

# Geochemical Character of Neogene Volcanic Rocks of the Central Beetle-Infested Zone, South-Central British Columbia (NTS 093B, C)

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**KEYWORDS:** alkaline magmatism, geochemistry, Neogene volcanic rocks, Chilcotin group, Anahim volcanic belt

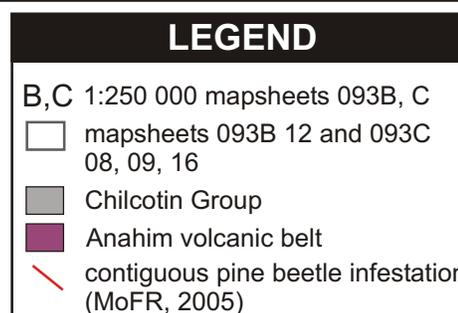
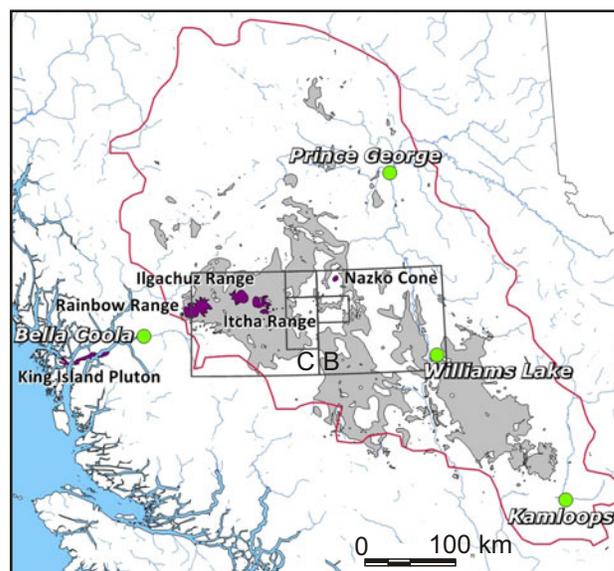
## INTRODUCTION

Volcanic rocks have formed along the length of the North American Cordillera during the Neogene and Quaternary periods. Volcanism of this age is dominantly alkaline and extends throughout central British Columbia and the southern Yukon Territory. Distinct provinces of alkaline magmatism have been identified in BC, including the northern Cordillera volcanic province, the Chilcotin Group and the Anahim volcanic belt (e.g., Edwards and Russell, 2000). Of these, the latter two occur within the Beetle-Infested Zone and represent the youngest units mapped by Mihalynuk et al. (2008). A major-element geochemical investigation of the contemporaneous Anahim volcanic belt and Chilcotin Group basalts suggests that their compositions are controlled by a common process.

## PREVIOUS WORK

### *Anahim Volcanic Belt*

The Anahim volcanic belt is a well-studied east-trending chain of Neogene shield volcanoes and intrusions extending from coastal BC west of Bella Coola, into the interior of the province (Figure 1). There is a well-defined age progression from west to east along the belt (Souther, 1986). The oldest expressions of magmatism in the Anahim volcanic belt are the coastal Bella Bella dikes and the King Island pluton, dated at 14.5–12.5 Ma and 13–10.3 Ma, respectively. Discrete volcanic centres occur east of Bella Coola, including the Rainbow Range, the Ilgachuz Range, the Itcha Range and the Nazko cones, which range systematically in age from 8.7 to 0.34 Ma (Bevier, 1978; Souther, 1986; Souther et al., 1987). With the exception of the Nazko lavas, all volcanic activity along the Anahim volcanic belt began with early shield-building felsic eruptions, which were later capped by mafic flows (Bevier, 1978; Souther, 1984; Charland and Francis, 1993). These late-stage mafic lavas range in composition from hawaiite through alkali olivine basalt to basanite, with a trend of progressive Si-undersaturation in later eruptions (e.g., Charland and Francis, 1993). Work on the Itcha Range has



**Figure 1.** Location of fieldwork area with respect to the Chilcotin Group and Anahim volcanic belt volcanic centres.

shown that the hawaiite can be produced through the fractionation of olivine+clinopyroxene from primitive alkali olivine basalt at the base of the crust (Charland et al., 1995), while the basanite requires a distinct parental magma based on trace-element ratios (Charland and Francis, 1993).

