cal</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ovancop Ridge</td>
<td>16-ARU-143B</td>
<td>Foliated, coarse-grained, porphyroelastic</td>
<td>Dol, Fap, Phl, Mgt, Cal</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ovancop Ridge</td>
<td>16-ARU-143B</td>
<td>Foliated, coarse-grained, porphyroelastic</td>
<td>Dol, Fap, Phl, Mgt, Cal</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
6. Upper Fir carbonatite complex
6.1 Geology

The Upper Fir carbonatite complex (Figs. 12, 13) contains tabular bodies up to 72 m thick that are in isoclinal recumbent similar (in part rootless) folds that are refolded by southwest-vergent, open folds with parallel geometries (Gorham et al., 2013). Fold (F3) hinges in the Upper Fir area have an average trend of 125° and an average plunge of 17° (Kraft, 2011). The carbonatite has been traced for about 1450 m north-south and about 800 m east-west.

Compositionally very similar to the Upper Fir, the Fir and Bone Creek carbonatites immediately to the west (Fig. 4) are probably part of a single complex. The Upper Fir complex consists of mainly dolomite carbonatite and subordinate (<4%) medium- to coarse-grained calcite carbonatite. Calcite carbonatites form lenses up to a few m thick within dolomite carbonatite bodies (Fig. 13); contacts between the two may be sharp or gradational (Fig. 14; Rukhlov and Gorham, 2007; Chudy, 2013). The carbonatites display a foliation that is concordant with penetrative schistosity and compositional layering in the enclosing gneisses, schists, and amphibolites. They are recrystallized, and display fabrics ranging from granoblastic (likely preserving magmatic protoliths) to porphyroclastic and medium- to coarse-grained gneissic, to fine-grained and well foliated (see below). Granoblastic varieties occur mainly at the margins of the carbonatite bodies, whereas the porphyroclastic and fine-grained well-foliated varieties typically make up the central portions of >10 m-thick bodies. This distribution correlates favourably with the composition of amphibole-group minerals and the Nb-Ta oxide phases (see below for details), strongly suggesting that primary compositional zoning within individual dolomite carbonatite
Fig. 12. Stops and positions of north-south and east-west cross sections in the Upper Fir area (NAD83 UTM, Zone 11 coordinates). a) Bedrock geology modified from Kraft (2011) and Gorham et al. (2013).
Fig. 12 continued. b) Road network and diamond drill-holes after Gorham et al. (2011b, 2013).
Fig. 13. Geological cross-sections through the Upper Fir carbonatite complex based on 3D view of drill-hole data (Commerce Resources Corp., pers. comm.) in 100-150 m wide corridor. Lithological logs are shown only for a 10 m wide corridor for clarity (NAD83 UTM, Zone 11 coordinates). a) East-west. b) North-south.
bodies controlled original rheology and thus strain partitioning.

Discontinuous zones (typically <1 m thick) of compositionally heterogeneous phlogopite schist with minor calcite, dolomite, fluorapatite, and amphibole (i.e. phlogopite fenite) mantle carbonatite sills (Fig. 13). These fenites range from phlogopitite or ‘glimmerite’, made up of mostly phlogopite, to banded schists with thin layers of dolomite carbonatite, amphibolite, and biotite-rich gneiss. They are generally strongly foliated and locally crenulated (Fig. 15).

Heterogeneous medium- to coarse-grained, equigranular and porphyroblastic, calcite-clinopyroxene-amphibole rocks are closely associated with carbonatites, typically enveloping glimmerites. These heterogeneous amphibolites, commonly interbanded with glimmerites and country gneisses, contain variable amounts of Ca to Na-Ca amphibole, hedenbergite, titanite, albite, calcite, almandine, potassium feldspar, phlogopite, and fluorapatite (Figs. 5i and 16) and probably represent an outer, Na-fenite aureole. Alternatively, the amphibolites could represent metamorphosed alkaline rocks of the Mica Creek assemblage and thus be unrelated to the Upper Fir carbonatite complex, but their association with glimmerites and carbonatites favours a metasomatic origin. The coarse-grained clinopyroxene-calcite amphibolites are also distinct from foliated amphibolites of the Mica Creek assemblage (Rukhlov and Gorham, 2007).

Carbonatites typically have sharp contacts with fenites, which tend to have gradational contacts with the country rocks. Strongly heterogeneous zones of fenitization also occur at a distance to carbonatites, where they can be up to tens of m thick and consist of massive phlogopite fenite intercalated with thin (up to 30 cm), coarse-grained dolomite carbonatite lenses, variably fenitized country gneisses, and clinopyroxene-calcite

---

**Fig. 14.** Drill core slabs (scale in mm and cm) in normal and UV light showing relationships between calcite and dolomite carbonatites, Upper Fir (Chudy, 2013). a) Abrupt fault contact between dolomite carbonatite (right) and calcite carbonatite (left). b) Gradational contact; calcite is red luminescent, fluorapatite is light-blue, and amphibole and magnetite are black.

**Fig. 15.** Examples of exocontact glimmerites (phlogopite fenites) at Upper Fir. a) Crenulated glimmerite with calcite (Cal) and amphibole (Amp). b) Lepidoblastic texture in glimmerite with biotite (Bt), richterite (Rct), dolomite (Dol), and accessory apatite (Ap), zircon (Zrn), and pyrochlore (Pcl); plane-polarized light photomicrograph (Chudy, 2013).
amphibolites (Na-fenites). Locally, a coarse-grained, amphibole-pyroxene ± phlogopite ± calcite skarn (usually <1 m thick) is developed between the carbonatites and the country rocks, where the glimmerite is lacking. Similar skarn, commonly exhibiting compositional zoning, also mantles carbonatite contacts with crosscutting syenitic and granitic pegmatites (up to 15 m wide). Sharp contacts between carbonatites and unaltered host rocks are likely faults. Ubiquitous veins (several cm wide) composed of aligned amphibole crystals up to 2.5 x 3.0 cm (±pyroxene±phlogopite±pyrrhotite) crosscut the carbonatites (Fig. 9; Appendix 2).

6.2 Textural evolution of the carbonatites

One of the most striking features of the Blue River carbonatites is their textural diversity, which likely records prolonged tectonometamorphic reworking. Compared to the smaller occurrences in the area, the Upper Fir carbonatites exhibit at least four major textural types. The historical nomenclature of the carbonatites recognized texturally distinct rauhaugite (medium- to coarse-grained dolomite carbonatite), beforsite (fine-grained dolomite carbonatite), sövite (medium- to coarse-grained calcite carbonatite), and alvikite (fine-grained calcite carbonatite), suggesting different intrusion phases within individual carbonatite complexes (e.g., Kresten, 1983). Below, we briefly summarize the fabric types and their characteristics, based on detailed textural analysis by Chudy (2013).

The most common fabric is a medium- to coarse-grain size and a shape-preferred orientation of dolomite grains that gives the rock a gneissic appearance. Minor and accessory phases such as elongated amphibole and apatite crystals are aligned within the gneissosity resulting in a penetrative foliation (Figs. 17a, b). Rarely, dolomite and minor phases lack a preferred orientation resulting in a granoblastic texture. Granoblastic rocks are generally restricted to the smaller (<30 cm thick), coarse-grained dolomite lenses that are completely embedded in glimmerite at the margins of the major sills or within extensive zones of fenitization.

Rocks with a porphyroclastic texture display medium- to coarse-grained milky-white dolomite grains set in a very fine-grained, bluish-grey matrix (Fig 17c). The porphyroclast-to-matrix ratio is highly variable and compositional banding is preserved. The fabric overprints the gneissic fabric due to dynamic recrystallization.

Fine- to very fine-grained dolomite carbonatites with no or only minor dolomite clasts set in a bluish-grey matrix represent zones of localized retrograde strain. Aligned dolomite clasts, minor and accessory phases, and pyrrhotite and magnetite define a well-developed foliation (Fig. 17d). The gneissic fabric is the result of re-equilibration under metamorphic conditions. The primary igneous fabric of the dolomite carbonatite cannot be unequivocally determined, but the occurrence of dolomite carbonatites with the granoblastic texture suggests that it was a coarse-grained to pegmatitic rock. Following the peak of metamorphism, continued ductile to brittle deformation resulted in the development of localized shear zones. These zones formed a continuum of fabrics ranging from variably porphyroclastic to very fine-grained foliated rocks. The transitions between the fabric types are commonly gradational, but sharp transitions juxtaposing a gneissic fabric against well-foliated rocks can also be observed. Furthermore, the retrograde shear zones are heterogeneous in terms of their vertical zoning and lateral distribution. They transect the Upper Fir carbonatite complex oblique to the regional foliation and result in the displacement of different mineralogical facies.

The syntectonic pegmatites, syenites and leucogranites crosscutting the carbonatites and related metasomatic rocks constrain the timing of retrograde shearing. The shearing must have occurred after emplacement of the syntectonic quartz veins and pegmatites because the shear zones contain fragments of skarn typically found at vein/pegmatite contacts with dolomite carbonatite. It is important to note that the syntectonic intrusions had a stabilizing effect on the carbonatite during shearing. The carbonatite zones that are intersected by thicker felsic dikes or pegmatite veins (>1 m wide) always preserve their gneissic fabric and the original compositional zoning is intact.
Fig. 17. Typical fabrics in the Upper Fir dolomite carbonatites: a) Foliated, coarse-grained granoblastic or gneissic texture (sample 16-ARU-104, Table 1). b) Foliated, coarse-grained pyrrhotite-actinolite segregation in a). c) Porphyroclastic texture. d) Foliated, fine-grained porphyroclastic texture. Act = actinolite, Dol = Fe-rich dolomite, Fap = fluorapatite, Fktp = ferrikatophorite, Pcl = pyrochlore, Po = pyrrhotite, Rct = richterite, Zrn = zircon. Scale in mm and cm.
6.3 Mineralogy

The Upper Fir carbonatite complex consists mainly of dolomite carbonatite with only subordinate calcite carbonatite. Ferroan dolomite is the main constituent of the dolomite carbonatite. Within the dolomite carbonatite, three major mineralogical facies are distinguished, based on minor and accessory phases: 1) anchimonomineralic facies, 2) ferriwinchite facies, and 3) ferrikatophorite facies (Chudy, 2013).

The anchimonomineralic facies contains very coarse-grained (up to 2 cm) Fe-rich dolomite and minor amounts of accessory phases such as amphibole, fluorapatite, and phlogopite (Fig. 18a). The accessory minerals are concentrated into layers, lenses, and veinlets a few cm thick. The amphiboles are fine grained (<1 mm), pale-green in hand specimen, and belong to the calcium group (actinolite with minor tremolite). They are situated predominantly between the coarse-grained dolomite. Fluorapatite is fine grained (<1 mm) and forms granular accumulations, commonly with substantial amounts of ferrocolumbite and minor amphibole. This assemblage typically forms foliation-parallel lenses or layers, and in some instances, vein-like features in granoblastic or slightly gneissic sections. Accessory pyrrhotite forms anhedral masses (<1 mm) in interstices between the dolomite grains. Isolated flakes of phlogopite, up to 5 mm long, are rare and might be xenocrystic. The anchimonomineralic dolomite carbonatite facies generally occurs within fenite selvages and at the margins of the carbonatite bodies.

