Anomalous diamond in the Atlin-Nakina region, British Columbia: insights from heavy minerals in stream sediments

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Abstract – The sources of rare anomalous diamonds reported in northwestern British Columbia, southwestern Yukon and parts of Alaska are enigmatic. We carried out a heavy mineral survey of seventeen streams draining bedrock in the Atlin-Nakina region in northwestern British Columbia, to determine if high pressure igneous and metamorphic rocks exhumed in the area could be potential sources for the diamond. Non-magnetic heavy mineral fractions returned flakes of gold but no diamond. The magnetic fractions were examined optically and by electron microprobe analysis of key indicator minerals – olivine, orthoproxene, clinopyroxene, garnet, spinel and titanite. Our results suggest that detritus from a horizon of coarse pebble conglomerate in the Jurassic Laberge Group south of Atlin, is the most likely source of the anomalous diamond. Garnets and pyroxenes in the latter sedimentary unit were derived by rapid erosion of peridotite and eclogite bodies that were exhumed from depths approaching the diamond stability field during collision and Pliensbachian uplift throughout the northern Cordillera. Our results show evidence for glacial transport of detritus derived from the Laberge coarse pebble conglomerate, as well as from Neogene volcanics, which evidently covered a much wider area before the last glaciation.

Introduction

Economic diamond deposits are usually hosted in kimberlite and lamproite emplaced in Archean crustal provinces or as placer deposits eroded from these sources (Helmstaedt & Gurney, 1994). Diamonds have also been documented in more non-traditional settings. For example, some placer diamonds appear to have a source in orogenic belts that border Precambrian crustal provinces (LeCheminant & Kjarsgaard, 1995) and there is growing recognition of diamonds in crustal metamorphic rocks, attesting to the burial of rocks to considerable depth during collisional orogenesis (Liou *et al*., 2002). Diamond has also been reported in ophiolite and 'Alpine" ultramafic rocks (Bai *et al*., 1993; Kaminsky & Vaganov, 1977).

Several occurrences of 'anomalous' diamonds have been reported in southeastern Alaska, southwestern Yukon and northwestern British Columbia (Casselman & Harris, 2002). The source of such diamonds is not obvious, for there are no known Archean crustal provinces in these regions of the Cordillera nor evidence of diamond-bearing igneous rocks which could have eroded to produce detrital diamond. Drainage patterns in the region are not obviously sourced in any of the major Precambrian crustal provinces to the east of the Cordillera, and so alluvial transport by rivers from the interior craton is unlikely. Glacial transport from the craton is also improbable given that many regions in southwestern Yukon north of the Denali fault are reported to contain alluvial diamonds, and yet have not been glaciated.

The occurrence of microdiamonds in anomalous settings in orogenic belts elsewhere (LeCheminant & Bedard, 1995) begs the question whether diamonds in southeastern Alaska, southwestern Yukon and northwestern British Columbia are sourced in ophiolites, high pressure metamorphic rocks (blueschists and eclogites) or other unexpected sources in these regions. Mantle tectonite sections in ophiolite have been recognized in this part of the northern Cordillera (Ash , 1994; Canil & Johnston, 2002; Foster *et al*., 1994; Mihalynuk *et al*., 2003; Monger, 1975; Shellnutt *et al*., 2001) as have occurrences of blueschist and eclogite (Erdmer *et al*., 1998; Mihalynuk *et al*., 2004b). In addition, garnets and pyroxenes from orogenic garnet peridotite, equilibrated at pressures approaching the diamond stability field, occur as detrital minerals in Jurassic sediments to the southeast of Atlin (MacKenzie *et al*., 2005).

To address the above question, and ultimately to explain the source of anomalous diamonds in the northern Cordillera, we sampled stream sediments for heavy minerals in the Atlin-Nakina area of northwestern British Columbia. The Atlin area has a long history of placer gold mining operations, and a diamond occurrence was reported in one placer operation (Casselman & Harris, 2002). We wished to test if diamonds or indicator minerals associated with diamond and of a high pressure origin could be found in the stream catchments that source large areas of bedrock in the region.

Regional Geology

The Atlin-Nakina area is situated in the northern Cordillera astride the Cache Creek and Stikinia Terranes (Fig. 1a). The Cache Creek terrane is typified by an oceanic assemblage of massive limestone, ribbon cherts and ophiolite dominantly of mantle harzburgite tectonite, serpentinite mélange, minor gabbro and volcanic rocks. Sequences of chert and limestone accumulated from Mississipian to early Jurassic age. Felsic intrusions in the ophiolite have Permian crystallization ages (Mihalynuk *et al*., 2003). The Cache Creek terrane is separated from strata of the Laberge Group to the west by the Nahlin fault. The Laberge Group is a sequence of early to middle Jurassic turbidites which stretches north into Yukon and is known there as Whitehorse Trough (Hart et al, 1995; Johannson *et al*., 1997; English *et al*., 2005;). Post-tectonic granitoid plutons of the Fourth of July and the Surprise Lake suites intruded the Atlin-Nakina region at 172 Ma and at 84 Ma, respectively (Mihalynuk, 1997). An Eocene (55 Ma) continental arc built upon the former rocks is represented by bimodal and coeval volcanic and hypabyssal intrusives of the Sloko Group to the south and west of Atlin (Aitken, 1959; Mihalynuk, 1997). Small isolated cones and flows of alkaline lava of Neogene age, some showing evidence for eruption into ice, occur mainly in the north and eastern sections of the study area (Fig. 1b) and are part of the Northern Cordilleran Volcanic Province (Edwards & Russell, 1999).

Methods

Fifteen streams were sampled in map areas NTS 104N and K, southeast of Atlin (Fig. 1b). Two samples were also collected from the tailings of sluices in commercial placer operations near Atlin on

McKee, Wilson, and Feather creeks. Control samples were also collected in Wilson and Feather creeks from below the placer workings.

Sediment samples were taken in streams where flow gradients would efficiently concentrate heavy minerals in the bedload. Sediments were sieved at the sampling site to grain sizes less than 2 mm and the resultant samples (4 – 8 kg) processed for heavy minerals at Vancouver Indicator Processors Inc. Samples were wet screened to less than 0.25 mm fraction, passed through a magnetic separator operating at 2.1 Tesla, and underwent two steps of heavy liquid separation to specific gravities greater than 3.33.

Magnetic heavy mineral fractions were hand-picked at the University of Victoria whereas non-magnetic heavy mineral fractions were examined for diamond and gold by I. & M. Morrison Geological Services. In each magnetic heavy mineral fraction, 100 grains of each mineral were picked, mounted in epoxy, and polished. Major element compositions of subsets of these grains were determined by electron microprobe analysis at the University of British Columbia. Operating voltage was 15 kV and beam current 40 nA with elements counted for 20 seconds on peaks.

Results

Non-magnetic Fraction

No diamonds occurred in any of the non-magnetic heavy mineral fractions. Gold flakes up to 1 mm in size were recognized in sediments from north of Hard Luck Peaks, Mt. Mt. Nimbus and in Feather Creek.

Magnetic Fraction

The major minerals identified in the magnetic heavy mineral fractions were olivines, pyroxenes, garnets, spinels, and titanites, with lesser rutile and ilmenite. The subdvision into different groups of the same mineral was based in part on optical examination, but mainly on mineral chemical observations that are described in more detail below. Mineral chemical data are given in Table 1 ***(in the final version Table 1 will be as a electronic database or data repository item)***.

Olivine

Olivine was identified optically in heavy mineral fractions by its light green colour and conchoidal fracture. In some cases, olivine preserved remarkably well-defined crystal habit and faceted faces, with spinel inclusions, indicative of an igneous origin, and little abrasion during transport from a proximal source rock.

Olivine in mantle tectonite from ophiolite, incldudign that from the Atlin area (Ash, 1994) has a high Mg/(Mg+Fe) (Mg# > 0.89) due to the fact it represents lithosphere formed as a residue remaining from basalt extraction. Olivine with Mg# greater than 0.89 was sampled at Goldbottom Creek, Sloko River, Peridotite Peak, Scarface Mountain and Nahlin Mountain (Fig. 2). With the exception of Sloko, all these locations are in drainages from massifs of mantle harzburgite (Fig. 1b). The Sloko location drains a ridge of the Laberge Group where mantle-derived garnets, pyroxenes and eclogite clasts are recognized in outcrop (MacKenzie *et al*., 2005). Most of the olivines sampled in this study from all other locations, however, were distinctly lower in Mg# than typical mantle olivine, and not unlike those from cumulate rocks in the lower crustal sections of ophiolite or as phenocrysts in basalts (Fig. 2).