### *Chilcotin Group*

The Chilcotin Group ranges in age from 34 to 1 Ma, with periods of high eruption rates at 16–14 Ma, 9–6 Ma and 3–1 Ma (Mathews, 1989). Extrusive rocks include subaerial and subaqueous basaltic lava and pyroclastic rocks, while minor gabbroic to basaltic plugs are interpreted to represent lava vents (Bevier, 1983). The basalts are transitional from alkali olivine basalt to quartz-normative basalt. The Chilcotin Group is broadly contemporaneous

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ous with both the Anahim volcanic belt and the calcalkaline Pemberton and Garibaldi volcanic belts to the southwest, and has been interpreted as having formed in a backarc basin behind the Cascadia-Garibaldi arc (Souther, 1977; Bevier, 1983). Once believed to cover up to 50 000 km<sup>2</sup> of the south-central interior of BC with an average thickness of 70 m (Bevier, 1983), recent mapping suggests a significantly smaller aerial extent (~36 500 km<sup>2</sup>), and topographic modelling shows that the basaltic cover is relatively thin (<50 m), except where the flows have filled paleochannels (Andrews and Russell, 2007; Mihalynuk, 2007).

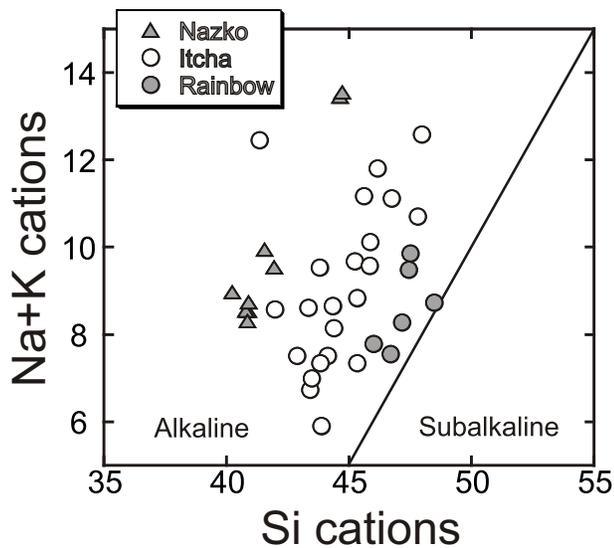
A number of important petrological observations were made by Bevier (1983) pertaining to the Chilcotin Group basalts, including that

- they are commonly aphyric, with olivine as the dominant phenocryst phase;
- the most magnesian olivine composition is Fo<sub>83</sub>, too low for the basalts to be primary melts of peridotite containing Fo<sub>89-92</sub>; and
- the major-element compositional array cannot be modelled by fractional crystallization.

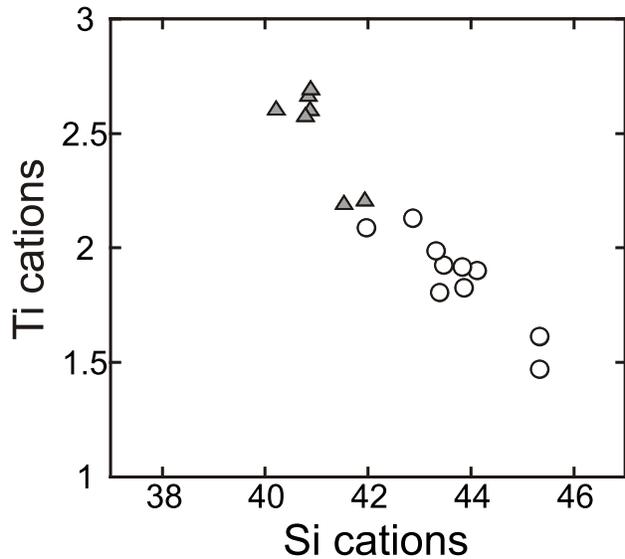
## GEOCHEMISTRY OF NEOGENE MAFIC LAVAS

### Anahim Volcanic Belt

Major-element geochemistry of the mafic lava flows of the Anahim volcanic belt reveals systematic changes in the degree of Si-undersaturation along the belt (Figure 2), paralleling changes in the geochemistry of late mafic flows at individual centres. The trends defined by individual volcanic centres are readily explained by fractionation of olivine+clinopyroxene±plagioclase, phases that are present as phenocrysts in mafic lava from the Anahim volcanic belt. If the data is filtered to minimize the effects of crystal sorting, however, a different story emerges. Lava flows having



**Figure 2.** Plot of Na+K versus Si (in cation units) showing trends among the Rainbow Range, Itcha Range, and Nazko cone volcanic centres of the Anahim volcanic belt. Data from Bevier (1978), Charland and Francis (1993), Charland et al. (1995) and Souther et al. (1987). The alkaline-subalkaline line is from Charland and Francis (1993).