Both the ferriwinchite and ferrikatophorite facies contain 5-10 vol.% of fine- to medium-grained (<4 mm), euhedral to subhedral amphiboles, 10-15 vol.% fluorapatite typically as 3-5 mm-long ovoid grains, and variable amounts of pyrrhotite, Nb-Ta oxide phases, zircon, baddeleyite, phlogopite, magnetite, olivine, ilmenite and monazite (Fig. 18b and c). These minerals are generally evenly disseminated and fully enclosed within...
Fe-rich dolomite, or form aggregates in layers or lenses up to 10s of cm thick, resulting in a compositional banding parallel to contacts of carbonatite bodies and to foliation in the carbonatite. Magnetite porphyroclasts typically form sheared and tightly folded lenses. Subhedral to anhedral ilmenite (<1 vol.%%) occurs as inclusions in dolomite and amphibole. Nickel-rich pyrrhotite (average 0.5 vol. %) is a ubiquitous accessory phase, forming anhedral disseminated grains and sideronitic segregations with rare pyrite and chalcopyrite. Molybdenite books (up to 1.7 cm long) occur sporadically in both amphibole-rich accumulations and fenites (Fig. 3).

The ferriwinchite-facies dolomite carbonatite contains the sodium-calcium amphibole ferriwinchite or rare magnesioferrihornblende transitional to actinolite, which forms fine- to medium-grained (5-10 mm), subhedral to euhedral prisms showing Na-enrichment (transitional to richterite) at their margins. Ferrocolumbite (Mg±Ta-rich) is the main Nb-Ta oxide phase in the ferriwinchite-bearing dolomite carbonatite. In contrast, the ferrikatophorite-facies dolomite carbonatite contains the sodium-calcium amphibole ferrikatophorite transitional to richterite and magnesio-arvedsonite at rims and predominantly U-Ta±Ti-rich pyrochlore. A 1-5 m wide transition zone between the two facies contains texturally variable pyrochlore-ferrocolumbite intergrowths (Fig. 19).

Volumetrically minor, calcite carbonatite at Upper Fir (Fig. 13; Appendix 1) is a heterogeneous rock with abundant (up to 20 vol. %), medium- to coarse-grained masses of anhedral Ti-bearing magnetite enclosing fluorapatite, richterite, pyrrhotite, and pyrochlore that are unevenly distributed within the calcite carbonatite matrix or form massive layers and lenses. These apatite-magnetite masses thus resemble phoscorite (Krasnova, 2004). Other than Ti-bearing magnetite with ilmenite exsolutions, the calcite carbonatite contains fluorapatite, ferrikatophorite transitional to richterite, pyrrhotite, ubiquitous orange, U-Ta-rich and rare Na-Ca-deficient, Fe-rich pyrochlore, Fe$^{3+}$-rich ilmenite, zircon, chalcopyrite, and Fe$^{2+}$-Na-rich phlogopite in similar amounts to those of the ferrikatophorite-bearing dolomite carbonatites. The carbonates

![Fig. 19. Examples of pyrochlore-ferrocolumbite composite grains in the Upper Fir dolomite carbonatites (a, b plane-polarized light photomicrographs, c, d corresponding BSE images; Chudy, 2013): a) and c) Ferrocolumbite (Cmb) and other mineral inclusions in pyrochlore (Pcl) enclosing matrix and partially mantled by pyrrhotite (Po). b) and d) Fine-grained ferrocolumbite intergrown with pyrochlore. Note enclosed matrix phases and associated pyrrhotite, ferrikatophorite (Ktp) and magnesio-arvedsonite (Marf).](image-url)
are mostly calcite, containing coarse to sub-µm, irregular and lamellar dolomite exsolutions (Fig. 18d), and minor amounts of Fe-rich dolomite, and typically form a heterogeneous porphyroclastic matrix. The minor and accessory phases tend to form aggregates.

6.4 Niobium-tantalum mineralization

The Nb-Ta mineralization at the Upper Fir complex consists predominantly of ferrocolumbite and pyrochlore-supergroup minerals. The proportion of these minerals generally depends on position relative to the margins of the carbonatite bodies or on mineralogical facies of the carbonatites. Rare fersmite replaces ferrocolumbite. The two main Nb-Ta oxide minerals vary widely in grain size (sub-mm to cm), with ferrocolumbite typically forming coarser grains than pyrochlore. Pyrochlore forms disseminations and clusters.

Ferrocolumbite occurs mainly in the ferriwinchite-facies dolomite carbonatite and forms massive, subhedral or xenomorphic, commonly skeletal to strongly poikilitic grains (up to 3 cm) that enclose all other phases in the dolomite carbonatite (Figs. 18b and 20). Ferrocolumbite concentrations occur in fluorapatite layers and lenses within the anchimonomineralic and ferriwinchite facies, and within the phlogopite fenite. Compositional variation in terms of Nb/Ta and Fe/Mn ratios is relatively minor compared to that of rare metal pegmatites (Fig. 21). The X_{Ta}(Ta/(Ta+Nb) in a.p.f.u.) ranges from <0.002 to 0.068, and the X_{Mn}(Mn/(Mn+Fe) in a.p.f.u.) ranges from 0.054 to 0.123. Titanium and iron show no systematic variation and range between 0.022–0.083 a.p.f.u. Ti and 0.703–0.932 a.p.f.u. Fe. Altered ferrocolumbite crystals have distinctly higher X_{Ta} values (up to 0.179) compared to those of primary ferrocolumbites and, are associated with secondary Ta-rich pyrochlore together with monazite and thorite inclusions (Fig. 20). Subordinate amounts of ferrocolumbite occur also as exsolution lamellae in ilmenite.

The transition zone between the two amphibole facies of dolomite carbonatites displays intergrown ferrocolumbite-pyrochlore aggregates (Fig. 19). The proportion of the two Nb-Ta oxide phases varies widely with unclear replacement relationships between these phases. The composition of ferrocolumbite in the aggregates is within the range of that for primary ferrocolumbrite and does not show elevated Ta concentrations. The composition of pyrochlore in the aggregates is also similar to that of free pyrochlore grains disseminated in the matrix of the ferrikatophorite-bearing dolomite carbonatite. Pyrochlore in the Upper Fir carbonatites (Fig. 22) shows a wide range of compositions extending from the Nb end member to the Ta-rich compositions attaining that of microlite (i.e. Ta\textsubscript{2}
\textsubscript{±}Nb and (Nb+Ta)\textsubscript{2}±2Ti) accompanied by an increase in Ti (Fig. 23). Chudy (2013) distinguished primary and secondary pyrochlores at Upper Fir. The primary pyrochlore forms sub-mm to cm-sized euhedral to subhedral grains, ranging from yellow through red-brown to black, that are disseminated in the carbonate matrix and form inclusions in fluorapatite, amphibole, magnetite, and zircon (Fig. 22a-d and f). Some pyrochlores contain inclusions of dolomite, calcite, and apatite. Clusters and seams of larger pyrochlore clumps are common, resulting in discontinuous high-grade zones with up to 1.06 wt.% Ta\textsubscript{2}O\textsubscript{5} and 17.57 wt.% Nb\textsubscript{2}O\textsubscript{5} across a single 1 m drill-core interval (Gorham et al., 2013). The secondary pyrochlore transitional to microlite is exclusively associated with altered ferrocolumbite, forming subhedral grains less than 100 µm (Fig. 22g). The A-site cations of primary pyrochlore are mostly occupied by Ca and Na in equal amounts, with a tendency of the enrichment in Ca and a small percentage of vacancies (Chudy, 2013; Rukhlov et al., 2018). Minor elements include Sr, Th, REE, Fe, and Mn.

Most pyrochlores in the dolomite carbonatite at Upper Fir have a Nb/Ta ratio between 6.0–6.5 with relatively high Nb (up to 1.659 a.p.f.u.) and fluorine (up to 0.767 a.p.f.u.), and relatively low Ti (<0.083 a.p.f.u.) and U (<0.043 a.p.f.u.) contents (Chudy, 2013). Rare pyrochlore grains that form a locally restricted paragenetic assemblage together with zircon and ferrikatophorite are enriched in Ta (Nb/Ta ratio = 2.8–5) and have higher Ti (up to 0.310 a.p.f.u.) and U (0.360 a.p.f.u.), and lower F (<0.55 a.p.f.u.) concentrations. They are compositionally similar to pyrochlores in the calcite carbonatite. Pyrochlore intergrown with ferrocolumbite in composite grains from the transition zone between the carbonatite facies has relatively high Nb/Ta ratios (up to 8.5) and low Ti (<0.14 a.p.f.u.) and U (<0.045 a.p.f.u.) contents (Chudy, 2013).

Secondary pyrochlore, which formed during the alteration of ferrocolumbite, is strongly enriched in Ta, resulting in the low Nb/Ta ratios of <3 (Fig. 22g). Some compositions with up to 0.840 a.p.f.u. Ta and 0.753 a.p.f.u. Nb are thus classified as microlite. Secondary pyrochlore is also typically enriched in Ti (0.34 – 0.57 a.p.f.u.), U (0.125 – 0.49 a.p.f.u.), Th (0.04 – 0.17 a.p.f.u.), La+Ce (up to 0.09 a.p.f.u.), and have detectable Mn and Fe (Chudy, 2013). Generally, pyrochlores from the Upper Fir are unique in possessing distinctly higher Ta contents compared to those from other carbonatites in the Blue River area and to most carbonatites elsewhere (Fig. 23).

6.5 Whole rock geochemistry

Based on whole-rock geochemistry, the Upper Fir carbonatites classify as magnesiocarbonatite and ferrocarbonatite (Fig. 24). The classification of the calcite carbonatites as ferrocarbonatite is due to abundant magnetite (up to 30 vol.% across 1 m intervals). Likewise, the compositionally heterogeneous dolomite carbonatites with clusters and bands of minor and accessory phases results in a considerable range of major- and trace-element concentrations. Appendix 1 provides examples of downhole geochemical variations in the Upper Fir carbonatites and related fenites for selected elements. Appendix 3 also shows 3D interpolated gridding of drill-core assay data (99th percentile) for selected elements by the inverse-distance method using GOCAD® software in the east-to-west cross section (see Fig. 13a for geological interpretation). The CaO/MgO ratio and total Fe expressed as Fe\textsubscript{2}O\textsubscript{3} contents discriminate between the calcite and dolomite carbonatites, whereas the Na\textsubscript{2}O/Si\textsubscript{O} ratio...
Fig. 20. Examples of paragenetic relationships of ferrocolumbite (Cmb) at Upper Fir (BSE images from Chudy, 2013). a) Poikilitic crystal with vermicular calcite and thorite (Thr) inclusions. b) Interstitial, zoned nioboaeschynite-(Ce) (Aes) and thorite. c) Secondary ferrocolumbite (Cmb 2) with lower Nb/Ta ratio and high-Ta pyrochlore (Pcl 2) and thorite replacing primary ferrocolumbite (Cmb). d) Altered ferrocolumbite with abundant pyrochlore (Pcl) inclusions. e) Poikilitic ferrocolumbite with pyrochlore inclusions. Note brighter rims with lower Nb/Ta ratio than that of the darker cores. f) Fersmite (Fsm) replacing poikilitic ferrocolumbite with thorite inclusions.
correlates with the main mineralogical facies described above (Appendices 1 and 3). To better assess the overall geochemical character of the Upper Fir carbonatites, the data from highly heterogeneous samples with large accumulations of non-carbonate minerals (i.e., >30 vol.% of compositional banding) were omitted from the following summary.