The CaO content of olivine varies mainly with temperature and bulk composition, and so can be a useful petrogenetic indicator. Olivines from mantle tectonites and mafic and ultramafic cumulate rocks have low CaO (< 0.1 wt%) due to the low temperatures and the Ca-poor nature of the source rock from which they crystallized and last equilibrated. In contrast, olivine phenocrysts in basalt, which are quenched from higher temperatures, have much higher CaO contents. The great majority of olivines in concentrates from this study overlap with the CaO content of volcanic olivines, as typified by those from the nearby Neogene Llangorse lavas (Fig. 3).

Orthopyroxene

Orthopyroxene was poorly represented in the heavy mineral fraction sampled, occurring at only five locations. The $Kd_{Fe\text{-}Mg}$ ol-opx is near unity (von Seckendorff & O'Neill, 1993) so orthopyroxene has the same Mg# of olivine in which it is in equilibrium. Orthopyroxene with typical Mg# of mantle tectonites was found only near larger massifs of mantle tectonite in the region, such as at Peridotite Peak, Nahlin Mountain, Scarface Mountain and Mt. Nimbus (Fig. 4). All other orthopyroxenes lower in Mg# appear to be derived from lower crustal cumulate rocks from ophiolite.

Clinopyroxene

Clinopyroxene was identified as grains with either bright emerald green or dark green colour, usually showing well-developed cleavage and a prismatic grain shape. Clinopyroxene forms in several igneous and metamorphic environments and its paragenesis can be difficult to distinguish in the absence of coexisting minerals. The $Kd_{Fe\text{-}Mg}$ ol-cpx is greater than one (Brey *et al*., 1990) such that clinopyroxene from mantle tectonite has a Mg# greater than 0.9. Clinopyroxenes from mantle residues of basalt extraction are also rich in Cr_2O_3 (> 0.5 wt%) and poor in Na (as jadeite component Jd - $NaAlSi_2O_6$) and Ti because these elements are compatible and incompatible, respectively, during melting. Clinopyroxene from mantle tectonite is Cr-diopside with variable Al as a Ca-Tschermaks component (CaTs - $CaAl_2SiO_6$).

Clinopyroxene with Mg# greater than 0.9, rich in Ca-Tschermaks component (CaTs - $CaAl_2SiO_6$) and poor in jadeite (Jd - $NaAlSi_2O_6$) component is present only in the southeast at Nahlin Mountain, Scarface Mountain and Peridotite Peak (Fig. 5, 6). The source rock was likely mantle peridotite tectonite from massifs in these areas. Clinopyroxenes with lower Mg# at all other sample locations, however, must have have been eroded from source rocks with lower bulk Mg# than mantle tectonite, such as cumulate ultramafic rocks or basalt. Many of these grains have high X_{CaTs} (Fig. 6) and so could be derived from cumulate rocks, but some of the clinopyroxenes in the Scarface Mountain and Sloko samples, are distinctly rich in jadeite (Jd - $NaAlSi_2O_6$). Clinopyroxenes from the Sloko locations are almost certainly derived from eclogite, because clasts of this rock type with identical Jd-rich clinopyroxene compositions are recognized in a distinct ridge of garnetiferous pebble conglomerate (referred to as Eclogite Ridge) to the west in the Laberge Group (MacKenzie *et al*., 2005). The Sloko River sample was taken from a creek draining Eclogite Ridge. An eclogite source rock for some of the Jd-rich pyroxenes is also suggested by the mineral chemistry of garnets described below. Na-rich clinopyroxenes at Scarface Mountain are similar to those in alkaline volcanics from the Neogene Llangorse volcanics (Fig. 6).

Several studies have applied the major and minor element compositions of pyroxenes to discriminate the tectonic environment in which they formed (Aggarwal *et al*., 1984; Krawinkel *et al*., 1999; Leterrier *et al*., 1982; Nisbet & Pearce, 1977; Schweitzer *et al*., 1979). We applied many of these tests on pyroxenes screened to have an Mg# less than mantle tectonites (< 0.9). The results show that most sample locations contain significant amounts of clinopyroxene from mafic volcanic or intrusive host rocks. A distinctive population in the southwesternmost part of the study area at Nahlin Mountain, Mt. Nimbus, Scarface Mountain and Peridotite Peak, however, have clinopyroxenes rich in Ti, Na or Fe (Fig. 7, 8) not unlike pyroxenes from alkaline volcanics. The latter grains are not well represented to the northwest, but are nearly identical to Na- and Ti- rich pyroxenes present in alkaline lavas from the Lhangorse volcanic center to the northeast of Atlin.

Garnet

Garnets have pink, purplish pink, green or orange in colour, and in some cases well-developed dodecahedral crystal faces. The dominant garnet in all samples is a pink or pale orange variety, poor in pyrope component ($X_{Pyr} < 0.05$) and rich in almandine component (Fig. 9). These are likely crustal garnets derived from metapelitic protoliths. Thermal-metamorphic aureoles around Middle Jurassic and younger plutons in the Atlin area are known to contain garnet.

A notable population of light green garnets with nearly 100% andradite component are common at Scarface Mountain and also present at Goldbottom Creek and Mt. Nimbus. The andradite grains are likely sourced from skarns in areas where limestone units are in contact with later felsic intrusive rocks or from meta-rhodingite blocks that are locally present within serpentinite mélange.

Garnets rich in pyrope component (X_{Pyr} = 0.2 to 0.6) are recognized at all sample locations (Fig. 9). Many of these garnets are red-orange or purplish pink in color and are identical to the garnets in eclogite clasts and concentrate from pebble conglomerate within the Laberge group to the west. These pyrope rich garnets dominate the populations in the Sloko, Jos'Alun and Hard Luck Peaks samples, but they are present everywhere. Their compositions would correspond to Group B and C (crustal) eclogites according to the classification scheme of Coleman et al (Coleman *et al*., 1965).

Spinel

Spinel occurs as black grains with conchoidal fracture, or as euhedral octahedrons and cubes. Most spinel is well–preserved but some grains are rounded or frosted with a dull grey sheen. The majority of the spinels in the population are Cr-rich spinel, and at each location these grains show a positive correlation between Cr/(Cr+Al) (Cr#) and Fe/(Fe+Mg) (Fe#) (Fig. 10), reflecting the interplay of cation exchange equilibrium between spinel and olivine as a function of temperature (Irvine, 1965;

Roeder, 1994). Magnetite and 'ferritchromit' (an alteration product of Cr-spinel) plot along the right hand side of Figure 10 and thereby can be distinguished from Cr-spinels.

The spectrum of Cr-rich spinel compositions in the heavy mineral samples is compared with trends for spinel in equilibrium at various temperatures with olivine having compositions that encompass the range expected in mantle tectonites (Mg# = 0.9) and that typical of cumulate rocks from ophiolites and layered intrusions, or as phenocrysts in basalts (Mg# = 0.80). Also shown for comparison are the range of spinel compositions from mantle harzburgite near Hard Luck Peaks, and from dunite and harzburgite bodies near Atlin. Application of Fe-Mg exchange thermometry applied to coexisting olivine and spinel in these rock samples show them to have last equilibrated between 600 and 850ºC.

Many sample sites contain Cr-spinel that could have been in equilibrium with mantle olivine (Mg# ~ 0.9) at temperatures measured using olivine-spinel geothermometry for the harzburgite samples in outcrop (Fig. 10). Some of the spinel grains at Mt. Nimbus, Goldbottom and Scarface Mountain, however, occur along on an array to the right of that expected for spinel in equilibrium with mantle olivine (Fig. 10), and are were in equilibrium with olivine that has a far lower Mg# than mantle olivine. These spinels must be derived from rocks in equilibrium with olivine having a Mg# between 0.9 and 0.8, and at temperatures higher than those recorded by olivine-spinel thermometry of the mantle massifs in the area. The preponderance of olivine compositions having Mg# between 0.86 and 0.75 in the study area (Fig. 2) supports an igneous rock source in which these spinels could have been in equilibrium. That some of the spinels must be igneous in origin is also borne out by the euhedral cube and octahedral forms of many of the grains, a shape expected for magmatic spinel in plutonic ultramafic rocks or volcanic rocks (Roeder & Campbell, 1985) but not for spinel recrystallized in the subsolidus in mantle tectonites (Mercier & Nicolas, 1975).