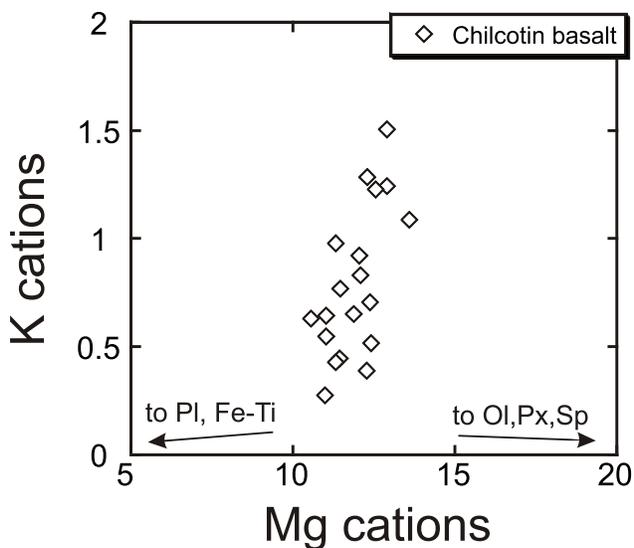
**Figure 3.** Plot of Ti versus Si, showing Itcha Range and Nazko cone lavas having >8 wt. % MgO (anhydrous).

>8 wt. % MgO form an array that cannot be explained by olivine+clinopyroxene fractionation (Figure 3).

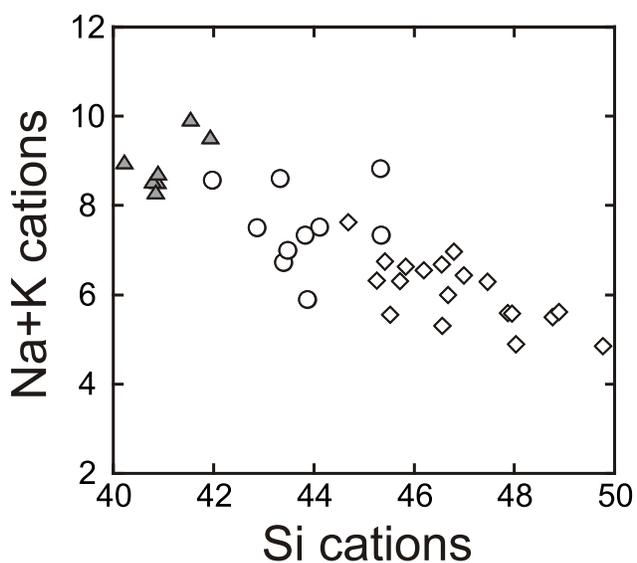
### Chilcotin Group

The Chilcotin Group basalts (also filtered for lava flows with >8 wt. % MgO) form poorly defined trends in most major-element space. Like the primitive lava of the Anahim volcanic belt, they display behaviour that is incompatible with olivine fractionation, or indeed any combination of olivine, clinopyroxene, plagioclase, spinel and/or Fe-Ti oxide fractionation (Figure 4). Likewise, the trend cannot be explained by varying degrees of partial melting of an anhydrous peridotite source, as the incompatible elements do not correlate with indices of partial fusion. Bevier (1983) took such behaviour to indicate chemical heterogeneity in the source region, requiring the Chilcotin Group basalts to have been generated by different degrees of partial melting of a source with a heterogeneous distribution of incompatible elements. The negative correlation of Mg (a proxy for the extent of partial melting) with K and Ti requires that regions that melted to the greatest extent (producing the highest Mg basalts) also had the highest incompatible-element concentrations (Figure 4). This trend could be explained by the melting of a hydrous K- and Ti-bearing phase, such as amphibole or phlogopite. Hydrous fluids have the effect of depressing the peridotite solidus, and the more water present, the greater the degree of partial melting at a given temperature (e.g., Hirose and Kawamoto, 1995). Regions of high amphibole or phlogopite concentration would have released more water upon melting, depressing the peridotite solidus to a greater extent, and yielding comparatively large melt fractions (those with the highest Mg contents). While there is little correlation between Pb isotopic ratios and K in the Chilcotin Group basalts, K and Sr isotopic ratios do correlate as expected. One possible explanation for the Pb isotopic ratios is that the process that created heterogeneous mantle metasomatism immediately predated the generation of Chilcotin Group melts without sufficient time for Pb isotopic differentiation.

A decreasing alkali concentration with an increasing degree of silica saturation is a geochemical characteristic that persists from the basalt array to the most primitive lavas of the Anahim volcanic belt, and suggests that a similar process governs their geochemistry (Figure 5). As with the Chilcotin subset, this greater trend cannot be explained by any reasonable fractionation assemblage, or by partial melting of an anhydrous peridotite source. Incompatible minor and trace elements also display colinearity, the further suggestion of a common process (Figure 6). With this in mind, the origins of the most Si-undersaturated members of the spectrum may shed light on the nature of the process responsible for this geochemical trend.