The three mineralogical facies of the magnesiocarbonatites have virtually identical average total Fe (7.82–8.12 wt% as Fe$_2$O$_3$), MgO (15.15–15.41 wt%), and CaO (29.93–30.27 wt%) contents but distinct minor- and trace-element contents (Figs. 25 and 26). The anichomonomineralic- and ferriwinchite-bearing carbonatites have similar average MnO (0.96 and 0.94 wt%), P$_2$O$_5$ (3.21 and 3.51 wt%), Sr (5310 and 5000 ppm), and total REE (608 and 630 ppm) contents. Reflecting mainly different modal abundances of amphibole, the anichomonomineralic carbonatites have generally lower average SiO$_2$ (2.0 wt%) and Na$_2$O (0.08 wt%) concentrations but similar Na$_2$O/SiO$_2$ ratios (0.04) as the ferriwinchite-bearing carbonatites, which have 4.03 wt% SiO$_2$, 0.2 wt% Na$_2$O, and Na$_2$O/SiO$_2$ ratios of 0.05 (Chudy, 2013). Likewise, the anichomonomineralic carbonatites have distinctly lower average Nb (791 ppm) and Ta (88 ppm) than those of the ferriwinchite facies, which have 1538 ppm Nb and 196 ppm Ta, but similar Nb/Ta ratios (8-9), consistent with the different modal abundances of ferroclumbite.

The ferrikatophorite-bearing dolomite carbonatites have slightly lower average MnO (0.8 wt%) and Sr (4225 ppm) but slightly higher P$_2$O$_5$ (4.22 wt%), Na$_2$O (0.4 wt.%), and total REE (660 ppm) contents than those of the anichomonomineralic and ferriwinchite-bearing carbonatites. The average Na$_2$O/SiO$_2$ value (0.14) is also distinctly higher, and the Nb/Ta ratio (ca. 3) is distinctly lower in the ferrikatophorite-bearing dolomite carbonatite compared with those of the ferriwinchite-anichomonomineralic dolomite carbonatites (Fig. 26a).

The calcite carbonatite has higher average total Fe (16.5 wt% as Fe$_2$O$_3$), Na$_2$O (0.57 wt%), P$_2$O$_5$ (5.13 wt%), and total REE (951 ppm) concentrations than those of the dolomite carbonatites. The calcite carbonatite has lower average MnO (0.67 wt%) but similar Sr concentrations (5162 ppm) relative to those of the ferriwinchite-bearing dolomite carbonatites. The average Nb$_2$O/SiO$_2$ ratio is very similar to that of the katophorite-bearing dolomite carbonatites, whereas the average Nb/Ta ratio of 2.4 is the lowest of all the carbonatites at Upper Fir (Figs. 25 and 26; Chudy, 2013).

Across the four carbonatite facies, the high-field strength elements (Ti, Nb, Ta, Zr, and Hf), U, and Th show systematic variations. Niobium decreases from the marginal facies (anichomonomineralic and ferriwinchite-bearing carbonatites) towards the center (calcite carbonatite), whereas Ta abruptly increases from the anichomonomineralic to the ferriwinchite- and ferrikatophorite-bearing dolomite carbonatites and calcite carbonatite, where it is at approximately the same level (195-204 ppm) resulting in a continuous decrease in the Nb/Ta ratio from ca. 10 to 2 (using average values). The Zr/Hf ratio shows the opposite behavior and increases from the marginal facies (26 – 30) towards the calcite carbonatite (59). A similar trend is observed for the U/Th ratio, which increases from ca. 1 to ca. 33 from the anichomonomineralic facies to the calcite carbonatite (Chudy, 2013). The calcite carbonatites also have much higher Y and slightly higher total REE contents than those of the dolomite carbonatites at Upper Fir (Fig. 26).

Compared to other carbonatites of the Blue River area, the Upper Fir carbonatites have distinctly higher MnO and Ta, lower median Cs, Rb, Ba, Hf, Zr, Ti, and REE, but generally similar median Th, K, and P contents, and similar slopes of chondrite-normalized REE patterns (Figs. 25-27). Some calcite carbonatites such as the Gum Creek occurrence immediately to the east of the Upper Fir have much higher Sr and total REE.
**Fig. 22.** Examples of paragenetic relationships of pyrochlore (Pcl) at Upper Fir (b, c, d plane-polarized light photomicrographs; e, f, g BSE images from Chudy, 2013). a) Euhedral U-Ta-rich pyrochlore crystal in weathered, porphyroclastic dolomite carbonatite containing fluorapatite (Ap) and ferrikatophorite (Ktp), BS-1 cut (Stop 4). Pen for scale. b) Dark-red pyrochlore coexisting with yellow pyrochlore in porphyroclastic dolomite carbonatite containing fluorapatite, ferrikatophorite, and pyrrhotite (Po). c) Primary pyrochlore inclusion in fluorapatite and poikilitic pyrochlore with dolomite and fluorapatite inclusions. d) Mg-Ta-rich ferrocolumbite (Cmb) replacing orange U-Ta-rich pyrochlore in granoblastic calcite-dolomite carbonatite containing fluorapatite, ferrikatophorite (transitional to richterite), Ti-bearing magnetite (Mgt) with exsolution lamellae of ilmenite, calcite (Cal) with sub-μm to coarse exsolution lamellae of dolomite, pyrrhotite, and accessory Fe<sup>2+</sup>-Na-rich phlogopite, Na-Ca-deficient and U-Fe-Ta-rich pyrochlore, zircon, and chalcopyrite (sample 16-ARU-101, Table 1). e) Ferrocolumbite forming a replacement rim on unzoned pyrochlore from a fluorapatite-rich layer in ferriwinchite-bearing dolomite carbonatite. f) Low-temperature alteration forming darker coloured dendritic features overprinting primary compositional zoning in a pyrochlore crystal with lower Nb/Ta ratio of the brighter coloured core compared to that of the darker coloured rim. g) Microlite associated with secondary ferrocolumbite (lighter grey) having lower Nb/Ta ratio relative to that of the primary ferrocolumbite (darker grey) with inclusions of Ta-rich pyrochlore (Pcl 4).
**Fig. 23.** Ternary Ti–Nb–Ta diagram (mol.%) after Atencio et al. (2010) for pyrochlore supergroup from the Blue River carbonatites compared with pyrochlore compositions from worldwide carbonatites. Data from Chakhmouradian (2006), Chakhmouradian et al. (2015), Mackay and Simandl (2015), Chudy (2013), and Rukhlov et al. (unpublished data from Blue River carbonatites).

**Fig. 24.** Ternary CaO–MgO–[Fe₂O₃(T) + MnO] diagram (in wt.%) after Woolley and Kempe (1989) for the Blue River carbonatites with <20 wt% SiO₂. Note that magnetite-bearing calcite carbonatites plot into the ferrocarbonatite field. Data from Rukhlov and Gorham (2007), Gorham (2008), and Chudy (2013).

**Fig. 25.** Whole-rock geochemical variation plots for the Blue River carbonatites: a) SiO₂ (wt%) vs Na₂O (wt%). b) P₂O₅ (wt%) vs total REE (ppm). c) Sr (ppm) vs MnO (wt%). d) MgO (wt%) vs Fe₂O₃(T) (wt%). Data from Rukhlov and Gorham (2007), Gorham (2008), and Chudy (2013).
contents, and higher \((\text{La/Yb})\text{CN}\) ratios compared to those of the Upper Fir carbonatites (Figs. 25 and 26). These geochemical characteristics directly reflect the mineralogical compositions of the Blue River carbonatites (Table 1), whereby \(\text{K}_2\text{O}, \text{Na}_2\text{O}\) and \(\text{SiO}_2\) are controlled by the amphibole, pyroxene and mica species, \(\text{MnO}\) and \(\text{Sr}\) by the composition of the carbonates, REEs mainly by apatite (Mitchell et al., 2017), and the high-field strength elements and \(\text{V}\) by complex oxide phases such as pyrochlore, ferro Columbite, aeshynite, zirconolite, ilmenite, magnetite, and baddeleyite, and by other accessory minerals such as zircon and thorite.

The Blue River carbonatites have very similar primordial mantle-normalized patterns to those of the ca. 344 Ma Aley carbonatite, except for much higher \(\text{Th}\) and lower \(\text{Ta}\) contents (Fig. 27). The normalized REE and multi-element patterns for regional marbles in the Blue River differ completely from those of the carbonatites and related alkaline, silica-undersaturated rocks (Fig. 27).
6.6 Geochronology and Sr-Pb-Nd isotopes

Summarizing radiometric dates from the Blue River carbonatites and related rocks, Pell (1994) interpreted that determinations of ca. 360 to ca. 325 Ma from the Verity and Mud Lake carbonatites and the Paradise Lake nepheline syenite represent magmatic crystallization ages. Similar dates were obtained by Rukhlov and Bell (2010) from the Verity dolomite carbonatite (ca. 350 Ma, $^{207}\text{Pb}/^{206}\text{Pb}$ zircon), and by Millonig et al. (2012, 2013) from the Serpentine Creek carbonatite (341 Ma, U-Th-Pb zircon). Millonig et al. (2012, 2013) also reported much older dates of ca. 498 Ma on zircon from the Felix carbonatite and ca. 496 Ma on baddeleyite from the Little Chicago carbonatite, which record the Cambrian pulse of carbonatite activity in western Laurentia (Millonig and Groat, 2013). Zircons from the Upper Fir carbonatite were analyzed in situ for U-Pb geochronology (Commerce Resources Corp. pers. comm.) using laser ablation, multi-collector inductively coupled plasma mass spectrometry (LA-MC-ICP-MS). Analyses (Table 2, Fig. 28) and data reduction were by A. Simonetti at the University of Alberta; the analytical method is described in Simonetti et al. (2005). Due to the relatively young age and the low Pb and U contents of the zircons, most analyses used a relatively large laser spot (~80 µm) instead of usual ~40 µm spot. Two results using a 40 µm spot are consistent with those obtained from the 80 µm analyses. The 21 analyses show variable degrees of Pb loss and yield a weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ date of 332.5±5.7 Ma (2σ), interpreted to establish the time of magmatic crystallization, and consistent with results from other Late Paleozoic carbonatites in the area.

In contrast, much younger K/Ar dates on phlogopite and amphiboles and U-Th-Pb dates on pyrochlore, monazite, zirconolite, and zircon (ca. 205–49 Ma) from the Blue River carbonatites overlap K/Ar mica and hornblende cooling ages and U-Pb zircon, monazite and xenotime ages from the country metamorphic rocks and syenitic granites (171-46 Ma; see Pell 1994; Rukhlov and Bell, 2010; Millonig et al. 2012, 2013; Gervais and Hynes, 2013; Mitchell et al., 2017). These data reflect open system behaviour and substantial resetting of the K/Ar and U-Th-Pb systems in primary carbonatite minerals during multiple episodes of metamorphism (e.g., Digel et al., 1998; Crowley et al., 2000; Ghent and Villeneuve, 2006; Gervais and Hynes, 2013). The young U-Th-Pb dates on zircon, pyrochlore, monazite, and zirconolite from the Blue River carbonatites are best explained by episodic Pb loss during amphibolite-facies metamorphism rather than by continuous diffusion (Rukhlov and Bell, 2010; Millonig et al., 2012, 2013; Mitchell et al., 2017).

Zircons from the Upper Fir carbonatite were analyzed in situ for U-Pb geochronology (Commerce Resources Corp. pers. comm.) using laser ablation, multi-collector inductively coupled plasma mass spectrometry (LA-MC-ICP-MS). Analyses (Table 2, Fig. 28) and data reduction were by A. Simonetti at the University of Alberta; the analytical method is described in Simonetti et al. (2005). Due to the relatively young age and the low Pb and U contents of the zircons, most analyses used a relatively large laser spot (~80 µm) instead of usual ~40 µm spot. Two results using a 40 µm spot are consistent with those obtained from the 80 µm analyses. The 21 analyses show variable degrees of Pb loss and yield a weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ date of 332.5±5.7 Ma (2σ), interpreted to establish the time of magmatic crystallization, and consistent with results from other Late Paleozoic carbonatites in the area.
Table 2. In situ U-Pb analyses on zircons from the Upper Fir carbonatite by laser ablation, multi-collector inductively coupled plasma mass spectrometry, obtained by A. Simonetti at the University of Alberta (Commerce Resource Corp., pers. comm.).

