THE LLANGORSE VOLCANIC FIELD: VOLCANOLOGY AND MANTLE PETROLOGY

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by

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A THESIS SUBMITTED IN PARTIAL FULFUILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF

MASTER OF SCIENCE in THE FACULTY OF GRADUATE STUDIES GEOLOGICAL SCIENCE DIVISON DEPARTMENT OF EARTH AND OCEAN SCIENCES

We accept this thesis as conforming to the required standard

THE UNIVERSITY OF BRITISH COLUMBIA

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Master & Science Year: 2004 Earth and Ocean Sciences British Columbia

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ABSTRACT

The Llangorse volcanic field (LVF) comprises Miocene to Holocene mafic alkaline volcanic rocks near Llangorse Mountain, in northwest British Columbia. Volcanic remnants of alkali olivine basalt, basanite, and nephelinite are exposed in the LVF. The largest of these volcanic remnants is the Llangorse Mountain locality, a 100 m thick basanite lava flow overlying debris flow sediments and containing peridotite xenoliths. Lavas at this locality are interpreted as a valley-filling flow that was substantially overthickened by ponding against a valley-wide barrier. The characteristics of the debris flow sediments and their temporal relationship with the overlying lavas suggest that they may derive from lahars or outburst floods related to melting of a snow pack or ice cap during the eruption. Palagonitized glass clasts in the debris flow sediments provide ancillary evidence of post-depositional conditions in the sediments and suggest that they were water-saturated when deposited and were heated to at least 80-120°C by the overlying flow.

Peridotite xenoliths collected from the lava flow and underlying debris flow sediments at Llangorse Mountain are the basis for a detailed study of the mantle lithosphere beneath this portion of the Cordillera. The xenoliths are spinel lherzolites and spinel harzburgites. Twopyroxene thermometry is applied to 44 xenoliths, recovering minimum temperatures of 800-850°C and maximum temperatures of 1050-1100°C for this xenolith suite. The minimum xenolith temperatures are taken to represent the maximum MOHO temperature, and the maximum temperatures are taken to approximate temperatures near the lithosphere / asthenosphere boundary. These data are combined with measured heat flow data in the northern Canadian Cordillera to produce a set of model geotherms, which are used to constrain

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the thickness of the mantle lithosphere in this area. These model geotherms indicate the mantle lithosphere in the northern Canadian Cordillera to be 18 - 39 km thick, corresponding to a depth to the lithosphere / asthenosphere boundary of 54 - 75 km. The implications of these geotherms are explored, defining three potential source regions for lavas in the LVF: a) the amphibole-bearing lithospheric mantle; b) the spinel-bearing base of the mantle lithosphere or upper asthenosphere; c) the deep, garnet-bearing asthenosphere. Variation in peridotite chemistry, mineralogy, and physical properties with depth are used to determine that the mantle lithosphere in the northern Canadian Cordillera is heterogeneous, with distinct layers of foliated, lherzolitic mantle versus non-foliated, harzburgitic mantle.

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ACKNOWLEDGMENTS

Funding for my M.Sc. was covered primarily by an NSERC PGS-A scholarship. Analytical costs were covered by an NSERC Discovery Grant to J.K. Russell. Field costs were covered by the Geological Survey of Canada, courtesy of Bob Anderson. Special thanks to my supervisor, Kelly Russell, this thesis would not have been accomplished without your enthusiasm and knowledge. This thesis also benefited greatly through discussions with Ben Edwards, Kirstie Simpson, Phil Hammer, and Roy Hyndman. Discussions (thesis and otherwise) with Melanie Kelman, Steve Quane, and Alison Rust have been invaluable, you three have formed a great 'IPL' support group. Electron microprobe and scanning-electron microscope analyses and images were possible through the help of Mati Raudsepp and Elisabetta Pani. Assistance in analytical work by Lorraine Tam, Marcia Wilson, Lisa Swinnard, and Krista Michol is greatly appreciated, providing many of the physical property and modal mineralogy results in this thesis. The past few years in grad school have been memorable (and often even enjoyable!) courtesy of the many wonderful graduate students at UBC, but particular thanks must go to Liane Boyer, Heidi Annell, Natalie Lefebvre, and Diane Mitchinson. Lastly, I would like to thank Aubrey and my family for their support throughout my M.Sc.

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Chapter 1

OVERVIEW

Volcanic rocks and associated peridotite xenoliths from the Llangorse Mountain area, near the town of Atlin, northwest British Columbia, are the focus of this thesis. This thesis comprises two distinct study areas. The first part of this thesis details field descriptions and volcanological interpretations of volcanic rocks near Llangorse Mountain. The primary focus is an overthickened basanite flow located immediately south of Llangorse Mountain, which is interpreted as a valley-filling lava flow that overrode synvolcanic debris flow sediments. The overthickened aspect of the lava flow is indicative of ponding against a valley-wide barrier, which may have been glacial ice. Palagonitized clasts of basanite glass are abundant in the debris flow sediments underlying the lava flow and are used to constrain the temperatures at which the basanite glass was altered. The second part of this thesis focuses on peridotite xenoliths recovered from lavas at a large basanite lava flow immediately south of Llangorse Mountain. Geothermometry on these peridotites is used to characterize the thermal profile of the mantle lithosphere beneath Llangorse Mountain, and ultimately to create model geotherms for this area of the northern Cordillera. The variation in peridotite properties with depth is used to determine if the mantle lithosphere is heterogeneous or homogeneous. I conclude by comparing the thermal profile of the mantle lithosphere beneath Llangorse Mountain with other volcanic localities in the Northern Cordilleran Volcanic Province. This thesis comprises three papers (Chapters 2, 3 and 4), as described below, followed by an examination of structure in the lithospheric mantle

(Chapter 5) and a summary (Chapter 6). Some repetition in the introductory sections of Chapters 2, 3 and 4 is unavoidable as each of these chapters is a separate manuscript for different publications.

Chapter 2 is based on a paper written for the Geological Survey of Canada Current Research Papers (Harder et al, 2003), with a section added detailing the chemistry of Llangorse volcanic field lavas. This chapter details field observations from mapping of numerous volcanic remnants near Llangorse Mountain. The volcanic remnants are Miocene to Holocene in age and are basanitic to nephelinitic in composition. The largest of the volcanic localities studied is the Llangorse Mountain locale, a 100 m thick basanite flow that is associated with debris flow sediments. Other volcanic localities studied include: Table Hill, a mound-shaped volcanic remnant with large, well-developed columnar jointing; Lone Point, a mound-shaped volcanic remnant showing erratic orientation of cooling joints and associated with boulders of hyaloclastite; and Hidden Ridge, a ridge of lava showing erratically oriented cooling joints and associated with debris flow sediments. Bulk chemistry results are provided for all the volcanic localities studied in the LVF, and appropriate rock classifications are provided.

Chapter 3 is a manuscript presenting volcanological interpretations based on field observations (provided in greater detail in Chapter 2) and is intended for publication. At the Llangorse Mountain locale, the lava flow is interpreted as a valley-filling lava flow associated with syn-volcanic debris flow sediments. Palagonitized basanite clasts, which are abundant in the debris flow sediments, are used to infer post-depositional conditions in the debris flow sediments. The morphology of the lava flow is indicative of ponding against a lower-elevation, valley-wide barrier, and the implications of this barrier are discussed.

Although the other localities in the LVF are not focused on in detail, the significance of erratically-oriented cooling joints at Lone Point and Hidden Ridge may indicate a glaciovolcanic origin of the lavas.

Chapter 4 is a manuscript which is currently in review for the journal Lithos, Special Volume "From Mantle to Magma" (Harder and Russell, in review). Although primarily written by myself, Chapter 4 contains sections contributed by J.K. Russell for the paper in review, including: section 4.8 (Model of the Lithospheric Mantle) and sections 4.9.1 and 4.9.2 (Discussion: Crustal Geotherms and Lithosphere - Asthenosphere Transition). These sections are included in this thesis as they are necessary for my interpretations in section 4.9.3 (Implications for NCVP Magmatism). In Chapter 4 the focus shifts from volcanology of the LVF to mantle petrology of peridotite xenoliths. A large (>80) suite of peridotite xenoliths were collected from lavas at the Llangorse Mountain locale. A total of 45 of these peridotites were studied for geothermometry, 19 of which are used for chemical, mineralogical, and physical property studies. The thermometry database created is used to create a thermal profile of the northern Canadian Cordillera. Geothermometry results are also combined with geophysical heat flow measurements to produce model geothermal gradients of the lithospheric mantle. These geothermal gradients are used to constrain the thickness of the mantle lithosphere in this area. The implications of these geotherms are discussed relating to the source area of the basanite lavas that entrain the xenoliths.

Chapter 5 provides a summary of various chemical, mineralogical, and physical properties with depth, resulting in a structural profile of the northern Canadian Cordillera. A distinct layering is observed, indicating that the lithospheric mantle in this area is chemically and mineralogically heterogeneous. In addition to a mantle stratigraphy, Chapter 5 provides

a comparison of geothermometric results from Llangorse Mountain with other volcanic localities in the NCVP.

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Chapter 2

THE LLANGORSE VOLCANIC FIELD

2.1 INTRODUCTION

The Northern Cordilleran volcanic province (NCVP) comprises Miocene to Holocene volcanic rocks distributed across northwestern British Columbia, the Yukon Territory, and Alaska (Edwards and Russell, 2000). The Atlin volcanic district (AVD), a subset of the NCVP, consists of several Miocene to Holocene volcanic localities in the Atlin area (see Fig. 2.1). These localities include Moose Bay and Anderson Bay (Bultman, 1979), the Llangorse volcanic field (Harder et al, 2003), and the Surprise Lake volcanic field (Edwards et al, 2003). The Surprise Lake volcanic field includes Cracker Creek, Ruby Mountain, Ruby Creek, and Volcanic Creek (Edwards and Bye, 2003; Edwards et al, 2003). The Llangorse volcanic field (LVF) is situated approximately 55 km southeast of Atlin, BC (Fig. 2.1), and comprises Miocene to Holocene volcanic rocks exposed at Hirschfeld Creek (Line Lake), Mt. Sanford, Llangorse Mountain (including four localities), and Chikoida Mountain. The geographical area included in the LVF is approximately 144 km² and is centred around Llangorse Mountain (Fig. 2.2).

The northern Canadian Cordillera comprises dominantly suspect terranes (i.e. allochthonous terranes) that were accreted onto the western margin of North America, beginning in the middle Jurassic (Coney et al, 1980; Monger et al, 1982). The suspect terranes welded to ancestral North America include Quesnellia, Stikinia, Cache Creek, and Wrangellia. Together, these terranes comprise the Intermontane Belt, which is bounded to



Figure 2.1

Figure 2.1: The northern Cordilleran volcanic province (NCVP) comprises Miocene volcanic centers distributed across northern British Columbia (BC), the Yukon territory (YK) and easternmost Alaska (AK). The Queen Charlotte transform fault (QCTF) defines the boundary between the Pacific and North American plates. The NCVP is shown relative to other major volcanic provinces in British Columbia, including: Wrangell volcanics (WV), Anaheim volcanic belt (AV), Wells-Grey volcanic field (WG), and Cascade volcanic arc (CVA) (after Edwards and Russell, 2000). Major centres (large black circles) include Hoodoo Mountain (H), Mount Edziza (E), Prindle Mountain (P), Alligator Lake (A), and Level Mountain (LM). Small black circles represent individual occurrences of Miocene to Holocene volcanic rocks. The occurrences near Llangorse Mountain (L) are within the Atlin volcanic district (dashed outline), which is situated immediately southeast of Atlin (At). Other population centres shown include Whitehorse (Wh) and Terrace (Te). A tectonic assemblage map is shown in the upper-right corner of the map, illustrating the tectonic belts that comprise that Canadian Cordillera.



Figure 2.2: Geological map for the Llangorse volcanic field (LVF; Map Sheet NTS 104N; after Aitken, 1959 and Harder et al, 2003). The LVF includes occurrences of Miocene – Holocene volcanic rocks (Tables 1, 2) at: 1) Llangorse Mountain, 2) Table Hill, 3) Lone Point, 4) Hidden Ridge, 5) Hirschfeld Creek, 6) Chikoida Mountain, and 7) Fire Mountain.

the east by the Omineca Crystalline Belt, comprising a complex suite of rocks with a dominantly North American affinity, and to the west by the Coast Plutonic Complex, comprising Cretaceous to Tertiary plutonic rocks considered to result from subductionrelated accretion of the suspect terranes (Monger et al, 1982). The Atlin area straddles the boundary between the Coast and Intermontane belts (Aitken, 1959); the Atlin volcanic district lies entirely within the Intermontane belt.

The regional geology of the Atlin area was first mapped by Aitken (1959). Subsequent work by Bloodgood and Bellefontaine (1990), Ash and Arksey (1990), Mihalynuk and Smith (1992), and Mihalynuk et al (2002) have provided greater insight into the geology of this area. The Atlin volcanic district (Fig. 2.1) is underlain by Paleozoic sedimentary and volcaniclastic rocks from the Cache Creek Group and by Jurassic and Cretaceous plutons (mostly granitic to granodioritic in composition) (Aitken, 1959). The Llangorse volcanic field (Fig. 2.2), with the exception of Hirschfeld Creek and Chikoida Mountain, is underlain primarily by the Llangorse Mountain Batholith, a Jurassic granodiorite intrusion, and possibly by Pennsylvanian/Permian cherts from the Cache Creek Group. Hirschfeld Creek volcanic rocks are underlain by Pennsylvanian/Permian cherts, and possibly by greywacke and amphibolites, from the Cache Creek Group. Chikoida Mountain volcanic rocks are underlain by a Jurassic granitic intrusion.

Mapping and sample collection of the LVF was undertaken in the summer of 2002, as part of the Geological Survey of Canada Atlin Targeted Initiative Project, to achieve three main goals. The first goal is to provide an inventory of Miocene volcanic rocks within the LVF and to establish their composition, volume, and age. The second goal is to study the volcanology of the individual deposits to elucidate the style of eruption and the

environmental conditions into which the lavas were erupted (e.g., hypabyssal, glaciovolcanic, subaerial). The third goal is to collect xenoliths entrained in these lavas, with particular emphasis on mantle peridotite xenoliths.

2.2 THE LLANGORSE VOLCANIC FIELD

Localities of volcanic rocks in the LVF (Table 2.1, 2.2) were mapped at a scale of 1:20 000. The localities nearest Llangorse Mountain are the focus of the work and are labeled the Llangorse Mountain, Table Hill, Lone Point, and Hidden Ridge locales (Fig. 2.1, 2.2). The volcanic rocks exposed nearest to Llangorse Mountain have been previously studied in detail by Nicholls et al. (1982), Nicholls (pers. comm.), Higgins and Allen (1985), Carignan et al. (1994), Francis and Ludden (1995), Edwards et al. (1996), Nicholls and Stout (1998), Shi et al (1998), Peslier et al. (2000; 2000b), and Abraham et al. (2001). The Chikoida Mountain volcanic rocks were not mapped by Aitken (1959), but were subsequently described by Mihalynuk and Smith (1992), and by Edwards et al. (1996) (Table 2.1). The LVF also includes the centres of Hirschfeld Creek / Line Lake (Aitken, 1959; Francis and Ludden, 1995), and Mt. Sanford (Higgins and Allen, 1985). Although all volcanic rocks are Miocene to Holocene in age, their relative ages and relationships to one another are unclear.

Three of the four sites nearest Llangorse Mountain (the Llangorse locale, Table Hill, and Lone Point) were originally identified by Aitken (1959) and subsequently studied by Francis and Ludden (1985), Edwards et al (1996), and Nicholls (pers. comm.). These centres have been interpreted by the various authors as hypabyssal feeder pipes, lava lakes,

sources that pro-	ovide descri	iptions or	petrologic data.		I		
Locality		JTM] ¹	Rock Types ²	Features	Previous	Volume	References ³
1	Northing	Easting			Interpretations	(m ³)	
⁴ Llangorse Mtn	6582950	625950	BAS, MX> CX	Valley-filling flow	^{3f, m} Valley-filling flow; ^{3e} Tuya; ^{3e} Lava lake	2.1×10 ⁷	a, b, d, f, g, h, i, j, k, l, m
⁵ Table Hill	6580122	619838	NE, MX < CX	Feeder Pipe	^{3e} Ponded lava flow	3.0×10°	a, e, k, l
⁶ Lone Point	6577850	625600	NE, MX < CX	Subglacial Mound	^{3e} Hypabyssal intrusion	3.040	a, e, k, l
⁷ Hidden Ridge	6585401	628415	BAS, MX> CX	Valley-filling flow	N/A	1.540	k, l
Hirschfeld Creek	6601100	616900	NE, BAS, AOB	Hypabyssal intrusion	^{3e} Hypabyssal intrusion	2.0 ± 0^7	a, d, e, i, j, k, l
Chikoida Mtn	6565200	612450	BAS, CX > MX	Valley-filling flow	^{3f} Feeder Pipe	>1.040	f, k, l
⁸ Fire Mountain	6592000	625600	NE, MX	Dykes	^{3c,e} Dykes	>5.0404	c, e, k, l
¹ Northing and eastin	ig in metres, relati	ive to North A	merican Datum 1927.				
² Basanite (BAS), nel	phelenite (NE), al	lkali olivine ba	salt (AOB), mantle (M	X) or crustal (CX) xenoliths			
³ a Aitken (1959); b	Nicholls et all (19	982); c Higgin:	s and Allen (1985); d C	Carignan et al (1994); e Franci	is and Ludden (1995);		

Table 2.1: Locations and attributes of Miocene to Holocene volcanic rocks in the Llangorse volcanic field, including

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f Edwards et al (1996); g Nicholls and Stout (1966); h Nicholls and Stout (1997); i Shi et al (1998); j Abraham et al. (2001);

k Harder et al (2003), l Edwards et al (2003), m Nicholls (pers. comm.)

⁴ Llangorse site 1 (Harder et al, 2003).

⁵ Llangorse site 2 (Harder et al, 2003).

⁶ Llangorse site 3 (Harder et al, 2003).

⁷ Llangorse site 4 (Harder et al, 2003).

⁸ Mt Sanford (Higgins and Allen, 1985; Edwards et al, 1996; Edwards et al, 2003; Harder et al, 2003).

localities are Lla	ngorse Moun	tain (LM), l	Lone Point ((LP), Table	Hill (TH), a	nd Hidden F	Ridge (HR).	
Sample	. 101	101r	101s	101s	112	160	173	213
Center	LM	LM	LM	LM	LM	LP	, TH ,	HR
SiO ₂	~ 45.13	44.99	44.91	45.00	44.11	41.47	40.57	45.57
TiO ₂	2.234	2.213	2.196	2.219	1.923	2.724	2.531	2.053
Al_2O_3	13.12	13.13	13.15	13.18	12.36	11.99	11.70	13.37
Fe ₂ O ₃	2.23	2.59	2.80	2.73	¹ 2.63	5.37	6.70	2.76
FeO	10.09	9.68	9.59	9.59	9.76	8.35	7.57	9.58
MnO	0.184	0.179	0.181	0.184	0.177	0.192	0.211	0.188
MgO	11.69	11.61	11.65	11.68	13.55	12.02	10.67	11.04
CaO	9.96	9.96	9.94	9.99	10.24	9.18	9.45	9.37
Na ₂ O	3.14	3.12	3.11	3.15	2.34	4.52	5.22	2.81
K ₂ O	1.23	1.22	1.22	1.22	1.07	1.79	1.76	1.01
P ₂ O ₅	0.501	0.501	0.503	0.507	0.429	1.004	1.298	0.383
Total	99.51	99.19	99.25	99.45	98.59	98.61	97.68	98.13
		N	ormative M	lineralogy ((wt%)			
Or	7.3	7.2	7.2	7.2	6.3	10.6	10.4	6.0
Ab	12.4	12.9	12.9	12.7	10.3	4.3	4.2	19.1
An	18.1	18.2	18.3	18.2	20.1	7.1	3.3	20.9
Ne	7.7	7.3	7.3	7.6	5.1	18.4	21.7	2.6
Di	22.7	22.5	22.3	22.6	22.5	25.3	27.8	18.6
Ol	22.8	21.9	21.9	21.8	25.8	17.6	12.8	22.2
Mt	3.2	3.8	4.1	4.0	3.8	7.8	9.7	4.0
II ·	4.2	4.2	4.2	4.2	3.7	5.2	4.8	3.9
Ар	1.2	1.2	1.2	1.2	1.0	2.3	3.0	0.9
¹ Rock Type:	Bas	Bas	Bas	Bas	Bas	Neph	Neph	AOB
		Thermod	ynamic Liq	uidus (1 at	m.) Propert	ies		
T(°C)	1301	1296	1296	1297	1331	1305	1263	1285
² Mg#	0.67	0.68	0.68	0.68	0.71	0.72	0.72	0.67
³ ∆QFM	0.24	0.66	0.86	0.79	0.77	2.43	2.99	0.92
⁴ X _{Fo}	86.9	87.2	87.4	87.4	88.5	90	90	86.7
⁵ a SiO ₂	0.27	0.34	0.34	0.34	0.34	0.26	0.25	0.38
⁵ a Na ₂ SiO ₃	9.74E-4	6.47E-4	6.40E-4	6.57E-4	5.75E-4	1.58E-3	1.74E-3	4.25E-4
⁵ a KAlSiO₄	2.86E-2	3.61E-2	3.65E-2	3.65E-2	3.64E-2	7.92E-2	8.65E-2	2.56E-2

Table 2.2: Major element chemical compositions of samples (MH-02-) of lavas from the Llangorse Mountain volcanic field including replicate (split and duplicate) analyses of individual samples. The localities are Llangorse Mountain (LM). Lone Point (LP). Table Hill (TH), and Hidden Ridge (HR).

¹Rock types: basanite (Bas), olivine nephelinite (Neph), alkali-olivine basalt (AOB).

²Mg# computed as Mg/(Mg+Fe), in cation proportions.

³ ΔQFM: log units above the Quartz-Fayalite-Magnetite buffer.

⁴XFo Predicted composition of olivine crystallized.

⁵Activities of components in melt at 1 atm and stated temperature; calculated with MELTS (see Chapter 4).

tuyas, or valley-filling lava flows (Table 2.1). During field mapping in the summer of 2002 a new volcanic remnant was discovered (Hidden Ridge, Fig. 2.2).

2.2.1 The Llangorse Mountain Locale

The sequence of basanite lavas at the Llangorse Mountain locale (Fig. 2.1, 2.2) is volumetrically the second largest Miocene volcanic locale in the LVF after Hirschfeld Creek (Francis and Ludden, 1995) (Table 2.1, 2.2). The outcrop is a large, cliff-forming bluff approximately 100 m high and 350 m wide (Fig. 2.3a). Vertical columnar joints extend throughout the flow; the columns are generally smaller at the top of the flow (~1 m) than at the base of the flow (~2 m). The lava groundmass is aphanitic and contains microphenocrysts of olivine and clinopyroxene (< 500 μ m; < 3 vol%), xenoliths and xenocrysts. Mantle-derived xenoliths are common; they are mostly concentrated in the base of the lava flow (1-5% by volume) and are rare (<<1% by volume) in the upper part of the flow. The peridotite xenoliths are mostly spinel lherzolite and are typically 1-2 cm in size, although xenoliths >20 cm in size are not uncommon. Xenocrysts are predominantly olivine and clinopyroxene, and are likely derived from the disaggregation of peridotite xenoliths; xenocrysts are not abundant (< 1%) and are small (< 1-2 mm). Some other volcanic localities near the Llangorse locale contain large black pyroxene xenocrysts with prominent reaction rims, however, these are not observed at the Llangorse locale. Crustal xenoliths are rare (<<1% by volume) and are generally < 5 cm in size; most of these appear to be derived from the Llangorse batholith and may represent surficial fragments entrained during flow of the lavas. At the base of the lava flow, a layer about 30 cm thick of highly vesicular,





Figure 2.3: Field photographs of the exposures of lava flow at Llangorse Mountain (No. 1, Fig 2). (A) South-facing view of bluff formed by an 100 m thick valley-filling, columnar jointed lava flow, looking up the paleo-valley; floor of paleo-valley is approximately coincident with top of scree slope in locale of photograph. (B) View of west margin of bluff (east-facing) in 3a. Massive lava flow is underlain by north-dipping, crudely bedded debris flow sediments (mid-upper right corner). (C) Detail of crude layering in debris flow sediments; scale given by two people on the sediment / lava contact (arrows). (D) Detailed view of debris flow sediments (person for scale) illustrating crude graded bedding and poorly sorted, matrix supported, heterolithic nature of sediments. Light coloured blocks are rounded boulders of granodiorite. quenched (but not glassy) basanite is present. Crustal xenoliths and xenocrysts are more common (\leq 5%) in this quenched basal lava.