Titanite

Titanites were recognized as equant red-brown grains. All of the titanites are low in Al_2O_3 (< 2wt%) indicative of protoliths from low P environments (Franz & Spear, 1985) most likely as accessory phases in felsic plutonic rocks from the Atlin-Nakina area.

Discussion

The Mg-rich compositions of olivine, clinopyroxene and spinel in samples from the study area suggest a mantle tectonite source for these heavy minerals. The several large masses of harzburgite tectonite that crop out throughout the study area are obvious sources (Fig. 1b).

Ironically, many sample sites contain significant proportions of olivine, clinopyroxene, orthopyroxene and spinel that are too Mg-poor to be sourced in mantle tectonite. The more Mg poor minerals could be sourced from cumulate rocks, perhaps from the crustal section of ophiolites in the Atlin – Nakina area. Cumulate rocks are known from the lower crustal section of many ophiolites (Moores, 1982) but they are not well represented proportionately in outcrops of the Atlin-Nakina area. Only a small section less than 100 m thick of peridotite with cumulus textures is recognized just north of Hard Luck Peaks. Gabbro in the same lower crustal section does contain clino- and orthopyroxene but is very poor in olivine (< 5%). One possibility is that the serpentinite melange throughout the Atlin-Nakina is an alteration product of cumulate rocks, perhaps harzburgite or wehrlite, rather than mantle tectonite. The mélange units, being less competent and easily eroded, could contribute in a large way to the stream sediments. If so, it remains unclear how fresh and euhedral igneous olivine and spinel grains would have been preserved in mélange that is heavily altered to serpentine and magnetite. More critical is that a cumulate igneous rock source for olivine in the heavy minerals from most of the samples is not supported by the regional geology, nor by the high CaO content of these olivines (Fig. 3a).

The source of much of the Mg-poor olivine in the concentrates is best explained given the presence of Neogene basaltic lavas in the region (Fig. 1b), at volcanic centres near Atlin, and in isolated outcrops on the east side of Atlin Lake (Anderson Bay) . The compositions of olivine from the Neogene Lhangorse volcanics match nearly exactly those present in sediments of many sample sites (Fig. 2, 3). The Neogene volcanics contain fresh eudral olivine phenocrysts, and once eroded would explain the presence of such grains in stream sediments. The array of some spinel compositions in the sediments is also best explained by equilibrium at high temperatures in volcanic rocks (Fig. 10) rather than slow cooling in cumulate or mantle tectonite. Many of these Neogene volcanic centres show evidence for eruption or interaction with ice (Harder, 2004), and glaciation was likely the source of erosion and dispersal. The presence of volcanic olivine and spinel in sample sites far from the current known centres to the north near Atlin, calls for significant transport of material southeastward by glaciation. Alternatively, there may be as yet unidentified Neogene volcanic centers in the southern parts of the study region near Peridotite Peak and Nahlin Mountain, although 1:50 000 scale mapping in this area has revealed two small diatremes near Chikoida Mountain, but none further south (Mihalynuk *et al*., 2004a).

The source of clinopyroxenes with Mg# < 0.9 throughout the study area is less obvious (Fig. 5). The pyroxenes in the stream sediments from the northern and middle parts of the study area (Figs. 6ab, 7ab, 8ab) cluster near compositions that are not unlike fresh pyroxene phenocrysts in the volcanic rocks (ankaramites) of the Triassic Stuhini Group, exposed on the western side of Atlin Lake. Clinopyroxene from gabbro bodies present as knockers in mélange in the Atlin-Nakina area may also be part of this array, but there are no compositional data for comparison. In the southern parts of the study area, however, the array of clinopyroxene compositions extends to far more Fe, Ti and Na-rich varieties, that are not observed in the Stuhini Group nor in gabbro. The very Fe, Ti and

Na-rich varieties in this part of the study area akin to alkaline rocks, and are likely sourced in Neogene alkaline volcanic rocks throughout the region.

The presence of eclogitic garnet and clinopyroxene in several heavy mineral samples is significant with respect to the anomalous diamond in the area. Exposures of this rock type are sporadic in the Cordillera, but known at localities on the eastern margin of Cache Creek Terrane in British Columbia and further north in Yukon Tanana Terrane of Yukon (Ghent et al, 1993; Erdmer *et al*., 1998; Mihalynuk *et al*., 2004b). The results of this study are intriguing in that they show widespread dispersal of grains from this rock type. No eclogite has been recognized in outcrop in the Atlin-Nakina area, except as clasts in a coarse pebble conglomerate of the Laberge Group to the west, near the Sloko sample. These rare clasts are less than 1 cm in size, but are likely the source of much of the garnet within this particular unit of the Laberge Group, as evidenced by the great abundance of eclogitic minerals in heavy mineral concentrate from a stream draining this region.

The Laberge Group pebble conglomerate is the only known bedrock containing eclogitic garnets and pyroxenes in the Atlin region, and is the most likely source for these heavy minerals throughout the entire study area. The Laberge pebble conglomerate crops out only in the western portion of the study area west of the Nahlin fault. Ice flow during the most recent glaciation in the Atlin area was west to east (Levson, 1992). If ice direction further south of Atlin Lake was also west to east, then glacial transport from the Laberge Group could be the source of eclogitic grains in stream sediments at Sloko site, Wilson and Feather Creeks, and as far as Scarface Mountain. Outside and further south of the Atlin area, however, the orientation of mega-rat-tail features on air photos suggests ice directions were southward (Lowe *et al*., 2003). Eclogitic garnets have been dispersed over 100 km to the southeast as far as Nahlin Mountain.

Summary

A yellowish white diamond, ~6 mm in diameter and with a rough rounded shape was originally reported by Marvin Sherman in a placer operation on Wilson Creek near Atlin. (Casselman & Harris, 2002). None of the heavy mineral samples from this study, including one on Wilson Creek, produced any diamond. The Wilson Creek diamond, if not a 'rumour', was not obviously sourced in mantle tectonite from ophiolite, nor in regional outcrops of high pressure metamorphic rocks. The Wilson Creek occurrence, as well as those to the north in Yukon, remain enigmatic, but evidence from the the distribution of key heavy minerals in this study suggests that the most likely source for the diamonds is by glacial transport of detritus from a coarse pebble conglomerate in the Laberge Group to the west, where diamond indicator minerals have been located in a specific horizon (Eclogite Ridge) a few hundred meters thick. Many of the garnets and pyroxenes in this unit show evidence to have been derived from depths approaching the diamond stability field. To explain their occurrence, it has been suggested that Pliensbachian uplift exhumed mantle peridotite and eclogite, which was quickly eroded and deposited as detritus in a forearc basin now represented by sediments in the Laberge Group (MacKenzie *et al*., 2005). The Pliensbachian uplift which resulted in this deep exhumation event is recognized to the north in Yukon in the Aishihik metamorphic suite (Johnston *et al*., 1996) and in northernmost British Columbia (Mihalynuk et al, 1999). Widespread Jurassic uplift is also recorded by $^{40}Ar/^{39}Ar$ cooling ages in large tracts of crystalline rocks to the northwest in Yukon and Alaska (Dusel-Bacon *et al*., 2002). We speculate this event may have exhumed deep-seated mantle rocks, similar to those south of Atlin, which were the source of anomalous diamonds elsewhere in southwestern Yukon and southeastern Alaska.

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Figure Captions

Fig. 1 – (a) Location of study area (shaded) in northernwestern British Columbia, with inset showing Stikine and Cache Creek terranes in the Cordillera (after (Mihalynuk *et al*., 2003))(b) Map of the study area (grey box in (a)) showing major stream drainages and geology of ophiolite massifs (black) and granitoid intrusives (grey shaded) in the Atlin-Nakina area. Triangles are the locations of Neogene volcanic rock occurrences, including the Llangorse lavas (Edwards *et al*., 2003). Diamonds show location of heavy mineral samples in this study.

Fig. 2 – Olivine Mg# (Mg/(Mg+Fe)) in heavy mineral samples. Locations are divided into the northern (a), and southern (b) parts of the study area. Also shown are (c) olivine analyses from the Neogene Lhangorse lavas within the study area (after (Harder, 2004)) and from cumulate rocks (d) in the Bay of Islands ophiolite (J. Bedard, pers. Comm..)

Fig. 3 – CaO content (in wt%) of olivines from the samples in this study (a) compared with those from cumulate rocks in the the Bay of Islands ophiolite (b) and Neogene Lhangorse lavas (c) in the study area. References for the latter two datasets are as in Figure 2.