<table>
<thead>
<tr>
<th>Grain</th>
<th>Spot size (µm)</th>
<th>206Pb counts</th>
<th>206Pb/238U Age (Ma)</th>
<th>207Pb/235U Age (Ma)</th>
<th>% discordance</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>40</td>
<td>83992 infinite</td>
<td>0.05315 0.00063 0.3671 0.0133 0.0498 0.0018 0.946 335 27</td>
<td>313 11</td>
<td>6.5</td>
</tr>
<tr>
<td>2</td>
<td>40</td>
<td>74043 infinite</td>
<td>0.05325 0.00067 0.3700 0.0140 0.0502 0.0019 0.943 340 29</td>
<td>316 12</td>
<td>7.0</td>
</tr>
<tr>
<td>3</td>
<td>80</td>
<td>266514 infinite</td>
<td>0.05362 0.00060 0.3932 0.0125 0.0528 0.0017 0.939 355 25</td>
<td>332 10</td>
<td>6.6</td>
</tr>
<tr>
<td>4</td>
<td>80</td>
<td>318560 6371</td>
<td>0.05329 0.00145 0.3683 0.0171 0.0498 0.0019 0.810 341 62</td>
<td>313 12</td>
<td>8.2</td>
</tr>
<tr>
<td>5</td>
<td>80</td>
<td>55691 11138</td>
<td>0.05347 0.00153 0.3912 0.0164 0.0531 0.0017 0.733 349 65</td>
<td>334 11</td>
<td>4.3</td>
</tr>
<tr>
<td>6</td>
<td>80</td>
<td>207733 infinite</td>
<td>0.05310 0.00058 0.3712 0.0121 0.0504 0.0016 0.943 333 25</td>
<td>317 10</td>
<td>4.8</td>
</tr>
<tr>
<td>7</td>
<td>80</td>
<td>136502 infinite</td>
<td>0.05276 0.00073 0.3212 0.0211 0.0438 0.0028 0.977 319 32</td>
<td>276 18</td>
<td>13.3</td>
</tr>
<tr>
<td>8</td>
<td>80</td>
<td>396317 infinite</td>
<td>0.05334 0.00057 0.3781 0.0124 0.0510 0.0017 0.946 343 24</td>
<td>321 10</td>
<td>6.6</td>
</tr>
<tr>
<td>9</td>
<td>80</td>
<td>154123 infinite</td>
<td>0.05304 0.00059 0.3672 0.0116 0.0502 0.0016 0.938 331 25</td>
<td>316 10</td>
<td>4.5</td>
</tr>
<tr>
<td>10</td>
<td>80</td>
<td>494479 infinite</td>
<td>0.05312 0.00056 0.3719 0.0120 0.0504 0.0016 0.948 334 24</td>
<td>317 10</td>
<td>5.0</td>
</tr>
<tr>
<td>11</td>
<td>80</td>
<td>162296 infinite</td>
<td>0.05281 0.00057 0.3658 0.0123 0.0498 0.0017 0.949 321 24</td>
<td>313 10</td>
<td>2.3</td>
</tr>
<tr>
<td>12</td>
<td>80</td>
<td>444894 infinite</td>
<td>0.05309 0.00056 0.3626 0.0122 0.0493 0.0016 0.950 333 24</td>
<td>310 10</td>
<td>6.9</td>
</tr>
<tr>
<td>13</td>
<td>80</td>
<td>180088 infinite</td>
<td>0.05296 0.00057 0.3720 0.0117 0.0506 0.0016 0.940 327 25</td>
<td>318 10</td>
<td>2.7</td>
</tr>
<tr>
<td>14</td>
<td>80</td>
<td>296279 infinite</td>
<td>0.05314 0.00057 0.3684 0.0119 0.0500 0.0016 0.944 335 24</td>
<td>315 10</td>
<td>6.0</td>
</tr>
<tr>
<td>15</td>
<td>80</td>
<td>292390 infinite</td>
<td>0.05308 0.00055 0.3820 0.0125 0.0521 0.0017 0.950 332 23</td>
<td>327 11</td>
<td>1.5</td>
</tr>
<tr>
<td>16</td>
<td>80</td>
<td>63212 infinite</td>
<td>0.05281 0.00075 0.3664 0.0123 0.0501 0.0016 0.907 321 32</td>
<td>315 10</td>
<td>1.8</td>
</tr>
<tr>
<td>17</td>
<td>80</td>
<td>499319 infinite</td>
<td>0.05306 0.00057 0.3642 0.0116 0.0494 0.0016 0.941 331 25</td>
<td>311 10</td>
<td>6.1</td>
</tr>
<tr>
<td>18</td>
<td>80</td>
<td>78756 infinite</td>
<td>0.05266 0.00068 0.3729 0.0132 0.0510 0.0018 0.932 314 29</td>
<td>321 11</td>
<td>-2.1</td>
</tr>
<tr>
<td>19</td>
<td>80</td>
<td>366099 infinite</td>
<td>0.05338 0.00057 0.3741 0.0131 0.0506 0.0018 0.953 345 24</td>
<td>318 11</td>
<td>7.7</td>
</tr>
<tr>
<td>20</td>
<td>80</td>
<td>478962 infinite</td>
<td>0.05303 0.00056 0.3738 0.0121 0.0511 0.0016 0.946 330 24</td>
<td>321 10</td>
<td>2.7</td>
</tr>
<tr>
<td>21</td>
<td>80</td>
<td>343648 infinite</td>
<td>0.05286 0.00055 0.3805 0.0122 0.0520 0.0017 0.947 323 24</td>
<td>327 10</td>
<td>-1.3</td>
</tr>
</tbody>
</table>

*counts per second.*
The British Columbia alkaline province has only sparse Sr-Nd isotopic data (Rich and Gower, 1968; Mumford, 2009; Tappe and Simonetti, 2013; Mitchell et al., 2017). Rukhlov et al. (2018) analyzed separated calcite, dolomite, and apatite fractions from 21 carbonatite and related rocks that were sampled at nine 360-330 Ma complexes in the Blue River area for Pb, Sr, and Nd isotopic compositions. The analysis was performed at the University of British Columbia and used thermal ionization mass spectrometry for Sr and solution MC-ICP-MS for Pb and Nd. Concentrations of Rb, Sr, Nd, Sm, Pb, Th and U were determined by high-resolution ICP-MS on the same dissolution. In addition, molybdenite from a coarse-grained ferrikatophorite-apatite-dolomite carbonatite at the Upper Fir (sample 16-ARU-198; Table 1; Fig. 3) was analyzed for Pb isotopic composition by MC-ICP-MS, along with Pb, Th and U concentrations by ICP-MS, and for Re/Os model age determination by negative thermal ionization mass spectrometry at the University of Alberta. The molybdenite has 14 ppm \(^{187}\)Re and 42 ppb \(^{187}\)Os, shows no common Os above blank levels, and is probably strongly decoupled with respect to Re and Os (Rukhlov et al., unpublished).

Initial \(^{87}\)Sr/\(^{86}\)Sr ratios (0.702875 - 0.704589) and the \(\varepsilon_{\text{Nd}}(T)\) values (-0.7 to 6.3; Fig. 29) for calcite, dolomite, and apatite fractions from the Blue River carbonatites and related rocks overlap the in situ data from carbonatites from the Blue River carbonatites obtained by Mitchell et al. (2017). The Upper Fir carbonatites have the lowest initial \(^{87}\)Sr/\(^{86}\)Sr ratios (0.702902 – 0.702963) and the highest \(\varepsilon_{\text{Nd}}(T)\) values (5.64 to 6.28) compared to other carbonatites and related rocks in the Blue River area, indicating a depleted mantle source similar to FOZO. Calcite from the Gume Creek carbonatite has the lowest \(\varepsilon_{\text{Nd}}(T)\) value of -0.7 and a higher initial \(^{87}\)Sr/\(^{86}\)Sr ratio of 0.704084 relative to most of the data from the Blue River carbonatites and related rocks. The data define a mixing array between the depleted end-member (FOZO) and the enriched end-member (EM1) in the Sr-Nd isotope correlation diagram, similar to the mixing trends for the 370 Ma Kola alkaline province, East African carbonatites and most other <200 Ma carbonatites worldwide (Fig. 29). However, calcites from the Paradise Lake nepheline syenite and from the Switch Creek silicocarbonatite (containing abundant olivine, plhlogipite, and zirconolite) have slightly higher initial \(^{87}\)Sr/\(^{86}\)Sr ratios (0.704357 - 0.704589) at \(\varepsilon_{\text{Nd}}(T)\) values of ca. +4, indicating involvement of a third mantle component with relatively high time-integrated Rb/Sr and moderately superchondritic (depleted) time-integrated Sm/Nd.

Despite the very low U/Pb ratios (<0.08) in all samples, the Pb isotopic compositions of apatites, carbonates, and molybdenite from the Blue River carbonatites and related rocks show an extreme range of values (\(^{206}\)Pb/\(^{204}\)Pb\text{ initial} = 19.24-238.57, \(^{207}\)Pb/\(^{204}\)Pb\text{ initial} = 15.65-24.42, \(^{208}\)Pb/\(^{204}\)Pb\text{ initial} = 34.39-114.23). They define a line on the \(^{207}\)Pb/\(^{204}\)Pb vs. \(^{206}\)Pb/\(^{204}\)Pb diagram, yielding a Pb-Pb isochron date (using robust regression) of 324±40 Ma (Fig. 29a) consistent with the emplacement ages of the Upper Fir and other late Paleozoic carbonatites in the Blue River area. The extremely radiogenic Pb isotopic compositions of the low-U molybdenite, carbonates and apatite from the

---

Fig. 29. Sr-Pb-Nd isotope correlation diagrams for the 360-330 Ma carbonatites and related rocks of the Blue River area (after Rukhlov et al., 2018). a) Initial \(^{206}\)Pb/\(^{204}\)Pb vs initial \(^{206}\)Pb/\(^{204}\)Pb for separated calcite (Cal), dolomite (Dol), apatite (Ap) and molybdenite (Mol) fractions (U/Pb <0.08 in all samples); 2σ error bars are less than the symbol size. b) \(\varepsilon_{\text{Sr}}(T)\) vs \(\varepsilon_{\text{Nd}}(T)\) for separated calcite, dolomite, and apatite fractions, East African carbonatite line. c) \(\varepsilon_{\text{Pb}}(T)\) vs \(\varepsilon_{\text{Nd}}(T)\) for separated calcite and dolomite fractions with the lowest initial \(^{206}\)Pb/\(^{204}\)Pb ratios.
Upper Fir carbonatites reflect various proportions of radiogenic Pb sequestered from coexisting U-rich pyrochlore during metamorphism, constrained by the molybdenite Re/Os model age of 175.2 ± 0.8 (2σ) Ma (Rukhlov et al., 2018).

In the Sr-Pb-Nd isotopic correlation diagrams using the lowest initial 206Pb/204Pb, the Blue River carbonatites and related rocks follow the mixing trend involving FOZO and EM1 mantle end-members found in oceanic island basalts, young (<200 Ma) carbonatites worldwide, and the plume-related Kola alkaline province (370 Ma; for an overview, see Rukhlov et al., 2015 and references therein). Thus, despite protracted deformation and upper amphibolite-facies metamorphism, the Blue River carbonatites are isotopically indistinguishable from worldwide carbonatites generated by deep-mantle plumes (Mitchell et al., 2017; Rukhlov et al., 2018).