**Figure 4.** Plot of K versus Mg, showing Chilcotin Group basalts having >8 wt. % MgO (anhydrous). The locations of common anhydrous minerals are also shown. Abbreviations: Fe-Ti, magnetite-ilmenite; Ol, olivine; Pl, plagioclase; Px, pyroxene; Sp, spinel. Data from Bevier (1983).

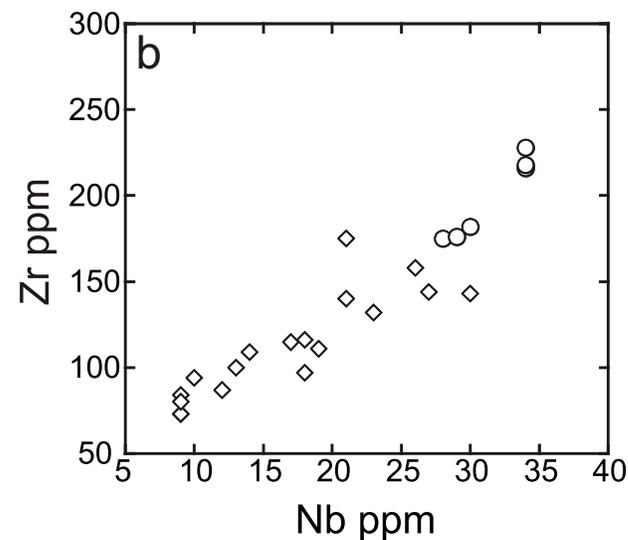
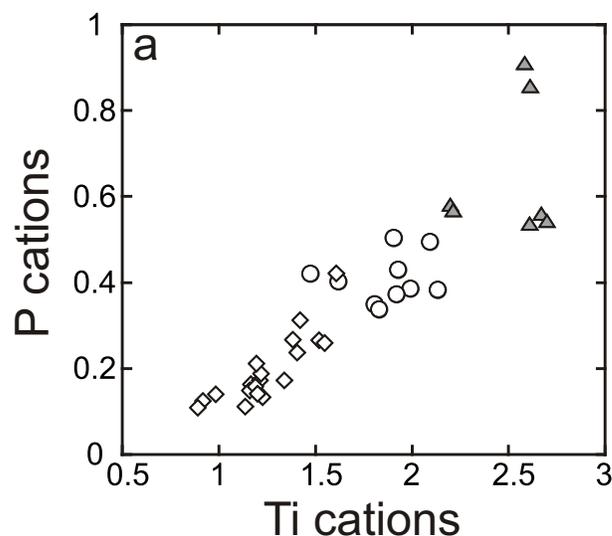


**Figure 5.** Plot of Na+K versus Si showing Itcha Range, Nazko cone and Chilcotin Group lavas with >8 wt. % MgO.

## GENERATION OF HIGHLY UNDERSATURATED MAFIC ALKALINE MAGMA

Strongly Si-undersaturated magmas such as olivine nephelinite and basanite have been attributed to a variety of processes, including low-degree melts of peridotite, melting of Si-poor garnet pyroxenite and melting of amphibole in metasomatized mantle (e.g., Francis and Ludden, 1995; Kogiso et al., 2003; Pilet et al., 2008).

Experimental work has demonstrated that low-degree partial melts of anhydrous peridotite can produce strongly Si-undersaturated liquids at elevated pressures (Hirose and Kushiro, 1993). Partial melting trends for anhydrous peridotite compositions, however, can neither replicate the Ca/Al ratios of highly undersaturated magma, nor the positive slope of Ca/Al in the olivine nephelinite-alkali olivine basalt array (Francis and Ludden, 1995). In addition, such



**Figure 6.** a) Plot of P versus Ti, showing Anahim volcanic belt lavas and Chilcotin Group basalts. Dataset as in Figure 5. b) Plot of Nb versus Zr. Dataset shows only Chilcotin and Itcha data because published Nazko data lacks trace elements.

source compositions cannot reproduce the elevated Ti content of alkaline magma in general (Pilet et al., 2008). Silica-deficient garnet pyroxenite has been suggested as the source of undersaturated magma, in part because of their elevated Ti contents. Below pressures of at least 50 kbar (~150 km depth), however, partial melts of such compositions do not reproduce the elevated K concentrations of alkaline magma (Pilet et al., 2008). Experimental investigations of the melting behaviour of amphibole at upper mantle conditions reveal that it melts incongruently to form clinopyroxene+spinel±olivine+liquid at pressures below 15 kbar, and to form clinopyroxene+garnet+liquid at pressures above 15–20 kbar (Ólafsson and Eggler, 1983). Experimental liquids are highly undersaturated, with nephelinitic compositions (Pilet et al., 2008). Francis and Ludden (1995) offer compelling evidence that the olivine nephelinite–basanite transition in the northern Cordillera corresponds to the exhaustion of amphibole in the source region. Experimental work by Pilet et al. (2008) shows that this model also explains the trace-element patterns of basanitic lava. As such, the melting of amphibole appears to best explain the major- and trace-element systematics of highly undersaturated mafic alkaline lava.