Fig. 4 - Orthopyroxene Mg# (Mg/(Mg+Fe)) in heavy mineral samples from this study at each locality.

Fig. 5 – Summary of Mg# in clinopyoxene in heavy mineral samples from this study. The histogram is divided into northern (Feather, McKee, Wilson), middle (Goldbottom, Sloko, Laberge) and southern (Mt. Nimbus, Scarface Mountain, Peridotite Peak, Nahlin Mountain) parts of the study area (see Fig 1b for locations).

Fig. 6 – Plot of the mole fraction of CaTschermaks (CaTs) component ($CaAl_2SiO_6$) vs jadeite ($NaAlSi_2O_6$) component in clinopyroxenes from heavy mineral samples for locations from the northern (a), middle (b) and southern (c) parts of the study area. Shaded fields for volcanic pyroxenes

from Neogene Llangorse alkaline centers and Triassic Stuhini group are from data in Harder (2004) and Canil (unpubl.), respectively. Eclogite pyroxenes in (b) and (c) plot to the far right of the dividing line, after (White, 1964).

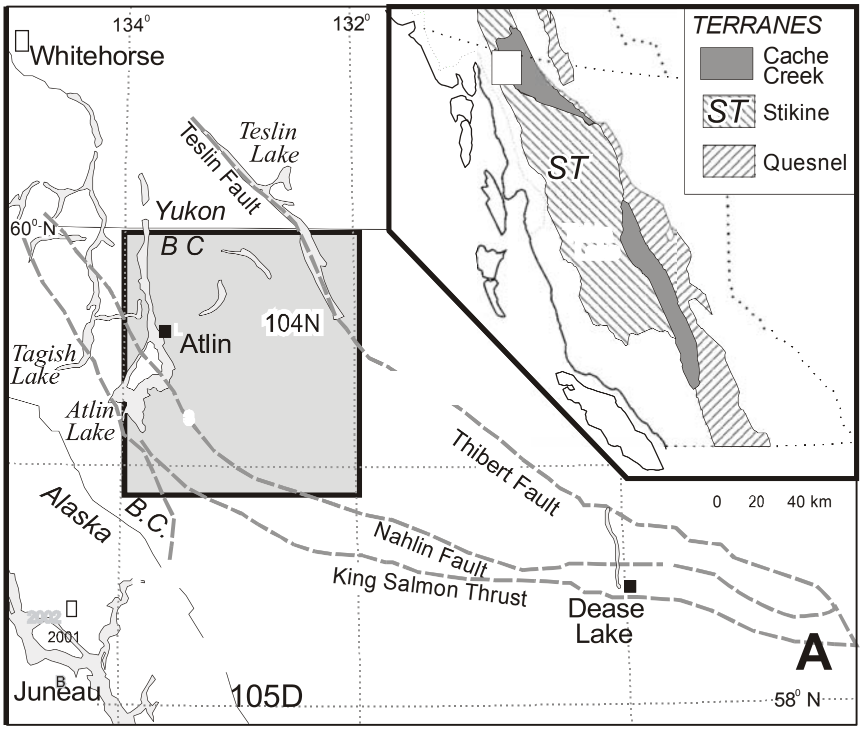
Fig. 7 – Ti versus Na + Ca in clinopyroxenes from heavy mineral samples at each locality in the northern (a), middle (b) and southern (c) parts of the study area. Grains from mantle tectonite (Mg# > 0.9) were screened from this plot. The line discriminating alkalic from tholeiitic pyroxenes is from (Leterrier *et al*., 1982). Data sources for pyroxenes from Neogene Llangorse lavas and Triassic Stuhini group in caption to figure 6.

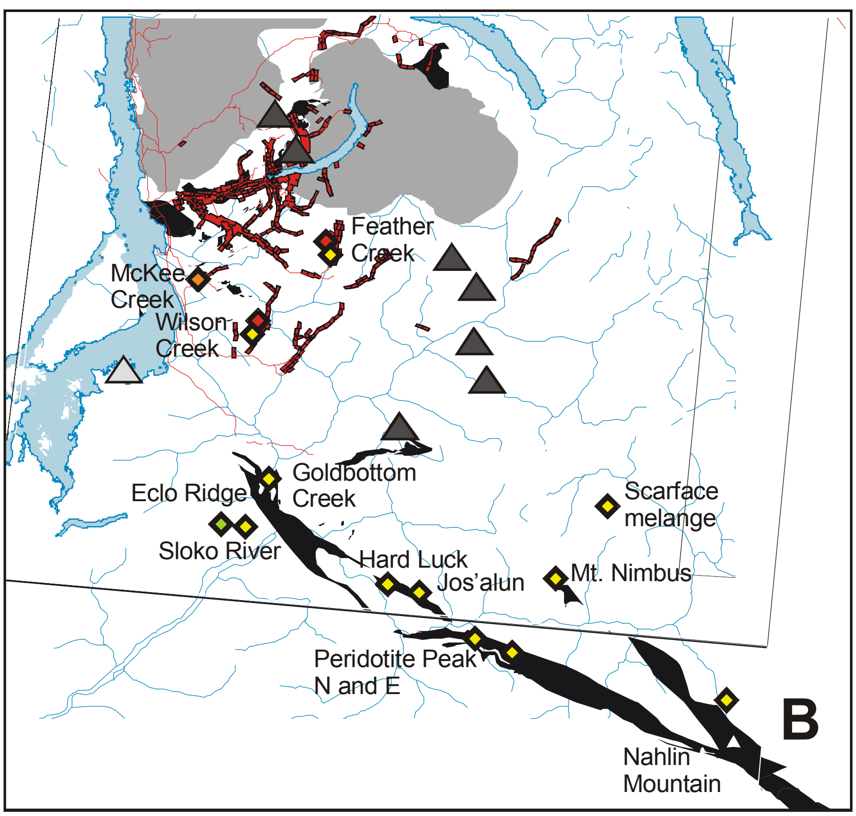
Fig 8 - Ti versus Fe/Fe+Mg in clinopyroxenes from heavy mineral samples in the northern (a), middle (b) and southern (c) parts of the study area. Grains from mantle tetconite (Mg# > 0.9) are not plotted. The fields distinguishing alkalic from tholeiitic pyroxenes are from (Schweitzer *et al*., 1979). Data sources for pyroxenes from Neogene Llangorse lavas and Triassic Stuhini group in caption to figure 6.

Fig. 9 - Histogram showing mole fraction of pyrope (X_{Pyr}) component ($Mg_3Al_2Si_3O_{12}$) in garnets from heavy mineral samples at each locality in the northern (a), middle (b) and southern (c) parts of the study area. The compositions of garnet in eclogite clasts and in heavy mineral concentrate from a pebble conglomerate from the Laberge Group ('Eclogite Ridge' in Fig 1b –(MacKenzie *et al*., 2005)) are shown for comparison in (d).

Fig. 10 – Plot of Cr# (Cr/Cr+Al) vs. Fe# (Fe/Fe+Mg) for spinel grains from heavy mineral concentrates. The lines show trends expected for spinel in equilibrium with various olivine compositions at different temperatures (after (Roeder, 1994)). The grey box encompasses the compositions of spinel grains from two harzburgite tectonite samples in outcrop north of Hard Luck Peaks. Also shown are spinels from a dunite body and harzburgite from near Atlin. These samples

equilibrated at 850 to 700ºC as determined by olivine-spinel Fe-Mg exchange thermometry (Ballhaus *et al*., 1991). Note the abundance of spinels that could not be in equilibrium with mantle tectonite olivine (Fo_{90}) but would be in equilibrium with more Mg-poor olivine at magmatic temperatures.