### 6.7 Stable isotopes

The C and O isotopic compositions of separated calcite and dolomite fractions (δ13C = -5.5 to -2.2 ‰ VPDB, δ18O = 4.0 to 9.3 ‰ VSMOW) are within the range of mantle values (e.g., Deines, 1989, 2002), scattering from the field of ‘primary (unaltered) carbonatite’ of Taylor et al. (1967) towards lighter δ13C and δ18O values. Regional marbles in the Blue River area have much heavier δ13C_VSMOW values (-0.15 to 6.55) and δ18O_VSMOW values (14.52 to 20.47), typical of continental crust. We do not think that the scattering of the C-O data from the Blue River carbonatites and alkaline rocks reflects seawater influence. Instead, the scatter possibly indicates a 13C-heterogeneous mantle source coupled with high-temperature alteration during metamorphism (Fig. 30a). Most of the coexisting dolomite-calcite pairs indicate isotopic disequilibrium for both δ13C and

![Stable isotope diagrams](image)

**Fig. 30.** Stable isotope diagrams for the Blue River carbonatites, related rocks, and regional marbles (after Rukhlov et al., 2018). **a)** δ18O_VSMOW (‰) vs δ13C_VPDB (‰) for separated calcite (open symbols) and dolomite (filled symbols) fractions, with ties shown for coexisting phases; 2σ error bars are less than the symbol size. Arrows indicate processes causing shifts in δ18O and δ13C (after Demény et al., 2004 and references therein). ‘Primary’ unaltered carbonatite field after Taylor et al. (1967) and mantle field after Deines (1989). **b)** δ34S_VCDT (‰) vs δ13C_VPDB (‰) for separated, coexisting pyrrhotite and molybdenite (δ34S data) and calcite and/or dolomite (δ13C data). Arrows indicate processes causing shifts in δ34S after Deines (1989) and Rollinson (1993). Average δ13C value for mantle after (Deines, 2002) and assumed δ34S value for primitive mantle after Hoefs (2009).
δ^18O, consistent with recrystallization and consequent open-system behaviour during metamorphism as recorded by the Pb isotopic data. The C-O isotopic data from the Blue River carbonatite show little fractionation of the parental carbonatite magmas and clearly rule out any assimilation of crustal materials, consistent with the Sr-Nd-Pb isotopic evidence discussed above. In contrast to the whole-rock geochemical composition, the C-O isotopic compositions from the Upper Fir carbonatites do not show systematic variations related to the mineralogical facies (Chudy, unpublished).

The range of S isotopic compositions of separated pyrrhotite and molybdenite fractions from the Blue River carbonatites and related rocks (δ^34S = -6.0 to 6.5 ‰ CDT) encompasses both heavier and lighter values than that of assumed primitive mantle (e.g., Hoefs, 2009), similar to the range of δ^34S values in carbonatites from the Canadian Shield (Farrell et al., 2010) and from the Kola alkaline province (Bell et al., 2015). All δ^34S values for pyrrhotite and molybdenite from the Upper Fir are heavier than assumed primitive mantle (1.55-6.47 ‰ CDT), with the molybdenite having δ^34S_CDT value of 2.41, slightly heavier than that of the coexisting pyrrhotite (1.55). Considering their depleted Sr and Nd isotopic compositions, crustal contamination is an unlikely cause for the heavy S isotopic composition of the Ta-rich Upper Fir carbonatites, which instead perhaps reflect 34S-enriched source or SO_2 degassing during their emplacement. On the other hand, the very light S isotopic compositions (δ^34S) of pyrrhotite from the Sr- and REE-rich Gum Creek calcite carbonatites (-6.00 to -5.40 ‰ CDT), probably reflect fractionation of the carbonatite magma accompanied by the increasing fO_2 (Deines, 1989; Rollinson, 1993; Farrell et al., 2010; Bell et al., 2015).

The stable isotope data indicate that the Blue River carbonatites and related silica-undersaturated rocks retained their primary mantle signature despite high-grade metamorphism and retrograde shearing, which contrasts with data from many carbonatites worldwide (e.g., Demény et al., 2004), including Cordilleran carbonatites such as the Aley (Chakhmouradian et al., 2015) and the Wicheeda Lake (Trofanenko et al., 2016).

7. Petrogenetic summary
7.1 Geothermobarometry

The Blue River carbonatites and related rocks show a range of temperatures inferred from chemical or oxygen isotope equilibrium between coexisting minerals (Figs. 31 and 32). Chudy (2013) used the calcite-dolomite solvus in the ternary CaCO_3-MgCO_3-FeCO_3 system after Anovitz and Essene (1987) to estimate crystallization temperatures of the Upper Fir carbonatites. The highest temperatures preserved by calcite, considered to represent magmatic crystallization temperature, were at least 50 °C above those reached during metamorphism in the Blue River area (up to ~700 °C; e.g., Digel et al., 1998; Ghent and Villeneuve, 2006). Distinctly lower temperatures (below 450 °C) were attributed to retrograde overprint (Chudy, 2013). New calcite-dolomite equilibrium data from 9 carbonatite complexes in the Blue River area obtained by Rukhlov et al. (2018) are consistent with these findings (Fig. 31). The carbonate solvus data generally show more restricted range of temperatures compared with those based on Fe-Ti equilibrium between coexisting magnetite – ilmenite solid solution series after Andersen and Lindsley (1985). Calcite from regional marble in this study yielded equilibrium temperatures of 561-626 °C or 667-709 °C with reintegrated 5 vol.% dolomite.

![Fig. 31. Summary of the geothermometry for the Blue River carbonatites, related rocks, and regional marbles (after Rukhlov et al., 2018), based on calcite-dolomite solvus, assuming 0-10 vol % reintegrated dolomite (after Anovitz and Essene, 1987), magnetite-ilmenite solid solutions (after Andersen and Lindsley, 1985), and oxygen isotope equilibrium between the co-existing mineral pairs (after Chacko et al., 2001; Valley, 2003). Abbreviations: CC = calcite-carbonatite, CNS = calcite-nepheline syenite, DC = dolomite-carbonatite, DCC = dolomite-carbonatite, MA = regional marble, PH = phoscorite, SC = silicocarbonatite.](image-url)
exsolution lamellae, which straddle the maximum calcite-dolomite and magnetite-ilmenite temperature estimates from the Blue River carbonatites (Fig. 31). In contrast, the oxygen isotope equilibrium temperatures for dolomite, magnetite, ilmenite, zircon, and amphiboles coexisting with calcite, using fractionation factors after Shackow et al. (2001) and Valley (2003) yield values between 689–1079 °C, which are clearly much higher than the regional metamorphic temperatures and hence reflect crystallization temperatures of relatively hot, oxidized (ΔQFM = -0.5 to +5) magmas (Figs. 31 and 32). The oxygen isotope equilibrium temperature for coexisting calcite-phlogopite pair from the Paradise Lake nepheline syenite (592 °C) is 330 °C lower than that for coexisting zircon-calcite pair from the same sample, and thus records metamorphism.

7.2 Parental magmas and magmatic evolution

Most carbonatites in the Blue River area lack silica-undersaturated rocks such as calcite-rich nepheline syenites, ultramafic, and phoscoritic rocks that are intimately associated with carbonatites at Howard Creek and Paradise Lake (Rich and Gower, 1968; Mariano, 1982; White, 1985; Pell, 1994). This suggests that the Blue River carbonatites could have formed from primary carbonatitic magmas with only a small proportion of silicate, or that rocks that could represent their unevolved parental magmas (e.g., melilitites to kamafugites) are not exposed (Mitchell, 2015; Mitchell et al., 2017). Although the disaggregated masses of heterogeneous phlogopite-titanite- amphibole-pyroxyene ultramafic and phoscorite-like, apatite-olivine-magnetite rocks at Howard Creek could represent cumulates, the magnetite-rich rocks at Upper Fir might have formed from a late-stage ferrocolumbite magma that crystallized at higher temperatures (>700 °C) than that of ankerite or siderite stability (<500 °C; Chudy, 2013; Mitchell et al., 2017).

Primary minerals in the Blue River carbonatites generally crystallized in the following order: Nb-Ta-Zr oxides ±zircon ± ilmenite-amphibole-apatite-olivine-pyroxyene-magnetite ± pyrrhotite ± chalcopyrite ± molybdenite ± pyrite ± phlogopite-dolomite ± calcite. However, phlogopite could also have crystallized relatively early, and magnetite, apatite and Nb oxides could have been the last phases to form (Mitchell et al., 2017; Rukhlov et al., 2018). Compositions of amphiboles from the Upper Fir carbonatites show a magmatic evolutionary trend from the Al- and Ti-rich Ca end member towards the Na-Ca and Na end member, which is commonly observed in carbonatites elsewhere (e.g., Hogarth, 1989), but the high F contents distinguish the Upper Fir from most occurrences (Chudy, 2013). The crystallization of magnesio-arfvedsonite and K-richrichtite in some carbonatites of the Blue River area reflects alkali-rich, late-stage fluids that altered earlier amphibole species and were responsible for the fenitization of the country rocks. Compositions of amphiboles in the exocontact fenites are distinct from those of regional amphibolites (Chudy, 2013).

The abundant ferrocolumbite and its paragenesis with pyrochlore supergroup minerals in carbonatites at Upper Fir (Figs. 20 and 22) are the hallmark of this complex (Chudy, 2013; Mitchell, 2015). The Nb/Ta, U, and Ti increase from the early U-Ta-rich pyrochlore towards the later pyrochlore, consistent with the uranpyrochlore evolutionary trend observed in carbonatites elsewhere (e.g., Hogarth, 1989; Lee et al., 2006; Chakhmouradian et al., 2015). However, microlite is clearly related to the metasomatic alteration of the primary ferrocolumbite at Upper Fir (Fig. 22g). Chudy (2013) proposed that such metasomatic alteration could result from recharge of a more evolved (high Nb/Ta, low Na, and possibly high OH/F) carbonatite magma, precipitating ferriwinchite with ferrocolumbite, by the more primitive magma (low Nb/Ta, high Na) crystallizing ferrikatophorite transitional to magnesio-arfvedsonite with the pyrochlore supergroup.

The lack of siderite and hydrothermal Na-, Sr- and REE-bearing (fluoro)carbonates in the Blue River carbonatites suggests that they formed from relatively unevolved parental magmas, which could be a consequence of their emplacement as small bodies such as sills as opposed to large zoned complexes made up of predominantly alkaline, silica-undersaturated rocks with highly differentiated, late-stage carbonatites (Mitchell et al., 2017). Based on the whole-rock geochemical and the mineralogical evidence, the Upper Fir carbonatites likely formed from a weakly fractionated, primary magnesio-carbonatite magma with high Ta, (sub)chondritic Ti, Zr, Hf contents and Nb/Ta ratio, and low REE, Ba and Sr contents, hence the geochemical attributes of its mantle source rather than protracted magmatic evolution (Chudy, 2013).

The distinct mineralogy, whole-rock geochemistry, Sr-Nd-Pb isotopic and trace-element signatures of minerals from individual carbonatite occurrences in the Blue River area rule out their origin from a single parental magma (Mitchell et al., 2017; Rukhlov et al., 2018). Instead, each complex likely represents weakly fractionated, low-volume, low-degree partial melt from isotopically heterogeneous, carbonated mantle peridotite (e.g., Harmer and Gittins, 1998; Wyllie and Lee, 1998; Wyllie and Lee, 2000).
7.3 Isotopic constraints on the mantle source and tectonic implications

The Sr, Nd and Pb isotopic compositions from the ca. 500 Ma and 360-330 Ma carbonatites in the Blue River area (Mitchell et al., 2017; Rukhlov et al., 2018) fit the evolutionary model for the depleted component in the sources of carbonatites worldwide with the present-day Sr-Pb-Nd-Hf isotopic attributes of FOZO, also found in OIBs and considered to represent ancient (>4 Ga), depleted, deep mantle (for an overview, see Rukhlov et al., 2015 and references therein).