Volcaniclastic debris flow sediments are directly in contact with the base of the lava flow along the west side of the massif (Fig. 2.3b). At the eastern contact the lava flow is in direct contact with weathered granodiorite (grus) of the Llangorse Mountain batholith. The volcaniclastic debris flow sediments are poorly sorted, crudely stratified, and contain subangular to subrounded fragments (Fig. 2.3c, 2.3d). The crude layering in the debris flow deposit dips approximately 25° to the northwest. The sediments are primarily matrixsupported, with minor clast-supported layers. The debris flow matrix has a grain size of 2-4 mm and comprises clasts of granodiorite, vesicular basanite, peridotite, and palagonitized basanite. The deposit is partly consolidated but still friable. The dominant clast type is granodiorite, which derives from the underlying Llangorse batholith. Granodiorite clasts vary from 2 cm to 5 m in diameter. Clasts of basanitic lava are also common and vary in size from 2 to 50 cm. Larger boulders up to 1 to 2 m in size are concentrated towards the bottom of the depositional sequence. An unusual feature of these volcaniclastic sediments is the occurrence of peridotite clasts, which are a few mm to >20 cm in diameter and represent mantle xenoliths originally entrained in the basanite lavas.

2.2.2 Table Hill

Table Hill (Fig. 2.1, 2.2) is a mound-shaped volcanic remnant approximately 5 km west of the Llangorse locale, and is approximately 400 m in diameter (Table 2.1, 2.2). Lavas at Table Hill are olivine-nephelinites and show moderately well-developed columns,

oriented vertically and ranging in size from 1 to 2 m (Fig. 2.4a). Bulbous, highly vesicular boulders that appear to be lava bombs are present as float boulders. Peridotite xenoliths are common (1-2% by volume) and range in size from 5-10 cm. At Table Hill, crustal xenoliths are rare (<<1% by volume) and appear to derive from the Llangorse batholith. Large (< 2-3 cm) black pyroxene xenocrysts are observed in Table Hill lavas; these xenocrysts are characterized by 1-2 mm reaction rims.

2.2.3 Lone Point

Lone Point (Fig. 2.1, 2.2) is located approximately 5 km south of the Llangorse locale. Lavas exposed at Lone Point are similar in chemistry and morphology to Table Hill, forming a mound-shaped volcanic remnant approximately 300 m in diameter and composed of olivine-nephelinite lavas (Table 2.1, 2.2). A distinctive feature of the Lone Point lavas is the highly variable size and orientation of cooling joints. Well-developed, 10-30 cm diameter column jointing shows highly variable fanning orientations on the east and west faces of the outcrop (Fig. 2.4b). The north face of the lava mound features larger (~30 cm), poorly formed, vertical joints. Highly vesicular lava and volcaniclastic float boulders are present around the outcrop. The volcaniclastic boulders have an orange-coloured, indurated, juvenile volcanic clast-rich matrix (clasts $\leq 1-2$ mm). Clast composition of the volcaniclastic material comprises predominantly vesicular and massive lava clasts (≤ 10 cm in size), as well as peridotite (≤ 2 mm), chert ($\leq 2-3$ cm) and granodiorite (≤ 5 cm) clasts. The volcaniclastic material in the boulder is poorly sorted and juvenile volcanic clasts are angular. The blocky, angular nature of the volcanic clasts and the matrix composition



Figure 2.4: Field photographs of lavas at Table Hill (A) and Lone Point (B). (A) Columnar jointed lavas from Table Hill; outcrop is approximately 6 m high. (B) Fanning of columnar jointing in lavas at Lone Point; photo height represents approximately 3 m.

suggests that these boulders may be hyaloclastites. Peridotite xenoliths are common at Lone Point (1-2% by volume), ranging in size from 5-10 cm. Crustal xenoliths at Lone Point are also rare (<<1% by volume) but appear to be composed of chert and therefore are likely derived from the Cache Creek Group. Lone Point, like Table Hill, contains black pyroxene xenocrysts with 1-2 mm reaction rims.

2.2.4 Hidden Ridge

The outcrops of lava at Hidden Ridge (Fig. 2.1, 2.2) are newly discovered and not shown on Aitken's map (1959). The outcrop occurs approximately 3 km northeast of the Llangorse locale, is approximately 350 m by 300 m, <50 m high, and forms two main ridges (an eastern and western ridge). The volcanic rocks are aphanitic basanites containing abundant xenoliths, similar to lavas exposed at the Llangorse locale (Table 2.1, 2.2). Columnar jointing is well developed in some areas and in particular in the western ridge. The columnar jointing exposed in the western ridge forms an A-shaped structure, with ~40 cm joints on the flanks plunging at approximately 30° away from the locale of the ridge (Fig. 2.5a); columnar joints exposed at the locale of the ridge are roughly vertical. On the southernmost region of the western ridge, horizontal platy columnar jointing is observed (Fig. 2.5b). Columnar jointing in the western ridge is poorly developed, but where it is observed the joints are small (< 20-30 cm), dip at approximately 30° into the locale of the flow, and fan slightly at the base of the flow following the contour of the paleovalley floor. Bulbous, highly vesicular lava float boulders are observed in lava scree and appear to be lava bombs. The contact between the lava and the granodiorite basement rock (the



Figure 2.5

Figure 2.5: Field photographs of lavas and debris flow sediments at Hidden Ridge. (A) Columnar jointing of lavas on the western ridge; photo is facing south. Cooling joints on the north side of the ridge dip at approximately 30° to the north (towards the viewer). Cooling joints on the top of the ridge are roughly vertical, and joints on the south (far) side of the ridge dip at approximately 30° to the south (away from the viewer). Arrow indicates location of photograph (B). (B) Platy columnar jointing on the easternmost area of the western ridge; arrow in (A) indicates location of (B). Platy jointing is roughly horizontal, and the outcrop exposing these joints faces south. (C) Debris flow sediments that underly the lava flow; person for scale. Sediments are crudely bedded, poorly sorted, matrixsupported, and contain clasts of granodiorite, vesicular basanite, palagonitized basanite, and rare peridotite. The volcaniclastic sediments dip down-valley to the northeast and define the shape of a narrow valley that is partially filled by the lava flow. Llangorse batholith) is clearly exposed. Lavas near the contact contain abundant granodiorite xenoliths, whereas granodiorite xenoliths are rare in the central part of the flow. The granodiorite xenoliths near the contacts appear highly deformed due to the flowing lava (flow deformation textures; e.g., Smith, 2002), and have reacted with the lava to a large degree resulting in porous, partly dissolved xenoliths. Xenocrysts of plagioclase, quartz and amphibole, generally 1-2 cm in size, are common within the basanite and gradually disappear towards the centre of the flow, suggesting that some melting of the granodiorite occurred during emplacement of the lava. Peridotite xenoliths are common (1-5% by volume) and are dominantly 1-10 mm in diameter, although some are as large as 15 cm. Large peridotite xenoliths are more common in the eastern ridge than the western ridge, and lava in one area in particular is extremely rich in xenoliths, comprising of up to 70% xenoliths. Unlike Table Hill and Lone Point, the Hidden Ridge lavas do not contain large black pyroxene xenocrysts.

Volcaniclastic debris flow sediments, very similar in composition to those found at the Llangorse locale, occur directly beneath the lava flow in the lowest-elevation areas (Fig. 2.5c). The volcaniclastic sediments dip down-valley to the northeast at approximately 20° and define the shape of a narrow valley that is partially filled by the lava flow. Very crude bedding is observed, with the largest boulders (up to several metres) concentrated at the bottom of the sediments. The volcaniclastic sediments are, in most areas, matrix-supported and poorly sorted, with subangular to subrounded clasts. The sediments are indurated but weather easily, and contain clasts of granodiorite, vesicular basanite, palagonitized basanite, and rare (<1% by volume) millimetre to centimetre-size peridotite. The matrix is composed

of sand-size particles similar in composition to the larger clasts, as well silt and clay-size particles.

The morphology of the lava flow at Hidden Ridge shows clear evidence that it is a valley-filling flow. The contact morphology suggests that granodiorite basement formed an irregular surface in the valley that was surrounded and partly to completely covered by the basanite lava and volcaniclastic sediments. Volcaniclastic sediments are observed only in the lowest-elevation areas (paleotopographic lows). Basanite lava also occurs in depressions across the top of the exposed granodiorite outcrop, and coats the granodiorite basement at exposed contacts. Cooling joint orientation varies across the flow. In the eastern ridge cooling joints are poorly developed; in the western ridge cooling joints are well developed with orientations varying from platy, subhorizontal joints to large (~40 cm), vertical joints.

2.2.5 Chikoida Mountain

The Miocene to Holocene volcanic rocks at Chikoida Mountain (Fig. 2.1, 2.2) were described by Mihalynuk and Smith (1992) and by Edwards et al. (1996), and were interpreted to represent hypabyssal feeder pipes to a volcanic locale subsequently removed by glaciation. The locality features several small (<1500 m²), highly weathered outcrops of basanite. Columnar joints in two outcrops are nearly horizontal, dipping at ~15° into the locale of the outcrop (Fig. 2.6a, 2.6b). Both crustal and mantle xenoliths are observed in the lava. Peridotite xenoliths are rare (<<1% by volume) and <5 mm in size. Crustal, granitic xenoliths are common (1-5% by volume), most likely derived from the surrounding granitic intrusion (Fig. 2.6a). They are up to 50 cm in size, and in some places are partly melted and



Figure 2.6: Field photographs of lavas at Chikoida Mountain. (A) Lava flow showing welldeveloped, roughly horizontal columnar jointing; person for scale. Granitic crustal xenoliths are observed (indicated by arrows) and are often highly disturbed and deformed by heat and flow of the lava. (B) Close up of well-developed horizontal columnar jointing in lavas; hammer for scale. can be significantly deformed by flow. The chemistry of the Chikoida Mountain lavas is unknown, but they are basaltic in composition, likely alkali-olivine basalts or basanites.

Small amounts of volcaniclastic sediments underlie the lava flow at Chikoida Mountain. The sediments are matrix-supported, and in most places there are <15% clasts; clasts are generally small (<10 cm in size) and composed of rounded granite and subrounded massive and vesicular basanite clasts. Peridotite clasts are rare, and all are <3 mm in size. The association of volcaniclastic sediments with the lava suggests that the Chikoida volcanic rocks were subaerially erupted as a valley-filling flow. The matrix-rich nature of the volcaniclastic sediments and absence of large clasts suggests that these deposits may be a substantial distance from their source. The volcaniclastic sediments and lava are present at a high elevation (1850 m), indicating that the source was considerably higher and has been eroded.

2.3 CHEMISTRY OF THE LLANGORSE VOLCANIC FIELD

The NCVP is host to a wide variety of lava types (Fig. 2.7a), ranging from silicic (e.g. peralkaline rhyolites) to alkaline-mafic (e.g. alkali olivine basalts) (Edwards and Russell, 1999, 2000). Silicic magmas are restricted to the few long-lived volcanic complexes (e.g. Level Mtn., Mt. Edziza), whereas alkaline-mafic magmas are ubiquitous throughout the NCVP. Lavas of the AVD (Fig. 2.7a) are all alkaline basalts, ranging from nephelinite to trachybasalt (Edwards et al, 2003).

Lavas in the LVF are all broadly characterized as alkaline basalts. Detailed classification of the lavas, however, shows a range of lava types. On a total-alkali versus


Figure 2.7: TAS diagrams (after Le Bas et al, 1986). (A) Miocene to Holocene volcanic centres in the northern Cordilleran volcanic province (NCVP; large, light-gray shaded region) and the Atlin volcanic district (AVD; smaller, darker-gray shaded region) (Edwards et al, 2003).
(B) The Llangorse volcanic field (LVF), including data from the Llangorse Mountain locale, Table Hill, Lone Point, Hidden Ridge, and Hirschfeld Creek.

silica plot (TAS; after Le Bas et al, 1986), LVF lavas range from alkali olivine basalt (AOB; Hidden Ridge) to basanite (Llangorse locale) to olivine-nephelinite (Lone Point, Table Hill) (Fig. 2.7b; Table 2.2). All three of these rock types are found at the Hirschfeld Creek locality (Francis and Ludden, 1995), where a single body grades from olivine-nephelinite on the margin to AOB in the centre. Francis and Ludden (1990) used normative-mineralogy to classify NCVP lavas as: AOB, defined as >0, <5 wt% normative nepheline (e.g. Hidden Ridge); basanites, defined as 5-15 wt% normative nepheline with dominantly olivine phenocrysts (e.g. Llangorse Mountain locale); olivine-nephelinites, defined as >15 wt% normative nepheline (e.g. Lone Point, Table Hill). Figure 2.7 shows distinct groups of lava types at the localities nearest Llangorse Mountain, rather than a gradation, suggesting that there are two chemically distinct groups: Lone Point and Table Hill are olivine-nephelinites, whereas the Llangorse locale and Hidden Ridge are basanites (or border-line AOB's). Based on the chemical, petrological, and field relationships of the four volcanic localities nearest Llangorse Mountain, I suggest that two distinct mantle melting events are represented: Lone Point and Table Hill represent nephelinite volcanism, whereas the Llangorse locale and Hidden Ridge represent basanite volcanism.

2.4 CONCLUSION

The Llangorse Volcanic Field (LVF) consists of seven localities of Miocene to Holocene volcanic rocks: Hirschfeld Creek/Line Lake, Mt. Sanford, the Llangorse Mountain locale, Table Hill, Lone Point, Hidden Ridge, and Chikoida Mountain. The four localities nearest Llangorse Mountain sites and Chikoida Mountain were mapped in detail during the

summer of 2002. All four sites near Llangorse Mountain are Miocene to Holocene in age. The Llangorse Mountain locale and Hidden Ridge are basanites associated with debris flow sediments, and are interpreted to represent valley-filling lava flows. Table Hill and Lone Point are mound-shaped, and Lone Point displays large variations in columnar jointing size and orientation; Table Hill may be an erosional remnant of a volcanic neck, whereas Lone Point appears to be a subglacial mound. Chemical distinctions between these localities lead us to postulate that there were at least two different eruptive events, nephelinite volcanism and basanite volcanism. Our interpretations of the lava flow morphologies and petrology are provided in Chapter 3.

2.5 ACKNOWLEDGMENTS

I would like to thank Bob Anderson for his field support and expertise, Lyle Hanson for his assistance in the field, and Kirstie Simpson for her helpful improvements to this manuscript.

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Chapter 3

PHYSICAL VOLCANOLOGY

3.1 INTRODUCTION

The northern Cordilleran volcanic province (NCVP) comprises Miocene to Holocene volcanic rocks distributed across northwestern British Columbia, the Yukon Territory, and eastern Alaska (Edwards and Russell, 2000). The province is dominated by small, mafic, alkaline centres that represent single eruptive events (Edwards and Russell, 1999, 2000), although long-lived volcanic centres with evolved lava types do occur in the NCVP (e.g. Mt. Edziza, Hoodoo Mtn., Level Mtn.). The Atlin volcanic district (AVD), a subdivision of the NCVP, consists of several fields of Miocene to Holocene volcanoes (Fig. 2.1; see Chapter 2), including the Llangorse and Surprise Lake volcanic fields (Edwards et al, 2003; Harder et al, 2003). All lavas from the AVD (Fig. 2.7a; see Chapter 2) are alkaline, ranging in composition from nephelinite to trachybasalt (Edwards et al, 2003). The Llangorse volcanic field (LVF) is approximately 144 km² and is centered around Llangorse Mountain (Fig. 2.2; see Chapter 2), which is located approximately 55 km southeast of Atlin, BC (Fig. 2.1).

In this paper we present field observations, petrographic descriptions, and chemical data on lava flows and associated volcaniclastic deposits from Llangorse Mountain. This information is used to constrain the volcanological origins of these deposits. A detailed study of valley-filling basanite lavas and their underlying debris flow deposits suggests that these two units are genetically related. The nature of the debris flow sediments and their association with the lava flows suggests that they may represent lahars or outburst floods

associated with melting of a snow-cap or glacial ice during the early stages of eruption. Subsequent basanitic lava flows derived from the same eruption were captured by drainage systems surrounding the volcanic vent, over-riding the sediments and partly filling the river valleys. The juxtaposition of hot basanitic lavas over water-saturated sediments containing a high proportion of basanitic clasts in the sediment matrix led to palagonitzation of the glassy clasts. Basanite lavas at the Llangorse Mountain locale ultimately were caused to overthicken by encountering a large valley-filling barrier, possibly glacial ice, at lower elevations. In summary, we infer that the lavas and sediments to result from a series of short-lived, possibly subglacial volcanic eruptions from a vent that has been subsequently removed by glaciation.

3.2 THE LLANGORSE VOLCANIC FIELD

The regional geology of the Atlin area was first mapped by Aitken (1959); during his mapping, numerous young volcanic remnants were noted. The area straddles the boundary between the Coast and Intermontane belts, however, the Atlin volcanic district (AVD) lies entirely within the Intermontane Belt. The AVD (Fig. 2.1) is underlain by Paleozoic sedimentary and volcaniclastic rocks from the Cache Creek Group and by Jurassic and Cretaceous plutons (mostly granitic to granodioritic in composition). The LVF (Fig. 2.2) is underlain primarily by the Llangorse Mountain Batholith, a Jurassic granodiorite intrusion, and possibly by cherts from the Cache Creek Group (Aitken, 1959).

Miocene to Holocene volcanic rocks in the LVF occur at Hirschfeld Creek, Mt. Sanford, Llangorse Mountain, and Chikoida Mountain (Table 2.1; see Chapter 2). Most of

these occurrences have been studied previously by Nicholls et al. (1982), Higgins and Allen (1985), Carignan et al. (1994), Francis and Ludden (1995), Edwards et al. (1996), Shi et al (1998), Peslier et al. (2000; 2000b), and Abraham et al. (2001). Lavas in the LVF include alkali olivine basalts, basanites, and olivine-nephelinites (Fig. 2.7; Table 2.2; see Chapter 2). The Llangorse Mountain locale, a Middle-Pleistocene volcanic remnant, comprises olivine porphyritic basanite lavas, having 5-15 wt% normative nepheline (e.g., Francis and Ludden, 1990). Other relevant volcanic centres include: Table Hill (Fig. 2.2; olivine nephelinite), Lone Point (Fig. 2.2; olivine-nephelinite), and Hidden Ridge (Fig. 2.2; alkali olivine basalt).

Lavas at the Llangorse locality have been studied in detail and interpreted by various workers as: a) a series of lava flows, some with slightly different chemistry, overlying debris flow deposits (Nicholls and Stout, pers. comm.); b) a vent plug or a vent-proximal flow (Edwards et al, 1996); c) a lava lake or tuya overlying non-syngenetic debris flow deposits (Francis and Ludden, 1995).

3.2.1 Llangorse Mountain Locality Basanite Lavas

The focus of this study is the Llangorse Mountain locale (Fig. 2.2), an erosional volcanic remnant situated immediately south of Llangorse Mountain. At this locality, there is a 100 m thick, 350 m wide bluff comprised of basanite lava. Vertical columnar joints extend throughout the basanite lava flow (Fig. 3.1a, 3.2a); the columns are generally smaller at the top of the flow (~1 m) than at the base (~2 m); at the base of the flow columns fan thereby defining the paleovalley floor (Fig. 3.1b, 3.2b). The lower margin of the lava flow is marked by a 30 cm thick zone of highly vesicular, quenched basanite (Fig 3.4a). The lava



Figure 3.1: Field photograph and interpretative sketch of outcrop exposures at the Llangorse Mountain locale. (A) South-facing view of the valley-filling basanite flow. (B) Sketch showing contact relationships between the basanite flow (BF), debris flow sediments (DS), and granodiorite (G) shown in (A). Lines in the basanite flow represent approximate orientations of columnar jointing, which fan at the base of the flow thereby defining the paleovalley floor. Lines in the debris flow sediments approximate dips of bedding.



Figure 3.2: Field photograph and interpretive sketch of outcrop exposures at the Llangorse Mountain locale. (A) East-facing view of the basanite flow, showing the contact and orientation of debris flow sediments (top right corner of photo). (B) Sketch showing contact, bedding, and columnar jointing orientations in (A). Abbreviations as in Fig. 3.1.

is aphanitic, although no glass is observed. At the base of the flow (≥ 10 cm), the basanite has a mottled appearance, with many small fractures in the lava forming approximately polygonal shapes with roughly straight faces; each polygon is ≥ 1 cm in diameter. Quartz xenocrysts (likely from granodiorite material) comprises up to 1 vol% of the basanite lava at the contact.

Mantle-derived xenoliths are common in the base of the lava flow (1-5 vol%). Xenoliths are rare (<<1% by volume) in the upper part of the flow and suggests one of two things: a) peridotite xenoliths were only present in the early phases of eruption; or b) the entire basanite bluff was molten at one time, allowing the dense peridotite xenoliths to settle out of the upper part of the flow and concentrate at the base. The peridotite xenoliths are spinel lherzolites and harzburgites. The xenoliths are typically \geq 5 cm in size, but xenoliths 10 cm in size are common and the largest xenoliths observed are approximately 25 cm in diameter. The maximum size of peridotite xenoliths implies ascent rates of between approximately 18-36 cm/s (Fig. 3.9, Appendix 3.A; after Spera, 1984). Crustal xenoliths are exceedingly rare (<<1% by volume) and are always small (< 5 cm) at this locality; most of these appear to be derived from the Llangorse batholith and may represent surficial fragments entrained during flow of the lavas.

Lavas at the Llangorse Mountain locale are fine-grained (200-500 μ m; Fig. 3.3a). The lavas contain small (<5%) amounts of xenocrystic and microphenocrystic olivine and pyroxene. Microphenocrysts are very rare and consist of small (\geq 150 μ m) clinopyroxene or olivine grains, and occur as unzoned, euhedral crystals. Xenocrysts (\geq 250 μ m) are slightly more abundant than phenocrysts and are characterized by prominent reaction relationships



Figure 3.3

Figure 3.3: Photomicrographs of volcanic rocks from the LVF. (A) Fine-grained basanite lavas from the Llangorse Mountain locale; sample is composed of lath-shaped plagioclase (~40%), clinopyroxene (30-40%), and subordinate olivine, nepheline, Fe-Ti oxides, and rare biotite. (B) Fine-grained basanite recovered from the chilled base of the lava flow, immediately overlying the debris flow sediments. Sample mineralogy is similar to (A) with one important distinction: considerably more biotite (Bt) is present (2-3%) in the chilled base of the lava flow. (C) Fine-grained basanite lavas from Hidden Ridge; sample mineralogy is similar to (A). (D) Very fine-grained nephelinite lava from Table Hill composed of approximately 25% plagioclase (~25%), clinopyroxene (~40%), nepheline (>15%), and subordinate olivine and Fe-Ti oxides. (E) Very fine-grained nephelinite lava from Lone Point; sample mineralogy similar to (C). (F) Orthopyroxene phenocryst (Opx) from Lone Point showing breakdown to a corona of clinopyroxene (Cpx) due to interaction with the high temperature melt.





with the host lava, commonly expressed as thin (< 50μ m) reaction coronas or resorbed margins. Orthopyroxene and olivine are the most common xenocryst grains, which is consistent with peridotite xenolith mineralogy; disaggregation of peridotite xenoliths is a likely source of these xenocrysts. Neither xenocrysts nor phenocrysts are visible with the naked eye, thus variations within the flow are not evident from our limited sampling of the lavas.

Lava samples from the highly fractured, quenched base of the flow (Fig. 3.3b) show the same mineralogy as the main bluff, with one important distinction. Primary magmatic biotite is observed at the very base of the flow, intergrown with other minerals. Biotite appears to be the last mineral to crystallize, and consequently is often subhedral to anhedral, filling spaces between other minerals. The presence of biotite only at the base of the flow is significant as it indicates that the chemistry or crystallizing conditions are unique in the quenched base.

The quenched basanite at the base of the Llangorse Mountain bluff directly overlies debris flow sediments (Fig. 3.5a). The contact is exposed along the west side of the massif. The debris flow sediments immediately below the chilled base of the lava flow do not show any erosional profile (i.e., there are no signs of vegetation or reworking of sediments). The debris flow sediments are not observed at similar elevations beneath the flow on the eastern side of the massif, where the contact juxtaposes lava with a highly weathered surface of granodiorite belonging to the Llangorse Mountain batholith. The highly weathered nature of the granodiorite (e.g., grus) at this contact indicates that, at this location, the granodiorite was not glaciated. The eastern contact is only observed near the top of the lava flow and is

nearly vertical in orientation, which explains the absence of debris flow sediments.