134°
132°
Whitehorse
Teslin Lake
Teslin Fault
Yukon
B C
60° N
104N
Atlin
Tagish Lake
Atlin Lake
Alaska
B.C.
Thibert Fault
Nahlin Fault
King Salmon Thrust
Dease Lake
2001
Juneau
105D
58° N
0 20 40 km
A
TERRANES
Cache Creek
ST Stikine
Quesnel
ST


Feather Creek
McKee Creek
Wilson Creek
Goldbottom Creek
Eclo Ridge
Sloko River
Scarface melange
Hard Luck
Jos'alun
Mt. Nimbus
Peridotite Peak N and E
Nahlin Mountain
B

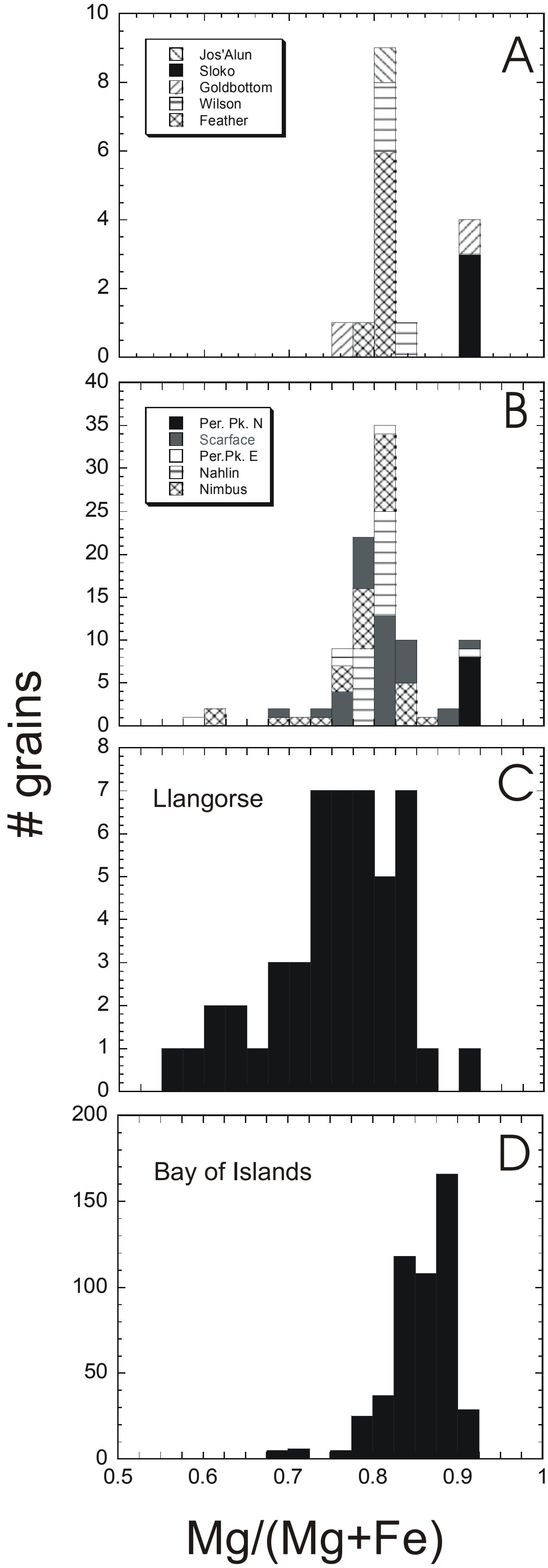
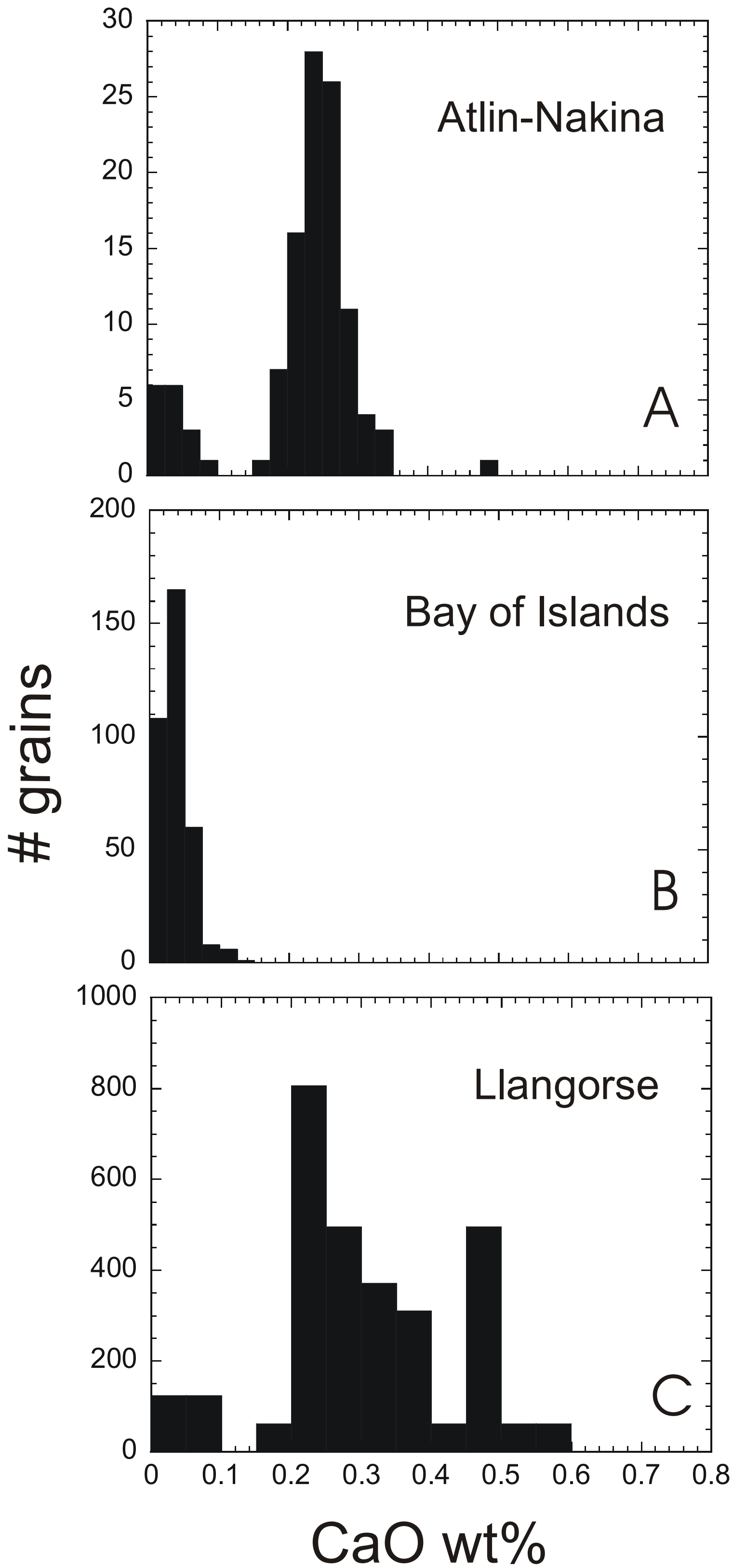

A
Jos'Alun
Sloko
Goldbottom
Wilson
Feather
B
Per. Pk. N
Scarface
Per.Pk. E
Nahlin
Nimbus
C
Llangorse
D
Bay of Islands
grains
Mg/(Mg+Fe)
0.5
0.6
0.7
0.8
0.9
1

Fig 2



30
25
20
15
10
5
0
Atlin-Nakina
A
200
150
100
50
0
Bay of Islands
B
1000
800
600
400
200
0
Llangorse
C
grains
0
0.1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
CaO wt%