Mitchell et al. (2017) noted that the Blue River carbonatites tapped a sub-lithospheric mantle source that is isotopically similar to that of carbonatites from cratonic interiors such as the Canadian Shield (e.g., Bell et al., 1982). Mitchell et al. (2017) also recognized that the compositional similarity of rocks in the British Columbia alkaline province implies partial melting of the same type of depleted mantle during a period of ca. 480 Ma, and suggested that lithospheric extension provided pathways for the low-degree, sub-lithospheric partial melts along reactivated, detachment faults resulting in sill-like intrusions distributed along the length of the continental margin. However, Mitchell et al. (2017) considered that the rocks were unlikely to record a mantle plume because associated flood basalts and large volumes of undersaturated silicate rocks are lacking, because the multiple episodes of carbonatite emplacement span ca. 480 Ma (ca. 810-700 Ma, ca. 500 Ma, ca. 360-330 Ma), and because the geometry of the British Columbia alkaline province defines a long (at least 1,000 km), narrow (ca. 200 km) orogen-parallel belt (Fig. 1), rather than one that could reflect plume-related regional doming.

Here we build on the Mitchell et al.’s (2017) findings and emphasize that the isotopic and elemental compositions of minerals from Blue River carbonatites and related rocks are indistinguishable from worldwide carbonatites generated by deep-mantle plumes (Fig. 29). The Late Paleozoic Cordilleran examples formed along the western margin of Laurentia while subduction was taking place immediately to the west. Lithospheric extension related to this Late Paleozoic subduction is considered responsible for rifing the continental margin and initiating the Slide Mountain ocean as a back-arc subduction is considered responsible for rifing the continental margin and initiating the Slide Mountain ocean as a back-arc setting for the Late Paleozoic carbonatites might be explained in terms of back arc-extension stretching the inboard continental lithosphere and thus triggering carbonatite magmatism by tapping a long-lived mantle plume. We also thank J.R. Chiarenzelli (St. Lawrence University) and S.P. Regan (United States Geological Survey and University of Alaska Fairbanks) for perceptive reviews.

References cited


Survey, Assessment Report 9566, 13 p., 5 maps (1:10,000 and 1:20,000 scale), 3 cross sections (1:500 and 1:1,000 scale), 4 appendices.


carbonatitic rocks: geochemistry, crystal chemistry and significance for constraining the sources of carbonatites. Chemical Geology, 235, 138-160.


Dawson, G.M. 1886. Geological and Natural History Survey of Canada, Annual Report 1, 122B-124B.


Ernst, R.E., and Bell, K., 2010. Large igneous provinces (LIPs) and carbonatites. Mineralogy and Petrology, 98, 55-76.


Hofmann, A.W., 2014. Sampling mantle heterogeneity through


Montana, 404 p.
Mitchell, R., Chudy, T., McFarlane, C.R.M., and Wu, F.-Y., 2017. Trace element and isotopic composition of apatite in carbonatites from the Blue River area (British Columbia, Canada) and mineralogy of associated silicate rocks. Lithos, 286-287, 75-91.
Pell, J., 1987. Alkaline ultrabasic rocks in British Columbia:
Scammell, R.J., 1993. Mid-Cretaceous to Tertiary tectonometamorphic history of former mid-crustal rocks, southern Omaneca belt,
Appendix 1. Downhole lithological and geochemical profiles for selected elements and elemental ratios at Upper Fir

- **Overburden**
- Regional gneiss and schist
- Regional amphibolite
- Calcite-pyroxene amphibolite (Na-fenite)
- Glimmerite (phlogopite fenite)
- Dolomite carbonatite (granoblastic or gneissic fabric)
- Dolomite carbonatite (granoblastic or gneissic fabric with calc-silicate veins)
- Dolomite carbonatite (foliated porphyroclastic fabric)
- Calcite carbonatite
Appendix 1 a) Diamond drill hole F10-198 (72.33-129.56 m; collar NAD83 UTM zone 11 coordinates: 352813 m E, 5796902 m N; elevation: 1201 m; plunge: 60° E).
Appendix 1 b) Diamond drill hole F10-208 (7.32-230.19 m; NAD83 UTM zone 11 coordinates: 352967 m E, 5796622 m N, 1268 m, 75° E).
Appendix 2. Road log

Fig. A2.1. Field trip stops and route (Highway 1 and Highway 5) from Vancouver to Blue River, east-central British Columbia.
Day 1. Vancouver to Blue River (Fig. A2.1)

0.0 km. Assemble at the front entrance to Vancouver Convention Centre West (Burrard St at Canada Pl) at 7:30 AM. We leave at 8:00 AM.

Travel east from Vancouver Convention Centre to get on Trans-Canada Hwy (BC-1E)

35.5 km. Hwy 17/15 overpass. Continue southeast on Trans-Canada Hwy (BC-1E) for about 36 km to the Hwy BC-17/15 overpass.

0.0 km. Reset odometer at the merging immediately past the overpass.

40.0 km. Coast belt. Heading northeast, our route follows the Fraser River valley, which is filled with unconsolidated deltaic, fluvial, and glacial deposits. The valley separates the rugged Coast Mountains to the north and the Cascade Mountains to the southeast (Fig. 2b). These mountains form a chain 100-200 km-wide and 1600 km-long along the mainland coast from the Yukon Plateau near latitude 60° to the North Cascades near latitude 48°30′ and coincide with the core of the Pacific or Coast-Cascade orogen (Monger, 2014). Their rugged topography reflects differential uplift of up to 4 km and glacial-fluvial erosion in the past 10 million years.

For the next 135 km, we pass through uplifted and dissected Middle Devonian to Miocene rocks comprising: 1) remnants of island arcs and oceanic lithosphere accreted to, and displaced along, the margin of the Ancestral North America, and 2) overlap basins and younger continental arcs built on older rocks due to convergence between the North American and Pacific plates. The core of the Coast-Cascade orogen straddles the Coast morphogeological belt between the Insular (western) and Intermontane (interior) belts of accreted offshore crust (Figs. 1 and 2a). The Coast Belt is a tectonic 'welt' resulting from Mesozoic collision of the Insular arcs with the Intermontane terranes and subsequent crustal thickening, plutonism, and differential uplift, related to eastward subduction, orthogonal shortening, northwest-southeast transcurrent displacement, and late transtension and extension. Middle Jurassic to Eocene granitoid plutons make up about 80% of the Coast Belt and, together with the high-grade metamorphic rocks (Fig. 2c), comprise the Coast Plutonic Complex, which developed as the Insular island arcs above west-dipping subduction during their accretion and collision with the Intermontane terranes at the leading edge of Ancestral North America, and subsequent continental front arcs above the east-dipping Farallon slab since ca. 155 Ma (Sigloch and Mihalynuk, 2017). Younger granitoid plutons and volcanic rocks (34-0 Ma) of the Coast Belt south of latitude 51° mark the northern end of the Cascade magmatic arc, which extends to northern California above the currently subducting Juan de Fuca and Gorda plates. The syntectonic plutons ‘stitch’ the accreted Insular superterrane to the west and the Intermontane superterrane to the east, along with several small terranes and overlapping basins wedged between the two superterranes in the southern Coast belt. These small terranes comprise remnants of Middle Devonian to Late Jurassic island arcs (Nooksack-Harrison and Chilliwack terranes) and Mississippian to Middle Jurassic ocean (Bridge River, Methow and Cadwallader terranes; Fig. 1). Late Jurassic to Early Cretaceous marine siliciclastic rocks of the Tyaughton-Methow basin overlap all three oceanic terranes.

88.3 km. Cascade magmatic arc. The western margin of a Miocene, quartz monzonite-granodiorite-tonalite pluton (Mount Barr batholith; 208 km³), related to the Cascade magmatic arc, is exposed alongside the highway southwest of the Harrison Lake. The pluton intrudes marine siliciclastic strata and arc-related volcanic rocks of the Gambier Group (Early Cretaceous) and ‘stitches’ the north-south Fraser fault, which accommodated ca.140 km of dextral displacement in the Eocene.

138.8 km. Oceanic terranes. We pass by Mississippian to Middle Jurassic rocks of the Bridge River, Cadwallader and Methow terranes, made up of disrupted oceanic crust, lenses of serpentinitized mantle and Late Triassic blueschist, argillite, radiolarian chert, and local arc-related volcanic and siliciclastic strata (Monger, 2014). These ocean floor remnants overlying Late Jurassic to earliest Cretaceous marine elastic rocks of the Tyaughton-Methow basin comprise an accretionary complex wedged between the Insular and Intermontane superterranes in the southern Coast belt (Sigloch and Mihalynuk, 2017). Synorogenic (ca. 95 to 45 Ma) granitic plutons related to eastward subduction of the Pacific plates under North America cut all three oceanic terranes and the overlapping siliciclastic strata.

161.0 km. Coquihala Pass. At the Coquihala Summit recreational area, an Eocene granodiorite pluton (208 km³) intrudes Permian ophiolite that is overlain by Late Triassic to Middle Jurassic arc-related volcanic and siliciclastic rocks of the Cadwallader terrane. These rocks are thrust over Late Jurassic to Early Cretaceous marine siliciclastic strata of the Tyaughton-Methow basin, and overlain by Eocene conglomerates and sandstones. The view from the rest area features Zopkios Peak to the north and Markhor and Needle peaks to the south.

165.0 km. Coast belt boundary. Heading northeast, our route passes through a northwest-southeast belt of Middle Jurassic to mid-Cretaceous granitoid plutons marking the transition from the Coast belt to the Intermontane morphogeological belt, which occupies the central part of the Canadian Cordillera (Fig. 2a). As we leave the rugged Cascade Mountains behind, this transition corresponds with the physiographic change to the more subdued, gently rolling topography of the Interior Plateau (Fig. 2b).

179.1 km. Intermontane Belt. We pass through volcano-sedimentary rocks of the Nicola Group (Late Triassic) at the leading edge of the Intermontane arc superterrane. Continuing northeast for the next 210 km, our route (Hwy 5) passes through Late Devonian to Middle Jurassic volcanic and sedimentary rocks that are cut by numerous Late Triassic to Early Jurassic granitoid plutons of the Quesnel island arc.
terranes produced some of the largest porphyry Cu-Mo-Au on the Paleozoic basement of the Stikine and Quesnel. Late Triassic to Early Jurassic magmatic arcs superimposed underlying the Omineca belt to the east.

Unlike the flanking Coast and Omineca tectonic welts, the Intermontane Belt is underlain mainly by subgreenschist facies grade rocks (Fig. 2c; Monger, 2014). The Slide Mountain, Yukon-Tanana, Stikine, Quesnel, and Cache Creek terranes (collectively the Intermontane superterrane; Fig. 1) comprise remnants of ocean basins, microcontinents, island arcs, and forearc accretory complexes that evolved outboard of the western margin of Ancestral North America in a setting perhaps similar to the modern western Pacific margin before they finally accreted onto the western edge of Laurentia by the Middle Jurassic (ca. 174 Ma). Middle to Late Devonian subduction along the western Laurentian margin was overprinted by Late Devonian to Mississippian widespread, plume-related carbonatite and alkaline magmatism and coeval rifting that led to the Slide Mountain ocean (up to 3000 km wide) between the continental margin and continental fragments (Yukon-Tanana terrane) detached from Laurentia by the Early Permian (e.g., Nelson et al., 2013; Monger, 2014). Juvenile intra-oceanic arcs (Quesnel and Stikine terranes) and those superimposed on pericratonic terranes formed above east-dipping subduction zones by westward drifting of the frontal arcs into the proto-Pacific ocean (Panthalassa) in the Mississippian through the Early Permian. The Slide Mountain ocean collapsed by west-dipping subduction by the end of the Permian, and the North American plate drifting westward from spreading of the new Atlantic ocean was colliding with the Intermontane arc terranes throughout the Triassic and Early Jurassic (Nelson et al., 2013).

