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Eocene melting of Precambrian lithospheric mantle: Analcime-bearing volcanic rocks from the Challis–Kamloops belt of south central British Columbia

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Abstract

Potassic silica-undersaturated mafic volcanic rocks form a minor portion of the predominantly calc–alkaline Eocene Challis–Kamloops volcanic belt, which extends from the northwestern United States into central British Columbia (Canada). Their major occurrence is in the Penticton Group in south central British Columbia, where they reach a thickness of up to 500 m and form the northwestern edge of the Montana alkaline province. These analcime-bearing rocks ($\sim$53–52 Ma old) are typically rhomb porphyries of ternary feldspar ($An_{28}Ab_{52}Or_{20}$). Additional phenocryst phases include clinopyroxene, analcime, phlogopite and rare olivine. The rocks are characterized by high total alkalis, particularly K_2O (> 4.5 wt%) as well as by a distinct enrichment of large-ion lithophile elements versus heavy rare-earth elements and high-field-strength elements. They have unusual isotopic compositions compared to most other rocks of the Challis–Kamloops belt, particularly high negative ε_{Nd} values and elevated but relatively uniform initial $^{87}Sr/^{86}Sr$ ratios ($\sim$0.7065). The potassic silica-undersaturated rocks overlie Precambrian crust and lithosphere and were at least in part derived from ancient metasomatized subcontinental mantle lithosphere, which was modified in a Precambrian subduction setting. The alkaline rocks of the Challis–Kamloops belt are related to a slab-window environment. In particular, they were formed above the southern edge of the Kula plate adjacent to the Kula–Farallon slab window, whereas the Montana alkaline province situated well to the southeast was formed directly above the Kula–Farallon slab window. Upwelling of the hotter asthenospheric mantle may have been the thermal trigger necessary to induce melting of fertile and metasomatized lithospheric mantle.
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1. Introduction

Potassic mafic volcanic rocks, including leucite- or analcime-bearing lavas, are small in volume and occur only rarely in the geological record. In recent years, they have attracted a great deal of interest, in part because of their unusual mineralogy and chemical composition. The origin of the rocks is still controversial, although they can provide additional insights into the compositional variability of subcontinental lithospheric mantle and the processes that cause these variations. Potassic silica-undersaturated mafic rocks are well exposed in south central British Columbia (Daly, 1912; Church, 1973), where they form a part of a belt of Early to Middle Eocene igneous rocks, which spreads from the northwestern United States through British Columbia (Challis–Kamloops belt of Souther, 1991; Fig. 1) and continues northward to the southern Yukon and Alaska (Coast Plutonic Complex). Igneous rocks of the Challis–Kamloops belt vary from calc–al-
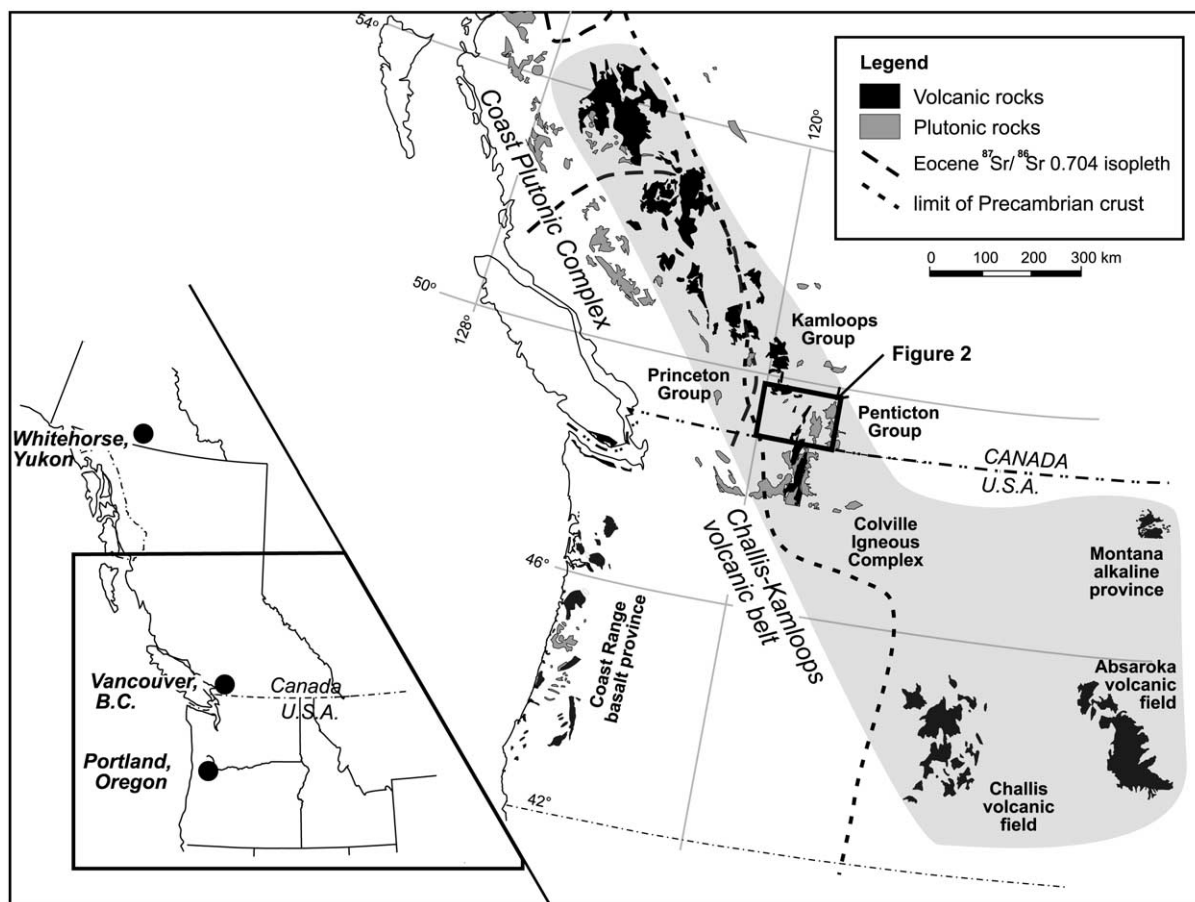