3.2.2 Debris Flow Sediments

The debris flow sediments underlying the basanite flow at Llangorse Mountain are consolidated but still friable; crude layering at the base of the deposit dips approximately 25° to the northwest and is most evident along the extreme southwest contact of the basanite lava flow (Fig. 3.2b; 3.5a). The sediments are poorly sorted, crudely stratified, and contain subangular to rounded fragments (Fig. 3.4b, c; Fig.3.5b). Crude bedding can be seen in the deposit, identifying numerous debris flow pulses (outlined in Fig. 3.2b). The sediments are primarily matrix-supported, with minor clast-supported areas. The dominant clast type is granodiorite, which derives from the surrounding Llangorse batholith. Granodiorite clasts vary from 2 cm to 5 m in diameter (see bottom right corner, Fig. 3.5a) and are generally well-rounded, suggesting that they were previously rounded by another process. Clasts of basanitic lava are also common, vary in size from 2 to 50 cm, and are generally angular. Palagonitized clasts of basanite vary from angular to rounded (Fig 3.4d, c); the rounded clasts, however, have a fluidal appearance which is likely a primary volcanic texture, and not related to transport in the debris flows (Fig. 3.6d). Peridotite clasts vary in size from < 1cm to approximately 25 cm. Some peridotite clasts have rims of basanite lava; conversely, many basanite clasts contain peridotite xenoliths. There is no discernible difference (e.g., chemistry, mineralogy, mineral chemistry) between peridotite samples collected from the basanite lava flow and those taken from the debris flow sediments.

The matrix of the debris flow has a typical grain size of 0.5-2.5 mm, and is composed



Figure 3.5: Field photographs of debris flow sediments at the Llangorse Mountain locale. (A) Detail of crude layering in debris flow sediments; scale provided by two people on the sediment / lava contact (arrows). (B) Detailed view of debris flow sediments (person for scale) illustrating crude graded bedding and poorly sorted, matrix supported, heterolithic nature of sediments. Light coloured blocks are rounded boulders of granodiorite.

primarily of disaggregated granodiorite clasts, with subordinate basanite fragments and disaggregated peridotite grains (Fig. 3.4c). The matrix lacks a silt to clay size fraction. Clastic grains of quartz, feldspar, hornblende and biotite are abundant throughout the matrix, and appear to derive from the granodiorite country rock (Fig. 3.6a, c, e). Olivine, clinopyroxene, and orthopyroxene grains are also common, deriving from peridotite clasts. Clasts of basanite (aphyric and therefore unpalagonitzed) are invariably vesicular and very fine grained (Fig. 3.6f). Basanite and palagonitized basanite clasts usually contain xenocrysts similar to the deposit matrix, with quartz, olivine, feldspar, and biotite as the most common xenocrysts. Palagonitized clasts of basanite comprise up to 30% of the debris flow sediments (Fig. 3.6a). The presence and nature of these palagonitzed clasts of basanite in the debris flow sediments provide provide unique constraints on the post-depositional conditions in the sediments.

Palagonitized clasts of basanite within debris flow sediments at the Llangorse Mountain locale are very heterogeneous, varying from a bright orange, transparent material (Fig. 3.6a, 3.6c), to a dark reddish-brown, nearly opaque material (Fig. 3.6b, 3.6d). No primary basanite glass is observed, although in some larger clasts, a rim (2-3 cm thick) of palagonitized basanite grades into an aphanitic (but crystalline) basanite core (Fig. 3.6d). Palagonitzed clasts of basanite vary in vesicularity from as low as 5% vesicles (Fig. 3.6c) to as high as 50-60% (Fig. 3.6b). In all clasts, however, primary porosity (vesicles and matrix pore space) is partly to completely filled by secondary zeolite cements (Fig. 3.6a-e). Secondary zeolite forms rims on all palagonite grains and often rims other clasts in the matrix, thereby cementing the entire deposit. The partially indurated nature of the sediments is critical as it has allowed the sediments to resist complete erosion.



Figure 3.6: Photomicrographs and scanning-electron microscopic (SEM) images of clasts and matrix within debris flow sediments. (A) Palagonitized basanite clasts (PB), containing numerous vesicles (Ve) and xenocryst grains (e.g., olivine (Ol)). Clasts derived from country rocks (e.g., quartz (Qz)) and peridotite (e.g., Ol, from peridotite xenoliths entrained by the lavas) are common as xenocrysts within palagonitized basanite grains and as individual grains in the sediments. Zeolite cements (Ze) are common in the deposit. (B) Highly vesicular palagonitized basanite clast (dark material). Most of the original vesicles (Ve) are filled in with zeolite cements (Ze). (C) Palagonitized basanite clast surrounding a large xenolithic quartz grain. (D) Palagonitized basanite clast showing gradation from an aphanitic, but crystalline, core (left side of photomicrograph) to a palagonitized rim (right side of photomicrograph). Vesicles in the palagonitized rim become increasingly fluidal approaching the clast boundary. (E) SEM image of a palagonitized basanite clast, illustrating the heterolithic nature of xenocrysts (Feldspar (Fe), Qz, Ol) and the level of induration caused by zeolite cements. (F) Extremely fine grained clast of basanite lava recovered from the debris flow sediments.

X-ray diffraction (XRD) analyses were performed on palagonitized basanite clasts and associated zeolite cements. XRD analyses were performed on the Siemens (Bruker) D5000 Bragg-Brentano X-ray powder-diffractometer at the University of British Columbia; operation conditions are detailed by Raudsepp et al (1999). The results indicate that the palagonitized material is composed of very fine grained clays, with a predominantly smectite component. The zeolite cements have strong phillipsite and chabazite signatures.

3.2.3 Other LVF Centres

Although not the focus of this study, several smaller mafic-alkaline centers within a 5 km radius of the Llangorse Mountain bluff contain features which are important to consider. These sites include: (a) a 200 m in diameter mound-shaped outcrop of nephelinite (Lone Point; Fig. 2.2; Fig. 2.7b); (b) a basanite ridge, approximately 250 m by 100 m (Hidden Ridge; Fig. 2.2; Fig. 2.7b). The Lone Point outcrop shows complex development of columnar jointing (Fig. 3.7). Well-developed, 10-30 cm diameter columns show highly variable fanning orientations on the east and west faces of the outcrop (outlined in Fig. 3.7b). The north face features larger (~30 cm), poorly formed, vertical joints. The morphology of Lone Point and the presence of erratically oriented columnar jointing at this locality suggests that it may have been erupted under glacial ice, forming a lava mound with some similarities to tuyas (Mathews, 1947; Jones, 1966, 1969; Lescinksy and Fink, 2000).

In overall morphology, although not scale, Hidden Ridge is very similar to the Llangorse Mountain bluff (Fig 2.5, Chapter 2). At both centres, lavas are valley-filling flows that overly a sequence of debris flow deposits containing clasts of granodiorite,



Figure 3.7: Field photograph and interpretive sketch of outcrop exposures at the Lone Point nephelinite locale (Fig. 2.2, 2.7; Table 2.1, 2.2). (A) Photograph of fanning columnar jointing in lavas; photo height represents approximately 3 m. (B) Sketch showing the orientations of columnar jointing shown in (A); arrows indicate columnar joints that plunge into the locale of the lava mound, away from the viewer.

basanite lava, peridotite, and palagonitized basanite. Hidden Ridge, however, displays variably oriented columnar joints, similar to Lone Point.

The petrology of these centres reveals important information regarding the association of eruptive events associated in the LVF. Fine grained basanite lavas from Hidden Ridge (Fig. 3.3c) have a similar mineralogy to the lavas at Llangorse Mountain, although the lavas have slightly higher silica contents (Table 2.2; Fig. 2.7b). The higher SiO₂ contents at Hidden Ridge may reflect greater contamination by country rock material as the lava directly overlies granitic bedrock (see Chapter 2). Very fine grained nephelinite lavas from Table Hill (Fig. 3.3d; Fig 2.4a) and Lone Point (Fig. 3.3e; Fig 2.4b) contain microphenocrysts of olivine and pyroxene and large (≥ 2 cm) xenocrysts of orthopyroxene (Fig. 3.3f). These xenocrysts are characterized by significant (up to 1.5 mm thick) clinopyroxene coronas likely caused by interaction with the high temperature lava. These large, distinctive xenocrysts are absent at both the Llangorse Mountain locale and Hidden Ridge. Although these orthopyroxene xenocrysts presumably source from disaggregated peridotite xenoliths, interestingly there are proportionally much fewer peridotite xenoliths at Table Hill and Lone Point (<< 5%) than at Llangorse Mountain and Hidden Ridge. Conversely, Table Hill and Lone Point have far greater amounts of crustal xenoliths (1-2%) than the other localities.

3.3 VOLCANOLOGICAL MODEL

3.3.1 Glaciovolcanism in the LVF

The valley-filling basanite lava flow exposed at the Llangorse Mountain locale is

approximately 100 m thick and 350 m wide (Fig. 3.1). These dimensions give an aspect ratio (flow thickness / flow width) of approximately 0.3. Basanite lava flows are typically characterized by effusive eruptions of sheet-like lava flows with aspect ratios on the order of 10⁻⁴ due to their low viscosity (Best and Christiansen, 2001). Clearly, the Llangorse Mountain bluff is overthickened by ~3 orders of magnitude. Moreover, the contact orientations of the lava flow change from a dip of approximately 25° at the base of the flow (parallel to the paleovalley floor) to horizontal at the top of the flow (Fig. 3.2). The wedgeshaped morphology and overthickened aspect ratio of these lavas suggests that the ponded against a lower-elevation, valley-wide barrier.

The present day valley which the Llangorse Mountain bluff faces is a wide, glacially sculpted U-shaped valley, with no evidence of any lithological barrier. The absence of any barrier combined with the morphology of the present-day valley suggests that glacial ice was a likely barrier against which the Llangorse Mountain lavas ponded (Fig. 3.8). The nearly vertical cliff of lava and absence of horizontal columnar joints suggests that the leading edge of the lava flow has been removed; however, platy columnar jointed lava boulders are not observed in the present-day scree apron. As lava displaying this jointing style is highly susceptible to mass wasting (i.e. landslides) and erosion, the absence of any platy columnar jointed lava at this locale is to be expected. A likely mechanism for removing the platy columnar joints is mass wasting accompanying the retreat of the glacier.

The debris flow deposits immediately underlying the Llangorse Mountain lava flow are interpreted to be syngenetic with the lavas, based on the presence of basanite and peridotite clasts in the debris flows that are chemically identical to the lavas and peridotites



Figure 3.8

Figure 3.8: Interpretive sketch of the impoundment and overthickening of the basanite lavas at the Llangorse Mountain locale. The present-day basanite flow (BF) and debris flow sediments (DS) are as represented in Fig. 3.2. Our interpreted paleo-basanite flow (PBF) is shown impounded against glacial ice, resulting in the overthickening of the lava flow. Horizontal lines perpendicular to the ice mass represent platy jointing; this jointing is not observed today as a result of high erosion rates of the bluff. Dashed lines indicate approximate orientations of columnar jointing. Columnar jointing near the ice sheet is expected to be horizontal; with increasing distance from the ice sheets the joint orientation will become vertical, consistent with subaerial cooling.

in the overlying bluff (see Appendix D). The clast composition of the debris flow, in particular the presence of very fine-grained, vesicular basanite clasts, palagonitzed mafic glass clasts, and peridotite xenoliths coated in lava, indicates that the debris flows sampled material from around the volcanic vent. The nature of the sediment / lava contact indicates that very little time passed between deposition of the debris flows and emplacement of the lava flow. Firstly, no erosional profile is observed in the sediments immediately below the sediment / lava contact. Secondly, the very base of the overlying lava flow is highly fractured, with small-scale (≥ 1 cm), roughly polygonal jointing. The nature of this jointing suggests that rapid quenching (possibly in the presence of water), and not basal flow brecciation, is the cause of the fracturing (Grossenbacher and McDuffie, 1995; Cas and Wright, 1987).

The fine-scaled, roughly polygonal jointing at the base of the lava flow (Fig 3.4a) implies rapid quenching against the underlying cool, water-saturated sediments (e.g., Grossenbacher and McDuffie, 1995). The granitic material present in the quenched base of the lava flow may source from disaggregated granodiorite xenoliths, or from sediments incorporated into the base of the flow from the underlying debris flow sediments. The presence of primary magmatic biotite within this highly jointed base of the lava flow (Fig. 3.3b) suggests higher water contents than found in the rest of the lava flow. If mixing of the underlying wet debris flow sediments occurred (thereby explaining the presence of granitic material in lavas at the base of the flow), the water content of lavas at the base of the flow could be increased, allowing biotite to be a stable phase during the latest stages of crystallization.

Debris flow sediments underlying the basanite lavas at Llangorse Mountain have many features suggesting the sediments were deposited proximal to their source. The clay and silt-poor nature of the debris flow matrix and the presence of angular basanite fragments indicates that the sediments are immature and have not traveled far from their source region; this is common in lahar-type deposits (Vallance, 2000). Peridotite clasts found in the sediments are highly friable and would be completely disaggregated by any significant transport. Their presence, therefore, severely limits the distance the debris flows traveled. The juxtaposition of rounded granodiorite clasts and angular basanite clasts excludes a fluvial origin of the sediments. Although the rounded granodiorite boulders could have been rounded by fluvial process pre-dating the basanite eruption, glacially-dropped granodiorite material is scattered across the landscape and is a potential source for the granodiorite material found in the debris flow sediments. The poorly sorted, matrix-supported deposit containing large lithic blocks (up to 5 m in diameter) is consistent with a high-energy, watersaturated debris flow (e.g., Knighton, 1998, Vallance, 2000). Therefore, the juxtaposition of 5 m granodiorite boulders with angular juvenile volcanic clasts in the debris flow sediments at Llangorse Mountain implies that the debris flows were high energy flood events that did not travel far from the source area.

Debris flows are commonly associated with volcanic eruptions (e.g. lahars, outburst floods) as a result of the highly friable and unstable nature of juvenile volcanic material and steep topography (Vallance, 2000; Björnsson, 1975). As well, many volcanic vents occur at high latitudes and therefore are associated with snow cover or glacial ice. Our interpretation of the debris flows at Llangorse Mountain as being syngenetic with the lavas and sampling volcanic material from around the vent is consistent with the early stages of eruption melting

a snow pack or glacial ice over the vent and rapidly producing large amounts of water. Early volcanism caused melting and several pulses of debris flows resulting from this melt water. Continued eruption led to the production of a basanite lava flow, which was captured by the same drainage as used by the debris flows.

Evidence of synglacial and subglacial eruptions are common throughout the NCVP. The Tuya-Teslin volcanic district (Edwards and Russell, 2002) contains abundant subglacial volcanic mounds (tuyas), including the type locality for tuyas (Mathews, 1947). This abundance reflects the coincidence of Pliocene to Pleistocene glaciers with NCVP volcanism (Edwards and Russell, 2000). In particular, evidence of glaciovolcanism is observed at volcanic localities in the Llangorse volcanic field. Erratic size and orientations of columnar jointing in lavas at Lone Point (Fig. 3.7) and Hidden Ridge (Fig. 2.5) require both vertical (i.e. adjacent to the lavas) and horizontal (i.e. above the lavas) cooling surfaces to explain the jointing patterns. Strong variations in size and orientation of cooling joints can be indicative of interactions between lava and ice masses (e.g. Mathews, 1947; Grossenbacher and McDuffie, 1995; Lescinksy and Fink, 2000), suggesting that lavas at these localities were erupted under or against glacial ice. Furthermore, Lone Point has many features indicating that it may be the remnant of a subglacial mound. The mound-shaped morphology of the lavas combined with columnar joint orientation indicates that these lavas erupted under ice. Boulders of hyaloclastite are also found at Lone Point, suggesting that standing water was present around the locale at some point during its eruption. Although hyaloclastite deposits or platy jointing (the jointing style found most proximal to the lava / ice contact; e.g. Mathews, 1947) are not observed at Lone Point, this is likely the result of erosion. Similarly, although hyaloclastite is not observed at Hidden Ridge, horizontal platy

columnar jointing is observed, which may represent a surface very near the lava / ice contact. The geographic proximity and the many textural, chemical, and morphological similarities between Hidden Ridge and the Llangorse locale suggest that the two localities are roughly coeval in age and could source from the same volcanic vent, which has subsequently been eroded away.

3.3.2 Palagonitzed Basanite

An unusual feature of the debris flow sediments at the Llangorse Mountain locale is the presence of palagonitized clasts of basanite. Palagonitized glass forms via the hydration and breakdown of volcanic glass (usually basaltic) in water-saturated environments, and is a common feature of hyaloclastite deposits (Cas and Wright, 1987; Schiffman et al, 2000). Palagonitization is generally only observed in subaqueous (submarine or subglacial) deposits (e.g. Thorseth et al, 1991; Furnes, 1984) or in deposits subjected to hydrothermal circulation of fluids (e.g. Jakobsson, 1978; Jakobsson and Moore, 1986). Rates of palagonitization are extremely low for meteoric alteration and, therefore, palagonite is generally not a product of weathering; temperatures of 50-100°C are generally required for rapid palagonitization (i.e., < thousands of years; Moore, 1966; Friedman and Smith, 1960). Although some previous workers have suggested that palagonitization can occur through weathering (e.g. Hay and Iijima, 1968), recent work suggests that palagonitization is not associated with pure weathering processes (Schiffman et al, 2000; 2002). Schiffman et al (2002) demonstrate the physical and mineralogical differences between palagonitized and pedogenically-altered basaltic tephra deposits, suggesting that fluids involved with

palagonitization are not simply associated with weathering, and that some elevation of fluid temperatures is necessary. Schiffman et al (2000; 2002) conclude that subaerial deposits that are subject only to weathering do not undergo palagonitization, only pedogenic alteration.

The colour of palagonitized glass can be indicative of the temperature of palagonitizing fluids; dark-brown to red-brown palagonite is associated with higher temperature fluids (80-120°C) than orange palagonite (Stroncik and Schmincke, 2001). The common occurrence of dark-brown to red-brown palagonite in the Llangorse locale debris flow sediments (see Fig. 3.6b, d) therefore suggests that fluids associated with palagonitization reached higher than ambient temperatures. The moderately-well cemented nature of the sediments as a result of zeolite precipitation (see Fig. 3.6 a-b) is also typical of well-palagonitzed deposits (Stroncik and Schmincke, 2001; 2002), and generally indicates advanced stages of palagonitzation associated with either high fluid temperatures or long periods of time (i.e. millions of years; Stroncik and Schmincke, 2001).

The association of palagonitized glass with secondary zeolite cements is typical of mafic volcanic rocks that have been palagonitized (Stroncik and Schmincke, 2001; 2002). Moreover, the slightly higher-temperature alteration resulting in palagonitization produces a suite of clay and zeolite minerals that are distinct from those produced by weathering (i.e., pedogenic alteration; Schiffman et al, 2000, 2002). Pedogenic alteration of basaltic glass results in the formation of kaolinite and gibbsite clays; palagonitzation results in smectite clays (e.g. saponite) (Schiffman et al, 2002). Our XRD analyses of palagonitized clasts from the Llangorse locale reveal a dominantly smectite clay signature that is indicative of

palagonitization, not weathering. XRD analyses of zeolite cement shows a dominantly phillipsite and chabazite signature. This combination of clay minerals (in palagonitized clasts) and zeolite cement is characteristic of other palagonitized deposits (Schiffman et al, 2001; 2002). Schiffman et al (2002) also demonstrate that there are clear textural distinctions between palagonitized deposits versus pedogenically altered deposits; palagonitization results in consolidation of deposits due to the precipitation of authigenic cements (e.g. zeolites), whereas pedogenically altered deposits result in degradation of basaltic glasses and unconsolidated deposits. As a result, palagonitzed deposits have a much higher preservation potential than pedogenically altered deposits.

The field location of the debris flow sediments at Llangorse Mountain (i.e. high elevations, steep topography; Fig. 3.2) indicates that the sediments must have been palagonitzed soon after deposition in order to be preserved. If the sediments were altered pedogenically, they likely would have been rapidly eroded away (see Schiffman et al, 2002). The field characteristics of the debris flow sediments at Llangorse Mountain clearly indicate that palagonitization of the sediments was not a subaqueous process. Therefore, the likely mechanism resulting in palagonitzation is hydrothermal circulation of heated waters. The debris flows may have been hot at the time of deposition, consisting of a mixture of hot volcanic particles and warm water derived from melting of ice. Conversely, if the debris flow sediments were initially saturated with cool waters when they were buried by the basanite flow, heating of the sediments by the base of the lava flow could easily heat the sedimentary deposit by 100°C or more, resulting in hydrothermal in-situ palagonitization. Our preferred interpretation is that, regardless of the initial temperature of the debris flow sediments, the basanite flow (> 1200°C, 100 m thick) that soon followed and covered the

sediments must have heated them to well above ambient temperatures. These elevated temperatures could be sustained for several months to years (e.g., Hort, 1997), allowing for very rapid (i.e. a few months to a few years) palagonitization of the debris flow sediments.

3.4 CONCLUSIONS

The Llangorse volcanic field, in northwestern BC, comprises Miocene-Holocene mafic, alkaline lavas. Within the LVF, volcanic rock types vary from alkali olivine basalts to olivine nephelinites. The focus of this study is a large, Middle Pleistocene, valley-filling basanite lava flow exposed south of Llangorse Mountain. We interpret syngenetic debris flow sediments underlying the lava flow to source from melting of a snow-pack or glacier around the volcanic vent in the early stages of eruption. Volcanic material was sampled near the vent and transported downslope in these high energy debris torrents that carried lithic blocks up to 5 m in size. An effusive basanite lava flow soon followed the debris flows; the lava ultimately entered a drainage system blocked by glacial ice, causing the flow to pond and overthicken, becoming at least 100 m thick. Following emplacement of the lava flow, hydrothermal circulation of fluids in the underlying, still water-saturated debris flow resulted in in-situ palagonitzation of the basanite glass shards composing up to 50% of the debris flow matrix. This rapid, high temperature (50-100°C) palagonitization produced authigenic zeolite cements which partly-indurated the debris flow sediments.

3.5 ACKNOWLEDGMENTS

Field costs for this study were covered by Dr. Bob Anderson at the Geological

Survey of Canada, courtesy of the Atlin Integrated Geoscience Project. All other costs were met through NSERC via a Discovery Grant to J.K. Russell and a postgraduate scholarship to Margaret Harder. Volcanological interpretations were greatly improved by discussions with Ben Edwards, Kirstie Simpson, Ian Skilling, and Melanie Kelman.

3.6 APPENDIX 3.A: Ascent Rate of Magmas

Spera (1984) applied Stokes Law of gravitational settling to determine ascent rates necessary to entrain peridotite xenoliths from the lithospheric upper mantle. The following equation was used to calculate the ascent rate of the lava:

$$U_{n} = 0.344 \left(\frac{\Delta \rho g}{\rho_{l}}\right)^{\frac{5}{7}} \left(\frac{\rho_{l}}{\eta_{l}}\right)^{\frac{3}{7}} \left(R_{n} - \frac{15\sigma_{0}}{4\Delta \rho g}\right)^{\frac{8}{7}}$$
(3.1)

where U_n = terminal settling velocity, ρ_l = lava density, $\varphi =$ density difference between peridotite nodule and lava, g = acceleration due to gravity, α_l = viscosity of lava, R_n = radius of peridotite nodule, and κ_0 = yield strength of lava. The values used for these variables to represent the Llangorse Mountain lavas and peridotites are provided in Table 3.1.

In order to entrain a xenolith of radius R_n with a density difference between xenoliths and lava of φ , the lava must be ascending at (or faster than) U_n , thereby giving a minimum ascent rate of the lava. This method assumes that the lavas are not ponding in the crust, but are ascending without pause from the moment that the peridotite xenoliths are entrained. Although this is difficult to prove, the sparsity of phenocrysts indicates that the lavas did not pond in the crust for a long period of time.
Calculations using Eq. 3.1 and the constants in Table 3.1 are shown in Fig. 3.9. Terminal settling velocities of approximately 18-36 cm/s are returned for realistic values of lava viscosity and yield strength. These values represent the minimum ascent rates of magma necessary to entrain the largest peridotite xenolith found, which is approximately 25 cm in diameter. These ascent rates correspond to ascent rates of 0.65-1.30 km/hr.