## PETROGENESIS

While there is compelling evidence for the role of amphibole in the generation of nephelinitic to basanitic magma, it remains to be seen how such highly undersaturated compositions are related to the alkali olivine basalts of the Anahim volcanic belt and the Chilcotin Group. It has been suggested that the basanite-alkali olivine basalt spectrum represents the two-step melting of a heterogeneous amphibole-bearing mantle assemblage, whereby initial amphibole-derived melts are progressively diluted with peridotite-derived basalt (Wilkinson, 1991; Francis and Ludden, 1995). More recently, it has been proposed that the spectrum largely reflects equilibration between amphibole-derived melts and subsolidus peridotite (Pilet et al., 2008). The silica content of primary melts is largely insensitive to the degree of partial melting and source composition, and instead is controlled by the pressure at which melting occurs (Hirose and Kushiro, 1993). Nephelinitic liquids produced by the melting of amphibole are out of equilibrium with surrounding peridotite at pressures below those at which nephelinitic melts would be generated from anhydrous peridotite (>50 kbar; extrapolated from data of Hirose and Kushiro, 1993). At pressures within the spinel stability field, nephelinitic liquids will essentially leach silica from orthopyroxene by transforming it into olivine and thereby increasing the Si content of the liquid to bring it into equilibrium with the surrounding mantle (Lundstrom et al., 2000). Experiments show that this reaction occurs on experimental timescales (Ólafsson and Eggler, 1983; Pilet et al., 2008). An important aspect of this reaction is that, given that the maximum pressure at which amphibole is stable in the mantle is ~30 kbar (Wallace and Green, 1991), strongly Si-undersaturated melts derived from amphibole must avoid equilibration with the surrounding mantle peridotite if they are to reach the surface unmodified.

If the compositional spectrum reflects the simple addition of silica to an originally nephelinitic melt, as required by the reaction with orthopyroxene to produce olivine, then it is to be expected that element ratios, such as K/Na and Mg#, should remain unchanged as the Si content increases.

For K/Na at least, this is not the case (Figure 7). The K/Na behaviour of the basanite–alkali olivine basalt (AOB) array indicates a progressive dilution of K by Na, most likely caused by the melting of clinopyroxene, suggesting that some degree of peridotite (or pyroxenite) partial melting has taken place. Likewise, the Na+K–Si and Ti–Si systematics of the basanite–AOB array cannot be explained by the addition of silica to nephelinite or basanite, but rather appear to reflect the dilution of nephelinite or basanite with relatively low-pressure (10–15 kbar) peridotite melts (Figure 8). The most suitable basaltic end members are near-solidus melts produced under hydrous conditions.

The basalt dilution model, however, has a major flaw: no ‘standard’ peridotite composition is capable of producing an end-member melt with the observed Fe contents (Figure 8c). The Mg# does not change significantly from nephelinite to Si-saturated basalt; in the dilution model, there ought to be a gradual increase in Mg# with increasing Si (Figure 8d). Perhaps coincidentally, the alkaline array clusters along a tie joining nephelinite with pure silica. Francis (1995) has shown that primitive intraplate lavas worldwide appear to require a source that is richer in Fe than the mantle source of mid-ocean ridge and subduction-related basalt. Therefore, either the mantle that is furnishing the Si-saturated end member is significantly richer in Fe than is indicated by Cordilleran peridotite xenoliths (e.g., Nicholls et al., 1982), or there is a decoupling of compatible and incompatible elements such that incompatible elements reflect dilution of the amphibole-derived melt with hydrous peridotite melts, while compatible elements reflect the reaction of nephelinitic melts with orthopyroxene.

This view provides an alternative explanation for why the Chilcotin Group basalts cannot be related by fractionation processes: their compositions may be unmodified by any crystal fractionation processes, and may instead be controlled either by dilution, by the addition of silica or both. The question naturally arises as to why there are no highly undersaturated compositions reported from the Chilcotin Group. This suggests that either the Chilcotin Group basalt source region had a low amphibole:peridotite melt ratio, or that amphibole-derived melts equilibrated with the surrounding peridotite at shallow depths. Conversely, the highly undersaturated AVB lavas must come from a source rich in amphibole, and must have travelled to the surface rapidly enough to avoid equilibration to higher silica concentrations.