Fig. 1. Distribution of Eocene igneous rocks in the Canadian Cordillera and adjacent United States, with emphasis on central and southern British Columbia (modified after Wheeler and McFeely, 1991). The dashed $^{87}Sr/^{86}Sr = 0.704$ isopleth (delimiting lower (primitive) values to its west and higher (evolved) values to its east) shows regional variation in Sr isotope values from Eocene rocks in south to central British Columbia (from Armstrong, 1988, except in southernmost British Columbia, where the isopleth was subsequently revised by Ghosh, 1995). The approximate extent of Precambrian crust/lithosphere (dotted line) is from Armstrong and Ward (1991), who defined the line from Mesozoic to Cenozoic Sr isotope values throughout the Cordillera. Inset on left shows location of the map on the west coast of North America. Rectangle in the map indicates the location of Fig. 2. Shaded area represents the extent of the Challis–Kamloops volcanic belt.

kaline to alkaline (Souther, 1991) and include units of both subduction-related and within-plate affinity (e.g. Ewing, 1980; Dudás, 1991). The proposed tectonic settings for various parts of the belt range from volcanic arc (Ewing, 1981a,b) and rifted arc (Dostal et al., 1998, 2001) to slab window (Thorkelson and Taylor, 1989; Breitsprecher et al., 2003) to extended continental crust (Morris et al., 2000).

The purpose of this paper is to present geochemical, isotopic, petrographic and K–Ar geochronological data on Eocene potassic silica-undersaturated volcanic rocks of the lower part of the Penticton Group (Yellow Lake Member of the Marron Formation) in south central British Columbia (Fig. 1). Additionally, this paper evaluates the processes involved in the geochemical evolution of these rocks, constrains the composition of their sources and proposes a tectonic model for their origin and emplacement. This information complements the original geological mapping and physical volcanology of Church (1973). The results of these investigations provide evidence for a slab-window-modified subduction environment near the British Columbia–United States border and highlight the role of ancient subcontinental mantle lithosphere in the genesis of potassic silica-undersaturated mafic rocks.

2. Geological setting

Most of the western Canadian Cordillera consists of terranes that accreted to the continental margin of Ancestral North America during the Mesozoic (Monger et al., 1982; Armstrong, 1988). Throughout and after these compressional events, much of this region and adjacent parts of the United States were affected by magmatism, with a voluminous and concentrated pulse of activity from Early to Middle Eocene. This Eocene magmatic event is recorded by the large semi-continuous volcanic successions of the Challis–Kamloops belt (Fig. 1), which were extruded in the final stages of Kula–Farallon ridge subduction (Engebretson et al., 1985), between ca. 55 and 45 or possibly as late as 42 Ma (e.g. Pearson and Obradovich, 1977; Ewing, 1980; Morris et

al., 2000). The Challis–Kamloops volcanic belt and the Tertiary component of the longer-lived Coastal Plutonic Complex to its west (Fig. 1) were both affected by strike-slip and extensional faulting, before and throughout the Eocene magmatic event (e.g. Ewing, 1980; Tempelman-Kluit and Parkinson, 1986; Parrish et al., 1988).

In southern British Columbia, successions of the Challis–Kamloops belt have been divided into three groups (Fig. 1): the typically calc–alkaline Kamloops and Princeton groups, and the calc–alkaline to alkaline Penticton Group. The Kamloops Group (Ewing, 1981a,b) and Princeton Group (Souther, 1991; Read, 2000) volcanic successions (Fig. 1) each exceed 1500 m in thickness and form a continuous series from basalt to rhyolite. The Kamloops Group is dominated by intermediate to mafic rocks, while the Princeton Group is typified by the more felsic lithologies. The Penticton Group (Church, 1973, 1985) occurs along the most inboard margin of the Challis–Kamloops belt in British Columbia (Fig. 1), east of the Princeton Group and southeast of the main axis of the Kamloops Group. The Penticton succession has an aggregate thickness of 2500 m. Eocene volcanic rocks throughout the region are typically preserved in grabens, half-grabens and strike-slip fault-bounded basins (Ewing, 1981a; Church, 1985). Ewing (1980) suggested that the arrangement of the volcanic units is typical of an arc setting, which exhibits an increase in alkalinity and potassium with distance from the trench, in this case from southwest to northeast.