The MOHO depth in the northern Canadian Cordillera has been identified at 36 km (Hammer and Clowes, 2004), and the depth to the source region of Llangorse Mountain lavas is estimated to be 54-75 km (Harder and Russell, in review; see Chapter 4). Therefore, the estimated travel time from the MOHO to the surface volcanic vent is 27-55 hours (1-2 days). Travel time from the lava source region of the surface volcanic vent is estimated to be between 42-116 hours (2-5 days).

Contants		
$^{1}\rho_{p}$	3300 kg/m ³	
$^{2}\rho_{1}$	2800 kg/m ³	
3 Δρ	500 kg/m ³	
⁴ g	9.81 m/s ²	
⁵ R _n	0.125 m	
Variables: σ_0 value	es and corresponding r	l ₁ values
$^{6}\sigma_{0}(N/m^{2})$	7 η_{1} (Pa s)	% phenocrysts
0	25	0
10	100	< 5
50	500	15
100	1000	25

Table 3.1: Constants and variables used in Eq. (3.1) to calculate the minimum ascent rates of magma that fed Llangorse Mountain lavas.

¹ Density of peridotite.

² Density of lava.

² Density of lava.
³ Density of peridotite nodule minus density of lava

⁴ Acceleration due to gravity.

⁵ Maximum radius of peridotite nodule.

⁶ Yield strength of lava.

⁷ Viscosity of lava.



Figure 3.9: Minimum ascent velocity of the Llangorse Mountain Lavas. Ascent rate (cm/s) is plotted against lava viscosity (Pa s). Calculations of the minimum ascent velocity are determined using Eq. 3.1 (Appendix 3.A); constants and variables used for these calculations are provided in Table 3.1. Realistic values of lava viscosity (< 500 Pa s)and yield strength (\geq 10 N/m²) return terminal settling velocities of approximately 18-36 cm/s.



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Chapter 4

MANTLE PETROLOGY

4.1 INTRODUCTION

The mantle lithosphere underlying the northern Canadian Cordillera has been the focus of geophysical and petrological research aimed at understanding the physical and chemical character of the Cordilleran lithosphere, the origins of the underlying mantle lithosphere and, ultimately, the mechanisms of crustal amalgamation (Clowes et al, 1999; Hyndman and Lewis, 1999; Hammer et al, 2000; Snyder et al, 2002; Lewis et al, 2003; Hammer and Clowes, 2004). For example, the northern Canadian Cordillera became welded to the western margin of North America beginning in the middle Jurassic (Coney et al, 1980; Monger et al, 1982), yet the nature and origin of the lithospheric mantle underlying these amalgamated crustal terranes remains a major uncertainty.

The Canadian Cordillera is ideal for direct study of the mantle lithosphere because of the presence of Miocene to Holocene mafic alkaline volcanic rocks (Fig. 2.1, Chapter 2; Edwards and Russell 1999; 2000). Lavas from these volcanic centres commonly contain peridotite xenoliths, providing primary samples of the mantle lithosphere underlying the northern Cordillera. There have been numerous studies of mantle xenoliths from Miocene to Holocene lavas within the Cordillera; many of these studies have been regional in nature and have relied on relatively small datasets collected from many different localities (Littlejohn and Greenwood, 1974; Ross, 1983; Brearley and Scarfe, 1984; Brearley et al, 1984; Shi et al, 1998; Peslier et al, 2000, 2000b).

The purpose of our study is to explore the utility of mineralogical and chemical

information within peridotite xenoliths for constraining the thermal state of the crustal and mantle lithosphere. Here, we present field observations, petrographic descriptions, and chemical data on a large suite of peridotite xenoliths collected from Llangorse Mountain. We use two-pyroxene thermometry on peridotite xenoliths (Brey and Kohler, 1990) to recover temperatures from the lithospheric mantle. These data constrain the temperatures at the base of the crust (MOHO) and at the base of the mantle lithosphere and are used, in conjunction with heat flow data, to infer the depth and thickness of the lithospheric mantle underlying the northern Cordillera.

4.2 LOCATION AND PREVIOUS WORK

The northern Cordilleran volcanic province (NCVP; Fig. 2.1, Fig. 2.2; see Chapter 2), as defined by Edwards and Russell, (1999; 2000), is one of the largest Miocene to Holocene volcanic provinces in western North America and comprises dominantly mafic, Miocene to Holocene alkaline volcanic rocks. The volcanism is proposed to result from extensional forces acting on the northern Cordilleran lithosphere (Edwards and Russell, 1999, 2000). The NCVP is mainly expressed as small volume, short-lived basaltic volcanoes or erosional remnants of centres (Fig. 2.1), in contrast to several longer-lived edifices (e.g. Mt. Edziza, Level Mtn., Hoodoo Mtn.; Fig. 2.1). The volcanism is distributed across northwestern British Columbia, the Yukon Territory, and eastern Alaska (Edwards and Russell, 2000), and is bounded by the Tintina fault system to the east and the Denali-Coast fault system to the west.

The Llangorse volcanic field (LVF; Table 2.1) covers approximately 144 km² and is

centred on Llangorse Mountain, approximately 55 km southeast of Atlin, BC (Figs. 2.1, 2.2). The LVF (Harder et al, 2003) is part of the Atlin volcanic district (Edwards et al, 2003), which straddles the boundary between the Coast and Intermontane belts (Aitken, 1959; Edwards et al. 1996). The LVF lies entirely within the Intermontane belt and is underlain mainly by the Llangorse Mountain pluton and by minor amounts of Cache Creek Group cherts (Aitken, 1959). There are seven volcanic localities within the LVF, consisting of eroded remnants of lava flows and exposed feeder dykes (Fig. 2.2). This study concerns material collected from the Llangorse Mountain locality (No. 1, Fig. 2.2). There are three chemical types of lava within the LVF (Table 2.2): alkali-olivine basalts (0 < normativenepheline <5 wt%), basanites, (5-15 wt% normative nepheline, containing only olivine phenocrysts), and olivine-nephelinites (>15 wt% normative nepheline; classification after Francis and Ludden, 1990). The lavas of the Llangorse Mountain centre are basanitic in composition with 5-8 wt% normative nepheline and on a Total Alkali-Silica (TAS) diagram plot in the basanite field, near the alkali-olivine basalt boundary (Samples MH-02-101 and MH-02-112; Fig. 2.7, Table 2.2). Volcanic rocks in the LVF have been studied by numerous groups (Table 2.1), including broad scale studies investigating chemical trends in the northern Cordillera (Carignan et al, 1994; Abraham et al, 2001) and smaller, detailed studies on lavas at one locality (e.g. Hirschfeld Creek; Francis and Ludden, 1995). Peridotite xenoliths, which represent sampling of the mantle lithosphere by these alkaline lavas, have also been the focus of many studies. Detailed petrology (e.g., petrography, chemistry and thermometry) of peridotite xenoliths provide direct evidence of the physical and chemical composition (Nicholls et al, 1982; Nicholls and Stout, 1996, 1997; Shi et al, 1998, Peslier et al, 2000b), the age and origin (Nicholls et al, 1982; Peslier et al, 2000), and

the thermal state of the mantle (Nicholls et al, 1982; Shi et al, 1998).

4.3 FIELD DESCRIPTION

The Llangorse Mountain centre is a large, cliff-forming bluff approximately 100 m high and 350 m wide (Figs. 2.3a, 2.3b). It is interpreted as an erosional remnant of a valley-filling basanite flow. Vertical columnar joints extend throughout the flow. The columns are generally smaller at the top of the flow (~1 m) than at the base of the flow (~2 m) and fan at the base of the flow, defining the paleo-valley floor. The base of the flow is marked by a 30 cm thick zone of highly vesicular, very fine-grained, quenched (but not glassy) lava. The lavas are aphanitic, non-porphyritic (< 5% phenocrysts), and contain abundant accessory mantle-derived peridotite xenoliths and xenocrysts. Samples from the centre of the flow are finely crystalline and are composed of approximately 40% lath-like plagioclase, 30-40% clinopyroxene, and subordinate amounts of olivine, nepheline, and Fe-Ti oxides.

At higher elevations, the eastern side of the massif comprises outcrops of highly weathered granite (e.g., grus) in direct contact with the lava. On the west side of the massif (Fig. 2.3b), the quenched base of the lava flow rests directly on syngenetic volcaniclastic debris flow sediments that exhibit crude bedding (Fig. 2:3c). The volcaniclastic debris flow sediments are poorly sorted, crudely stratified, and contain mainly subrounded to rounded blocks of granodiorite, which appear to derive from the Llangorse batholith, and subordinate angular to subangular (2-50 cm) fragments of lava (Fig. 2.3d). The sediments are primarily matrix-supported, but have minor domains that are clast-supported.

Our analysis of the underlying volcaniclastic debris flow sediments indicates that

their deposition may be directly tied to the volcanic eruption responsible for the overlying basanite lava flow. The debris flow sediments contain clasts of vesicular lava that are similar in mineralogy and mineral chemistry to the basanite. Furthermore, the volcaniclastic sediments contain lava-coated clasts of peridotite ranging in diameter from a few mm to >20 cm. These clasts of peridotite were probably sampled and transported by the same magma that fed the lava flow. The nature of the sediment / lava contact suggests that very little time passed between deposition of the two units. The close temporal and spatial association between the effusion of the lava flow and the slightly earlier deposition of water-rich debris flows suggests that the latter may represent lahars or sediments associated with outburst flood events originating at the volcanic vent. The substantially overthickened character of the basanite lava flow is indicative of impoundment by a barrier in the lower elevations of the valley (e.g., glacial ice).

Mantle-derived peridotite xenoliths are concentrated at the base (< 5 m) of the flow where they make up 1-5 % of the lava by volume, whereas they are exceedingly rare in the upper part of the flow. Crustal xenoliths are virtually absent throughout the lava flow; the few crustal xenoliths (< 5 cm in size) observed appear to be derived from the Llangorse batholith and may represent material entrained into the base of the lava flow.

4.4 PETROGRAPHY OF PERIDOTITES

A suite of more than 80 peridotite xenoliths was collected from the basanite lava flow and the underlying volcaniclastic sediments at Llangorse Mountain (Fig. 2.3; Table 4.1). Most xenolith samples are < 5 cm in size, but many exceed 10 cm in diameter and the

samples.	Averages	s or repo	ortea 10	r mulup	me(m)	replicat	e analys	ses; see	Appen	aix D.	
Label	19	100	104	108	110	111	113	114	115	117	118
¹ SiO ₂	43.55	43.98	43.87	43.80	44.42	44.14	43.20	42.97	43.73	44.05	43.56
TiO ₂	0.075	0.024	0.040	0.050	0.070	0.152	0.039	0.028	0.032	0.024	0.014
Al_2O_3	1.00	0.88	1.83	1.77	3.31	3.37	0.74	0.65	0.87	0.90	0.63
Fe ₂ O ₃	1.47	1.45	1.53	1.74	1.99	1.76	1.69	1.32	1.40	1.81	1.57
FeO	6.37	6.10	6.04	6.47	6.13	7.13	6.02	6.35	6.21	6.11	6.10
MnO	0.12	0.11	0.12	0.12	0.12	0.13	0.12	0.11	0.12	0.12	0.12
MgO	45.49	45.94	43.47	43.39	39.43	38.91	46.57	46.82	45.93	45.17	46.81
CaO	0.83	0.66	1.59	1.49	3.25	3.25	0.61	0.58	0.70	0.75	0.53
Na ₂ O	0.096	-0.002	0.235	0.103	0.277	0.238	0.125	0.036	0.084	0.048	0.026
K ₂ O	0.09	0.02	0.17	0.01	0.01	0.03	0.09	0.05	0.05	0.02	0.01
P ₂ O ₅	0.027	0.014	0.055	0.011	0.011	0.013	0.018	0.016	0.015	0.014	0.009
Total	99.12 [°]	99.18	98.95	98.95	99.02	99.12	99.22	98.94	99.14	99.01	99.37
								• •			
$\operatorname{Fe}_{2}O_{3}(T)$	8.55	8.23	8.24	8.93	8.80	9.68	8.38	8.38	8.30	8.60	8.35
LOI	0.37	0.03	0.29	0.12	0.13	0.26	0.09	0.16	0.07	0.30	-0.03
20	4412	1650	4042	4120	4502	2010	2001	2000	1217	4204	2070
Cr	4413	4650	4042	4129	4505	3010	2001	2009	4547	4294	3970
Ni	2471	2517	2332	2429	2169	2063	2576	2632	2545	2482	2584
Sc	6	6	13	8	16	17	8	8	6	4	8
V	32	22	32	47	70	82	23	17	20	28	18
Mo#	0.88	0.88	0.88	0.87	0.87	0.85	0.89	0.88	0.88	0.88	0.88
Fe2O3/Fe	$\begin{array}{c} 0.00\\ 0.23\end{array}$	0.24	0.25	0.27	0.32	0.25	0.28	0.21	0.23	0.30	0.26
10205/10	0.25	0. <i>L</i> F	0.25	0.27	0.52	0.20	0.20	01	0.20	0.00	0.20

Table 4.1. Whole rock chemical analyses of samples of peridotite xenolths collected from basanite lavas at Llangorse Mountain. Sample labels have prefix MH-02 except for LM samples. Averages of reported for multiple (N) replicate analyses: see Appendix D

.

¹Oxide values in wt%.

²Trace values in ppm.

14010 4.1.	Jonuna	lou						
Label	120	121 (N=4)	122	123 (N=4)	125	112	19b	95LM-1 (N=6)
SiO ₂	43.41	43.97	43.75	43.84	43.76	43.76	43.81	43.56
TiO ₂	0.025	0.035	0.018	0.021	0.047	0.016	0.020	0.013
Al ₂ O ₃	0.69	2.34	1.80	1.68	2.59	0.79	0.85	0.68
Fe ₂ O ₃	2.02	1.88	1.49	1.60	1.62	1.75	1.79	1.65
FeO	5.89	6.27	6.58	6.42	6.97	5.82	6.14	5.87
MnO	0.12	0.12	0.12	0.12	0.13	0.112	0.120	0.11
MgO	45.92	42.09	42.98	43.64	40.90	45.85	45.50	46.46
CaO	0.65	2.27	1.81	1.64	2.73	0.60	0.73	0.52
Na ₂ O	0.010	0.094	0.036	0.035	0.103	0.320	0.270	0.300
K ₂ O	0.04	0.01	0.01	0.02	0.01	0.02	0.02	0.01
P_2O_5	0.032	0.010	0.009	0.009	0.011	0.020	0.012	0.010
Total	98.81	99.08	98.60	99.01	98.87	99.07	99.26	99.18
$Fe_{2}O_{3}(T)$	8.57	8.84	8.80	8.73	9.37	8.23	8.62	8.18
LOI	0.09	0.03	0.15	0.15	0.17	<d l<="" td=""><td><d l<="" td=""><td>0.00</td></d></td></d>	<d l<="" td=""><td>0.00</td></d>	0.00
Cr	5581	4339	4141	4186	4084	4685	4448	4662
Ni	2527	2377	2396	2449	2312	2513	2509	2528
Sc	6	14	15	8	13	11	<d l<="" td=""><td>7</td></d>	7
V	26	58	50	44	58	29	32	19
Mo#	0.89	0.87	0.87	0.87	0.85	0.89	0.88	0.89
Fe2O3/FeC	0.34	0.30	0.23	0.25	0.23	0.30	0.29	0.28

Table 4.1. Continued

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largest are approximately 25 cm in diameter. All peridotites are spinel-bearing; we observed no garnet or spinel-garnet peridotite xenoliths. Approximately 60% of the xenolith samples are harzburgite and 40% are lherzolite (Fig. 4.1; Table 4.3). A single sample was collected in which the host peridotite was cut by a vein of spinel websterite (pyroxenite).

With the exception of the websterite sample, most xenoliths are coarse grained and equigranular. Coarse-grained peridotite samples usually have equilibrium textures with abundant triple point grain-grain contacts (Figs. 4.2a, 4.2b). Under the scanning-electron microscope (SEM), mineral grains show no discernible chemical zoning (Fig. 4.2b). Most peridotite samples are unfoliated to weakly foliated. Where peridotite samples are strongly foliated, the fabric is typically defined by spinel and occasionally by pyroxene. A total of 5 samples show mineralogical banding manifest as bands of bright green chromium-diopside.

Although equilibrium textures dominate (Figs. 4.2a, 4.2b), some samples show signs of disequilibrium (Figs. 4.2c-f). These textures are most common in smaller samples (\leq 50 grams) or near xenolith / lava contacts and appear to result from interactions with the highertemperature host lava. The most common expression of this thermal disturbance is the development of sieve-textured coronas on apparently pristine orthopyroxene cores (Fig. 4.2c). In three samples orthopyroxene has reacted completely to become entirely sievetextured (Fig. 4.2d); some orthopyroxene grains have developed coronas of clinopyroxene and olivine. Other disequilibrium features include the breakdown of spinel and the production of melt veinlets containing neoblasts of olivine (Figs. 4.2e, 4.2f). Similar textures have been interpreted by Shaw et al (1998) as dissolution of orthopyroxene resulting from interaction between peridotite xenoliths and entraining lava. Shaw et al's (1998) experiments showed that orthopyroxene, when in contact with silica-undersaturated



Figure 4.1: Modal mineralogy of peridotite samples from Llangorse Mountain (black circles), across the NCVP (gray circles; Shi et al, 1998) and specific localities in central and southem British Columbia (unfilled circles; Littlejohn and Greenwood, 1974). Modes for Llangorse samples are based on a combination of calculations based on bulk chemistry, Rietveld X-ray diffraction analysis, and visual estimation from hand-samples and thin sections; approximately . 60% of the samples are harzburgite and 40% are lherzolite. A single sample (MH-02-124) of lherzolite contains a ~10 cm wide vein of websterite.



Figure 4.2: Photomicrographs and SEM images of peridotite samples. (A) Microscope and (B) SEM images of coarse grained peridotite dominated by triple-point grain boundaries between olivine (Ol), orthopyroxene (Opx), and clinopyroxene (Cpx) suggesting attainment of equilibrium. (C) Photomicrograph and (D) SEM images of coarse grained spinel (Sp) peridotite showing sieve-textured coronas of orthopyroxene (Opx_s) suggesting a slight thermal disturbance; cores of orthopyroxene are generally unmodified. (E) Photomicrograph showing break-down of spinel in some samples; relict grains of spinel are surrounded by neoblasts of olivine (Ol_n). (F) Photomicrograph showing veinlet of melt cross-cutting an orthopyroxene grain showing exsolution lamellae (Opx_e). Melt contains neoblasts of olivine.

melts such as basanites, will dissolve incongruently to form olivine, clinopyroxene, and a Sirich melt.

4.5 ANALYTICAL TECHNIQUES

Mineral compositions were analysed using a fully automated CAMECA SX-50 electron microprobe (EMP) at the University of British Columbia. Operating conditions in the wavelength-dispersion mode included: excitation voltage, 15kV; beam current, 10 nA; peak count time, 20s; background count-time, 10s; spot diameter, 10 μ m. Data reduction was done using the 'PAP' $\phi(\rho Z)$ method (Pouchou and Pichoir, 1985).

Mineral chemical data were collected for all major phases in 44 samples: olivine (ol) orthopyroxene (opx), clinopyroxene (cpx), and spinel (sp) (Table 4.2). The compositions of coexisting pyroxenes were measured for geothermometric purposes. For each sample, 6 to 10 mineral pairs were analyzed, including several core and rim pairs to assess the extent of chemical zoning. Zoning was absent in all samples except in those showing strong disequilibrium textures. Temperature calculations were based on EMP analyses of adjacent orthopyroxene and clinopyroxene grains.

Olivine and pyroxene grains have compositions consistent with mantle peridotite; magnesium numbers (Mg# = Mg / (Mg + Fe)) for olivine, orthopyroxene and clinopyroxene vary between 0.887-0.917, 0.887-0.924, and 0.894-0.944, respectively. Spinel compositions are the most varied; the average chromium number (Cr# = Cr / (Cr + Al)) for spinel in peridotite is 0.352, but the range is 0.0181 (websterite) to 0.727 (lherzolite).

4.6 GEOTHERMOMETRY

Two-pyroxene geothermometry, based on diopside and enstatite solubility in coexisting ortho- and clinopyroxene, provides a reliable means of estimating equilibrium temperatures in rapidly transported mantle rocks (e.g., Wood and Banno, 1973; Wells, 1977). The Brey and Kohler (1990) geothermometer [BK90] is a revised calibration of previous estimations of the two pyroxene thermometer (Wood and Banno, 1974; Wells, 1977) that accounts for more realistic pyroxene compositions. Temperatures calculated for the same mineral pairs using the Wells (1977) thermometer are consistently lower than BK90 temperatures below approximately 900°C and higher than BK90 temperatures above approximately 980°C (Table 4.2). We elected to use the BK90 thermometer because it is more recent and has been calibrated over a wider range of temperatures and mineral compositions.

The BK90 thermometer requires an assumption of pressure; our calculations assume a pressure of 12 kbar for all samples. Based on the temperature range of our sample suite, a pressure difference of about 5 kbar is estimated between the highest and lowest temperature xenoliths. When the lowest and highest temperature xenoliths are re-calculated using a 'corrected' pressure the corresponding temperature difference is $\leq 15-20^{\circ}$ C; the inherent uncertainty (1p) associated with the BK90 thermometer is $\pm 15^{\circ}$ C.