Fig 3

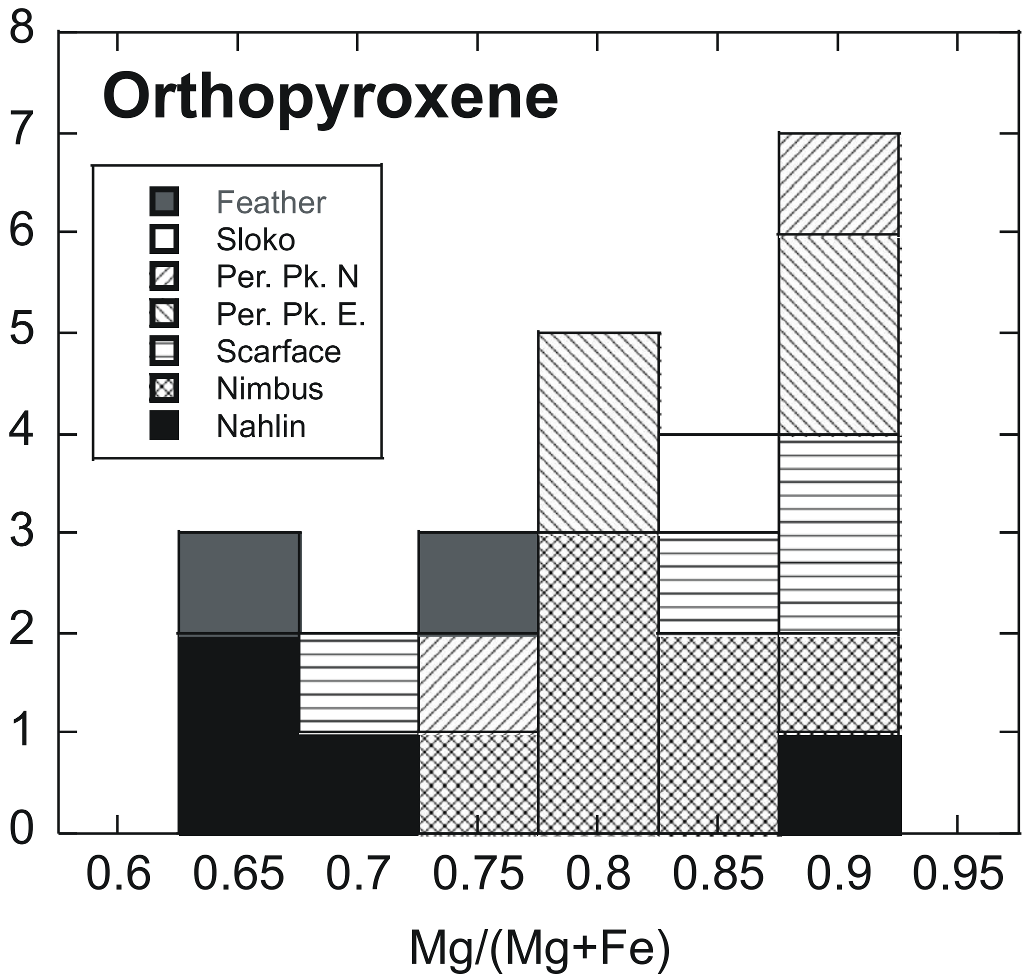
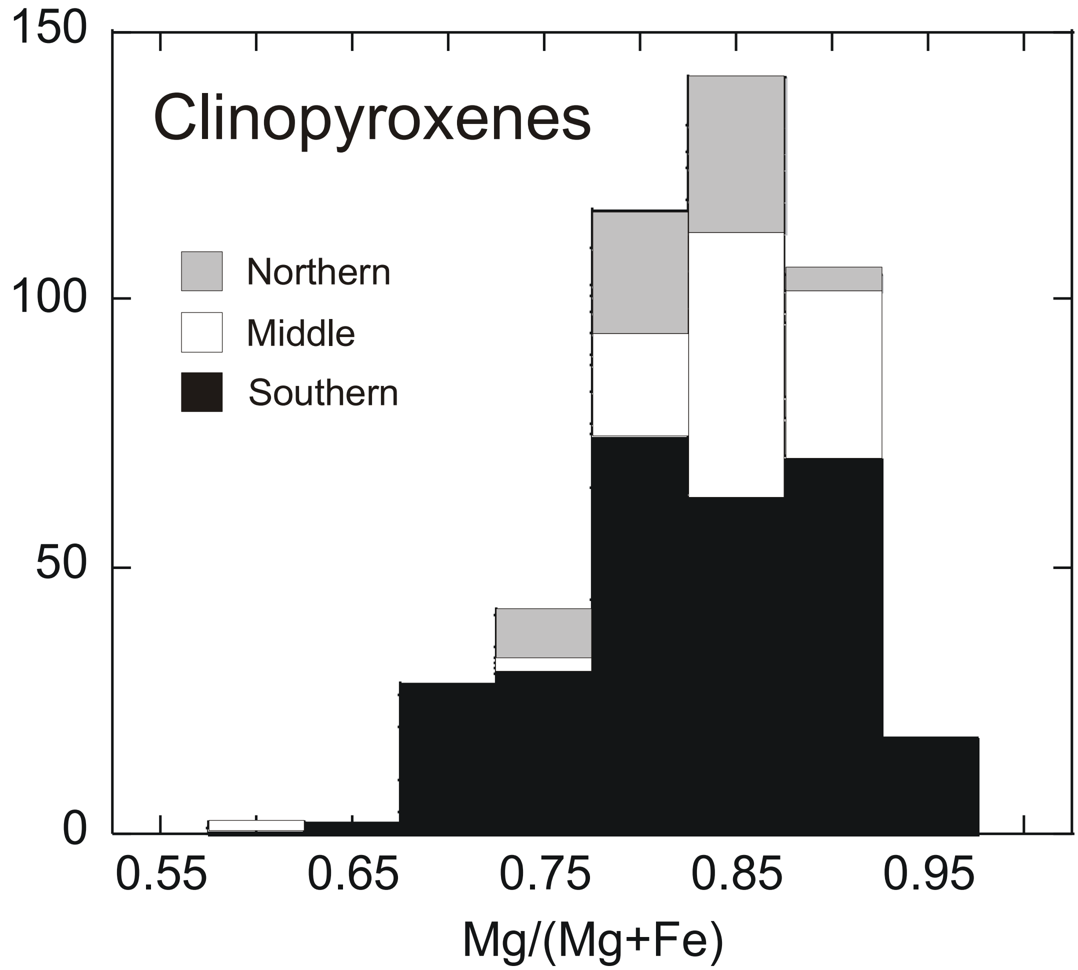

Orthopyroxene
Feather
Sloko
Per. Pk. N
Per. Pk. E.
Scarface
Nimbus
Nahlin
0
1
2
3
4
5
6
7
8
0.6
0.65
0.7
0.75
0.8
0.85
0.9
0.95
Mg/(Mg+Fe)

Fig 4



150
100
50
0
Clinopyroxenes
Northern
Middle
Southern
0.55
0.65
0.75
0.85
0.95
Mg/(Mg+Fe)