The most comprehensive section of the Penticton Group occurs at the type-area south of Penticton (Fig. 2) in the White Lake basin (Church, 1973). The extensional White Lake basin is 23 km × 16 km in size, and is bounded to the east by the Okanagan Valley Fault, a major crustal detachment of Eocene age (Parrish et al., 1988; Tempelman-Kluit and Parkinson, 1986). The Eocene strata in the basin rest non-conformably on Cretaceous and Early Tertiary (Paleocene) granitoids and metamorphosed Mesozoic sedimentary and volcanic rocks. Across its eastern fault boundary is high-grade Precambrian Monashee gneiss containing micas with Eocene cooling
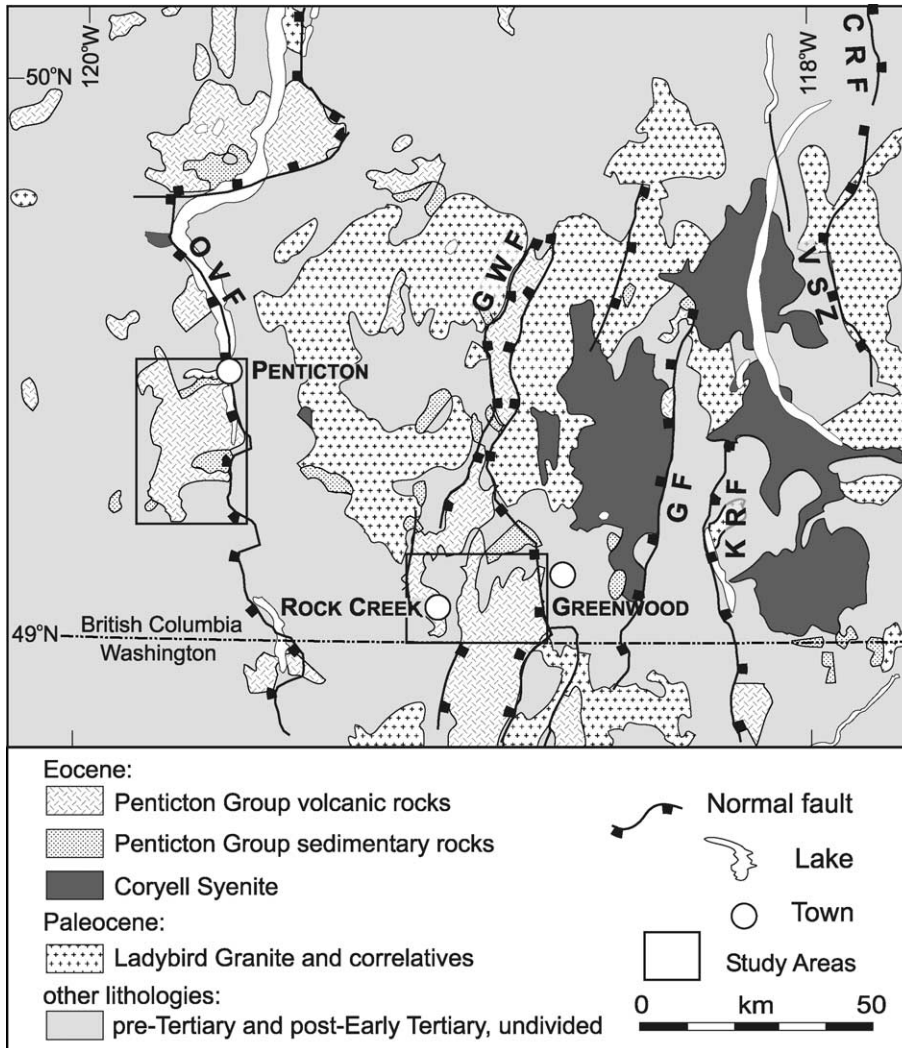
Fig. 2. Distribution of Penticton Group volcanic and related igneous rocks as well as sedimentary rocks in the vicinity of the NTS 82E (Penticton) map-sheet (modified after Tempelman-Kluit, 1989 and Wheeler and McFeely, 1991). OVF, Okanagan Valley Fault; GF, Granby Fault; GWF, Greenwood Fault; KRF, Kettle River Fault; VSZ, Valhala Shear Zone; CRF, Columbia River Fault.

ages (Parrish et al., 1988). Geological and geophysical studies indicate that attenuated Precambrian crust lies in the subsurface beneath the Penticton Group (Clowes et al., 1998). The Penticton Group occurs in other isolated areas on the Penticton map-sheet (Fig. 2), including the Rock Creek outlier and the north end of the Toroda graben, southwest of Greenwood. Rocks from these outliers are located approximately 60 km ESE of the White Lake basin and are, in addition

to the White Lake basin (Penticton Tertiary oulier), also included in the present study.

The Penticton Group was divided by Church (1973) into five formations. The lowermost of these is the Springbrook Formation, a discontinuous basal polymictic conglomerate unit up to 230 m thick containing clasts of the underlying pre-Tertiary basement rocks and those of the adjacent eroded highlands. Its equivalent in the Greenwood area is the lower part of the Kettle

River Formation of Daly (1912). It is overlain by alkaline to subalkaline lavas and pyroclastic rocks (mainly trachyandesites, trachytes and phonolites) of the Marron Formation. The Marron Formation represents a period of intense volcanism during which thick lava flows and pyroclastic activity produced more than 1500 m of volcanic strata (Church, 1973) including its basal unit, alkaline lavas of the Yellow Lake Member, which are the focus of this paper. The unconformably overlying Marama Formation is composed of as much as 300 m of dacites to rhyolites. This unit is overlain by pyroclastic rocks and lacustrine to fluvial sediments of the White Lake Formation, which reaches a maximum thickness of 1300 m. The uppermost part of the Tertiary sequence is represented by conglomerates and epiclastic volcanic breccia of the Skaha Formation, which reaches 300 m in thickness. Slide breccia and coarse fanglomerate deposits of this uppermost formation are the product of pronounced post-magmatic normal faulting, particularly along the eastern edge of the formation and basin (Church, 1973). The Tertiary fill of the White Lake basin was intruded by several generations of dioritic, lamprophyric and augite-phyric dikes and sills, some of which were probably feeders to the Eocene volcanic members.

In summary, development of the White Lake basin began with extensional faulting, followed shortly by extrusions of appreciable volumes of mafic alkaline to subalkaline mafic magmas (Marron Formation). The next significant phase was felsic magmatism, signifying extensive fractionation, crustal contamination, crustal melting or some combination thereof (Marama Formation). The final phase was post-magmatic normal faulting, suggesting that the extensional tectonics persisted beyond the initial extensional stage.

The Penticton Group volcanic successions straddle the Canada–United States international border into northeastern Washington, where they are assigned to the Paleocene to Eocene Colville Igneous Complex (Pearson and Obradovich, 1977; Carlson et al., 1991; Morris et al., 2000). Units that appear to be correlatives of the Penticton Group include the sedimentary to volcaniclastic O'Brien Creek Formation (correlative to White Lake Formation), the felsic lavas of the Sanpoil Formation (Marama Formation), and volcaniclastic breccias and sediments of the lower Klondike Mountain Formation (lower Skaha Formation) (Pearson and Obradovich, 1977; Carlson and Moye, 1990). The upper Klondike Mountain Formation consists of a diverse suite of lavas ranging from basalt to rhyolite, for which there is no direct equivalent in the Penticton Group. Conversely, the Marron Formation has no correlative in the Colville sequence, because it pinches out less than 10 km south of the international border, where it is included as an unnamed unit at the base of the Sanpoil Formation (Pearson and Obradovich, 1977).