These arc-continent collisions trapped an oceanic accretionary complex (Cache Creek terrane) between the Stikine and Quesnel arc terranes, which form an orocline fold hinged in the Yukon-Tanana terrane (Mihalynuk et al., 1994). The Cache Creek accretionary complex contains slivers of blueschist, abundant Pennsylvanian to Late Triassic carbonate reef caps (Fennell Formation; Mississippian to Permian). diabase, gabbro, and local chert with basal siliciclastic volcaniclastic rocks and minor carbonate rocks of the Harper Peaks Jasper). We pass by mainly intermediate to felsic volcanic rocks of the Spences Bridge Group (Early Cretaceous).

235.5 km. Exit 290 to Merritt. Take exit 290 toward Merritt City centre. Turn left on Princeton-Kamloops Hwy (BC-5A N)/Voght Street (signs to Merritt). Continue 600 m and turn right onto Belshaw St, turn right again at next intersection (50 m) onto De Wolf Way. Park on the left side at gas stations and fast food restaurants.

302.6 km. Iron Mask batholith. The view ahead is of the Iron Mask monzonite-diorite-syenite batholith (Late Triassic to Early Jurassic).

312.9 km. Kamloops basin. The hills on the left (north) are underlain by Eocene volcanic and sedimentary rocks deposited in the Kamloops basin, which is bounded by northwest-trending normal faults.

321.5 km. Take exit 374 to Yellowhead Hwy 5 North (Sun Peaks Jasper).

337.1 km. Quesnel basement. We drive by deep-water sedimentary rocks interbedded with mafic to felsic volcanlastic rocks and minor carbonate rocks of the Harper Ranch Group (Devonian to Permian; possible Paleozoic basement to Quesnel terrane; Mathews and Monger, 2005), that are overlain by rocks that may be part of the Nicola Group (Triassic) and are cut by Late Triassic to Early Jurassic monzonite to syenitic plutons.

386.7 km. Slide Mountain terrane. Dark grey cliffs seen alongside the highway (to the east) immediately north of the town of Barriere are underlain mainly by pillow basalts, diabase, gabbro, and local chert with basal siliciclastic rocks of the Fennell Formation (Mississippian to Permian). They represent the Slide Mountain terrane that makes up discontinuous slivers along the eastern edge of the Quesnel and Yukon-Tanana arc terranes thrust over the continental margin (e.g., Roback et al., 1994; Paradis et al., 2006). This
oceanic suture marks the boundary between the accreted
Intermontane superterrane to the west and the margin of
Ancestral North America to the east. Continuing north, our
route follows along the eastern edge of the Quesnel terrane
and passes through the Slide Mountain terrane just south of
the town of Clearwater. We will stop to look at basaltic rocks
of the Fennell Formation.

420.4 km. Stop 1. Fennell Formation. 51°28'8.30"N, 120°10'40.18"W. Park on the right side of the highway.

Pillow and massive basalts of the Fennell Formation (Fig
A2.2). Chloritization gives the lavas a mottled, maroon to
light green colour. Minor phlogopite is present. Pillows
(24 x 17 cm) are brecciated and are crosscut by dark grey,
fine-grained to aphanitic diabase dikes (0.5 m wide; 285°
/47° NE) and quartz veins (1 cm wide). An aplitic dike (1-4
cm) with quartz veinlets cores a subvertical shear zone
(up to 12 cm wide) striking 75°-83° NE. The rocks are also
cut by subvertical cleavage (041° /70° SE) and fractures
(003° /43° SW and 296° /60° NE). These ophiolitic rocks
represent remnants of the floor of a 3000 km-wide ocean that
initiated with rifting along the active margin of Ancestral
North America (then Laurentia) and coeval, widespread
carbonatite and alkaline magmatism in the Late Devonian to
Mississippian. The Slide Mountain ocean has been interpreted
as a back-arc basin formed by lithospheric extension during
east-dipping subduction of the Proto-Pacific plate under the
western margin of Laurentia (e.g., Nelson et al., 2013).

The carbonatites of the Blue River area that we will see
on this field trip have isotopic signatures identical to those
of worldwide carbonatites formed by deep-mantle plumes
(Mitchell et al., 2017; Rukhlov et al., 2018). We suggest
that Slide Mountain ocean back-arc extension also triggered
emplacement of the Blue River carbonatites, and that these
rocks were sourced from a deep-level mantle plume that was
tapped episodically since the Neoproterozoic (at ca. 810-700,
500, and 360-330 Ma).

445.1 km. Proceed north on the Yellowhead Highway 5. At
roundabout turn left (third exit) and proceed northwest on the
Clearwater Valley Rd.

Laurentian margin. As we continue north, our route leaves
behind the Intermontane terranes and enters the Shuswap
Highland and then the high Cariboo and Monashee mountains
of the Omineca belt, the inner tectonic welt of the Cordillera,
which is separated from the Foreland belt to the east by the
Rocky Mountain Trench (Monger et al., 1982). Both the
Omineca and Foreland belts are underlain by Proterozoic to
Paleozoic successions formed in intracratonic basins before
the breakup of the supercontinent Rodinia and as passive
margin deposits on the flank of Laurentia after breakup
(Goodfellow et al., 1995; Cecile et al., 1997; Colpron et
al., 2002; Mathews and Monger, 2005). Mostly distal shelf
to deep-water strata in the Omineca belt have undergone
predominantly ductile deformation and greenschist- to upper
amphibolitic-grade metamorphism from the Early Jurassic

Fig. A2.2. Pillow basalt of the Fennell Formation (Mississippian to Permian), Slide Mountain terrane.
through to the Paleocene, whereas the equivalent platform and shelf facies of the Foreland belt were deformed by thin-skinned folding and thrusting with little metamorphism mainly in the Cretaceous through the Paleocene (e.g., Bally et al., 1966; Price and Mountjoy, 1970; Evenchik et al., 2007; Simony and Carr, 2011; Panà and van der Pluijm, 2014). Synorogenic intrusions and local volcanic rocks are widespread in the Omineca belt, which was close to the collision zone between the offshore terranes and the Laurentian craton in the Early Jurassic. Crustal thickening, transpression, and uplift in the Omineca belt continued from the Middle Jurassic to the Paleocene. This was followed by transtension and uplift in the Eocene, which exhumed mid-crustal assemblages and, locally, crystalline cratonic basement in the southern Omineca belt (Parrish et al., 1988; Simony and Carr, 2011).

455.3 km. Turn left at Y intersection about 100 m past the bridge across Spahats Creek and drive 90 m. Turn left again and drive 130 m. Proceed right at Y intersection for 170 m to the Spahats Falls parking lot. **Stop 2: 51°44’9.92”N; 120°0’39.30”W. Clearwater Valley/Spahats Falls.**

Walk along a trail to the viewing platform to see Spahats Falls (73 m high), which cascade from alkali basaltic lava flows of the Clearwater unit (~400 Ky) that overlie unconsolidated sands and gravels seen near the base of the falls (Fig. A2.3; Hickson and Vigouroux, 2014). The lack of erosion or soil development between successive lava flows suggests rapid filling of the paleo-Clearwater valley by the lava. The volcanic activity of the Wells Gray field spans from about 3.5 Ma to recent times (Cannings et al., 2011). Along with the vast outpourings of basaltic lava across the Interior Plateau since the Miocene, it is attributed to a mantle hot spot below western North America (Cannings et al., 2011). At Wells Gray, basaltic magma probably used crustal-scale dextral faults along the eastern edge of the Quesnel terrane (Hickson and Vigouroux, 2014). Stop for **45 min.**

Return east 290 m from Spahats Falls parking lot. Turn right and continue southeast 86 m, turn right on the Clearwater Valley Rd heading south. Drive 10.2 km back to Clearwater. At roundabout, take the 3rd exit to rejoin Yellowhead Hwy 5 heading north.

**Drive to Blue River.** We continue past Devonian to Mississippian volcanic rocks and Neoproterozoic to Paleozoic metamorphic rocks (Eagle Bay assemblage, Snowshoe Group and Mica Creek succession) formed on the Laurentian margin and intruded by syntectonic Jurassic to Paleocene granitic plutons (Mathews and Monger, 2005).

573.8 km. **Arrive at Blue River.** Turn right (SE) on Harrwood Dr. and proceed 71 m, turn left (NE) on Harrison Rd. and proceed 78 m to the Mike Wiegele Helicopter Skiing main lodge.

We will check in at the front desk and get some rest before dinner at the Saddle Mountain restaurant across the highway at **7:00 PM.**

**Day 2. Blue River to Upper Fir deposit and back to Blue River.**

7:30 AM: breakfast at Saddle Mountain restaurant 150 m from Mike Wiegele Helicopter Skiing main lodge across the Hwy 5 (about 5 min walk).

8:45 AM: assemble by the vehicles in front of the main lodge. We depart at **9:00 AM.**

0.0 km. From the main lodge, turn right and drive northwest for 71 m. Turn right and proceed northeast on Yellowhead Hwy 5.

19.2 km. Turn right (to the southeast) onto Bone Creek Forestry Service Rd, and cross the bridge across the North Thompson River (Fig. A2.4).

21.1 km. Proceed southeast on Bone Creek Forestry Service Rd; keep left at Y-junction to continue northeast on Gum Creek Forestry Service Rd.
21.7 km. Continue NE on Gum Creek Forestry Service Rd (keep straight/left) at Y-junction with forestry permit 1830 Rd.

23.5 km. Turn right turn at Y-junction and proceed northeast on forestry permit L600 Rd.

25.9 km. This is an alternate landing site (52°17'49.38"N, 52°17'49.38"N); turn right (NE) to continue upslope on L600 Rd (see Fig 12).

25.98 km. Go left (NE) on L600 Rd at Y-jct with Shortcut Rd.

Fig. A2.4. Forestry service roads from Highway to Upper Fir, north of Blue River, east-central British Columbia.
26.06 km. Turn left (NE) on Upper Bulk Sample Rd. at Y-jct with 2900 Connector Rd.

26.33 km. Turn right on a connector road heading gently upslope to the northeast.

26.45 km. Continue straight (NE) on the connector road at Y-jct with 2900 Rd South Spur.

26.50 km. Turn right (NE) at Y-jct with 2900 Rd North and continue upslope on the connector road.

26.53 km. Arrive at Landing site (52°18'9.22"N, 119°9'27.03"W).

This is the start and end point of the Upper Fir hiking loop. We will leave vehicles here and hike to stops along deactivated trails and at excavation site BS-1 (Fig. 12). Take everything you will need during the day with you, i.e. backpacks, warm and waterproof clothing, footwear with ankle support, hard hats (to be worn at excavated sites), personal gear, lunch, snacks, water.

0.04 km. Walk southwest (downslope) on the connector road from Landing site. Turn right (north) on the deactivated 2900 Rd North. Proceed northeast on 2900 Rd North.

0.20 km. Stop 1  52°18'13.94"N, 119°09'26.88"W.
Contact between dolomite carbonatite and biotite gneiss (S_foliation = 323°/34° NE) cut by a pegmatite vein. Note abundant reddish-brown weathered carbonatite clasts in colluvium and grey till at the top of the road cut (Fig. A2.5). Stop for 25 min.

As we continue walking north to Stop 2, note biotite gneiss offset by a local-scale, chloritized plane (005°/60° W) with slickenfibres (214° SW) about 15 m from the Stop 1 (Fig. A2.6). This late brittle structure is probably related to Eocene extension marked by the steep, west-side down North Thompson normal fault, which is traced by the valley immediately to the west (Digel et al., 1998).