## TECTONIC IMPLICATIONS

Neogene to Quaternary volcanism along the length of the Cordillera appears to be controlled by plate boundary interactions between the North American Plate and its outboard oceanic counterparts (Thorkelson and Taylor, 1989; Edwards and Russell, 2000). Paleosubduction models suggest that a spreading ridge flanked by the Kula and Farallon plates began to subduct beneath the North American Plate during the Late Cretaceous, since which time a slab window has developed beneath virtually the entire northern Canadian Cordillera (Figure 9; Thorkelson and Taylor, 1989). The Kula Plate has been completely subducted, while the Juan de Fuca and Explorer plates are all that is left of the Farallon Plate. A southern slab window formed during the Miocene, when the Farallon-Pacific Ridge subducted, and underlies much of the Basin-and-Range Province of the southwestern United States (Thorkelson

and Taylor, 1989). For the past 5 Ma, volcanism in the Basin-and-Range Province has been dominantly Si-undersaturated (Fitton et al., 1991). The spatial and temporal coincidence of alkaline magmatism and slab windows is therefore highly suggestive.

Because of its linear expression, the Anahim volcanic belt has been variably interpreted as the product of a mantle plume, a propagating rift and a slab window-related edge-effect (e.g., Bevier et al., 1979). Its spatial coincidence with the southern edge of the Cordilleran slab window makes the latter interpretation attractive (Figure 9). The trend of the Anahim volcanic belt does, however, parallel the projected trend of the Yellowstone plume track (Johnston and Thorkelson, 2000), implying that the Anahim trend is consistent with the North American Plate's motion over a steady-state hotspot. The Chilcotin Group has been interpreted as an ensialic backarc basin, forming behind the Pemberton and Garibaldi volcanic belts as a result of subduction of the Juan de Fuca Plate beneath North Amer-

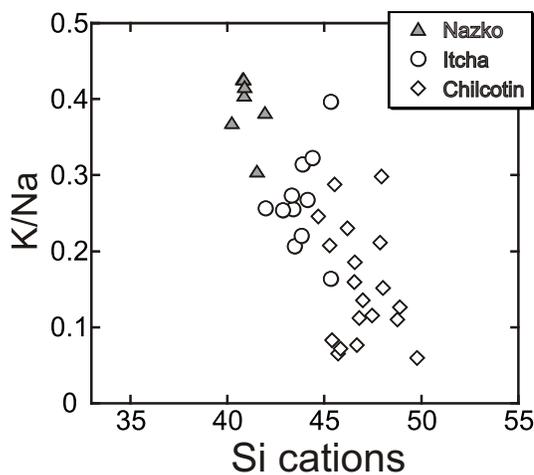


Figure 7. Plot of K/Na versus Si. Dataset as in Figure 5.