3. Marron Formation

The Marron Formation consists mainly of lava flows 3–60 m thick, locally, accompanied by pyroclastic rocks. This study focuses on rocks of the Yellow Lake Member, which is the oldest and most alkaline unit of the Marron Formation. In the White Lake basin, the Yellow Lake Member varies in thickness from 200 to 500 m (Church, 1973) and in the Greenwood/Rock Creek area, it comprises as many as 20 lava flows, locally attaining a composite thickness of 500 m (Church, 1985). The Yellow Lake Member underlies a cumulative area of about 600 km^2. It is distinct from all of the subsequent members, which are only mildly alkaline to subalkaline rocks and are similar to most of the Challis–Kamloops volcanic rocks. Normal-faulting was coincident with, and continued after, the Marron volcanism (Church, 1973; Tempelman-Kluit and Parkinson, 1986).

The Marron Formation is considered to be the volcanic equivalent of the coeval Coryell syenitic plutons (Church, 1973; Souther, 1991). The Coryell plutons that crop out just east of the study area, in the southern Okanagan region of south central British Columbia (near the Canada–United States border) are high-level, syenitic intrusions unroofed by the genetically related Early Tertiary extensional faulting (Tempelman-Kluit and Parkinson, 1986; Parrish et al., 1988).

4. Petrography and mineral chemistry

The Yellow Lake Member is predominantly composed of massive to amygdaloidal lavas; the amygdules are filled by calcite and zeolites, particularly natrolite, analcime and thomsonite. The rocks are typically porphyritic (especially in the lower part of the unit) or microporphyritic (in the upper parts of the unit) with phenocrysts of ternary feldspar ($An_{28}Ab_{52}Or_{20}$), clinopyroxene and analcime set in a holocrystalline matrix (Table 1). These unusual ternary feldspars resemble anorthoclase with gridiron twinning and oscillatory zoning. In the porphyritic rocks, ternary feldspar forms distinctive rhomb-shaped phenocrysts up to 2 cm in size (5–15 vol%) accompanied by euhedral prismatic phenocrysts of clinopyroxene (10–20 vol%) up to 5 mm long, minor phlogopite and rare Fe–Ti oxide and apatite. The groundmass is composed of alkali feldspars, clinopyroxene, phlogopite, apatite microlites, Fe–Ti oxides and glass. In the microporphyritic types, ternary feldspar (15–25 vol%) and subhedral stubby clinopyroxene ($\sim$5 vol%) are typically $<$2 mm in size and have the same groundmass assemblage as the porphyritic types. Patches of serpentine-like material with relics of fresh olivine ($Fo_{\sim74}$) are present in some rocks. Phenocrysts of phlogopite with Mg/Mg+Fe$\sim$0.76 (Table 1) occur most commonly as strongly pleochroic (brown–orange brown) tabular euhedral to subhedral crystals reaching up to 2 mm. The phlogopite is distinctly titaniferous ($\sim$7.4 wt% TiO_2) and rich in BaO (Table 1). Clinopyroxene is slightly pleochroic (pale to light green) and is in the diopsidic augite compositional range (Table 1) with an average composition of $Wo_{46}En_{42}Fs_{12}$. It displays well-developed oscillatory and sector zoning. Compared

Table 1A

The average compositions of mineral phases from the volcanic rocks of the Yellow Lake Member

	Olivine		Phlogopite		Analcime		Clinopyroxene				Sanidine		Plagioclase		Ternary feldspar	
							Yellow Lake		Buck Creek							
n:	12		8		11		63		99		4		6		29	
	Average	S.D.	Average	S.D.	Average	S.D.	Average	S.D.	Average	S.D.	Average	S.D.	Average	S.D.	Average	S.D.
SiO_2 (wt%)	38.36	0.22	36.79	1.22	54.95	0.43	50.11	0.73	50.88	1.05	60.12	1.97	59.03	1.12	58.14	0.79
TiO_2			7.37	0.41			1.09	0.15	1.09	0.36						
Al_2O_3			13.91	0.78	22.08	0.37	4.26	0.76	3.01	0.75	21.54	0.44	24.46	0.77	24.62	0.54
FeO	21.93	0.94	9.65	1.02	0.38	0.35	7.43	0.42	9.12	0.99	0.19	0.24	0.12	0.19	0.40	0.10
MnO	0.66	0.12					0.08	0.14	0.20	0.13						
MgO	38.62	0.75	17.19	0.99	0.07	0.23	14.23	0.56	14.99	1.02						
CaO	0.23	0.04			0.17	0.30	22.21	0.43	20.21	0.69	1.68	0.08	5.27	0.85	5.26	0.49
BaO			2.05	0.83							4.17	1.56	1.43	0.31	1.79	0.38
Na_2O	0.10	0.14	0.84	0.06	12.18	0.60	0.87	0.08	0.52	0.09	4.30	0.78	6.44	0.31	5.46	0.33
K_2O			8.48	0.54	0.30	0.14					7.22	0.43	1.75	0.45	3.08	0.68
Σ	99.90		96.28		90.14		100.28		100.04		99.23		98.50		98.75	
X_{Fe}	0.24	0.01	0.24	0.03												
X_{Mg}	0.76	0.01	0.76	0.03												
Wo							46.4	0.8	41.8	1.6						
En							41.3	1.3	43.1	2.5						
Fs							12.3	0.8	15.1	1.8						
Or											47.8	4.7	10.9	2.8	19.5	4.2
Ab											42.9	5.3	61.3	2.1	52.5	3.0
An											9.4	0.9	27.7	4.4	28.0	2.6

n, number of analyses; S.D., standard deviation. End-member components: Wo, wollastonite; En, enstatite; Fs, ferrosilite; Or, orthoclase; Ab, albite; An, anorthite. Buck Creek – Eocene high-K calc–alkaline lavas; central British Columbia – Dostal et al. (2001). Typical modal composition of the Yellow Lake volcanic rocks (Church, 1973): feldspars 50%; analcime 15%; clinopyroxene 25%; biotite 4%; Fe–Ti oxides 4%; apatite 2%.