Fig 5

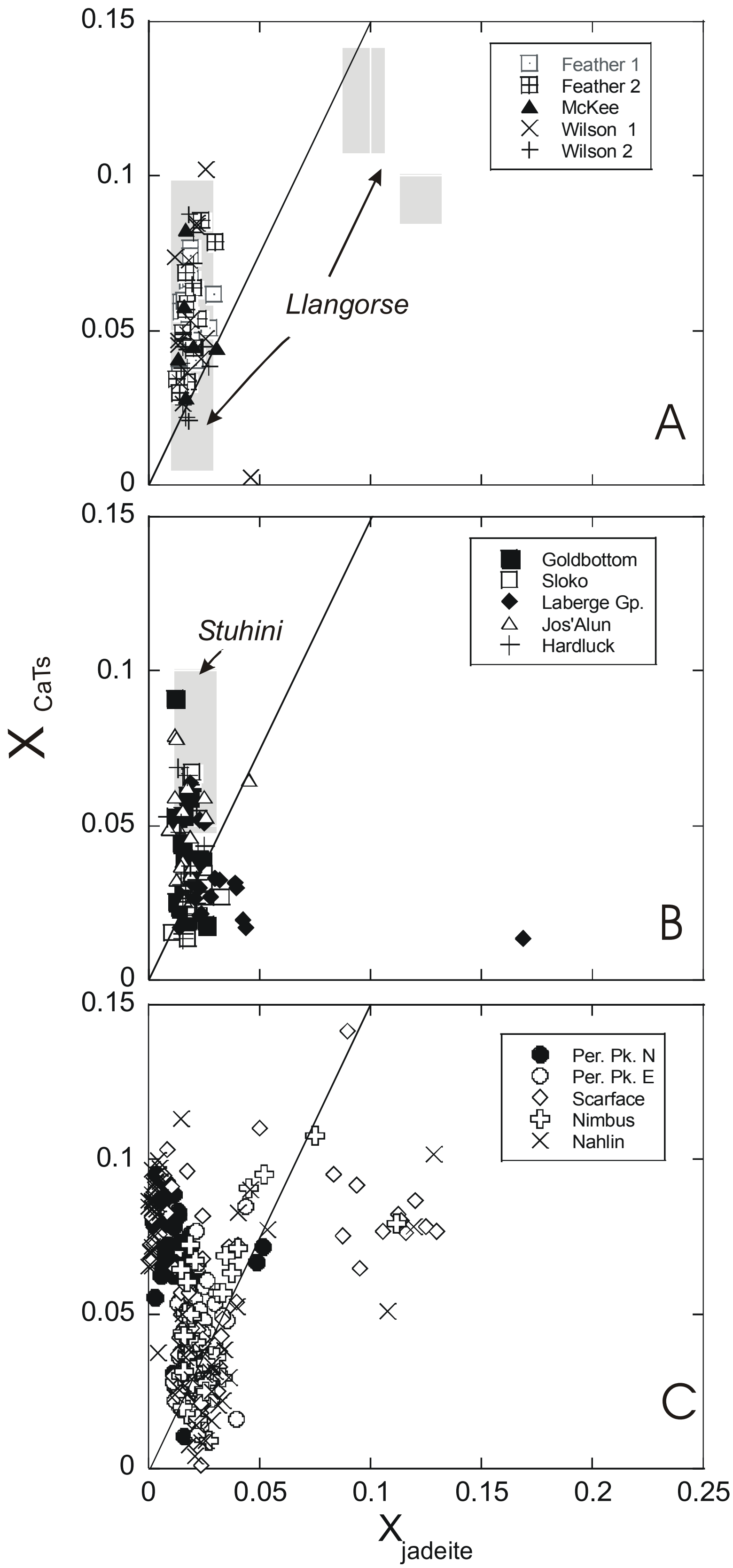

0.15
0.1
0.05
0
Feather 1
Feather 2
McKee
Wilson 1
Wilson 2
Llangorse
A
Goldbottom
Sloko
Laberge Gp.
Jos'Alun
Hardluck
Stuhini
B
Per. Pk. N
Per. Pk. E
Scarface
Nimbus
Nahlin
C
X CaTs
0.05
0.1
0.15
0.2
0.25
X jadeite

Fig 6

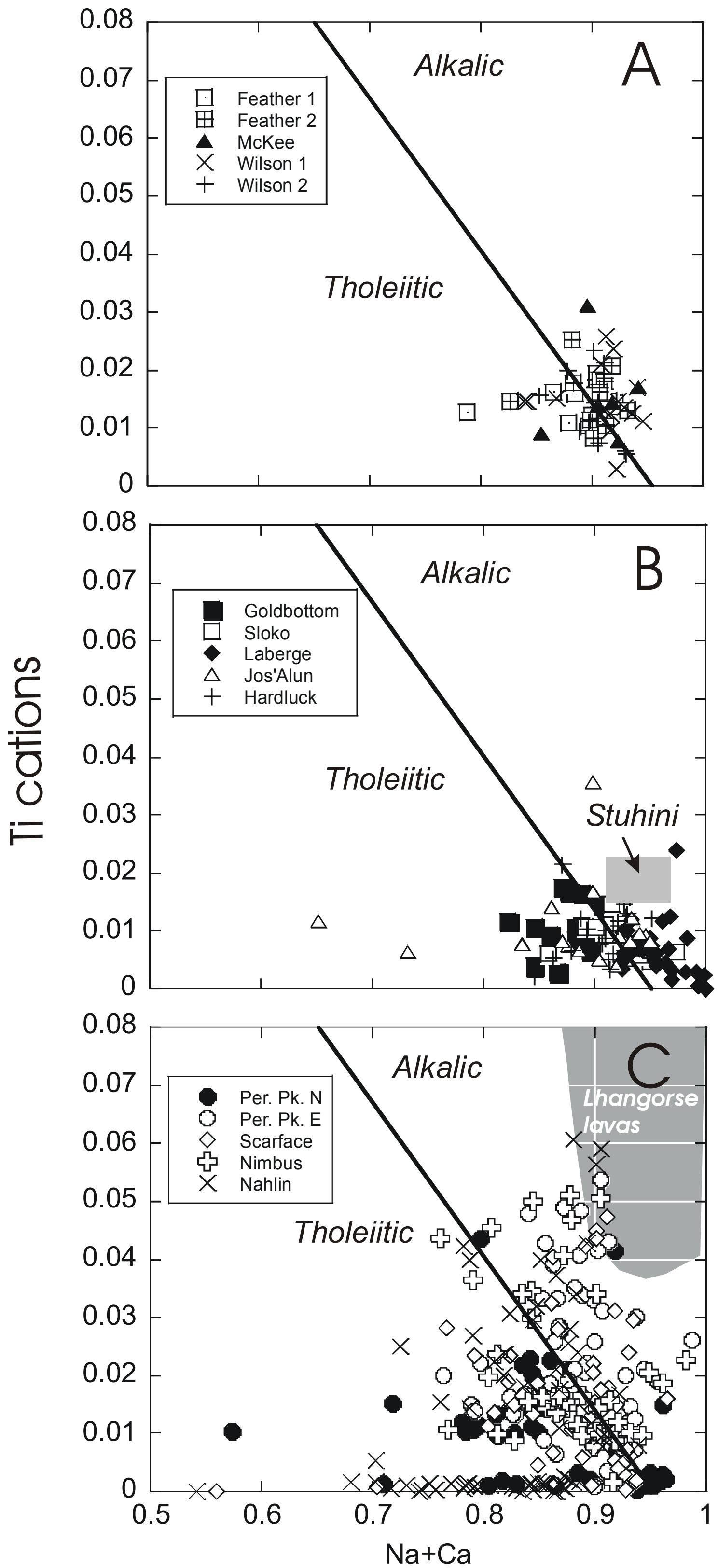

Ti cations
Na+Ca
A
Alkalic
Tholeiitic
Feather 1
Feather 2
McKee
Wilson 1
Wilson 2
B
Goldbottom
Sloko
Laberge
Jos'Alun
Hardluck
Stuhini
C
Per. Pk. N
Per. Pk. E
Scarface
Nimbus
Nahlin
Lhangorse lavas

Fig 7

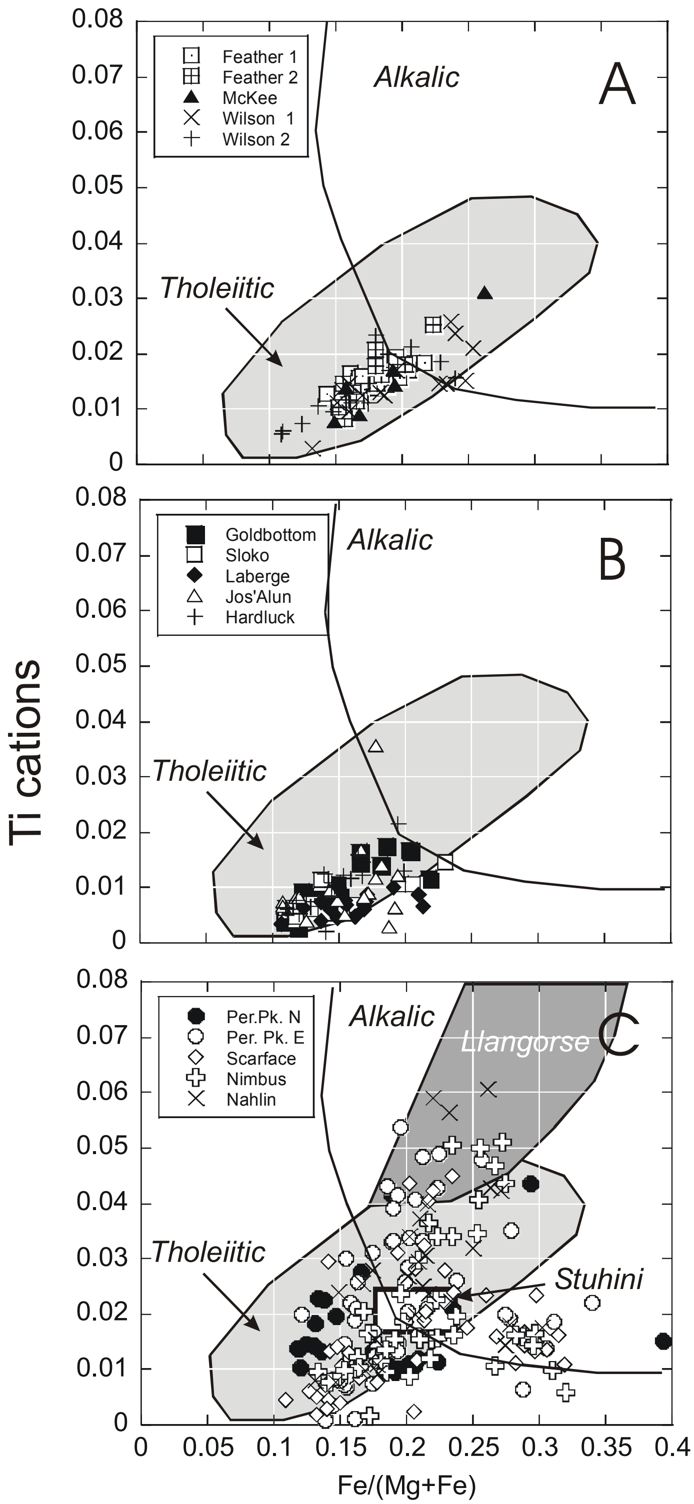

0.08
0.07
0.06
0.05
0.04
0.03
0.02
0.01
0
Feather 1
Feather 2
McKee
Wilson 1
Wilson 2
Alkalic
A
Tholeiitic
Goldbottom
Sloko
Laberge
Jos'Alun
Hardluck
B
Ti cations
Per.Pk. N
Per. Pk. E
Scarface
Nimbus
Nahlin
Llangorse
C
Stuhini
0
0.05
0.1
0.15
0.2
0.25
0.3
0.35
0.4
Fe/(Mg+Fe)