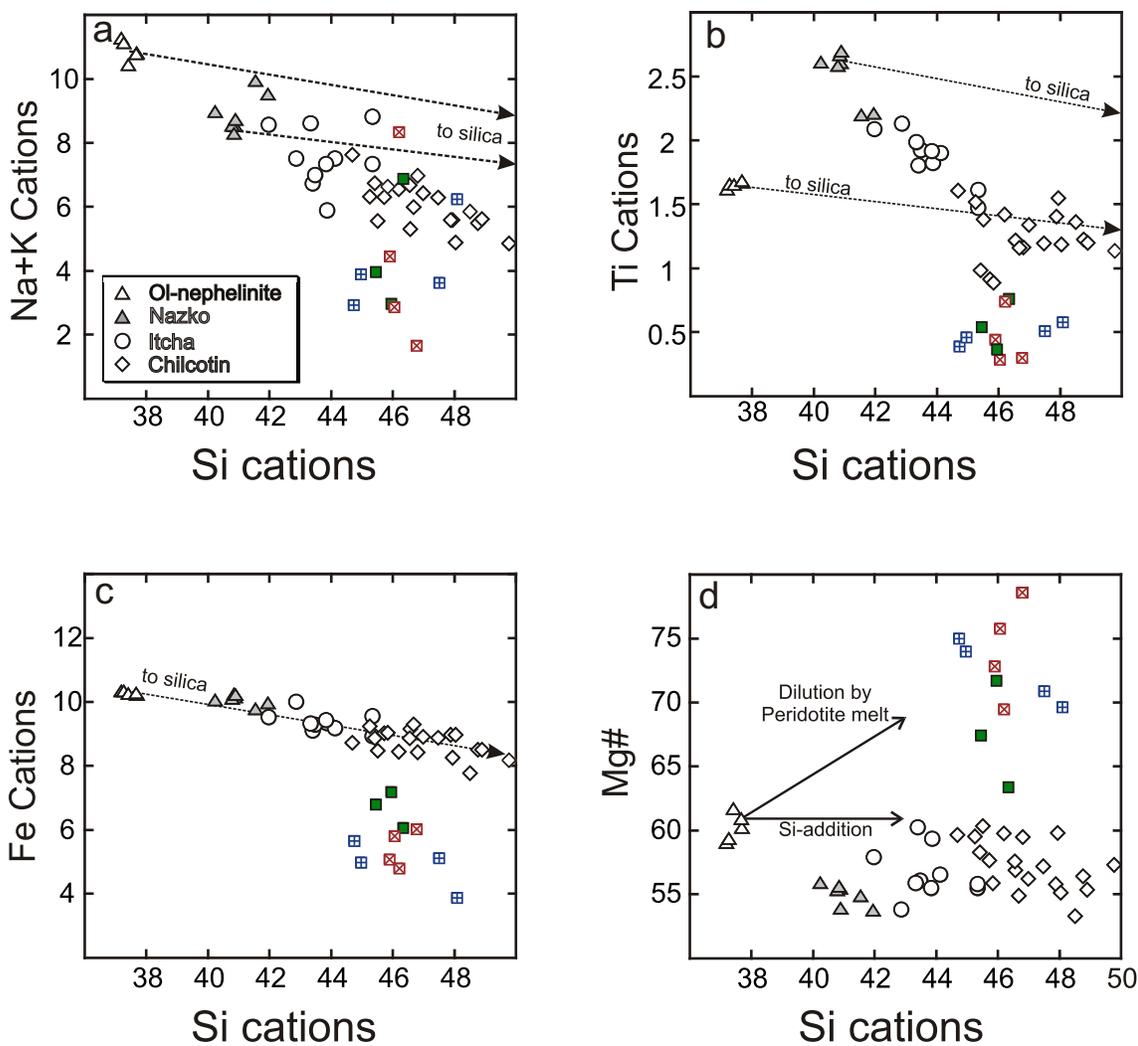


Figure 8. a) Plot of Na+K versus Si. b) Plot of Ti versus Si. c) Plot of Fe versus Si. d) Plot of Mg# versus Si (Mg# =  $100 \cdot \text{Mg}/(\text{Mg} + \text{Fe}_{\text{T}})$ ), with solid arrows showing trends expected from the dilution of nephelinite with peridotite melts, and by the addition of silica to nephelinite. Dataset as in Figure 5, with olivine nephelinite from Francis and Ludden (1995). Partial melts of spinel lherzolite at 10 kb also shown: solid green square, HK-66 (Hirose and Kushiro, 1993); open blue square with cross, KLB-1 (Hirose and Kushiro, 1993); open red square with x, KLB-1 hydrous (Hirose and Kawamoto, 1995). Dashed tie lines link olivine nephelinite and basanite with pure silica (100Si), indicating a Si-addition trend.

ica (Souther, 1977; Bevier, 1983). Thorkelson and Taylor (1989) suggest that, like the Anahim volcanic belt, the Chilcotin Group is also related to slab window edge-effects, although it is difficult to explain why the basalts should extend so far south of the projected slab window.

Because of the slope of the amphibole-out curve in P-T space, amphibole melting cannot be induced directly by decompression, and requires a thermal perturbation to drive the ambient mantle temperature out of the amphibole stability field (Figure 10). This can be accomplished by the upwelling of 'normal' asthenospheric mantle, without needing to infer the existence of an anomalously hot plume. In the Anahim volcanic belt, this upwelling is most likely generated by perturbation in the asthenosphere due to the foundering of the Farallon plate edge. The systematic age progression along the belt would then reflect the migration of this perturbation with continued subduction. If the interpretation of the Chilcotin Group as having formed in a backarc setting is correct, then frictional drag of the mantle wedge by the subducting Juan de Fuca Plate may create a minor convection cell in which the Chilcotin Group melt was produced above the upwelling arm. In either case, hot upwelling asthenospheric mantle comes into contact with amphibole-bearing lithospheric mantle, and in the course of thermal re-equilibration, the amphibole melts, generating nephelinitic to basanitic liquids and hydrous fluids (Figure 10).