We collected EMP data on peridotite samples in batches. Our first batch contained 10 samples in which mineral compositions were measured and processed; subsequent samples were analyzed in batches of 5. The geothermometry results were plotted sequentially in groups as acquired (Fig. 4.3). The goal of plotting the sequential distribution

Table 4.2:	Represe	ntative (chemica	l compo	sitions (wt%) of	f orthop)	yroxene	(OPX) a	and cline	pyroxe	ne (CPX) miner	al pairs t	from eac	ch samp	le (MH-	-02-).		
Calculated	tempera	itures are	s reporte	ed for ea	ich mine	sral pair	using be	oth the E	trey and	Kohler	(1990)	and Wel	ls (1977) thermo	ometers.					
OPX	110	108	111	10	<u>9</u> c	6	10b	109b	18b	104	100	115	122	117	5b	8b	4A	þ6	10c	10d
SiO ₂	54.42	55.64	55.08	54.90	56.94	56.55	55.36	54.85	56.51	55.33	56.92	56.56	55.15	55.50	56.28	56.01	54.78	56.22	56.03	54.47
TiO_2	0.08	0.09	0.14	0.12	0.02	0.00	0.01	0.00	0.05	0.01	0.02	0.00	0.07	0.04	0.00	0.00	0.01	0.07	0.00	0.07
Al_2O_3	3.99	3.43	4.27	4.61	2.03	1.70	3.27	2.49	1.50	3.52	2.12	1.90	3.81	2.36	1.83	2.29	2.89	0.80	2.49	4.67
Cr_2O_3	0.34	0.38	0.27	0.44	0.58	0.46	0.39	0.64	0.74	0.46	0.55	0.44	0.38	0.59	0.47	0.43	0.50	0.44	0.47	0.43
FeO	6.26	5.89	6.99	6.19	5.28	5.12	5.67	5.49	5.28	5.58	5.28	5.23	6.64	5.53	5.27	5.66	5.73	5.38	5.14	5.93
0uM	0.19	0.17	0.14	0.16	0.13	0.10	0.14	0.10	0.13	0.16	0.08	0.15	0.18	0.13	0.12	0.13	0.12	0.17	0.13	0.12
MgO	33.41	33.94	32.98	32.85	34.89	34.62	34.07	34.06	34.29	34.01	35.22	35.06	33.62	34.82	35.46	34.93	34.61	34.87	34.80	33.71
NiO	0.06	0.10	0.08	0.05	0.10	0.08	0.01	0.07	0.10	0.05	0.07	0.09	0.13	0.07	0.07	0.07	0.08	0.08	0.10	0.05
CaO	0.50	0.49	0.49	0.98	0.67	0.69	0.55	0.67	1.31	0.51	0.71	0.73	0.51	0.76	0.66	0.53	0.51	1.62	0.72	0.51
Na_2O	0.03	0.06	0.07	0.10	0.05	0.01	0.03	0.04	0.05	0.02	0.00	0.01	0.04	0.04	0.03	0.00	0.00	0.07	0.05	0.04
Total	99.28	100.19	100.52	100.40	100.70	99.33	99.51	98.41	96.66	99.66	1 70.00.	00.18	00.51	99.83 1	00.20 1	00.06	99.22	12.66	99.94 1	00.01
CPX	110	108	111	10	9c	96	10b	109b	18b	104	100	115	122	117	5b	8b	4A	P6	10c	10d
SiO_2	51.85	52.06	51.55	51.95	53.55	53.73	52.57	53.18	53.86	52.09	53.29	53.22	51.38	52.91	53.82	53.25	51.94	51.57	52.37	52.09
TiO_2	0.27	0.31	0.79	0.45	0.12	0.02	0.03	0.03	0.00	0.06	0.03	0.00	0.23	0.01	0.00	0.04	0.02	0.11	0.00	0.16
Al_2O_3	6.20	5.58	6.63	6.67	3.08	1.47	3.52	2.44	2.46	5.06	1.94	1.93	4.87	2.53	1.73	2.45	3.17	4.76	2.58	5.52
Cr_2O_3	0.79	0.15	0.67	0.91	1.47	0.67	0.86	0.81	0.83	1.15	0.69	0.69	0.69	0.81	0.70	0.82	0.97	0.72	0.75	0.76
FeO	2.47	2.45	2.88	2.58	1.96	1.86	1.95	1.94	2.21	2.17	1.97	1.88	2.47	2.27	1.91	2.14	2.33	2.20	2.28	2.28
MnO	0.05	0.08	0.09	0.10	0.08	0.07	0.11	0.05	0.06	0.10	0.09	0.08	0.08	0.15	0.08	0.09	0.09	0.10	0.06	0.06
MgO	15.16	15.23	14.99	14.72	16.72	18.20	16.59	17.36	17.58	15.98	18.35	18.16	16.08	17.73	18.37	17.64	17.12	16.29	17.74	16.12
NiO	0.03	0.00	0.04	0.06	0.02	0.00	0.08	0.05	0.00	0.07	0.08	0.08	0.03	0.04	0.05	0.04	0.00	0.02	0.05	0.03
CaO	20.78	20.78	20.94	20.73	21.19	23.30	22.81	23.21	21.24	22.07	23.18	23.15	21.96	21.78	23.21	23.19	22.89	22.04	21.92	21.92
Na_2O	1.56	1.54	1.40	1.81	1.23	0.15	0.78	0.34	1.02	1.03	0.15	0.15	0.92	0.87	0.19	0.31	0.53	0.85	0.75	0.94
Total	99.17	98.18	99.97	96.66	99.42	99.49	99.30	99.40	99.27	99.78	77.66	99.35	98.71	99.11	00.08	96.66	99.05	98.66	98.49	99.87
T _{wells} (°C)	913	894	926	889	953	953	863	915	991	902	696	962	200	956	963	931	896	606	962	938
Т _{В&К} (°C)	926	903	945	006	955	606	809	870	1001	884	936	928	883	956	938	868	859	917	946	931
I∆TI	13	6	18	11	7	4	55	45	10	18	33	34	24	1	25	33	37	6	16	9

	1 19b	5 56.29	0.00	3 2.30	0.51	ə 5.37	3 0.16	34.86	1 0.13	1 0.65	3 0.04	3 100.31	19b	7 53.32	0.00	3 2.12	0.69	5 2.08	7 0.02	t 17.91	1 0.02	7 22.88	2 0.27	3 99.31	1 967	2 03K	200
	95LM1	56.35	0.0	1.93	0.50	5.25	30.0	35.08	0.11	0.61	0.03	36.66	95LM1	53.77	0.0	1.68	0.61	1.96	0.07	18.24	0.01	23.47	0.12	<u> 99.95</u>	947	500	Š
	125	56.00	0.03	2.37	0.50	5.38	0.08	34.21	0.13	0.72	0.02	99.45	125	52.69	0.02	2.30	0.78	2.03	0.07	17.99	0.08	22.82	0.17	98.95	982	070	
	123	55.35	0.02	3.68	0.46	5.94	0.18	33.63	0.07	0.55	0.02	06.66	123	52.10	0.07	4.17	1.00	2.39	0.03	16.55	0.06	22.96	0.56	99.89	906	858	0000
	121	54.13	0.02	3.83	0.36	5.77	0.14	33.59	0.13	0.51	0.03	98.50	121	51.41	0.16	4.84	0.86	2.29	0.09	15.92	0.04	22.22	0.00	98.73	881	857	100
	120	56.25	0.01	2.00	0.46	5.26	0.17	34.72	0.08	0.65	0.02	99.62	120	52.92	0.02	1.97	0.73	1.92	0.07	17.94	0.04	23.03	0.21	98.86	952	012	111
	119	55.90	0.03	2.41	0.53	5.47	0.08	35.22	0.12	0.62	0.00	100.38	119	53.03	0.02	2.18	0.79	2.01	0.07	18.27	0.06	23.26	0.17	98.66	958	770	11
	118	56.45	0.00	1.58	0.45	5.33	0.13	35.30	0.06	0.67	0.00	76.66	118	52.93	0.00	1.57	0.68	2.03	0.12	18.59	0.10	23.25	0.07	99.33	965	033	000
	117b	56.46	0.04	2.36	0.55	5.45	0.13	35.36	0.11	0.68	0.04	101.17	117b	53.29	0.04	2.45	0.85	2.24	0.09	17.68	0.02	21.97	0.77	99.40	963	050	200
	116	56.73	0.01	1.90	0.51	5.07	0.10	35.56	0.06	0.66	0.00	100.61	116	53.61	0.00	1.77	0.70	1.92	0.06	18.53	0.11	22.92	0.17	99.78	992	071	112
	115b	55.97	0.00	2.15	0.52	5.37	0.09	34.88	0.07	0.76	0.02	99.83	115b	53.47	0.00	1.94	0.70	1.98	0.06	18.26	0.08	23.17	0.16	99.82	968	027	102
	114	57.33	0.13	0.63	0.51	5.16	0.11	34.62	0.07	1.69	0.08	100.33	114	53.35	0.00	1.89	0.85	2.08	0.05	18.22	0.04	23.23	0.10	99.80	779	020	222
	113	56.63	0.02	1.87	0.51	5.12	0.13	35.11	0.06	0.67	0.01	100.13	113	53.64	0.01	1.85	0.76	2.02	0.06	18.09	0.02	23.43	0.10	66.66	954	010	710
	102	55.42	0.06	3.52	0.40	6.05	0.10	34.20	0.09	0.53	0.04	00.42	102	51.74	0.25	5.30	0.95	2.26	0.05	15.87	0.02	21.89	1.05	99.38	906	600	100
	19	55.37	0.15	2.46	0.61	5.35	0.13	34.99	0.07	0.72	0.04	99.89	19	51.43	0.25	2.92	1.10	2.17	0.05	17.46	0.04	22.06	0.56	98.04	962	018	
	20	53.95	0.10	4.08	0.56	6.43	0.11	33.94	0.00	0.57	0.05	99.80	20	51.20	0.34	5.79	1.15	2.46	0.07	15.49	0.00	21.14	1.33	98.96	911	100	111
	124	53.19	0.09	6.79	0.20	7.82	0.17	31.73	0.07	0.60	0.09	00.75	124	51.22	0.59	8.91	0.26	2.87	0.06	14.26	0.02	20.16	1.80	00.16	934	070	
	112	56.58	0.00	2.00	0.47	5.30	0.15	34.88	0.06	0.62	0.08	00.16 1	112	53.58	0.00	3.19	1.75	1.91	0.09	16.98	0.05	20.97	1.22	99.73	983	005	270
ned.	109	55.68	0.02	3.05	0.72	5.50	0.10	34.39	0.10	0.90 J	0.10	00.56 1	109	52.49	0.06	3.70	1.31	2.45	0.11	17.24	0.07	20.36	1.07	98:85	1034	1065	2001
: Contin	12	56.83	0.05	0.83	0.48	5.84	0.17	34.22	0.11	1.72	0.07	00.31	12	52.02	0.23	5.21	0.98	2.29	0.08	15.98	0.05	21.58	1.05	99.47	934	050	101
able 4.2	15	54.65	0.03	3.97	0.45	6.00	0.16	33.80	0.08	0.55	0.02	99.69	15	52.45	0.14	4.44	0.85	2.15	0.07	16.47	0.13	22.20	0.82	99.73	930	017	212



Figure 4.3: Results of two-pyroxene geothermometry for 44 peridotite samples (Brey and Kohler, 1990). Data is plotted in the sequence that it was collected. (A) Mean temperatures (°C) from all samples. Dotted tie lines connect temperatures associated with core (filled squares) and rim (open squares) compositions of minerals. (B) Distribution of ambient mantle temperatures illustrating the running maximum and minimum values recovered from the entire suite. Filled circles represent ambient mantle samples, open squares represent high temperature samples having textures indicating they are thermally disturbed (Figs. 4.2b, 4.2c). The running maximum and minimum temperatures are used to determine when T_{max} - T_{min} no longer changes (see text).

of temperatures was to ascertain when enough samples had been analysed to ensure that the full range or temperatures had been captured. We tracked this by plotting the running values of T_{max} , T_{min} and T_{max} - T_{min} against the sequential sample number (Fig. 4.3). Plotted in this way we sought the situation where the value of T_{max} - T_{min} no longer changed as new samples were processed. As shown in Fig. 4.3b, a constant value was reached between samples 25 and 30; additional data collection did not change the values of T_{max} , T_{min} or T_{max} - T_{min} . This analysis suggests that suites with fewer than 25-30 samples may not provide statistically valid limits on the temperature distribution in the lithospheric mantle. Conversely, we would argue that our suite of samples records the maximum and minimum temperatures of the mantle lithosphere sampled by the basanite magma.

4.7 RESULTS

Geothermometry results for all 44 xenoliths sampled are reported in Table 4.3 and shown graphically in Figs. 4.3a and 4.3b. The minimum temperature calculated from our peridotite suite (assuming a pressure of 12 kbar) is 810°C, with a sample variance of 26°C. The maximum ambient mantle temperature is more difficult to constrain because some xenoliths have been thermally disturbed during transport. Samples with sieve-textured orthopyroxene grains generally give higher equilibration temperatures and greater variances; in some samples, sieve-textured rims provide temperatures up to 100°C higher than apparently pristine cores (Fig. 4.3a). For these samples, the high-temperature rim analyses were not included as ambient mantle temperatures. Three samples with strong development of sieve-textured orthopyroxene record temperatures of 1100-1150°C (Fig. 4.3b), which are

		Density	_	¹ Mode	s (%):		² Rock	X _{Fo} :	X _F	ie i	Cr#:		
No.	Wt. (g)	(g/cm^3)	Ol	Орх	Срх	Sp	Type:	Ol	Орх	Срх	Sp	T_A	T _H
110	369	3.24	59	23	15	3	Lh	0.899	0.088	0.071	0.11	934	
108	2079	3.31	71	21	7	1	Lh	0.905	0.084	0.065	0.21	902	
111	471	3.33	61	21	15	3	Lh	0.887	0.103	0.089	0.10	936	
10	196	3.17	60	20	18	2	Lh	0.901	0.091	0.079	0.14	907	
9c	156	3.27	58	30	10	2	Lh	0.916	0.075	0.059	0.52	961	1015
9b	50	3.26	65	30	3	2	Hz	0.915	0.075	0.050	0.56	910	
10b	40	3.24	58	30	10	2	Lh	0.913	0.078	0.043	0.25	810	
109b	21	3.23	60	30	8	2	Lh	0.912	0.075	0.051	0.49	872	
18b	53	3.24	60	30	8	2	Lh	0.915	0.074	0.057	0.59	998	1095
104	411	3.33	68	23	6	3	Lh	0.913	0.079	0.054	0.21	879	1087
100	1461	3.29	74	23	2	1	Hz	0.915	0.071	0.046	0.48	936	
115	48		<u>7</u> 0	25	3	2	Hz	0.916	0.071	0.043	0.47	935	
122	593	3.30	63	24	10	3	Lh	0.895	0.090	0.065	0.13	889	
117	54	3.18	70	20	8	2	Lh	0.912	0.072	0.045	0.47	955	
5b	81		77	17	4	2	Hz	0.916	0.067	0.046	0.52	926	
8b	161	· ·	73	22	`3	2	Hz	0.906	0.075	0.052	0.42	897	i.
4A	48	3.21	65	24	8	3	Lh	0.910	0.074	0.043	0.32	851	
9d	16		65	23	10	2	Lh	0.907	0.073	0.052	0.16	914	1097
10c	22	3.23	62	33	3	2	Hz	0.914	0.069	0.044	0.45	952	
10d	22	3.13	71	20	7	2	Lh	0.907	0.080	0.054	0.12	931	1102
15	63	3.25	60	30	8	2	Lh	0.907	0.082	0.058	0.18	907	
21	40		65	23	10	2	Lh	0.907	0.080	0.064	0.17	951	1128
109	237		70	21	7	2	Lh	0.915	0.069	0.057	0.42	1064	
112	245		75	22	2	1	Hz	0.912	0.072	0.050	0.51	995	1100
124	1036		0	40	45	15	We		0.108	0.092	0.02	984	
20	329	3.25	65	18	15	2	Lh	0.903	0.080	0.061	0.21	923	
19	578	3.24	71	25	3	1	Hz	0.915	0.065	0.037	0.46	933	
102	184		60	24	15	1	Lh	0.903	0.084	0.059	0.17	898	
113	366		78	20	1	1	Hz	0.915	0.074	0.046	0.53	909	
114	374	3.23	77	20	2	1	Hz	0.917	0.071	0.049	0.52	950	
115b	386	3.23	74	23	2	1	Hz	0.913	0.072	0.048	0.46	937	
116	256		66	25	8	1	Lh	0.917	0.065	0.047	0.52	970	
117b	340	3.21	71	26	2	1	Hz	0.913	0.071	0.049	0.47	966	
118	215		76	21	2	1	Hz	0.914	0.068	0.040	0.80	939	
119	297	3.19	62	34	2	2	Hz	0.914	0.069	0.044	0.41	930	
120	621	3.33	76	21	2	1	Hz	0.915	0.074	0.049	0.50	918	
120	2830	3.34	65	23	10	2	Lh	0.905	0.081	0.053	0.17	865	
123	981	3.25	68	22	8	2	Lh	0.903	0.085	0.057	0.20	867	
125	298	0.120	79	18	2	- 1	Hz	0.911	0.077	0.051	0.44	948	
951 M1	3385	3 31	76	20	2	2	Hz	0.917	0.071	0.045	0.51	907	
195	30/	3 24	73	20 24	2	1	H7	0.911	0.074	0.053	0.46	937	
126	204	J. L T	60	27	2	3	112 Th	0.000	0.078	0.055	0.48	101	1142
120	127	3 20	60	29 20	Q Q	2	Th	0.202	0.078	0.004	0.40		1143
12(2)	127	3.20	70	47 76	2	1	니 니~	0.000	0.078	0.000	0.17		1178
12(2)	15/	5.20	/0	20	5	1	F12	0.204	0.004	0.070	0.10		1120

Table 4.3: Mineralogical, chemical, and thermal traits of all samples (MH-02-), including: weight, mineral modes and chemistry, and equilibration temperature (T_A : ambient samples; T_H : high-T rim; ΔT : T_{max} - T_{min})

¹Olivine (OI); orthopyroxene (Opx); clinopyroxene (Cpx); spinel (Sp);

²Lherzolite (Lh); harzburgite (Hz); websterite (We)

interpreted as re-equilibration temperatures approaching the lava temperature. The maximum ambient temperature of this sample suite is 1065°C with a sample variance of 15° C. In summary, these temperatures provide the minimum (810°C), maximum (1065°C), and total temperature difference (> 255°C) recorded by samples of peridotite mantle from beneath Llangorse Mountain. The temperature difference ($T_{max}-T_{min}$) relates directly to the minimum thickness of the mantle lithosphere; the exact value (e.g., thickness) depends on the geothermal gradient assumed.

4.8 MODEL OF THE LITHOSPHERIC MANTLE

Our ultimate goal is to constrain the thermal state of the lithosphere underlying the northern Canadian cordillera by combining petrological information gathered from mantlederived peridotite xenoliths with simple models for conductive heat flow in the lithosphere. The basis of our model lithosphere for the northern Cordillera is summarized in Fig. 4.4 and comprises a crust of specified thickness (Z_M) and a constant surface temperature (T_0) and surface heat flow (q_0). Seismic reflection data, collected as part of the Lithoprobe program (Clowes et al, 1999), is used to define the depth to the MOHO in the northern Cordillera (transect line 22; Hammer et al, 2000; Hammer and Clowes, 2004). Beneath the Atlin region, the MOHO depth is estimated to be 36 km (Z_M ; Hammer and Clowes, 2004). The crustal layer has a specific range in heat production (A_0) values and a constant thermal conductivity (K_1) and overlies a mantle lithosphere having a uniform thermal conductivity (K_2) and no heat production. Heat transfer in the mantle lithosphere is purely conductive. The depth to the base of the mantle lithosphere (Z_A) is unknown but is taken as marking the



Figure 4.4: Schematic representation of heat transfer in the crust, mantle lithosphere and asthenosphere (see Figs. 4.5-4.9). Variables include: temperature and heat flow at the Earth's surface (T_0 , q_0), at the MOHO (T_{moho} , q_{moho}), and at the top of the asthenosphere (T_a , q_a). Additional variables include thermal conductivity and heat production of the crust (K_1 , A_0) and the mantle lithosphere (K_2) and the depth to the MOHO (Z_M) and the top of the asthenosphere (Z_A). The MOHO is set at 36 km (Hammer and Clowes, 2004).

top of the convecting asthenospheric mantle. Our aim is to investigate the temperatures and heat fluxes at the two major boundaries in this schematic model (e.g., T_{moho} , q_{moho} , T_a and q_a).

The research efforts of Hyndman and Lewis (1999), Lewis et al (2003), and Hyndman et al (in press) have provided bounding estimates on the values of surface heat flow (q₀) and crustal heat production (A₀) for the northern Cordillera (Table 4.4). Much of their work has been directed at explaining heat flow variations between the Cordillera and the North American craton to the east. We have used their results to set some of the geophysical properties for the crust (e.g., Fig. 4.4) by adopting values that are most relevant to the northern Canadian Cordillera at latitudes consistent with Atlin, British Columbia (Table 4.4). Readers are directed to Hyndman and Lewis (1999) and Lewis et al (2003) for detailed discussions of the acquisition, distribution and quality of the heat flow data.

4.8.1 Surface Heat Flow

Relative to the Canadian craton, the Cordillera is characterized by high surface heat flow (Hyndman and Lewis, 1999). Recent work by Lewis et al (2003) has provided a series of new measurements showing that surface heat flow across the entire northern Cordillera is approximately twice as high as observed in the craton (80-100 mW m⁻² versus 40-60 mW m⁻², respectively). In addition, there are differences in q₀ between domains north and south of 59°N latitude. The mean values of q₀ derived from between 59°N and 64°N is 105 ±22 mW m⁻², whereas between 57°N and 59°N the surface heat flow is 73 ±11 mW m⁻². The Llangorse volcanic field is situated between 59°19'N and 59°25'N. Thus, we have used the

latter range of values but have extended the calculations over a slightly larger range (60-100 mW m⁻²) to account for the LVF being on the boundary between these two surface heat flow domains (Table 4.4).

4.8.2 Heat Production

Lewis et al (2003) provide new measurements of radiogenic heat production in crustal rocks across the northern Cordillera. Crustal rocks, especially younger intrusions situated in the northernmost portion of the Cordillera (59 – 64°N), have substantially higher values of heat production than similar rocks exposed between $57 - 59^{\circ}$ N. The geological units to the north of 59°N have a mean heat production of 4.6 ±2.4 µW m⁻³ versus 1.6 ±0.8 µW m⁻³ for rocks south of 59°N. For the purposes of our modeling we use heat production values of 1.6 ±0.8 µW m⁻³ and we compare our model results against the range of surface heat flow values expected for the same latitudes (e.g., 73 ± 11 mW m⁻²). Lewis et al (2003) used these data to argue that the high heat flow observed for the Cordillera is largely due to the unusually high values of heat production found in near surface crustal rocks of the northern Cordillera, and not to higher temperatures at the base of the crust.

4.8.3 Lower Crust Temperatures

For the northern Canadian Cordillera between 59 and 64°N, a reduced heat flow of 59 mW m⁻² is predicted, which is similar to the southern Canadian Cordillera (49 - 53°N) (Hyndman and Lewis, 1999; Lewis et al, 2003; Hyndman et al, in press). Depending on the model parameters adopted, Lewis et al (2003) suggest a MOHO temperature (T_{moho}) of 950

Property	Layer 1	Layer 2	Layer 3
	Crustal Lithosphere	Mantle Lithosphere	Asthenosphere
T ₀ [°C]	10	-	
q ₀ [mW m ⁻²]	73 ± 11	-	-
Z _{moho} [km]	36	-	-
T [°C] xenolith ^a	800 - 850		
T [°C] xenolith ^b		1050 - 1100	
A [W m ⁻³]	1.6 ± 0.8	0	0
K [W m ⁻¹ K ⁻¹]	2.5	3.2	3.2
[kg m ⁻³]	2700	3200	3300
[K ⁻¹]	-	-	3.0E-5
$[m^{-2} s^{-1}]$	-	-	1.00E-006

Table 4.4. Physical parameters used in modelling of geotherms for Cordilleran lithosphere.

^a Range of minimum temperatures recovered from mantle peridotite samples ^b Range of maximum temperatures recovered from mantle peridotite samples

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 ± 100 °C. Lewis et al (2003) also used the values of compressional seismic wave velocities at the MOHO (e.g., Pn) to estimate the temperatures at the MOHO for the northern Cordillera using the strong negative correlation between Pn seismic wave velocities and MOHO temperature developed by Black and Braile (1982). Adopting the Pn wave velocities for the northern Cordillera reported by Hammer et al (2000; e.g., 7.8 – 7.9 km s⁻¹), they estimate a T_{moho} for the northern Cordillera of 800-1000 °C.

Our suite of mantle xenoliths provide a means to constrain the value of T_{moho} using geothermometry results. Our rationale is that the coolest temperature recovered from the suite represents the shallowest mantle sampled by the magma and that the shallowest mantle should be coincident with the MOHO (Table 4.5). There is always the possibility that another sample exists with a lower temperature that was not sampled for EMP analysis, or not sampled by us in the field, or not sampled by the basanite magma. However, given our extensive thermometric database for this single locality, we suggest that our minimum temperatures (800 - 850 °C) are fair estimates of the temperature at the top of the mantle lithosphere and base of the crust (e.g., T_{moho}).

4.8.4 Crustal and Mantle Lithosphere

The steady-state conductive temperature distribution for the one-dimensional crustal lithosphere discussed above is given by:

$$T(Z) = T_0 + \frac{q_0 Z}{K_1} - \frac{A_0 Z^2}{2 K_1} \qquad 0 < Z < Z_{moho} \qquad (4.1)$$

where K_1 is a constant and q_0 and A_0 have a preferred range of values based on Lewis et al (2003). For each pair of values of q_0 and A_0 there will be a unique crustal geotherm that

T _{moho} [°C]		800		850
A ₀	0.8	1.67	0.8	1.48
q ₀ [mW m ⁻²]	69.3	85	72.7	85
$q_{moho} [mW m^{-2}]$	40.5	24.9	43.9	31.5
$(dT_1 / dZ) [^{\circ}C \ km^{-1}]^a$	16.18	9.95	17.6	12.67
q _A [mW m ⁻²]	40.5	24.9	43.9	31.5
Z _A [km] at 1300 °C	75.5	100.3	68.8	81.4
	T _A =	:1100 °C	$T_A = 1$	1100 °C
Z _A [km]	59.7	74.6	54.2	61.2
Z _{ML} [km] ^b	23.7	38.6	18.2	25.2
;			:	- *** .
		Convectio	on Conditions	
T [°C]°	266	197	280	230
$\log (T_{ave})^d$	19.45	19.55	19.43	19.52
T ave [°C] ^e	1233	1198	1240	1215
T [°C] ^f	1366	1297	1380	1330
D [km] ^g	47	57	46	53

1 able 4.5. Summary of model resul

^a Thermal gradient at the base of the crustal lithophere.

^b Thickness of mantle lithosphere.

^c Minimum critical difference in temperature for convection.

^d Implied viscosity of convecting mantle given average temperature of layer.

^e Average temperature in convecting boundary layer.

^fTemperature at top of asthenosphere.