Table 1B
Representative composition of mineral phases from the volcanic rocks of the Yellow Lake Member

(wt%)	Olivine		Phlogopite		Analcime		Clinopyroxene				Sanidine		Plagioclase	Ternary feldspar	
	1	2	1	2	1	2	1	2	3	4	1	2	1	1	2
SiO_2	38.55	38.15	36.61	37.82	55.24	54.15	49.93	49.82	50.53	50.21	58.55	58.38	57.84	58.75	58.58
TiO_2	0.00	0.00	7.39	7.42	0.00	0.00	1.08	1.04	1.00	1.19	0.00	0.00	0.00	0.00	0.00
Al_2O_3	0.00	0.00	14.28	12.88	21.97	22.45	4.11	4.34	3.92	4.37	21.76	21.99	25.24	24.00	24.00
FeO	22.12	22.27	10.27	8.76	0.47	0.00	7.05	7.15	7.55	7.69	0.00	0.30	0.37	0.32	0.38
MnO	0.58	0.78	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
MgO	38.58	38.12	15.82	18.25	0.00	0.00	14.38	14.20	14.12	13.88	0.00	0.00	0.00	0.00	0.00
CaO	0.22	0.22	1.68	0.00	0.00	0.45	22.40	22.40	22.06	21.86	1.72	1.75	6.12	4.93	5.44
BaO	0.00	0.00	0.00	1.51	0.00	0.00	0.00	0.00	0.00	0.00	5.67	5.36	2.00	1.60	1.24
Na_2O	0.00	0.00	0.75	0.86	12.14	13.16	0.76	0.72	0.97	0.99	3.72	3.65	6.01	5.14	5.46
K_2O	0.00	0.00	9.25	8.67	0.18	0.62	0.00	0.00	0.00	0.00	7.27	7.44	1.43	4.23	3.66
Σ	100.05	99.54	96.06	96.17	90.00	90.83	99.71	99.67	100.15	100.19	98.69	98.87	99.01	98.97	98.76
X_{Fe}	0.24	0.26	0.25	0.21											
X_{Mg}	0.76	0.74	0.75	0.79											
Wo								46.7	46.9	46.3	46.3				
En								41.8	41.4	41.3	40.9				
Fs								11.5	11.7	12.4	12.7				
Or												50.6	51.4	9.1	26.1	22.1
Ab												39.4	38.4	58.1	48.3	50.2
An												10.0	26.8	32.7	25.6	27.7

Olivine: sample 13A; 1 – core, 2 – rim; phlogopite: 1 – sample 39, 2 – sample 116; analcime: euhedral crystals in groundmass; 1 – sample 144, 2 – sample 248; clinopyroxene: sample 157: 1 – core, 2 – rim; sample 92A: 3 – core, 4 – rim; sanidine: 1 – sample 248, 2 – sample 39; plagioclase: sample 92A; ternary feldspar: sample 244, 1 – core, 2 – rim.

to clinopyroxene from the high-K, calc–alkaline volcanic rocks in the northernmost part of the Challis–Kamloops belt (e.g. Dostal et al., 1998, 2001), this Yellow Lake clinopyroxene is higher in Ca and lower in Mg and Fe (Table 1).

Phenocrysts of ternary feldspar (Table 1) have elevated Ba abundance along with high An component ($\sim$ 28), suggesting high-T crystallization (Nekvasil, 1990, 1992). Plagioclase (andesine-oligoclase) and sanidine (with high An contents) mantle ternary feldspar or occur in groundmass as microphenocrysts. Alkali feldspars, particularly sanidine, are also rich in Ba. Analcime mostly occurs in the matrix, where it forms rounded or polyhedral crystals about 0.1 mm in size. In some samples, analcime is recognizable as a phenocryst phase as large as 5 mm in size and reaches 15 vol%. Minor minerals include Fe–Ti oxides and apatite, which occur in both the groundmass and as inclusions in the phenocrysts, including within the analcime.

Some of these analcime-bearing lavas from the Rock Creek area were described and named shackanite by Daly (1912). Shackanite is a vesicular lava with augite and rhomb-shaped feldspar phenocrysts also containing euhedral crystals of analcime with sharp outlines in the groundmass. The analcime-bearing alkaline lavas have petrographic similarities with Eocene potassic volcanic rocks from the Highwood Mountains of the Montana alkaline province (e.g. Pirsson, 1905; Larsen, 1940; O'Brien et al., 1991).

The origin of the analcime phenocrysts in volcanic rocks is controversial. Two main hypotheses are either primary analcime or replacement pseudomorphs of leucite (e.g. Pearce, 1993; Karlsson and Clayton, 1993; O'Brien et al., 1988; Luhr and Kyser, 1989). In many cases in the Yellow Lake lavas, analcime appears to be formed during some late-stage magmatic reactions, possibly as a pseudomorph of leucite (although leucite relics have not been found). The analcime is fresh, and commonly contains primary inclusions of other mineral phases, particularly apatite, Fe–Ti oxides, pyroxene and glass. Church (1978, 1979) inferred the following crystallization sequence: clinopyroxene

Table 2
Major and trace-element analyses of volcanic rocks of the Yellow Lake Member