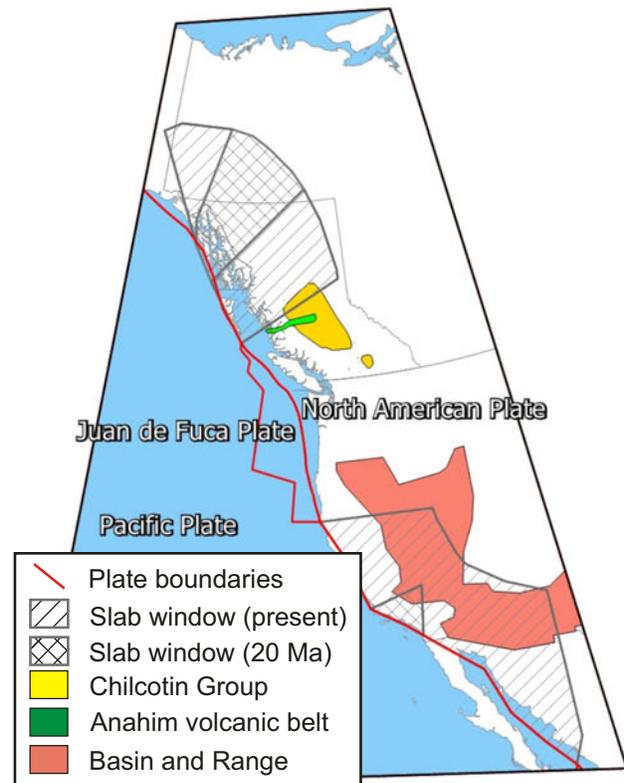
## CONCLUSIONS

Neogene alkaline volcanism in the central interior of BC requires a heterogeneous mantle source. The behaviour of incompatible elements such as K and Ti are best explained by varying concentrations of amphibole in the source regions of alkaline lava. Highly undersaturated alkaline compositions, such as nephelinite and basanite, appear to represent the spectrum of incongruent melts of amphibole (Francis and Ludden, 1995; Pilet et al., 2008). The coherent geochemical behaviour of primitive lavas ranging in composition from basanite to quartz-normative basalt in the Anahim volcanic belt and the Chilcotin Group suggests that a common process controls this compositional array. Incompatible element behaviour seems to require the dilution of amphibole-derived nephelinite and basanite with hydrous peridotite melts. Compatible elements, by contrast, suggest the reaction of nephelinite and basanite with orthopyroxene in order to attain equilibrium with mantle peridotite. The reason for the disconnection between compatible and incompatible element behaviour is unclear; an avenue for future work.

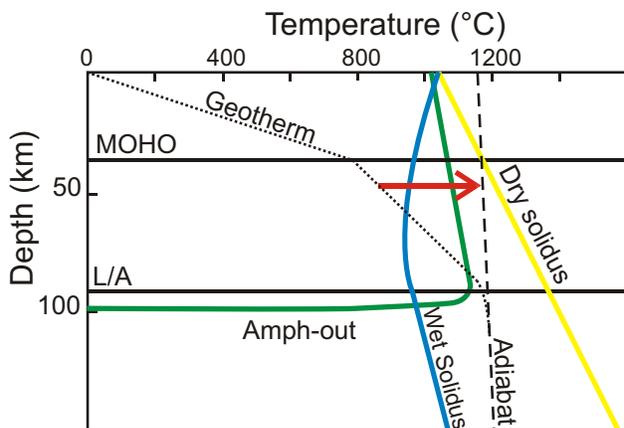
The breakdown of amphibole in the lithospheric mantle requires a thermal push. In the case of the Chilcotin Group basalts, this is most likely accomplished by extension and asthenospheric upwelling due to subduction-induced convection in the mantle wedge. In the case of the Anahim volcanic belt, this may be a result of asthenospheric upwelling caused by the sinking of the Farallon Plate edge, as suggested by Thorkelson and Taylor (1989).

## ACKNOWLEDGMENTS

Discussions with Graham Nixon, Dante Canil and Don Francis have helped to organize the ideas presented in this contribution. Graham Nixon also kindly reviewed the



**Figure 9.** Western North America, showing the approximate locations of slab windows at 20 Ma and at present, with respect to Neogene alkaline provinces mentioned in the text. Projected slab windows are from Edwards and Russell (2000) and Thorkelson and Taylor (1989).



**Figure 10.** Pressure-temperature phase diagram for the crust and upper mantle beneath central BC. Crustal geotherm, crust-mantle boundary and lithosphere-asthenosphere boundary from Harder and Russell (2006). Dry and water-saturated peridotite solidi and amphibole stability field adapted from Francis and Ludden (1995). Adiabāt gradient assumed to be 1.2°C/kbar. The red arrow shows the thermal effect of upwelling asthenospheric mantle impinging on cooler amphibole-bearing lithospheric mantle. Abbreviations: Moho, Mohorovicic discontinuity; L/A, lithosphere-asthenosphere boundary.

manuscript, although we absolve him of any association with the views of alkaline magmatism expressed herein.

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