Fig 8

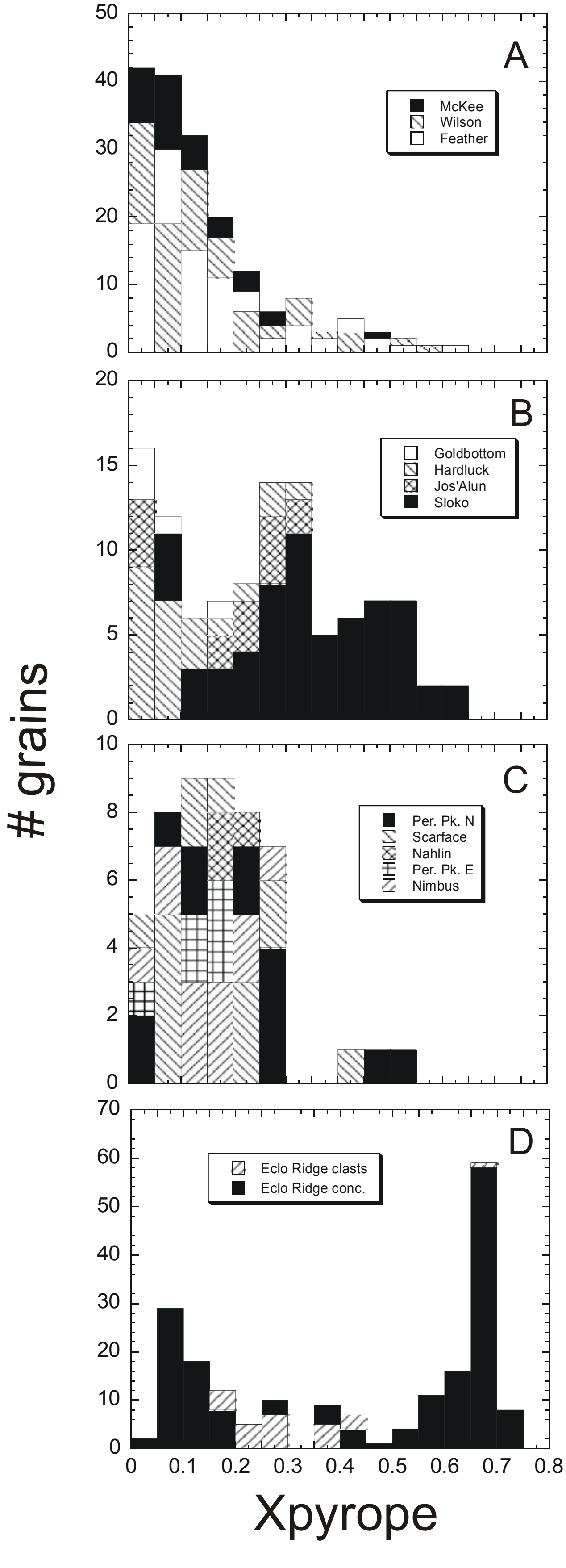

A
McKee
Wilson
Feather
B
Goldbottom
Hardluck
Jos'Alun
Sloko
C
Per. Pk. N
Scarface
Nahlin
Per. Pk. E
Nimbus
D
Eclo Ridge clasts
Eclo Ridge conc.
grains
Xpyrope
0 0.1 0.2 0.3 0.4 0.5 0.6 0.7 0.8

Fig 9

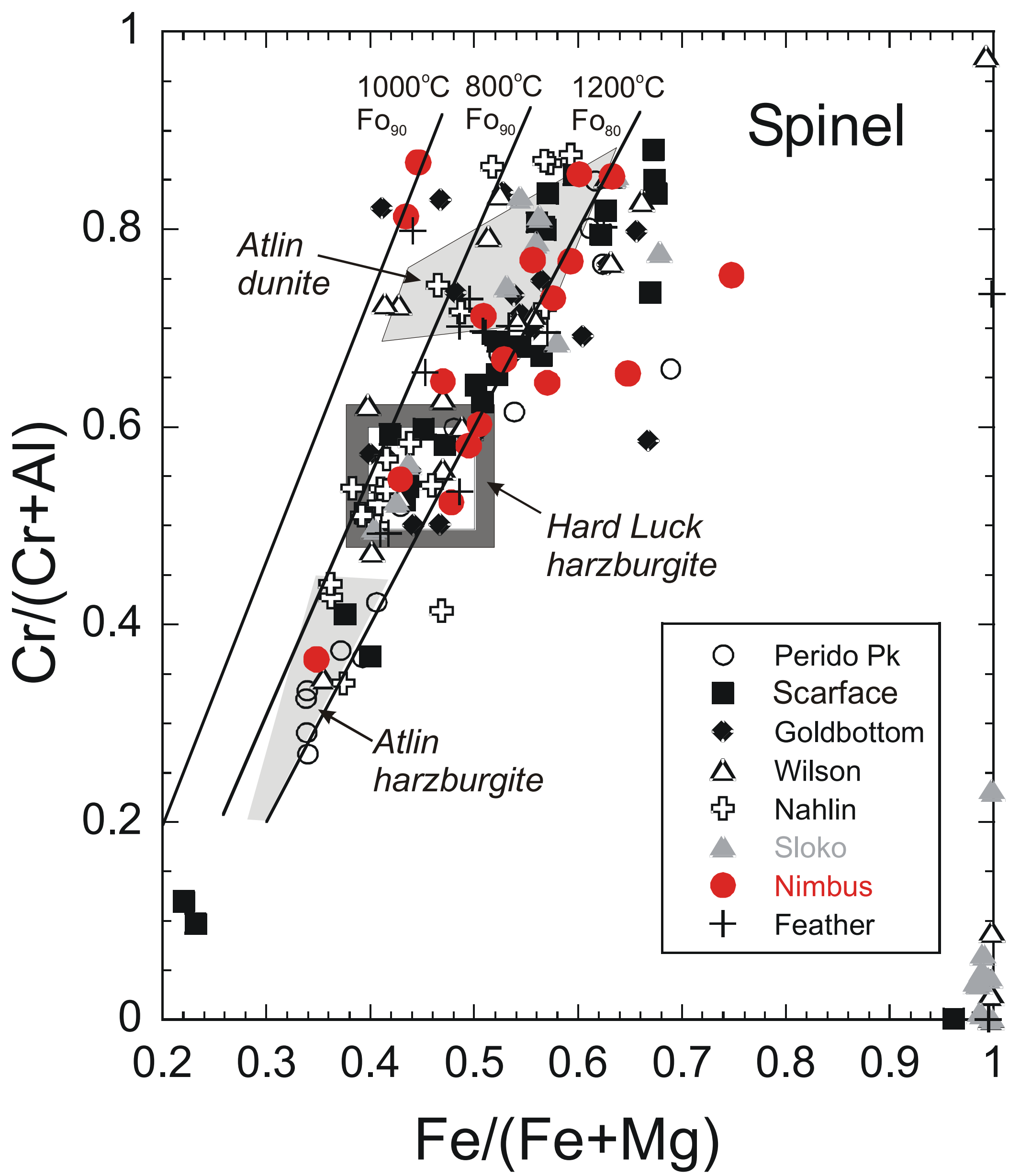

Spinel
1000°C Fo90
800°C Fo90
1200°C Fo80
Atlin dunite
Hard Luck harzburgite
Atlin harzburgite
Perido Pk
Scarface
Goldbottom
Wilson
Nahlin
Sloko
Nimbus
Feather
Cr/(Cr+Al)
Fe/(Fe+Mg)
0
0.2
0.4
0.6
0.8
1
0.2
0.3
0.4
0.5
0.6
0.7
0.8
0.9
1

Fig 10