Sample:	13A	39	71	83	84	92A	116	144	157	167	226	230	244	248	260
SiO_2 (wt%)	51.50	51.95	52.84	52.61	53.55	52.91	47.95	53.07	51.14	52.30	52.87	52.47	50.82	56.41	52.43
TiO_2	0.91	0.92	0.87	0.81	0.71	0.82	0.99	0.94	1.00	0.85	0.87	0.98	0.96	0.76	0.85
Al_2O_3	17.23	18.69	17.36	18.08	19.53	18.82	16.94	18.32	17.11	19.02	19.05	14.96	17.24	19.69	18.93
Fe_2O_3*	6.65	5.60	6.44	5.91	4.76	5.81	6.85	5.79	7.17	5.74	5.95	6.70	7.09	4.50	5.74
MnO	0.13	0.06	0.11	0.11	0.09	0.11	0.12	0.10	0.14	0.10	0.10	0.10	0.11	0.07	0.10
MgO	3.07	1.90	3.01	2.56	1.49	1.95	3.09	2.39	3.01	2.05	2.10	4.75	2.90	1.61	2.26
CaO	4.86	3.89	6.30	5.23	4.01	4.94	7.42	4.93	6.83	4.51	4.60	7.69	6.46	4.22	4.56
Na_2O	5.40	4.10	3.45	4.31	5.80	4.17	2.83	4.11	4.58	5.44	5.56	3.56	3.76	4.67	4.87
K_2O	4.47	6.12	5.49	5.70	5.09	5.99	6.25	5.83	4.62	5.73	5.77	4.66	5.82	5.61	6.03
P_2O_5	0.91	0.74	0.80	0.68	0.46	0.68	0.96	0.74	1.01	0.65	0.65	1.01	1.00	0.40	0.66
LOI	4.80	4.50	3.10	5.80	3.40	2.40	5.90	2.00	3.50	2.50	2.30	2.70	3.40	2.80	2.80
Total	99.93	98.47	99.77	101.80	98.89	98.60	99.30	98.22	100.11	98.89	99.82	99.58	99.56	100.74	99.23
Cr (ppm)	20	22	29	19	0	17	41	6	16	0	6	95	14	0	9
Ni	19	27	25	19	7	21	33	13	24	10	10	73	23	11	11
V	107	112	97	90	61	97	131	99	107	88	94	130	100	67	99
Cu	47	92	47	43	42	51	50	75	53	54	56	50	65	38	53
Pb	41	43	37	48	53	52	44	50	45	50	51	41	47	43	48
Zn	87	75	79	79	82	94	101	86	91	85	95	82	102	75	94
Rb	93	157	145	80	66	136	155	130	66	162	155	132	145	104	161
Ba	3772	5752	4469	3805	4499	3959	5799	4623	4247	4164	4360	3468	4378	4156	4489
Sr	2617	5060	3558	3887	6467	6422	3483	4021	3647	4237	4183	2661	5237	3896	4409
Nb	117.0	104.0	106.0	116.0	130.0	121.0	93.0	124.0	117.0	128.0	133.0	83.0	114.0	106.0	129.0
Hf	7.82	6.76	7.71	7.76	7.90	7.99	8.06	7.37	8.06	7.57	8.57	7.67	7.99	7.36	7.60
Zr	415	367	406	433	448	434	382	388	373	416	449	361	404	382	411
Y	25	19	22	21	19	22	25	22	25	22	22	25	24	18	21
Th	50.06	48.72	46.11	51.25	53.38	50.85	39.77	51.45	46.93	51.83	51.91	39.94	47.10	41.17	49.97
La	216.6	197.4	188.0	197.5	202.5	197.4	180.6	207.3	214.3	204.2	204.0	184.3	207.1	158.6	194.1
Ce	373.1	325.4	319.7	327.7	330.3	330.6	312.4	353.6	377.7	344.0	343.7	328.1	359.9	256.7	325.7
Pr	42.06	35.51	35.47	36.00	35.24	35.87	35.79	38.62	42.39	37.57	37.63	38.29	40.35	27.59	35.49
Nd	142.0	116.4	118.1	116.4	111.4	117.2	121.9	125.9	142.6	122.5	122.2	134.5	136.6	88.20	115.1
Sm	18.98	14.98	15.79	15.20	14.13	15.35	16.90	16.61	19.47	16.14	16.08	19.19	18.47	11.58	15.12
Eu	4.44	5.19	3.78	3.66	3.58	3.67	5.29	3.91	4.58	3.84	3.84	4.36	4.31	2.96	3.59
Gd	10.73	8.25	8.95	8.45	7.54	8.72	10.14	9.42	11.17	9.13	8.99	11.27	10.64	6.70	8.56
Tb	1.20	0.89	1.01	0.98	0.88	0.96	1.14	1.03	1.22	1.03	1.02	1.21	1.17	0.77	0.95
Dy	5.68	4.22	4.86	4.59	4.23	4.66	5.29	4.83	5.70	4.88	4.83	5.66	5.40	3.87	4.50
Ho	0.90	0.69	0.79	0.76	0.69	0.76	0.89	0.79	0.89	0.79	0.80	0.93	0.87	0.66	0.74
Er	2.26	1.79	2.04	1.93	1.77	1.98	2.20	1.94	2.29	1.99	2.11	2.30	2.34	1.75	1.95
Tm	0.28	0.24	0.27	0.27	0.25	0.26	0.30	0.27	0.31	0.27	0.28	0.32	0.30	0.23	0.27
Yb	1.83	1.51	1.75	1.76	1.64	1.80	1.93	1.70	2.02	1.83	1.72	1.94	1.88	1.66	1.73
Lu	0.29	0.23	0.26	0.24	0.23	0.26	0.28	0.26	0.28	0.28	0.28	0.31	0.26	0.28	0.23

Fe_2O_3*, total Fe as Fe_2O_3; LOI, loss on ignition.

followed sequentially by ternary feldspar and analcime. Some analcime is clearly secondary, occurring in amygdules together with carbonates and natrolite.

5. Analytical methods and alteration

Thirty-eight representative samples were selected from a suite of several hundred samples collected by Church (1973, 1985) during the detailed mapping of the White Lake basin and Greenwood mining camp conducted by the British Columbia Geological Survey. Major and some trace (Rb, Sr, Ba, Zr, Nb, Y, Cr, Ni, Pb, V, Cu and Zn) elements in these samples were analyzed by X-ray fluorescence at the Geochemical Centre of the Department of Geology of Saint Mary's University, Halifax, Nova Scotia. Additional trace elements (the rare-earth elements (REE) Hf, Nb and Th) were analyzed in 23 samples (Table 2) by inductively coupled plasma-mass spectrometry (ICP-MS) using a Na_2O_2-sintering technique at the Department of Earth Sciences of Memorial University of Newfoundland. The precision for the trace elements is between 2 and 8% of the values cited (Dostal et al., 1986; Longerich et al., 1990).