^g Minimum critical thickness required for convection.

predicts unique, but correlated, values of T_{moho} and reduced heat flow (qr). Fig. 4.5a illustrates the relationship between surface heat flow and crustal heat production embedded in Eq. 4.1. The lightly shaded box (Fig. 4.5a) denotes the range of values of q_0 and A_0 observed for the northern Cordillera near Atlin, B.C. The dashed lines are contours of reduced heat flow (qr = qmoho) for all possible crustal geotherms; the contours show the specific values of reduced heat flow dictated by the range of observed surface heat flow and heat production values. Each geotherm also dictates a unique value of T_{moho} . However, our geothermometry on mantle peridotites constrains the value of T_{moho} to between 800 and 850 °C. By fixing the value of T_{moho} , the range of possible values of q_0 and A_0 are severely restricted (solid lines, Fig 4.5a) and described by:

$$q_0 = 0.5 Z_{moho} A_0 + \frac{[T_{moho} - T_0] K_1}{Z_{moho}}$$
 (4.2)

Combining the T_{moho} constraints with the restricted range of values of q_0 and A_0 (Fig. 4.5a; Lewis et al 2003) greatly limits the range of model crustal geotherms (Fig. 4.5b). The complete solution space is denoted by the dark shaded quadrilateral in Fig. 4.5a which also defines the range of possible values of q_{moho} . For a fixed T_{moho} of 800 or 850 °C the values of reduced heat flow vary from 25 to 40 and 32 to 44 mW m⁻², respectively (Fig. 4.5c; Table 4.5). In general, higher values of T_{moho} dictate higher values of reduced heat flow; this has substantial implications for the temperature distributions in the mantle lithosphere and ultimately the depths to the thermally controlled base of the lithosphere (e.g., Hyndman et al, in press).

The lithospheric mantle (e.g., Layer 2; Fig. 4.4) is coupled to the crust by the MOHO temperature and the reduced heat flow (Figs. 4.5b, 4.5c). The expected steady-state



Figure 4.5: Model results for the temperature distribution in the crust. (A) The range of values for surface heat flow (q_0 ; mW m⁻²) and crustal heat production (A_0 ; μ W m⁻³) that are consistent with T_{moho} values of 800 or 850°C (solid lines). The corresponding model values of reduced heat flow $q_{r=moho}$ are contoured as dashed lines. Light grey shaded box is the range of values of q_0 and A_0 reported for the northern Cordillera (Hyndman and Lewis, 1999; Lewis et al, 2003). (B) The family of feasible geotherms generated for the crustal lithosphere assuming values of T_{moho} of 800 or 850°C. (C) The feasible solution space shown in Fig. 8a (dark grey shaded box) is expanded and labeled in terms of corresponding values of reduced heat flow. The solution space is defined by the intersection of the two solid lines (model) and the light grey shaded box (observed) in Fig. 4.5a.

temperature distribution is described by:

$$T(Z) = T_{moho} + \frac{[q_0 - A_0 Z_{moho}]}{K_2} [Z - Z_{moho}] \qquad Z_M < Z < Z_A \qquad (4.3)$$

The total thickness of the mantle lithosphere is not known *a priori*, but we can rearrange Eq. 4.3 to explore the two unknowns: the depth to the asthenosphere (Z_A) and the temperature at that transition (T_A):

$$T_{A} = \frac{q_{moho}}{K_{2}} Z_{A} + \langle T_{moho} - \frac{q_{moho}}{K_{2}} Z_{moho} \rangle \qquad (4.4)$$

The distribution of model geotherms for the mantle lithosphere is shown in Fig. 4.6a for two different values of T_{moho} . The higher values of reduced heat flow associated with the 850°C MOHO support higher geotherms and a potentially thinner mantle lithosphere. For example, if the 1300°C isotherm was arbitrarily adopted as the thermally controlled base of the lithosphere, then the model geotherms using $T_{moho} = 800$ °C would indicate depths to the asthenosphere of 75-100 km versus 68-81 km for $T_{moho} = 850$ °C. Conversely, a cooler MOHO implies a deeper boundary to the asthenosphere (Fig. 4.6b).

We propose an alternative constraint on the thickness of mantle lithosphere underlying this portion of the northern Cordillera based on the maximum temperatures recovered from our suite of mantle peridotite xenoliths (T_A ; Table 4.3). The maximum temperature clearly associated with ambient mantle lithosphere is close to 1100°C and weakly disturbed samples record temperatures as high as 1150°C (Table 4.3). These temperatures are over 200°C lower than the estimated liquidus temperature for the host basanite (bl; Fig. 4.6a). The maximum xenolith temperatures are indicative of the hottest and deepest portions of the mantle lithosphere and provide estimates of the minimum


Figure 4.6: Model geotherms for mantle lithosphere beneath the northern Canadian Cordillera. (A) The predicted temperature distributions for the mantle lithosphere are plotted against depth (km); model geotherms are for T_{moho} values of 800°C (light shading) and 850°C (dark shading). Vertical dashed lines indicate the range of maximum temperatures (1050-1100°C) recovered from peridotite xenoliths. (B) The same array of geotherms is represented as values of reduced heat flow (q_{moho}) and depth to the asthenosphere (Z_A) assuming T_A of 1100°C and fixed values of T_{moho} (800 and 850°C; see Table 7).

thickness of mantle lithosphere (Fig. 4.6a). Clearly, the higher the temperatures of ambient mantle recovered from the xenoliths, the thicker the mantle lithosphere.

The range of implied depths to the base of the mantle lithosphere (Z_A) are explicitly mapped out in terms of T_{moho} and the maximum xenolith temperature (T_A) in Fig. 4.6b. The thickest mantle lithosphere (60-75 km) is supported by the coolest T_{moho} and higher values of T_A ; the thinnest lithosphere (54-61 km) is associated with high values of T_{moho} and lower values of T_A (Table 4.5).

4.9 DISCUSSION

4.9.1 Crustal Geotherms

We have used petrological information recovered from mantle-derived peridotite xenoliths in combination with steady-state conductive heat transfer arguments to create a model set of geotherms for the lithosphere underlying the northern Cordillera. These geotherms define the thermal state of the lithosphere and there are attendant petrological implications. Firstly, the high MOHO temperatures implied by the coolest mantle peridotite temperatures demand elevated crustal geotherms. Conventional crustal geotherms usually indicate lower MOHO temperatures (e.g., 400-500°C) and have paths that lie well within the kyanite stability field (Fig. 4.6a). The model geotherms for the northern Cordillera describe paths that, at depth, become tangential to the kyanite-sillimanite univariant curve. Kyanite remains the most likely stable aluminosilicate in pelitic lower crustal rocks; however, small thermal perturbations could easily allow for development of sillimanite. If the MOHO were any hotter (e.g., ~900°C), then sillimanite would be the stable phase at the base of the crust

and there would be a kyanite - sillimanite isograd within the lower portion of the crustal column.

The elevated crustal geotherm requires consideration of partial melting of crustal rocks. For the purposes of discussion we have plotted the water-saturated and dry solidi for mafic crustal rocks (gabbro; Lambert and Wyllie, 1972) against the model geotherms (Fig. 4.7). Melting of gabbro would only proceed under conditions of full water saturation and such conditions are unlikely to be met in the deep crust. Granitoids with lower melting points could perhaps be induced to melt, but these rocks are more likely to be concentrated in the upper to middle crust where temperatures are substantially lower. If melting is occurring at the base of the crust, it is most likely under conditions of water-undersaturation and derived from the dehydration-breakdown of muscovite, biotite and hornblende in metamorphic and granitic rocks (e.g., Whitney 1988; Patino-Douce et al, 1990).

4.9.2 Lithosphere – Asthenosphere Transition

The maximum xenolith temperatures (≤ 1100 °C) are used as estimates of the temperature at the base of the mantle lithosphere; the implication is that the hotter asthenospheric mantle underlying this boundary is capable of sustained convection. The highest temperature xenoliths suggest that the base of the mantle lithosphere is close to 1100°C. Cooler MOHO temperatures (800°C) suggest values of q_A between 40.5 and 24.9 (Table 4.5), whereas a hotter MOHO (e.g., 850°C) requires higher mantle heat flow values (43.9 – 31.5; Table 4.5).

We use the heat flow calculated in the mantle lithosphere to investigate whether the



Figure 4.7: Model geotherms for the combined crustal and mantle lithosphere. Model geotherm temperatures are plotted against depth (km) and compared to stability fields of common metamorphic minerals (K: kyanite; S: sillimanite; A: andalusite; Holdaway, 1971), anhydrous (ag) and hydrous (hg) solidi for gabbro (Lambert and Wyllie, 1972), and 1 atm liquidus for host basanite (bl). Conductive mantle lithosphere geotherms terminate against maximum temperatures from mantle xenoliths ($1050 - 1100^{\circ}$ C; solid dots) defining a quadrilateral in T-Z space containing the minimum depth (and temperature) to the convective mantle asthenosphere.

minimum conditions for convection are met in the underlying asthenosphere. The problem is summarized as a layer of asthenosphere of unknown thickness (D) being heated from below and ultimately conducting all heat (q_A) into the base of the mantle lithosphere (Fig. 4.8; after Turcotte and Schubert, 2002, and Poirier, 2000). The heat flow (q_A) into the base of the lithosphere is equivalent in magnitude to the MOHO heat flow (q_{moho}) because of the lack of heat sources and sinks in the mantle lithosphere. Our question concerns the rise of convective instabilities in a layer of asthenosphere as it is heated from below: given the observed mantle heat flow (q_A) , under what conditions will convection in the underlying layer be sustained?

The Nusselt number (Nu; Poirier, 2000) is a measure of convective heat flow across the layer ratioed to the heat flow that conduction would transport:

$$Nu = \frac{q_A D}{K_2 \Delta T} \tag{4.5}$$

where ΔT is the difference in temperature between the top (T_A) and bottom of the convecting layer (T~) (Fig. 4.8). Where Nu <1, heat transfer is dominantly by conduction; conversely, convection requires Nu >1. The Rayleigh number (Ra) measures the relative importance of buoyancy forces favouring convection versus the viscous drag forces hindering convection:

$$Ra = \frac{\rho g \alpha \Delta T D^3}{\eta \kappa}$$
(4.6)

where α is thermal expansivity, κ is thermal diffusivity, and η is the viscosity of the convecting fluid (Tables 4.4 and 4.5; Fig. 4.8a). Theory and experimentation show the Nu number to be proportional to the Ra number to the one-third power as given by:

$$Nu = 0.225 Ra^{1/3}$$
 (4.7)



Figure 4.8: (A) conceptual model for convective instabilities in a layer of asthenosphere being heated from below; (B) Calculated viscosity in heated layer; (C) The ΔT and the value of D implied by range of heat flow values for the Cordilleran mantle lithosphere.

(Turcotte and Schubert, 2002). Rearranging Eq. 4.7 allows us to write an expression relating the mantle heat flow (q_A) to a set of physical constants for the layer of asthenosphere and the temperature difference across that layer (ΔT):

$$q_{A} = 0.225 K_{2} \left[\left\langle \frac{\rho g \alpha}{\kappa} \right\rangle^{1/3} \right] \frac{\Delta T^{4/3}}{\eta^{1/3}} \qquad (4.8)$$

Specifically, Eq. 4.8 relates the reduced heat flow $(q_r = q_A)$ to the minimum critical temperature difference needed to sustain convection in the underlying asthenosphere. We treat the viscosity of the fluid as a constant but allow its value to reflect the average and unknown temperature of the convecting layer $(T_f = 0.5 * [T_A + T_\infty])$. The viscosity of the mantle is approximated by the rheology of wet dunite (Chopra and Paterson, 1981) and the temperature dependent function is:

$$\ln \eta = \frac{11505}{T_f} + 35.45 \qquad (4.9)$$

The results of these calculations are summarized in Figs. 4.8b & 4.8c and Table 4.5. The known values include the heat flow (q_A) and the temperature (1100°C) on the upper boundary representing the base of the mantle lithosphere. The higher the mantle heat flow, the lower the viscosity of the mantle layer needed for convection (Fig. 4.8b). The viscosity of the convecting mantle is calculated by this method to be between 10^{19.7} to 10^{19.4} Pa s (Fig. 4.8b). Values of ΔT increase with increasing mantle heat flow and, for our range of q_A 's, we calculate values of 120 to 280°C (Fig. 4.8c). The minimum critical thickness of the convecting layer (D) decreases with increasing mantle heat flow (Fig. 4.8c).

Our incorporation of xenolith temperature data into simple geotherm models suggests a mantle lithosphere extending to depths of 54 - 75 km depth depending on the temperature



Figure 4.9: Implications of geotherm modeling for thickness and temperature distributions in Cordilleran mantle lithosphere. Coolest and hottest mantle lithosphere geotherms terminate at depths of 75 km (LM₂) and 54 km (LM₁), respectively. Other reference curves include the solidi (heavy lines) for fertile (F: Hirschmann, 2000) and depleted (D: Robinson and Wood, 1998) peridotite, and the approximate positions of the corresponding spinel (S) / garnet (G) transition for each peridotite (after Robinson and Wood, 1998). Calculated temperatures in convecting mantle asthenosphere (dashed lines with open circles or squares) are consistent with an average mantle model viscosity of $10^{19.5}$ Pa s (see Table 4.5). Basanite liquidus temperatures (bl) from Orlando et al (2000; dashed line, experimentally determined) and from the host basanite magma (solid line, sample MH-02-101; Table 2.2) as calculated from MELTS (Ghiorso and Sack, 1995).

of the MOHO (Table 4.5; Fig. 4.9). This corresponds with a mantle lithosphere as thin as 18 km or as thick as 39 km (Table 4.5). Over this depth interval the mantle lithosphere comprises only spinel peridotite (Fig. 4.9; Robinson and Wood, 1998), rather than garnet-bearing peridotite, which is consistent with the petrography of all peridotites recovered from Llangorse. The convecting asthenospheric mantle would have an average viscosity of 10^{19.5} Pa s, average temperatures of between 1200 and 1240 °C, and minimum temperatures towards the base of the convecting layer would be 1300 – 1380°C (Fig. 4.9). The minimum critical thickness of the convecting mantle layer varies between 46 and 57 km.

4.9.3 Implications for NCVP Magmatism

The lithospheric mantle in the northern Canadian Cordillera has been studied indirectly by geophysical crustal-scale surveys (Clowes et al, 1999; Hyndman and Lewis, 1999; Hammer et al, 2000; Snyder et al, 2002; Lewis et al, 2003; Hammer and Clowes, 2004) and directly via studies of mantle-derived xenoliths from volcanic rocks (Nicholls et al, 1982; Francis and Ludden, 1995; Shi et al, 1998; Peslier et al, 2000, 2000b). Geophysical surveys include measurements of heat flow, GPS surveys of tectonic displacements, gravimetric studies, and seismic P-wave refraction and wide-angle reflection surveys. These studies provide a wealth of information, including rates of plate motion and deformation and estimates of the strength, thickness and temperature of the lithosphere (Hyndman et al, in press). Peridotite xenoliths represent primary samples of material from the lithospheric mantle, and thus petrological studies of these rocks provide direct information about the nature of the upper mantle (e.g. Nicholls et al, 1982; Nicholls and Stout, 1996, 1997; Shi et al, 1998, Peslier et al, 2000, 2000b). Here, we summarize previous models advanced for the lithospheric mantle in the northern Cordillera and compare them to our preferred model.

Fig. 4.10a illustrates the model lithosphere from Nicholls et al (1982), based on their studies of alkaline basalts and peridotite xenoliths from across the southern and central Canadian Cordillera. They showed that the lavas could not be partial melts of the xenoliths they carried, but were derived from a deeper source. Peridotite xenoliths were interpreted as sourcing from the lithospheric mantle and the lavas as deriving from the underlying 'low velocity zone' (e.g., asthenospheric mantle). They suggest that the depth to the 'low velocity zone' is between 45 and 100 km, based on seismic data. They also use two-pyroxene thermometry (Wells, 1977) on peridotite xenoliths from many localities and show that the peridotite solidus. An average geotherm was adopted based on observed heat flow (Pollack and Chapman, 1977) which implied a MOHO temperature of 765°C at 36 km depth. The maximum xenolith temperature (1200°C) on this geotherm predicts the asthenosphere / lithosphere boundary at 90 km (Eq. 4.3). This estimate of the depth to the base of the lithosphere was shown to be consistent with the geophysical estimates of the low velocity zone (45 - 100 km).

Fig. 4.10b depicts the corresponding model of Francis and Ludden (1995) and Shi et al (1998). Francis and Ludden (1995) invoke the presence of amphibole in the source region of alkaline lavas in order to explain the trace chemistry patterns of the most primitive lavas observed in the NCVP, olivine nephelinites (NEPH). Their model places the lithosphere / asthenosphere boundary considerably deeper than the amphibole stability field, and on this







Figure 4.10

Figure 4.10: Models of the lithosphere underlying the northern Canadian Cordillera: (A) Nicholls et al (1982), showing the minimum and maximum temperatures calculated for peridotite xenoliths from diverse volcanic centres (shaded box from approximately 900-1200°C). They use a simple heat flow model to calculate a geotherm which constrains: $T_{moho} = 765^{\circ}C$, L/A (lithosphere / asthenosphere boundary) = 90 km. LVZ (low velocity zone) constrained by geophysical measurements in south-central BC at 45 - 100 km. (B) Francis and Ludden (1995) and Shi et al (1998), showing the potential melting zone of mantle peridotite (shaded area) from dry peridotite (P(D)) to wet peridotite (P(W)). The melting curve for amphibole (Amph) and the spinel / garnet transition zone (S/G) are also shown. Based on these relationships they argue that some lavas are derived from melting in the amphibole-bearing lithospheric mantle. (C) Compilation of geophysical work (Lewis et al, 2003; Hammer and Clowes, 2004; Hyndman et al, in press), showing two possible geotherms for north and south of 59°N and deep mantle reflectors (R; Hammer and Clowes, 2004). (D) Our preferred model for temperature distributions for the northern Cordilleran lithosphere contains the following elements: i) a family of crustal geotherms which are consistent with our calculated T_{moho} values of 800-850°C and the observed values of q_o and A₁ (Hyndman and Lewis, 1999; Lewis et al, 2003); ii) two limiting geotherms for the mantle lithosphere based on average values of crustal heat production and consistent with the maximum xenolith temperature (1100°C). These geotherms terminate against the lithosphere / asthenosphere boundary at depths of 58 and 73 km, respectively. The temperature along the geotherm is below the dry solidus temperatures of peridotite (P(D)) and the basanite liquidus (bl; see Fig. 4.9), but above the amphibole (Amph) and phlogopite (Phl) melting curves (see hatched box); iii) three zones of possible melting are illustrated (numbered regions): (1) mantle lithosphere melting in the amphibole stability field; (2) melting at the base of the lithosphere due to high temperatures sustained by the underlying asthenosphere; (3) melting within the convecting, garnet-bearing asthenosphere.

basis they argue that NEPH lavas must source from the lithospheric mantle rather than the asthenospheric mantle (e.g., Nicholls et al, 1982). However, their lithosphere / asthenosphere boundary is probably too deep for the active Cordillera and their geotherm is too cool, placing T_{moho} at too low a temperature (e.g., $T_{moho} = 400$ °C, rather than $T_{moho} > 800$ °C; Hyndman et al, in press). Nevertheless, the possibility that alkaline lavas can be sourced from the lithospheric mantle is still valid. The source of the peridotite xenoliths is further constrained to be above the spinel / garnet transition (e.g., 80 km) due to the lack of textural or geochemical evidence of garnet (Shi et al, 1998).

An integrated geophysical model of the northern Cordilleran lithosphere is summarized in Fig. 4.10c. This model includes seismic and heat flow data, GPS motion measurements, and gravimetric measurements (Lewis et al, 2003; Hammer and Clowes, 2004; Hyndman et al, in press). The MOHO depth in this portion of the Cordillera is set at 35 - 36 km (Hammer and Clowes, 2004). Detailed studies of heat flow, heat generation, and crustal temperatures in the northern Cordillera (Lewis et al, 2003) are used to generate lithosphere geotherms and provide MOHO temperatures of 800-1000°C (Lewis et al, 2003) and 950°C ± 100 °C (Hyndman et al, in press). Differences in crustal heat production result in slightly different crustal geotherms for areas north and south of 59°N latitude, with a slightly higher geotherm favoured for north of 59°N.

This integrated geophysical model places the lithosphere / asthenosphere boundary at between 50 and 60 km depth in the northern Cordillera. A series of upper mantle reflectors are observed at this depth (Hammer and Clowes, 2004). Though not well constrained, these reflectors are located at approximately 60 - 75 km depth and dip at shallow angles to the northeast. These reflectors may result from heterogeneities in the mantle associated with the

base of the lithosphere, as might arise from interleaving of distinct lithospheric material. Conversely, the heterogeneities may be related to partial melting processes. It is interesting to note that the depth of these mantle reflectors corresponds with Hyndman et al's (in press) estimate of the depth to the lithosphere / asthenosphere boundary.

Our model of the mantle lithosphere (Fig. 4.10d) combines petrographic and geophysical observations. Ideal geotherms are constrained to pass through the MOHO at temperatures of 800 and 850°C and terminate at 1100°C, corresponding with the minimum and maximum temperatures recovered from our peridotite xenolith suite. These geotherms are compared with other petrographic constraints to delimit the potential source regions for xenoliths and lavas. Two basanite liquidus curves are provided; one is from Orlando et al (2000) and is experimentaly determined, the other is calculated from Llangorse centre lavas using MELTS (Ghiorso and Sack, 1995). These liquidus curves indicate that the minimum temperature necessary to produce the basanite lavas is approximately 1375°C. Liquidus temperatures for melting of dry (P(D)) and wet (P(W)) peridotite (Hirschmann, 2000) define the region where melting of mantle material can occur (shaded gray). Superimposed on this region are the melting curves of phlogopite (Phl) and amphibole (Amph) (Kushiro et al, 1968; Millhollen et al, 1974; Moderski and Boettcher, 1973), which define the stability limits of these hydrous minerals in the mantle. These constraints allow inferences to be made about the depth, temperature, and composition of the lithospheric mantle. The shaded column on the left of Fig. 4.10d shows our model depth to the lithosphere / asthenosphere boundary (L/A). Our model lithosphere (light gray shaded box) is as thin as 18 km (L/A = 54 km) or as thick as 39 km (L/A = 75 km), depending on the geotherm used. We envision

the base of the mantle lithosphere to be at a temperature ≥ 1100 °C, immediately underlain by an asthenosphere composed of spinel-bearing peridotite; the depth to the spinel / garnet transition is estimated at > 80 km (Robinson and Wood, 1998).

Our interpretation of this model denotes three possible scenarios where melting of the mantle can occur. The first scenario involves melting of amphibole-bearing lithospheric mantle. Francis and Ludden (1995) interpret the geochemical diversity of nephelinite (± basanite) lavas as resulting from partial melting of amphibole-bearing peridotite (e.g., lithospheric mantle; Fig. 4.10b). Melting scenario 1 shows the restricted region in which this could occur. Given our model geotherms, amphibole is only stable in the mid-levels of the mantle lithosphere and should be absent at the base of the lithosphere. Thus, the Francis and Ludden (1995) model is possible only if melting occurs in the mid-lithosphere. However, temperatures at these depths are well below (> 200°C) the calculated liquidus temperature of Llangorse Mountain basanite lavas. Scenario 1 is also too shallow to sample the deepest xenoliths observed at Llangorse Mountain. Melting scenario 2 involves melting of two potential sources: the base of the mantle lithosphere or the uppermost, spinel-bearing asthenosphere. Temperature gradients at the lithosphere / asthenosphere transition, although high, do not bring the mantle temperature close to that of the basanite liquidus (e.g., within 50°C). Melting scenario 3 allows for partial melting of the deep asthenosphere in the garnet stability field; this scenario is also at temperatures well below the basanite liquidus. Some previous workers (e.g., Francis and Ludden, 1990, 1995) have argued that there is a conventional 'garnet signature' in trace chemistry of some NCVP lavas, suggesting a garnetbearing asthenospheric source region. However, the model lithosphere presented by these authors (Fig. 4.10b) also requires amphibole in the source region; these two requirements

cannot be reconciled in our model. Recent improvements in our understanding of the spinel / garnet transition indicate that this transition occurs much deeper than originally thought (Robinson and Wood, 1998; Klemme and O'Neill, 2000), calling into question the validity of the 'garnet signature.' If the additional constraint of the garnet signature is lost, the possibility that alkaline lavas in the NCVP may source from the deep garnet-bearing asthenosphere must be permitted in our model.

Our preferred interpretation is that the Llangorse Mountain lavas were produced by melting scenario 2 for the following reasons: a) scenario 2 would allow sampling of the full mantle lithosphere observed in peridotite xenoliths at Llangorse Mountain, which is not possible in scenario 1; b) unlike scenarios 1 and 3, two potential sources are involved (mantle lithosphere and asthenosphere); this could explain some of the chemical diversity observed in NCVP lavas (Edwards and Russell, 2000); c) scenario 2 is consistent with the rarity of garnet or amphibole-bearing peridotite xenoliths throughout the northern Cordillera (Littlejohn and Greenwood, 1974; Ross, 1983; Brearley and Scarfe, 1984; Brearley et al, 1984; Nicholls et al, 1982; Shi et al, 1998). There is, however, a discrepancy between the temperatures observed in source region 2 and the basanite liquidus curves. If the base of the mantle lithosphere is sightly hotter (i.e., 100°C), melting zone 2 would be displaced to both higher temperatures and pressures, possibly bringing the region up to the basanite liquidus curves. This would imply, however, that the highest-temperature mantle lithosphere is not represented in the Llangorse Mountain xenolith suite.