26.53 km. Arrive at Landing site (52°18'9.22"N, 119°9'27.03"W).

0.27 km. Stop 2:  52°18'16.24"N, 119°09'27.09"W.
Sheared glimmerite (phlogopite fenite) with dm vermiculite books separating porphyroclastic apatite-dolomite carbonatite below from amphibolite-biotite±garnet gneiss above. Banded, migmatitic semipelite (S_foliation: 65°/46° SE to 290°/6° NE) forms a 12 m-long recumbent isoclinal fold (F_hinges: 99°/26° to 104°/14°), cut by syntectonic pegmatite (10 cm wide) subparallel to axial surface (Fig. A2.7). Stop for 20 min.

Fig. A2.5. Stop 1. Weathered dolomite carbonatite clasts in colluvium.

Fig. A2.6. Late brittle fault surface with chlorite slickenfibres in biotite gneiss. The now-eroded block moved down dip, to lower right.

Fig. A2.7. Stop 2. Deformed and fenitized country rocks. a) Migmatitic amphibolite-biotite±garnet gneiss of the Mica Creek assemblage in a recumbent isoclinal fold, with coarse-grained glimmerite separating dolomite carbonatite (below) from country rocks (looking east). b) Dm vermiculite books from the hinge zone.
Proceed north from the end of 2900 Rd North to the upper bench of BS-1 cut. Watch your step and look out for falling rocks. Do not rush, and do not climb cliffs at the excavation site.

**0.30 km. Stop 3: 52°18’17.49”N, 119°09’27.23”W.**
Upper contact of a 20 m-thick carbonatite body mantled by calcite-bearing glimmerite (phlogopite fenite) interfingering with fenitized migmatitic amphibolite and biotite+garnet augen gneiss (S2, foliation: 342°/18° NE) forming a mesoscopic recumbent isoclinal fold with parasitic cm- to dm-size isoclinal leucosome folds (Fig. A2.8; F3, hinges: 124°/11’ to 136°/5’). Garnet porphyroblasts are up to 3 cm in diameter (Fig. A2.9). Note a series of carbonatite ‘eyes’ (10 cm wide by 20-30 cm long lenses), mantled by the inner diopside, intermediate coarse-grained amphibole, and outer apatite-calcite glimmerite zones, which are widest along the principal axis, in the fenitized country rocks. Stop for 20 min.

![Fig. A2.8](image1.png)
**Fig. A2.8.** Stop 3. Carbonatite-country rock relationships, BS-1 cut, Upper Fir. **a)** Highly strained, migmatitic semipelite with crumbly layers of carbonate-rich glimmerite above a 20 m-thick carbonatite body (looking north). Note readily oxidized Fe-rich dolomite in the carbonatite under a person and hinges of dm-scale isoclinal folds (closing south) gently plunging (5-11°) southeast. **b)** Close-up of the fenitized country rocks with coarse-grained, dolomite carbonatite ‘eyes’ (immediately above the hammer) in the isoclinal folds (looking east).

![Fig. A2.9](image2.png)
**Fig. A2.9.** Stop 3. Garnet-biotite-muscovite schist of the Mica Creek assemblage, BS-1 cut, Upper Fir.

**0.38 km.** Continue north (down) on ramp road past backslope outcrop of the dolomite carbonatite and a pile of carbonatite blocks, some of which contain segregations of very coarse-grained fluorapatite megacrysts (up to 3.0 x 5.5 cm) set in recrystallized ferroan dolomite matrix that readily oxidizes brown-red. Note aligned dark-green ferrikatophorite prisms locally forming segregation layers (Fig. A2.10). At the end of the ramp, turn left. Continue southwest on the BS-1 Rd.

![Fig. A2.10](image3.png)
**Fig. A2.10.** Very coarse-grained fluorapatite (Fap) megacrysts in recrystallized, ferrikatophorite (Fktp)-bearing dolomite carbonatite, BS-1 cut, Upper Fir. Ferroan dolomite readily oxidizes reddish-brown on the surface (compare with fresh dolomite in Fig. 17).
**0.41 km.** Turn left at Y jct. Continue south on the lower ramp road past calcite carbonatite in the backslope outcrop from the northern end of a side ditch (Fig. A2.11). Note undulating subhorizontal fractures and gravelly weathering of the calcite carbonatite lens (3 m thick), which pinches out about 6 m to the south. It contains 5-10% anhedral magnetite clumps (up to 8 x 6 cm), 15-20% dark-green ferrikatophorite prisms (2-3 mm), and 5-10% fluorapatite. In contrast, more competent dolomite carbonatite underneath has strongly aligned (117°/4°) ovoid apatite porphyroclasts (3 x 10 mm; 20%) and light-green richterite (15%), with accessory black pyrochlore octahedra (2 mm) set in a dark-grey, fine-grained, recrystallized matrix. Segregation layer (298°/55° NE) of aligned (135°/17°) amphibole (50%) and apatite (25%) contains magnetite porphyroclasts (2 x 0.7 cm) and euhedral brown-red pyrochlore crystals (5 mm).

**Fig. A2.11.** Gravelly weathered, foliated calcite carbonatite above more competent dolomite carbonatite.

**0.46 km Stop 4:  52°18'17.18"N, 119°9'29.65"W.  (Lunch, 90 minutes).**

Approximate lower contact of the carbonatite and subhorizontal pegmatite vein with biotite crystals up to 2 x 15 cm, underlain by country gneiss.

The pegmatite can be traced for at least 13 m from north to south. Note massive garnetiferous amphibolite with folded black amphibole segregations and late clinopyroxene-amphibole veins (Fig. A2.12). Sodium-rich amphibolites associated with carbonatites at Upper Fir contain abundant titanite, Mg-Fe³⁺-Na-rich hedenbergite, calcic to sodic-calcic amphibole, calcite, albite, and minor K-feldspar and fluorapatite (see Figs. 5i and 16). They represent Na-fenites on the outer metasomatic envelop around the inner glimmerite (phlogopite fenite) mantle of the carbonatite bodies.

Examine various dolomite carbonatites exposed in BS-1 (Figs. 9, 10a, A2.13), with fresh olive-green ferroan dolomite porphyroclasts, aligned amphibole (70%; up to 2 x 4 cm) and common segregation bands (342°/32° NE) with abundant ovoid fluorapatite (116°/20’, 119°/26’, 120°/16’) + pyrochlore + ferrocolumbite, S₂ foliation (25°/41° SE, 15°/42° SE, 0°/23° E) overprinted by high-strain zones (35°/50° SE), coarse-grained calc-silicate skarn veins (55°/40° NW), pyrrhotite veins (3-13 mm), and strained magnetite porphyroclasts (1.5 x 5 cm).

**Fig. A2.12.** Stop 4. Examples of Na-fenites, BS-1 cut, Upper Fir.  
**a)** Garnetiferous amphibolite with Na-Ca amphibole, albite and Na-rich clinopyroxene.  
**b)** Holomelanocratic hornblendeite made up of magnesiohastingsite with minor interstitial carbonate.
Fig. A2.13. Examples of textures and mineralization in carbonatites at Upper Fir. a) Folded amphibole segregation in dolomite carbonatite, BS-1 cut (Stop 4). b) Coarse euhedral U-Ta-rich pyrochlore (Pcl) in foliated, porphyroclastic dolomite carbonatite, containing ferrikatophorite (Fktp) and fluorapatite (Fap), BS-2 cut. c) Coarse ferrocolumbite (Fcl) in gneissic dolomite (Dol) carbonatite containing abundant fluorapatite and ferriwinchite (Fwnc), Landing site 2. d) Ferrocolumbite concentrated in fluorapatite segregation in granoblastic, anchimonomineralic dolomite carbonatite, Landing site 2.

0.53 km. Retrace north along the lower ramp road. Turn left (SW) on BS-1 Rd at Y jct.

0.58 km. Proceed southwest on BS-1 Rd and turn left (south) on Upper Bulk Sample Rd at Y jct.

1.00 km. Continue south on Upper Bulk Sample Rd. Turn left (NE) on the connector road at Y jct.

1.12 km. Proceed northeast (upslope) on the connector road. Keep straight (left) on the connector road at Y jct with 2900 Rd South Spur.

1.16 km. Proceed northeast. Turn right (NE) on the connector road at Y jct with 2900 Rd North.

1.19 km. Continue northeast to vehicles parked at Landing site. We leave for Blue River at 4:00 PM.

Retrace to Yellowhead Hwy 5 by driving southwest down on L600 road, on Gum Creek FSR, and then northwest on Bone Creek FSR. Turn left (SW) on Yellowhead Hwy 5 and drive 19.1 km towards Blue River. Turn left (SE) on Harrwood Dr. Proceed southeast 71 m, then turn left (NE) on Harrison Rd and proceed 78 m to MWHS main lodge. Dinner at 6:30 PM at Saddle Mountain restaurant.
Day 3. Core shack in Blue River and return to Vancouver

7:30 AM: breakfast at Saddle Mountain restaurant.

Check out and assemble by the main lodge. We depart for Commerce Resources Corp.’s core facility in Blue River at 9:00 AM.

0.0 km. From Mike Wiegele Helicopter Skiing main lodge, turn right (NW) on Harrwood Dr.

0.07 km. Proceed northwest on Harrwood Dr. Turn left on Blue River East Fronting at intersection.

Drive 350 m southwest on Blue River East Fronting and continue 600 m straight south on Cedar St. to the Commerce Resources Corp. blue board in front of two white and blue buildings (field office) and a beige hangar (core facility) on the left side.

1.02 km. Stop 1: 52°6’17.76”N, 119°18’26.89”W. Commerce Resources Corp. core facility.

We will examine representative core intervals from the Commerce Resources Corp.’s Upper Fir drillholes F10-198 (72.33-129.56 m; collar NAD83 UTM zone 11 coordinates: 352813 m E, 5796902 m N; elevation: 1201 m; plunge: 60° E) and F10-208 (7.32-230.19 m; NAD83 UTM zone 11 coordinates: 352967 m E, 5796622 m N, 1268 m, 75° E). Note texturally and mineralogically diverse dolomite carbonatites and calcite carbonatites containing up to 0.87 wt% Ta, 12.29 wt% Nb, and 6.7 wt% P_2O_5, amphibole-diopside± calcite± phlogopite skarn veins, exocontact-metasomatic calcite-amphibole glimmerites (phlogopite fenites) and calcite-pyroxene amphibolites (outer Na-fenites) with up to 68 ppm Mo, unaltered country gneissess, schists, amphibolites of the Mica Creek assemblage, and late syntectonic pegmatites. See downhole profiles for details (Appendix 1). Stop for 1 hour.

We depart for Vancouver at 10:00 AM for arrival at Vancouver Convention Centre West by 7:30 PM.
Appendix 3. Upper Fir east-west cross sections (same location as Fig. 13a).

These cross sections show the 99th percentile assay results for selected elements from drill-core samples (typically 1 m intervals), gridded by the inverse-distance 3D interpolation method using GOCAD® software. (a, b, c, d, e by LiBO$_4$ fusion - HNO$_3$ digestion and inductively coupled argon plasma emission spectroscopy (a – c) or inductively coupled argon plasma mass spectrometry (d and e); f by 2HNO$_3$:2HCl:2H$_2$O (v/v) acid solution and inductively coupled argon plasma mass spectrometry): a) CaO/MgO (wt.%). b) total Fe expressed as Fe$_2$O$_3$ (wt.%). c) P$_2$O$_5$ (wt.%). d) Nb (ppm). e) Ta (ppm). f) Mo (ppm); also showing Mo assay results for individual drill-core intervals.
Fig. A3 b)
Fig. A3 d)