Six samples were subsequently selected for Nd and Sr isotope analyses. Sm and Nd abundance and Sr- and Nd-isotope ratios were determined by isotope dilution mass spectrometry at the Department of Earth Sciences of Memorial University

(Kerr et al., 1995). Measured $^{143}Nd/^{144}Nd$ values were normalized to a $^{146}Nd/^{144}Nd$ ratio of 0.7219. An average value for $^{143}Nd/^{144}Nd = 0.511849 \pm 9$ resulted from replicate analyses of the LaJolla standard, which was analyzed repeatedly throughout. The $^{87}Sr/^{86}Sr$ was corrected using $^{86}Sr/^{88}Sr = 0.1194$. The NBS 987 Sr standard gave an average value of $^{87}Sr/^{86}Sr = 0.710250 \pm 11$ on replicate analyses. Initial Sr- and Nd-isotope ratios and epsilon values (ε_{Nd}) were calculated using the age of 50 Ma (Table 3). Depleted mantle model ages (T_{DM}) were calculated assuming values of $^{143}Nd/^{144}Nd = 0.513114$ and $^{147}Sm/^{144}Nd = 0.213$ for modern depleted mantle.

K–Ar analyses of separated mica were done at the University of British Columbia by J.K. Harakal. Potassium concentrations were determined by replicate atomic absorption analyses. Argon analyses were performed by isotope dilution using an AEI MS-10 mass spectrometer operated in the static mode for mass measurement, a high-purity ^{38}Ar spike and conventional gas-extraction and purification procedures. The precision of the data in Table 4 is the estimated analytical uncertainty at one standard deviation. Mineral compositions were determined at the Department of Earth Sciences of Dalhousie University (Halifax, Nova Scotia) using a JEOL Superprobe 733 equipped with four wavelength-dispersive spectrometers and one energy-dispersive spectrometer, and operated with a beam current of 15 kV at 5 nA.

The Yellow Lake rocks were affected by zeolite/

Table 3
Nd and Sr isotopic composition of volcanic rocks of the Yellow Lake Member

Sample	Nd (ppm)	Sm (ppm)	$^{143}Nd/^{144}Nd_m$	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd_{init}$	ε_{Nd}	T_{DM} (Ma)	Rb (ppm)	Sr (ppm)	$^{87}Sr/^{86}Sr_m$	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr_{init}$	Ba/Nb
71	118.46	16.38	0.512381 ± 67	0.08537 ± 2	0.512353	−4.3	890	145	3558	0.706571 ± 14	0.118	0.706487	42.2
84	108.55	13.77	0.512307 ± 23	0.07830 ± 2	0.512281	−5.7	917	66	6467	0.706509 ± 17	0.030	0.706488	34.6
230	142.26	19.24	0.512332 ± 24	0.08348 ± 3	0.512305	−5.3	922	127	2560	0.706552 ± 31	0.144	0.706450	41.8
244	145.35	19.25	0.512317 ± 25	0.08176 ± 4	0.512290	−5.5	926	138	4999	0.706550 ± 22	0.080	0.706493	38.4
248	86.14	11.35	0.512308 ± 24	0.08132 ± 8	0.512281	−5.7	932	101	3798	0.706549 ± 21	0.077	0.706494	39.2
260	114.23	14.99	0.512288 ± 56	0.08102 ± 2	0.512262	−6.1	947	154	4226	0.706538 ± 22	0.105	0.706463	34.8

$^{143}Nd/^{144}Nd_m$ and $^{87}Sr/^{86}Sr_m$ are measured Nd and Sr isotopic ratios, respectively, and ε_{Nd} is the fractional difference between the $^{143}Nd/^{144}Nd$ of rock and the bulk Earth at the time of crystallization. Concentrations of Nd and Sm were determined by isotope dilution, and those of Rb and Sr by X-ray fluorescence. $^{87}Sr/^{86}Sr_{init}$ is initial Sr isotopic ratio. $^{143}Nd/^{144}Nd_{init}$ is initial Nd isotopic ratio. ε_{Nd}, $^{143}Nd/^{144}Nd_{init}$ and $^{87}Sr/^{86}Sr_{init}$ assume an age of 50 Ma. Values of ε_{Nd} were calculated using modern $^{143}Nd/^{144}Nd_{CHUR} = 0.512638$ and $^{147}Sm/^{144}Nd_{CHUR} = 0.1967$. Depleted mantle (DM) model ages (T_{DM}) were calculated assuming modern DM $^{143}Nd/^{144}Nd = 0.513114$ and $^{147}Sm/^{144}Nd = 0.213$. Precision of concentrations of Nd and Sm is $\pm 1\%$.

subgreenschist facies metamorphism but their chemical composition did not appear to be affected by secondary processes, with the exception of some hydration and high LOI (loss-on-ignition) in a few samples. Compositional similarities between the rocks and recent volcanic suites, as well as the positive correlation among incompatible trace elements, suggest that most major and trace elements retained their original concentrations. Furthermore, consideration of petrogenesis is based primarily on high-field-strength elements (HFSE) and REE, which are usually assumed to be immobile during the low-grade metamorphism.

Although K–Ar data are susceptible to alteration, a very good agreement of our results with a number of other age determinations of the Penticton Group rocks (e.g. Bardoux, 1993; Breitsprecher, 2002) and other suites of the Challis–Kamloops belt suggest that the mica K–Ar values in Table 4 were not modified by secondary processes. Likewise, the initial Sr isotopic ratios are consistent with other rocks from those geographic positions (Fig. 1) relative to the Sr isopleths of Ghosh (1995) and Armstrong (1988), implying that the original characteristics were retained.