4.10 CONCLUSIONS

A refined model of the lithospheric mantle for the northern Canadian Cordillera is presented. This model combines both petrological and geophysical data and observations, allowing us to estimate the minimum thickness of the lithospheric mantle. Heat flow measurements are used to calculate model geotherms appropriate to the northern Cordillera; geothermometry of peridotites allows us to constrain the temperatures in the lithospheric mantle to be between 800°C (T_{moho}) and 1100°C (T_A). Our model indicates the minimum thickness of the lithospheric mantle to be between 18-39 km, corresponding to depths at the base of the lithosphere of 54-75 km. We consider three melting scenarios in the upper mantle; our preferred scenario involves melting at the base of the lithospheric mantle or uppermost asthenosphere, at depths of approximately 54-75 km and at temperatures around 1300°C.

4.11 ACKNOWLEDGMENTS

Field work costs were derived from the Geological Survey of Canada through the Atlin Integrated Geoscience Project, courtesy of Bob Anderson. All other costs were met through NSERC via a Discovery Grant to J.K. Russell and a postgraduate scholarship to Margaret Harder. Our work benefited greatly through discussions, though not necessarily in agreement on all issues, with Ben Edwards, Phil Hammer, and Roy Hyndman.

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Chapter 5

THE MANTLE LITHOSPHERE BENEATH THE CANADIAN CORDILLERA

5.1 STRUCTURE IN THE LITHOSPHERIC MANTLE

The chemical, mineralogical, and physical properties compiled for Llangorse Mountain peridotite samples (Table 4.3; Chapter4) can be plotted against temperature as a proxy for depth. This establishes a structural profile for the mantle lithosphere, which can be used to show whether the mantle lithosphere beneath the northern Canadian Cordillera is heterogeneous or homogeneous. Fig. 5.1 shows variations in modal mineralogy, density, and mineral chemistry with depth. The "depth" of each sample is calculated from model geotherms derived from geothermometry of the peridotite suite. In Chapter 4 we provide two extreme geotherms; the geotherm assumed depends on the temperature assumed for the MOHO. A lower MOHO temperature (e.g. 800°C) corresponds to a deeper asthenosphere / lithosphere boundary (e.g. 75 km), and consequently a lower geothermal gradient (6.4° C/km). Conversely, a higher MOHO temperature (e.g. 850°C) corresponds to a shallower asthenosphere / lithosphere boundary (e.g. 54 km) and consequently a higher geothermal gradient (13.9°C/km). For the purposes of comparing peridotite properties with depth, I have used an average geothermal gradient of 10.2°C/km. Error bars shown in the first column represent uncertainty in depth corresponding to the variance in temperatures for each sample.

The shaded regions in Fig. 5.1 reflect the presence of foliated lherzolite samples only at certain depth intervals. In all cases, foliated peridotite samples are lherzolites, and



Figure 5.1: Variation in peridotite mineralogical, chemical, and physical properties with depth. % olivine and % clinopyroxene represent modal mineralogy of peridotites determined from calculations based on bulk chemical results; results are comparable with modal mineralogy obtained using the Rietveld method (Appendix B). Depth values for each sample are based on thermometry results and an average model geothermal gradient (10.2°C/km) detailed in Chapter 4; error bars shown in the first column represent the depth uncertainty corresponding with temperature variance in each sample. Dark circles represent foliated lherzolites, light coloured circles represent non-foliated harzburgites and rare lherzolites.

although non-foliated lherzolites are observed, most non-foliated peridotites are harzburgites. Moreover, foliated lherzolites are common only in the shallowest lithospheric mantle (approximately 36-40 km), and over a narrow depth interval of 42-44 km. Conversely, non-foliated harzburgites are rare in the shallow lithospheric mantle, but are the dominant rock type at depths in excess of 40 km. The variation in density of peridotite samples with depth is somewhat counter-intuitive in that density appears to decrease with increasing depth in the lithospheric mantle. However, density typically decreases with increasing temperature, and therefore temperature, rather than pressure, probably is the major control of density in the lithospheric upper mantle.

It is apparent from the location of foliated lherzolite samples that the lithospheric mantle in the northern Cordillera is heterogeneous with distinct layers of lherzolites and harzburgites. Although large uncertainties in depth for many samples make the mid-level (i.e., 40-43 km) harzburgite layer ambiguous, the deeper mantle (>44 km) is clearly harzburgitic. The presence of distinct harzburgite and lherzolite regions has important implications for partial melting. It is commonly accepted (e.g., Harte and Hawkesworth, 1986) that lherzolites represent less depleted mantle, whereas harzburgites represent residual mantle that has undergone extensive partial melting. Therefore, the apparent layering defined by the proportions of relatively depleted harzburgite implies that partial melting was more extensive at certain depths in the upper mantle; partial melting appears to have been most extensive in the deeper lithospheric mantle (i.e. > 40 km).

5.2 COMPARISON TO OTHER NCVP LOCALITIES

Figures 5.2 and 5.3 display a thermal profile along the northern Cordilleran volcanic province (NCVP). Geothermometry results, based on mineral chemistry compiled from various authors, is shown for numerous Miocene to Holocene volcanic localities throughout the NCVP. The thermal profile extends in a roughly north-northwest to south-southeast direction, from Prindle Volcano in eastern Alaska to Castle Rock in north-central British Columbia (after Edwards and Russell, 2000 and Nicholls et al, 1982). Also shown are the various tectonic terranes through which the lavas traversed to reach the surface.

Although the highest temperature samples in the Llangorse Mountain profile are those which show the greatest signs of interaction with lava (e.g., sieve-textured orthopyroxene; Chapter 4) and may not be reliable temperatures, they are nonetheless included in Figures 5.2 and 5.3 for two reasons. Firstly, these temperatures must be less than or equal to the temperature of the entraining basanite. As we interpret the basanite to source from the asthenosphere / lithosphere boundary, these potential disequilibrium temperatures may reflect temperatures close to those found at the boundary, but not higher. Therefore, they may still be valid as constraining temperatures. Secondly, the Llangorse Mountain, Fort Selkirk and Castle Rock thermal profiles look similar, with most peridotite temperatures clustered between 800 and 1000°C, a temperature gap between 1000 and 1050° C, and a cluster of higher-temperature (1050-1200°C) samples. The similarities in these thermal profiles may indicate that the highest temperature samples from these other localities are also disequilibrium samples.

In Figures 5.2 and 5.3, the minimum temperature from each peridotite suite is taken to represent the maximum MOHO temperature (T_{MOHO}), and the maximum temperature





Figure 5.3

Figure 5.2: Thermal profile across the northern Cordilleran volcanic province (NCVP). Temperatures are calculated using a two-pyroxene thermometer (Brey and Kohler, 1990 calibration) on orthopyroxene and clinopyroxene mineral chemistry determined by other workers. Abbreviations and authorship are as follows: D: West Dawson (Shi et al, 1998); P: Prindle Volcano (Prescott, 1983; Shi et al, 1998); FS: Forth Selkirk (Prescott, 1983; Shi et al, 1998); AL: Alligator Lake (Francis, 1987); R: Ruby Mountain (Nicholls et al, 1982); H: Hirschfeld Creek (Shi et al, 1998); Llang: Llangorse Mountain (see Table 4.4; Appendix A, F); CR: Castle Rock (Prescott, 1983; Shi et al, 1998). T_{MOHO} = maximum temperature of the MOHO; T_{Asthenosphere} = temperature at the lithosphere / asthenosphere boundary.

Figure 5.3: Thermal profile across the northern Cordilleran volcanic province (NCVP); temperatures (see Fig. 5.2) are shown for centres with > 5 analyses. The depth to the MOHO is set at 36 km (Hammer and Clowes, 2004). A constant mantle geotherm across the NCVP is assumed at 10.2°C/km, but T_{MOHO} is allowed to vary. Lines represent isotherms from 800°C to 1200°C at 100°C intervals. Abbreviations and references are as in Fig. 5.2. from each suite is taken to approximate the minimum temperature at the asthenosphere / lithosphere boundary (T_A). Fig. 5.2 shows the variations in apparent T_{MOHO} and T_A across the NCVP. In the northernmost part of the NCVP (Prindle Volcano), T_{MOHO} values are similar to Llangorse Mountain (~800°C), but T_A values are somewhat lower (~1000°C at Prindle Volcano versus 1050-1100°C for Llangorse Mountain). Slightly farther south, in the Yukon Territory, T_{MOHO} values are considerably higher than at Llangorse Mountain (~900° C), whereas T_A values are similar. In the southern portion of the NCVP (Castle Rock), T_{MOHO} values are the highest in the NCVP (-950°C), and T_A values may be as high as 1200° C. Without assuming a specific crustal geothermal gradient, no depth values are assigned to T_{MOHO} at each locality. However, if a constant crustal geotherm is assumed across the NCVP, Fig. 5.2 would imply significant topography in the MOHO of the northern Canadian Cordillera.

The depth to the MOHO, however, is well-constrained in the northern Canadian Cordillera at 35-37 km ±2 km (Hammer and Clowes, 2004). Therefore, the minimum temperatures recovered for the NCVP localities shown in Fig. 5.2 can be placed at the same depth (e.g., 36 km). The mantle geothermal gradient used in Fig. 5.3 is based on an average model geotherm of 10.2°C/km (see Chapter 4). Fig. 5.3 shows a depth versus latitude profile of the NCVP assuming a uniform geothermal gradient in the mantle and MOHO depth across the NCVP. Isotherms define variations in the thermal profiles from each NCVP locality. Fig. 5.3 shows elevated MOHO temperatures in the Stikinia Terrane (both in the Yukon Territory and British Columbia), with lower MOHO temperatures in the Cache Creek (Llangorse Mountain) and Yukon-Tanana (Prindle Volcano) Terranes. The depth to

the lithosphere / asthenosphere boundary varies from 52 km to 71 km.

The thermal profiles (i.e., $T_{MOHO} - T_A$) shown in Figures 5.2 and 5.3 indicate that the greatest thickness of mantle lithosphere in the NCVP is present beneath Llangorse Mountain. Figures 5.2 and 5.3 also show substantial variations in T_{MOHO} (and possibly T_A) across the NCVP. However, these observations must be interpreted cautiously. As shown in Chapter 4, a small suite of peridotite samples is not likely to accurately represent the true thermal profile of the mantle lithosphere that the entraining lavas pass through. Because the thickest mantle lithosphere is apparently beneath Llangorse Mountain, it is possible that the peridotite suite used in this study is simply the most complete suite of all the localities shown in Figures 5.2 and 5.3. Similar studies on a sufficient number of peridotites from other localities is necessary in order to provide meaningful comparisons of the thermal state of the mantle lithosphere across the NCVP.

5.3 ACKNOWLEDGMENTS

I would like to thank Lorraine Tam, Marcia Wilson, Lisa Swinnard, and Krista Michol for their assistance in various analytical procedures used to characterize these peridotites.

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Chapter 6

CONCLUSION

The Llangorse volcanic field (LVF) is located near the town of Atlin in northwest British Columbia, and belongs to the larger northern Cordilleran volcanic province (NCVP). The LVF comprises numerous alkaline mafic, Miocene to Holocene volcanic remnants. In particular, four volcanic localities within a 6 km radius of Llangorse Mountain have been studied in detail. The largest of these sites, the Llangorse Mountain locale, is a large, cliffforming basanite flow containing abundant peridotite xenoliths and is associated with volcanic-generated debris flow sediments. Another of these localities, Hidden Ridge, is also associated with debris flow sediments and is characterized by irregular development of cooling joints. Immediately south of the Llangorse Mountain locale is Lone Point, a moundshaped volcanic remnant that displays highly irregular columnar jointing and is associated with hyaloclastite boulders.

These volcanic remnants are interpreted to represent valley-filling lava flows associated with syn-volcanic debris flows (the Llangorse locale and Hidden Ridge) and subglacial volcanism (Lone Point; possibly part of Hidden Ridge). The characteristics of debris flows at the Llangorse locale suggest that they are contemporaneous with the overlying lava flow and sampled material from the volcanic vent during early stages of basanitic volcanism. Accordingly, the debris flows may represent lahars or outburst floods due to melting of snow or ice around the volcanic vent. Another interesting feature of the Llangorse Mountain debris flows is the extensive palagonitization of all basanite glass

clasts. Studies of the palagonitized glass fragments suggest that pore fluids associated with palagonitization were heated to 80-120°C, possibly by the overlying lava flow.

The morphology and aspect ratio of the valley-filling flow at Llangorse Mountain is indicative of ponding against a barrier in the valley. Erratic orientation of cooling joints at other volcanic localities near Llangorse Mountain (Lone Point, possibly part of Hidden Ridge) are indicative of cooling against ice. The evidence of glacial ice coinciding with volcanism in the LVF therefore implies that the barrier against which the Llangorse Mountain lavas ponded could have been glacial ice. Furthermore, glaciation is a likely mechanism for the complete erosion of the volcanic vent(s) from which the Llangorse Mountain and Hidden Ridge lavas were sourced.

The lavas and debris flow sediments at Llangorse Mountain are host to abundant peridotite xenoliths. These xenoliths represent accessory fragments of the lithospheric mantle entrained by the Llangorse Mountain basanite lavas. The xenoliths comprise spinel lherzolite and harzburgite and range from unfoliated to moderately foliated. Detailed mineral chemistry is used for two-pyroxene thermometry, based on the Brey and Kohler (1990) calibration. Thermal data for 44 peridotite xenoliths is used to define the minimum (800-850°C) and maximum (1050-1100°C) temperatures of the xenolith suite. These temperatures constrain the thermal profile of the northern Cordilleran mantle lithosphere. The minimum temperatures represent the maximum MOHO temperature, and the maximum temperatures approximate the minimum temperature at the lithosphere / asthenosphere boundary. The geothermometry data are combined with measured heat flow data from Lewis et al (2003) to produce a set of model geotherms for the northern Cordillera. These
geotherms constrain the thickness of the mantle lithosphere in the northern Cordillera to be between 18 - 39 km, which corresponds to depths to the lithosphere / asthenosphere boundary of 54 - 75 km. The thermal profile created by these geotherms is used to infer potential source regions of lavas in the NCVP.

A comparison of various mineralogical, chemical, and physical properties of peridotites collected at Llangorse Mountain provide a structural profile of the mantle lithosphere in this part of the Canadian Cordillera. Distinct layers of lherzolite versus harzburgite peridotite corresponds with denser and chemically different regions in the mantle lithosphere. In short, the mantle lithosphere beneath Llangorse Mountain is heterogeneous. Comparison of the thermal profile of the mantle lithosphere at Llangorse Mountain with other centres in the NCVP implies one of two things: a) the mantle lithosphere in the NCVP has large variations in temperature and, presumably, depth; or b) peridotite suites from other localities in the NCVP are not complete, and further comparison studies require more detailed work at these localities.

Appendix A

GEOTHERMOMETRY METHODS

A.1. GEOTHERMOMETRY CALCULATIONS

To convert weight percent chemistry to cation proportions the following formula is used:

$$\frac{\left[4 \operatorname{cations} * \left(\frac{(\operatorname{cations} \operatorname{per} \operatorname{oxide})}{(MW \operatorname{oxide})}\right)\right]}{\left[\Sigma\left(\frac{(\operatorname{cations} \operatorname{per} \operatorname{oxide})}{(MW \operatorname{oxide})} * \operatorname{wt} \operatorname{oxide}\right)\right]}$$
A.1

where MW is the molecular weight of each oxide. Eq. A.1 assumes an ideal pyroxene formula of 4 cations. Cationic proportions are necessary for geothermometry calculations.

The following equations are used for the geothermometry calculations of both the Brey and Kohler (1990) and Wells (1977) geothermometers:

A.1.1. Wells (1977) Geothermometer

The Wells geothermometer is based on the premise that diopside and enstatite solubility in coexisting pyroxenes is dependent only on temperature. The Wells (1977) thermometer is based on the Wood and Banno (1973) thermometer, but is better calibrated to include realistic pyroxene compositions of the mantle. The Wells (1977) thermometer defines the equilbrium coefficient as the activity of enstatite in coexisting ortho- and clinopyroxene pairs:

$$K = \frac{a_{Mg}^{cpx}}{a_{Mg}^{opx}} \qquad A.2$$

where K is the equilbrium coefficient and a is the activity of Mg in pyroxene.

The empirically experimentally determined thermometry formula is:

$$T = \frac{7341}{[3.355 + 2.44 * X_{Fe}^{opx} - lnK]} \qquad A.3$$

where T is temperature in degrees Kelvin.

Because random distribution of Mg and Fe between the M1 and M2 sites is assumed, the activity of Mg in each pyroxene depends on the amount of other cations in each site. The cation distribution for each site (according to Wood and Banno, 1973) is:

M1: Fe, Mg, Ti, Al(vi)

M2: Fe, Mg, Ca, Na, Mn

Table A.1 provides an example of the Wells (1977) thermometer for sample MH-02-

121, with values provided for each of the above variables for each pyroxene pair.

A.1.2. Brey & Kohler (1990) Geothermometer

The Brey and Kohler (1990) thermometer is also based on the premise that diopside and enstatite solubility in coexisting ortho- and clinopyroxene grains is dependent on temperature. This thermometer, however, takes into consideration more realistic pyroxene compositions. The activity of Fe in each pyroxene is defined as:

$$X_{Fe}^{Pyx} = \frac{Fe}{(Fe + Mg)} \qquad A.4$$

where Fe and Mg are the cationic proportions of iron and magnesium, respectively.

The Ca activity in the M2 pyroxene site is dependent on the total Na content:

$$Ca^{@} = \frac{Ca^{M2}}{(1 - Na^{M2})} \qquad A.5$$

where Ca and Na are the cationic proportions of calcium and sodium, respectively, and all Ca and Na is assumed to be in the M2 pyroxene site.

The equilibrium distribution of Ca between clinopyroxene and orthopyroxene is therefore:

$$K_D^{@} = \frac{(1 - Ca^{@})^{cpx}}{(1 - Ca^{@})^{opx}}$$
 A.6

where $K_D^{@}$ is the equilibrium coefficient and Ca[@], as defined in Eq. A.5, is the Ca activity in the M2 site.

Equations A.4-A.6 are applied to the experimentally determined thermometer of Brey and Kohler (1990):

$$T_{BKN} = \frac{\left[23664 + (24.9 + 126.3 * X_{Fe}^{cpx}) * P\right]}{\left[13.38 + (lnK_{D}^{@})^{2} + 11.59 * X_{Fe}^{opx}\right]} \qquad A.7$$

where T_{BKN} is temperature in degrees Kelvin and P is pressure in kilobars.

Table A.1 shows an example of the Brey and Kohler (1990) thermometer, with values provided for each for the variables defined above for each pyroxene pair.

A.2. COMPARISON OF THE THERMOMETERS

The mineral chemical data collected for Llangorse Mountain peridotite xenoliths was applied to both the Wells (1977) and Brey and Kohler (1990) thermometers (Fig. A.1a). The results indicate that temperatures calculated using the Wells (1977) thermometer are consistently lower than temperatures calculated for the same samples using the Brey and Kohler (1990) thermometer to approximately 900°C (Fig A.2b). The Wells (1977) thermometer produces consistently higher temperatures than the Brey and Kohler (1990) thermometer above approximately 980°C. Temperature differences between the two thermometers are most pronounced at low temperatures (< 880°C). The temperature discrepancies at high temperatures may in part result from a constant pressure (10 kbar) used in all calculations with the Brey and Kohler thermometer, whereas temperature disagreements at low temperatures must result from differences inherent in the thermometers. When the lowest and highest temperature xenoliths are re-calculated to assume a 'corrected' pressure for increasing depth, the corresponding temperature difference is approximately 15-20°C; the inherent uncertainties (1≥) associated with the Brey and Kohler (1990) thermometer are $\pm 15^{\circ}$ C. The highest temperatures could be up to 20°C higher than reported, which is just outside the thermometer uncertainty.

Fig. A.2 shows the equilibrium constants of ortho- and clinopyroxene (of all samples) plotted against each other; an approximate straight line is observed, showing that the equilibrium constants of the disturbed samples (open squares) do not deviate from those of the ambient mantle samples. This indicates that the Fe-Mg exchange reaction is not disturbed by these higher temperatures. In, Fig. A.2b, the molar fraction of Ca in ortho- and clinopyroxene is plotted. Fig. A.2b clearly demonstrates that the disturbed samples (open squares) have orthopyroxene grains enriched in Ca, and clinopyroxene grains depleted in Ca, relative to the ambient mantle samples. Fig. A.2 illustrates that Ca is mobilized by the higher temperatures of the surrounding lava, whereas the Fe-Mg exchange appears unaffected.

Because the Brey and Kohler (1990) thermometer is more recent and apparently better calibrated, we have chosen to use only the Brey and Kohler (1990) thermometer for our model geotherm studies.

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Cationic pre	portions for	eacn pyro	xene anal	ysis is can	culateu va	sed on an	lucal pyrc		nula ui 4	cauous.				
Mineral	opx	cpx	vdo	cpx	opx	cpx	xdo	cpx	xdo	cpx	xdo	cpx	xdo	cpx
T set	1	-	2	2	З	ε	4	4	S	S	9	9	7	7
SiO ₂	54.13	51.41	54.20	51.46	55.15	51.92	55.61	51.55	54.38	52.58	54.74	52.11	54.46	50.93
TiO_2	0.02	0.16	0.05	0.09	0.03	0.12	0.02	0.13	, 0.02	0.11	0.03	0.10	0.02	0.15
Al_2O_3	3.83	4.84	3.72	4.42	3.67	4.82	3.50	4.77	3.55	4.00	3.41	4.30	3.71	4.69
Cr_2O_3	0.36	0.86	0.42	0.76	0.38	0.93	0.33	0.87	0.34	0.68	0.36	0.70	0.37	0.88
FeO	5.77	2.29	5.79	2.15	5.96	2.23	5.69	2.23	5.76	2.03	5.98	2.22	5.93	2.15
MnO	0.14	0.09	0.15	0.10	0.13	0.05	0.16	0.07	0.09	0.09	0.15	0.09	0.12	0.08
MgO	33.59	15.92	33.50	16.33	33.86	15.97	33.69	16.07	. 33.58	16.42	33.90	16.44	33.84	16.09
NiO	0.13	0.04	0.02	0.07	0.10	0.06	0.10	0.00	0.09	0.03	0.08	0.03	0.11	0.00
CaO	0.51	22.22	0.49	22.27	0.48	21.92	0.53	21.90	0.49	22.20	0.54	22.51	0.53	22.21
Na_2O	0.03	06.0	0.02	0.87	0.07	0.93	0.03	0.92	0.05	0.80	0.04	0.85	0.01	0.87
Total	98.50	98.73	98.35	98.52	99.82	98.93	99.65	98.51	98.34	98.94	99.25	99.35	90.08	98.04
Formula bas	sed on 4 cati	ons:												
Mineral	xdo	cpx	xdo	cpx	xdo	cpx	xdo	cpx	xdo	cpx	opx	cpx	opx	cpx
T set	-	1	2	5	ŝ	ŝ	4	4	5	5	9	9	7	7
Si4+	1.89	1.88	1.89	1.89	1.90	1.90	1.92	1.89 [`]	1.90	1.92	1.90	1.89	1.89	1.88
Ti4+	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Al3+	0.16	0.21	0.15	0.19	0.15	0.21	0.14	0.21	0.15	0.17	0.14	0.18	0.15	0.20
Cr3+	0.01	0.02	0.01	0.02	0.01	0.03	0.01	0.03	0.01	0.02	0.01	0.02	0.01	0.03
Fe2+	0.17	0.07	0.17	0.07	0.17	0.07	0.16	0.07	0.17	0.06	0.17	0.07	0.17	0.07
Mn2+	0.00	0.00	0.00	0.00	0.00	00.0	0:00	0.00	00.0	0.00	0.00	0.00	0.00	0.00
Mg2+	1.75	0.87	1.75	0.89	1.74	0.87	1.73	0.88	1.75	0.89	1.75	0.89	1.75	0.88
Ni+2	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Ca2+	0.02	0.87	0.02	0.87	0.02	0.86	0.02	0.86	0.02	0.87	0.02	0.88	0.02	0.88
Nal+	0.00	0.06	0.00	0.06	0.00	0.07	0.00	0.07	0.00	0.06	0.00	0.06	0.00	0.06
Sum	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000
OXY's	5.971	5.972	5.978	5.964	5.979	5.985	5.995	5.977	5.977	5.991	5.971	5.969	5.970	5.964

Table A.1, coi	tinued:	Values fo	r all varia	bles neces	ssary to cal	culate ter	nperatures	using bo	th the Bre	y and Ko	hler (1990) and We	(1977) (1977)			
thermometers	(see A.1.).						I	•		ı						
Brey & Kohlei	r, 1990 (10	kbar):														
Mineral	xdo	cpx	xdo	cpx	xdo	cpx	xdo	cpx	xdo	cpx	vdo	cpx	xdo	cpx		
T set	1	-	7	7	б	3	4	4	5	S	9	9	7	7		
Ca*	0.019	0.931	0.018	0.932	0.018	0.919	0.019	0.921	0.018	0.921	0.020	0.933	0.020	0.935		
K_{D}^{*}	0.070	_	0.069	•	0.083		0.081		0.080		0.069		0.067			
${\rm X}_{{ m Fe}}^{ m pyx}$	0.079	0.056	0.081	0.044	0.083	0.062	0.085	0.057	0.080	0.059	0.081	0.050	0.080	0.045		
T _{BKN (K)}	1122		1118		1168		1159		1161		1115		1108	A	/erage	ΔT
T _{BKN (C)}	849		845		895		886		888		842		835		863	60
Wells, 1977:											••.					
Mineral	vdo	cpx	opx	cpx	xdo	cpx	xdo	cpx	xdo	cbx	vdo	cpx	xdo	cpx		
T set	1	-	2	2	ю	ε	4	4	5	ŝ	9	9	7	7		
Al (iv)	0.11	0.12	0.11	0.11	0.10	0.10	0.08	0.11	0.10	0.08	0.10	0.11	0.11	0.12		
Al (vi)	0.05	0.09	0.05	0.08	0.05	0.10	0.06	0.10	0.05	0.09	0.04	0.08	0.04	0.08		
¹ Mg (M1)	0.87	0.84	0.87	0.86	0.86	0.83	0.86	0.83	0.87	0.85	0.88	0.85	0.87	0.85		
² Mg (M2)	0.89	0.06	0.89	0.06	0.89	0.07	0.89	0.07	0.89	0.07	0.89	0.06	0.89	0.05		
¹ Fe (M1)	0.08	0.07	0.08	0.06	0.09	0.06	0.08	0.06	0.08	0.06	0.09	0.06	0.09	0.06		
² Fe (M2)	0.09	0.00	0.09	0.00	0.09	0.01	0.08	0.01	0.09	0.00	0.09	0.00	0.09	0.00		
X(Enstatite)	0.77	0.05	0.77	0.05	0.77	0.06	0.76	0.06	0.77	0.06	0.78	0.05	0.78	0.05		
X (Fe)	0.09	0.07	0.09	0.07	0.09	0.07	0.09	0.07	0.09	0.06	0.09	0.07	0.09	0.07		
T (K)	1154		1159		1190		1188		1187		1154		1150	Av	erage	ΔT
T (C)	881		886		917		915		914		, 881		876		896	40

•



Figure A.1: Two-pyroxene geothermometry results from peridotite xenoliths based on the Wells (1977) and Brey and Kohler (1990) (BK) thermometers; filled circles represent ambient mantle samples, open squares represent samples recording anomalously high temperature rims. (A) Temperatures for all samples calculated using the two geothermometers compared to a 1:1 line. Errors bars represent sample variance with each thermometer. (B) Difference in apparent temperatures between the BK and Wells thermometers are plotted against the apparent BK temperature. Temperature differences between the two thermometers are most pronounced at temperatures < 880 °C. Temperatures calculated from Wells (1977) are generally higher than BK temperatures over the temperature range 800 to 950 °C. The two methods are within error of each other over the range 900 – 1150 °C.



Figure A.2: Compositions of pyroxenes in mantle peridotite samples representing ambient mantle (filled circles) and high temperature disturbed samples (open squares). (A) Equilibrium constant (K_D) for coexisting pyroxenes. (B) Mole fractions of Ca (X_{Ca}) in coexisting pyroxenes; high temperature rims have distinctly higher Ca contents in orthopyroxene and lower Ca contents in clinopyroxene.

Appendix B

RIETVELD METHOD AND RESULTS

B.1. RIETVELD TECHNIQUE

The Rietveld technique (Rietveld, 1967; 1969) is a tool for determining quantitative phase abundances of crystalline rocks based on X-ray powder-diffraction (XRD) data. This technique is based on the refinement of physical structures with neutron diffraction data and has been applied to crystalline structures (e.g. Young et al, 1977; Hill, 1984; Baerlocher and Shicker, 1987, Raudsepp et al, 1990; Raudsepp et al, 1999). Theoretical crystallographic site-occupancies of observed minerals are compared with the XRD data to determine modal proportions of each mineral present. This technique is ideal for rocks such as these peridotite samples, as they are coarse-grained with a simple mineralogy (olivine, örtho- and clinopyroxene, and spinel).

Samples were prepared by an initial crushing in a tungsten-carbide ring-mill reducing each sample to <200 mesh; particle size was further reduced to $\leq 10 \ \mu m$ by a corundum mill. For XRD analysis, samples were pressed onto a textured glass plate in an aluminum sample holder. The surfaces of samples were 'roughened' with razor blades to break up any preferred orientation; one sample was prepared both with and without surface roughness to test the effect, if any, of preferred orientation. Rietveld analyses were performed on the Siemens (Bruker) D5000 Bragg-Brentano X-ray powder-diffractometer at the University of British Columbia. The long fine-focus Cu X-ray tube was operated at 40 kV and 40 mA, with a take-off angle of 6°. Data was obtained with a scanning step of 0.04° 20 and a

counting time of 2s per step over a range of $3-70^{\circ}$ 2 θ with each step taking 55 minutes. XRD data were refined with the Rietveld program 'Topas 2.1' (Bruker AXS).

B.2 RIETVELD RESULTS

The full results of Rietveld analyses, including estimated accuracy, are provided in Table B.1; results are normalized to 100%. Modal variation in these peridotites shows a trend from relatively fertile peridotites (lherzolite) to depleted peridotites (harzburgites, which are considered to represent residues following melting of lherzolites; Francis and Ludden, 1995). Rietveld analyses were performed on the 19 peridotite samples for which bulk and trace chemistry is available. Modal percent olivine varies from 59-81%, orthopyroxene from 15-24%, clinopyroxene from 1-16%, and spinel from <1-2%. The largest mineral variations are observed in olivine and clinopyroxene, reflecting the depleted versus fertile nature of the samples; samples with the highest amount of olivine contain the lowest amount of clinopyroxene, and therefore are the most depleted (harzburgites). In some samples, a small amount of quartz was reported (<0.3%); this appears to result from contamination during preparation for Rietveld analysis (L. Tam, pers. comm.).

Figure B.1 compares modal analyses determined using the Rietveld method with modal analyses calculated from peridotite bulk chemistry. Both methods produce very similar results for olivine abundance, and reasonably similar results for clinopyroxene. Orthopyroxene and spinel abundances, however, are routinely higher with the calculation method than the Rietveld method. These discrepancies could result from errors in either method. For instance, the mineral formulae used for orthopyroxene and spinel in the

calculation method may not accurately represent the true mineral compositions. Conversely, it is possible that development of preferred orientation in orthopyroxene and spinel grains during preparation for Rietveld analysis has resulted in inaccurate abundances for these minerals.

B.3. REFERENCES

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		¹ Modes	(%):		² Rock
Sample (MH-02-)	Ol	Орх	Срх	Sp	Type:
121	67	23	8	2	Lh
123	70	22	6	1	Lh
104	75	17	6	1	Lh
122	65	20	13	2	Lh
108	70	21	7	1	Lh
95LM1	77	20	2	1	Hz
113	80	17	2	<1	Hz
120	78	19	2	1	Hz
19	79	17	2	1	Hz
110	59	23	16	2	Lh
100	74	24	2	<1	Hz
111	61	22	15	2	Lh
19b	75	20	3	1	Hz
115b	76	20	2	· 1	Hz
118	76	19	3	1	Hz
125	79	18	1	1	Hz
114	81	15	2	1	Hz
- 117	74	22	3	1	Hz
112	76	21	2	1	Hz
Olivine (OI): orthonyroxene (Opx): clinon	vroxene	(Cnx) sn	inel (Sp)		

•

 Table B.1: Modal mineralogy of peridotite samples determined using the Rietveld method.

¹Olivine (OI); orthopyroxene (Opx); clinopyroxene (Cpx); spinel (Sp);

²Lherzolite (Lh); harzburgite (Hz)

. .



Fig. B.1



Figure B.1: Modal mineralogy of peridotite samples from Llangorse Mountain. The Rietveld method of determining modal mineralogy is compared with modal mineralogy calculated from bulk chemistry. Modal mineralogy is shown for all four major phases: (A) olivine; (B) orthopyroxene (OPX); (C) clinopyroxene (CPX); (D) spinel.

Appendix C

TRACE ELEMENT CHEMISTRY

Trace element chemistry on 19 peridotite and 4 lava samples (Llangorse Centre, Table Hill, Lone Point, and Hidden Ridge) was performed at the Memorial University of Newfoundland. The acid dissolution process (HF/HNO₃) is described in Jenner et al (1990). Samples were analyzed on the VG PQ2+S ICP-MS in St. John's, Newfoundland. Trace element chemistry data for lavas and peridotites from the Llangorse volcanic field are provided in Tables C.1. (lavas) and C.2. (peridotites). Fig. C.1. shows a primitive mantlenormalized plot (McDonough and Sun, 1995) with elements arranged in order of decreasing compatibility.

C.1. REFERENCES

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Figure C.1: Trace element chemistry of lavas and peridotites from the Llangorse volcanic field, normalized to primitive mantle values (McDonough and Sun, 1995). (A) Lavas; diamonds represent nephelinites (Lone Point, Table Hill), circles represent basanites (Llangorse locality, Hidden Ridge). (B) Peridotites from the Llangorse locality lavas.

uue Liangoise	Mountain (L	M), Lone I	Point (LP),	Table Hill	(TH), and	Hidden Ri	dge (HR).	
Sample	101	160	173	173	213	ď213		
Center	LM	LP	TH	TH	HR	HR		
Li	5.99	9.00	11.60	11.71	14.12	1.98		
Rb	17.4	60.0	19.7	25.3	20.1	16.4		
Sr	588	2285	966	1234	595	481		
Y	17.5	35.6	17.6	22.3	20.0	16.0		
Zr	156	320	264	314	144	132		
Nb	41.3	103.5	91.3	110.6	34.9	32.2		
Мо	2.46	8.19	4.44	5.46	2.16	1.94		
Cs	0.219	0.402	0.454	0.469	0.451	0.406		
Ba	425	571	689	692	364	350		
La	26.4	56.9	73.8	72.6	19.9	19.8		
Ce	56.0	124.7	152.5	158.3	42.3	41.0		
Pr	6.57	14.16	17.07	17.19	5.37	4.84		
Nd	29.0	59.7	66.6	72.9	22.9	22.8		
Sm	6.03	10.65	12.89	12.89	5.20	4.83	· · ·	
Eu	1.94	3.41	4.00	4.02	1.85	1.66		
Gd	5.27	8.74	10.07	9.94	5.29	4.63		
Tb	0.784	1.085	1.233	1.206	0.731	0.666		
Dy	4.32	5.64	5.92	6.14	4.76	. 4.15		
Но	0.767	0.833	0.873	0.855	0.770	0.699	•	
Er	1.90	1.95	1.89	1.75	1.92	1.65		
Tm	0.258	0.226	0.198	0.203	0.246	0.230		
Yb	1.53	1.09	0.93	0.98	1.39	1.33		
Lu	0.203	0.137	0.111	0.114	0.186	0.171		
Hf	4.12	6.84	7.33	7.55	3.76	3.31		
Та	2.24	4.73	5.55	6.00	1.80	1.69		
Tl	0.022	0.026	0.043	0.046	0.041	0.034		
Pb	2.09	3.27	4.75	4.96	2.68	2.09		
Bi	0.006	0.011	0.015	0.021	0.004	0.004		
Th	2.77	5.61	8.17	9.59	2.65	2.03		
U	0.991	1.982	2.671	2.902	0.758	0.611		

Table C.1: Trace element chemical compositions of samples (MH-02-), including replicate (^dduplicate) analyses, of lavas from the Llangorse volcanic field. All results in ppm. The localities include Llangorse Mountain (LM), Lone Point (LP), Table Hill (TH), and Hidden Ridge (HR).

uupiicate) analyst	es, or per	nuonies	nom ui	- Liangu	ise voic	ame nei	u. All le	suns m	ppm.		
Sample	19	19b	100	104	108	110	111	112	113	114	115b	^d 115b
Li	4.05	3.03	n/a	3.48	1.95	1.73	2.02	3.29	3.51	3.09	2.47	2.37
Rb	1.414	0.271	0.189	2.342	0.139	0.100	0.473	0.252	1.145	0.549	0.592	0.518
Sr	18.63	6.37	3.41	18.02	3.94	2.98	13.08	7.41	16.75	23.36	8.93	6.92
Y	0.692	0.152	0.137	0.923	0.935	2.201	2.720	0.133	0.269	0.251	0.267	0.249
Zr	10.85	0.82	1.09	2.68	1.21	2.30	5.14	: 3.21	2.72	2.36	. 1.76	1.70
Nb	1.15	0.25	0.24	1.15	0.02	0.06	0.04	0.36	0.87	0.61	0.50	0.44
Мо	0.109	0.066	0.071	0.238	0.339	0.116	0.064	0.075	0.095	0.087	0.086	0.084
Cs	0.014	0.014	0.011	0.019	0.013	0.011	0.024	0.014	0.019	0.005	0.032	0.029
Ba	6.43	1.16	1.18	19.79	0.10	0.47	0.86	0.86	4.87	`3.22	<u>5</u> .35	3.96
La	1.36	0.25	0.34	2.75	0.04	0.12	0.16	0.61	0.62	0.55	0.55	0.54
Ce	2.92	0.44	0.58	5.05	0.07	0.14	0.53	1.23	1.35	0.85	0.95	0.86
Pr	0.387	0.052	0.070	0.569	0.030	0.032	0.116	0.144	0.143	0.103	0.132	0.104
Nd	1.35	0.16	0.27	1.96	0.14	0.20	0.67	0.59	0.53	0.39	0.43	0.42
Sm	0.252	0.036	0.057	0.336	0.062	0.128	0.277	0.106	0.105	0.068	0.085	0.082
Eu	0.058	0.006	0.012	0.078	0.022	0.051	0.106	0.024	0.017	0.015	0.020	0.018
Gd	0.208	0.025	0.039	0.237	0.085	0.222	0.381	0.055	0.077	0.054	0.071	0.063
Tb	0.025	0.004	0.005	0.033	0.019	0.051	0.074	0.007	0.012	0.008	0.010	0.009
Dy	0.142	0.025	0.029	0.198	0.156	0.389	0.529	0.034	0.067	0.044	0.054	0.050
Но	0.025	0.006	0.005	0.035	0.035	0.089	0.102	0.005	0.010	0.008	0.009	0.009
Er	0.067	0.023	0.016	0.103	0.104	0.288	0.346	0.018	0.035	0.018	0.027	0.029
Tm	0.010	0.009	0.007	0.015	0.024	0.046	0.051	0.012	0.009	0.003	0.009	0.012
Yb	0.063	0.030	0.018	0.100	0.118	0.279	0.307	0.014	0.024	0.022	0.027	0.026
Lu	0.011	0.005	0.003	0.015	0.016	0.042	0.045	0.002	0.003	0.003	0.004	0.003
Hf	0.429	0.029	0.026	0.132	0.047	0.111	0.256	0.066	0.050	0.155	0.057	0.042
Та	0.096	0.068	0.048	0.070	0.030	0.027	0.019	0.062	0.055	0.060	0.059	0.055
Tl	0.002	0.009	0.006	0.004	0.011	0.005	0.000	0.008	0.009	0.003	0.005	0.009
Pb	0.102	0.091	0.126	0.224	0.103	0.203	0.124	0.059	0.161	0.092	0.100	0.110
Bi	0.003	0.012	0.007	0.002	0.011	0.009	0.003	0.010	0.006	0.002	0.007	0.007
Th	0.143	0.037	0.045	0.267	0.008	0.014	0.022	0.073	0.088	0.071	0.068	0.071
U	0.032	0.017	0.017	0.057	0.008	0.026	0.029	0.024	0.024	0.018	0.021	0.025

Table C.2: Trace element chemical compositions of samples (MH-02-), including replicate (^ssplit and ^dduplicate) analyses, of peridotites from the Llangorse volcanic field. All results in ppm.

•	Table C.	2, cont'd	:									
•	Sample	117b	118	120	121	°121b	122	123	125	^d 125	95LM1	^{\$95} LM1
	Li	3.64	3.22	1.50	1.55	1.56	1.53	1.72	2.26	1.88	1.57	1.24
	Rb	0.374	0.611	0.100	0.047	0.047	0.190	0.137	0.166	0.140	0.099	0.082
	Sr	9.85	13.71	1.24	1.58	1.50	4.02	1.99	3.03	2.78	1.80	1.82
	Y	0.295	0.264	0.078	0.862	0.956	1.328	0.471	0.108	0.082	0.034	0.034
	Zr	2.56	1.61	1.10	0.21	0.33	0.69	0.11	0.81	0.80	0.79	0.39
	Nb	0.33	1.21	0.09	0.04	0.03	0.03	0.02	0.24	0.23	0.12	0.12
	Мо	0.073	0.139	0.080	0.071	0.075	0.238	0.077	0.055	0.066	0.075	0.088
	Cs	0.052	0.013	0.022	0.009	0.012	0.027	0.016	0.009	0.006	0.015	0.015
	Ba	0.96	4.31	-0.23	-0.05	-0.21	0.34	-0.04	1.28	0.25	-0.46	-0.48
	La	0.51	1.37	0.08	0.03	0.03	0.12	0.03	0.18	0:18	0.13	0.15
	Ce	1.01	2.35	0.14	0.03	0.01	0.19	0.02	0.27	• 0.28	0.27	0.28
	Pr	0.147	0.265	0.014	0.008	0.013	0.022	0.013	0.036	·0.036	0.030	0.032
	Nd	0.57	0.93	0.06	0.01	0.02	0.11	0.02	0.13	0.13	0.09	0.11
	Sm	0.118	0.150	0.010	0.015	0.020	0.052	0.013	0.024	0.019	0.012	0.022
	Eu	0.030	0.034	0.002	0.006	0.005	0.024	0.000	0.002	0.002	0.002	0.002
	Gd	0.088	0.101	0.009	0.049	0.049	0.126	0.017	0.017	0.014	0.009	0.008
	Tb	0.011	0.012	0.001	0.013	0.014	0.028	0.005	0.002	0.002	0.001	0.001
	Dy	0.062	0.058	0.011	0.116	0.131	0.218	0.060	0.015	0.012	0.006	0.005
	Ho	0.010	0.009	0.003	0.032	0.035	0.053	0.016	0.003	0.002	0.001	0.001
	Er	0.030	0.027	0.011	0.114	0.123	0.168	0.068	0.012	0.009	0.004	0.007
	Tm	0.008	0.010	0.002	0.024	0.025	0.026	0.019	0.004	0.004	0.001	0.006
	Yb	0.029	0.020	0.016	0.144	0.160	0.181	0.090	0.022	0.019	0.009	0.008
	Lu	0.004	0.002	0.003	0.023	0.023	0.029	0.015	0.003	0.002	0.001	0.001
	Hf	0.100	0.065	0.110	0.019	0.015	0.106	0.012	0.020	0.015	0.097	0.014
	Та	0.055	0.071	0.031	0.040	0.035	0.025	0.044	0.051	0.051	0.054	0.053
	Tl	0.005	0.007	-0.002	0.006	0.008	0.008	0.007	0.002	0.000	-0.003	0.002
	Pb	0.109	0.129	0.068	0.081	0.116	0.039	0.045	0.055	0.041	0.027	0.049
	Bi	0.005	0.007	0.001	0.007	0.011	0.028	0.011	0.004	0.004	0.001	0.006
	Th	0.051	0.131	0.029	0.008	0.008	0.012	0.002	0.023	0.020	0.025	0.019
	U	0.017	0.041	0.004	0.004	0.008	0.004	0.005	0.009	0.007	0.003	0.005

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Fig. C.1.

APPENDIX D

STATISTICS ON MAJOR ELEMENT CHEMISTRY

Major element chemistry was performed at the McGill University Geochemical Laboratory. Results were obtained throuth X-ray flourescence on the PW2400 spectrometer. Accuracy for silica is within 0.5%, and for all other major elements is within 1%.

Table D.1: Statistics on major element contents of lavas (Sample MH-02-101) from the Llangorse volcanic field.

Sample	101	'101b	² 101c	² 101d			
Center	Site 1	Site 1	Site 1	Site 1	mean	variance	1≥
SiO2	45.13	44.99	44.91	45.00	45.01	0.0083	0.091
TiO2	2.23	2.21	2.20	2.22	2.22	0.0002	0.016
Al2O3	13.12	13.13	13.15	13.18	13.15	0.0007	0.026
Fe2O3	2.23	2.59	2.80	2.73	2.59	0.0644	0.254
FeO	10.09	9.68	9.59	9.59	9.74	0.0570	0.239
MnO	0.18	0.18	0.18	0.18	0.18	0.0000	0.002
MgO	11.69	11.61	11.65	11.68	11.66	0.0013	0.036
CaO	9.96	9.96	9.94	9.99	9.96	0.0004	0.021
Na2O	3.14	3.12	3.11	3.15	3.13	0.0003	0.018
К2О	1.23	1.22	1.22	1.22	1.22	0.0000	0.005
P2O5	0.50	0.50	0.50	0.51	0.50	0.0000	0.003
Total •	99.51	99.19	99.25	99.45		•	
						•	

¹Split.

²Replicate.

Table D.2: Statistics on major element contents of peridotites (Sample MH-02-121)

 from the Llangorse volcanic field.

	0						
Sample	121	'121b	² 121c	² 121d	mean	variance	1≥
SiO2	43.98	44.01	43.95	43.94	43.97	0.0010	0.032
TiO2	0.029	0.029	0.030	0.050	0.03	0.0001	0.010
Al2O3	2.18	2.26	2.31	2.60	2.34	0.0335	0.183
Fe2O3	2.02	1.59	1.91	1.99	1.88	0.0389	0.197
FeO	6	6.35	6.09	6.62	6.27	0.0780	0.279
MnO	0.121	0.119	0.119	0.127	0.12	0.0000	0.004
MgO	42.73	42.43	42.39	40.82	42.09	0.7427	0.862
CaO	2.09	2.04	2.16	2.77	2.27	0.1158	0.340
Na2O	0.109	0.062	0.093	0.110	0.09	0.0005	0.022
K2O	0.01	0.01	0.01	0.01	0.01	0.0000	0.000
P2O5	0.009	0.009	0.009	0.011	0.01	0.0000	0.001
Total	99.28	98.91	99.07	99.05			

'Split.

²Replicate.

Appendix E

MICROPROBE ANALYSES OF BASANITE GROUNDMASS MINERALS

Microprobe analyses of basanite groundmass minerals from the lava bluff, the chilled base of the lava flow, and a basanite clast from the underlying debris flow sediments are provided in Appendix E.2 (electronic; see attached cd). Fig. E.1. compares mineral chemistry of pyroxene and olivine grains in basanite from these three different lava regions. As Fig. E.1 shows, pyroxene and olivine grains from the lava bluff, the chilled base of the lava flow, and the underlying debris flow sediments appear to be chemically similar. However, the mineral compositions provided only demonstrate that basanite clasts from the debris flow sediments are similar in chemistry to the overlying lava flow. In order to determine if the basanite clasts and overlying lava flow formed from the same parent magma, bulk and trace chemistry of basanite clasts from the debris flow sediments would be necessary.

Figure E.1: Mineral chemistry of clinopyroxene (A, B) and olivine (C) grains in basanite lava from the lava bluff, the chilled base of the lava flow, and a basanite clast from the debris flow sediments underlying the lava flow. (A) Cations of Ti plotted against cations of Al in clinopyroxene grains. (B) Cations of Ti plotted against cations of Na in clinopyroxene grains. (C) Cations of Ni plotted against the magnesium number (Mg#) in olivine grains.







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