6. Geochemical composition

The SiO_2 content of rocks of the Yellow Lake Member varies between 48 and 57 wt%, although most of the samples are between 50 and 55 wt%. The rocks have high contents of alkalis (> 9 wt%) and according to Le Maitre (1989), most of them can be classified as tephriphonolites (Fig. 3). They are also high in K_2O (> 4.5 wt%) and most of them can be classified as ultrapotassic (Le Maitre,

1989). Additionally, they usually have a K_2O/Na_2O ratio > 1. The lavas have elevated contents of Al_2O_3 (18–20 wt%) and have nepheline and olivine in their norms. Their low TiO_2 (< 1.2 wt%; Fig. 4) and FeO_{tot} are typical of volcanic arc lavas. Overall, the rock compositions are comparable to potassic lavas from the Montana alkaline province (e.g. O'Brien et al., 1991, 1995; MacDonald et al., 1992), central Italy (e.g. Rogers et al., 1985) and Indonesia (Stolz et al., 1988, 1990; Edwards et al., 1994). However, the Yellow Lake rocks differ from high-TiO_2 alkaline rocks such as basaltic rocks of the Cameroon Line, West Africa (Fitton, 1987) and the Gregory Rift, East Africa (Latin et al., 1993), which have compositional characteristics of within-plate basalts (Fig. 4).

The low contents of MgO (< 3.5 wt%), Ni (< 40 ppm) and Cr (< 40 ppm) of the rocks indicate that the Yellow Lake lavas underwent extensive differentiation. When plotted against MgO (Fig. 5), FeO_{tot}, TiO_2, P_2O_5, CaO and CaO/Al_2O_3 display positive correlations, while Al_2O_3, Na_2O and SiO_2 correlate negatively. The compatible trace elements Ni, Co, Cr and V correlate positively with MgO, although trends are not well defined. These major and trace-element trends are broadly consistent with fractional crystallization of clinopyroxene, Fe–Ti oxides, apatite and possibly phlogopite, all of which are present as phenocrysts or their inclusions in the Yellow Lake rocks.

As might be expected from their potassic character, the rocks have high abundances of Sr (1500–6500 ppm), Ba (3500–6000 ppm) and Th (40–60 ppm). While Sr tends to increase with differentiation, Ba shows no systematic variation

Table 4
K–Ar data for biotite from rocks in the Penticton Group

Rock unit	Location		Host rock	K	Radiogenic Ar	Radiogenic Ar	Date
	Lat.	Long.		(%)	(cm³/10⁶)	(%)	(Ma ± 1σ)
Kettle River Formation	49°28.52′	119°38.83′	Porphyry	7.00 ± 0.04	14.458	89.6	52.4 ± 1.8
Marron Formation (1)	49°33′	119°52′	Lava	6.87 ± 0.04	14.279	79.9	52.7 ± 1.8
Marron Formation (2)	49°49.65′	119°44.4′	Lava	3.395 ± 0.025	7.082	80.5	52.9 ± 1.5
White Lake Formation	49°18′	119°37′	Pyroclastic	6.29 ± 0.04	13.02	72.8	52.5 ± 1.8

Ages were calculated using $\lambda = 4.962 \times 10^{-10}$ yr^{-1}; $\lambda_e + \lambda_e^- = 0.581 \times 10^{-10}$ yr^{-1}; ^{40}K/K$_T$ = 1.167×10^{-4} mol/mol. Marron Formation: (1) – Yellow Lake Member; (2) – Kitley Lake Member.
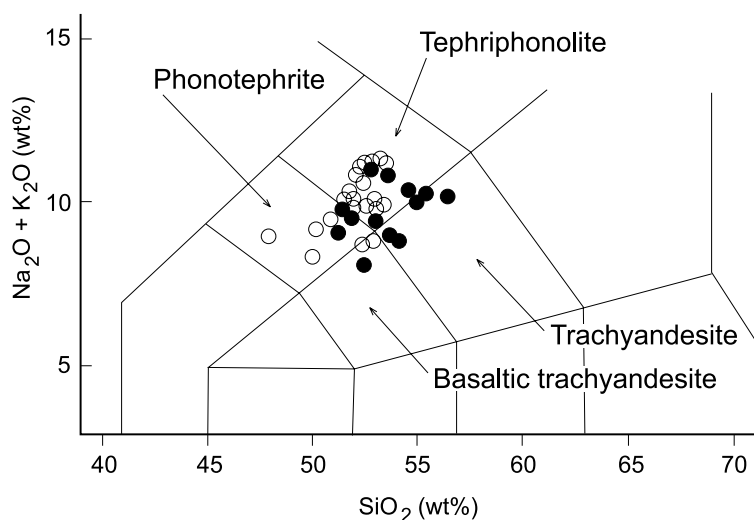
Fig. 3. Chemical classification of the Yellow Lake volcanic rocks using the total alkalis vs. silica diagram of Le Maitre (1989). Shoshonitic rocks (solid circles); ultrapotassic rocks (open circles) as determined from K_2 vs. silica plot (Le Maitre, 1989).

with MgO, implying that fractionation of phlogopite and possibly ternary feldspar prevented the buildup of Ba, a normally incompatible element, during the fractionation process. The rocks have high K/Rb ratios (300–400) compared to typical continental crustal values of 230 (Shaw, 1968). They are also depleted in Rb relative to Ba and Sr. Most basalts, including mid-ocean ridge basalts (MORB), ocean island basalts (OIB) and non-orogenic lamproites (Hawkesworth et al., 1985), have Rb/Ba values around 0.1, whereas the Yellow Lake rocks have lower ratios (0.02–0.04). The low Rb/Ba and Rb/Sr (0.01–0.1) ratios accompanied by high Ba and Sr contents suggest enrichment of mantle peridotite by silicate melts rather than by hydrous fluids (Miller et al., 1999).

The REE patterns for all Yellow Lake rocks are steeply sloping, with chondrite-normalized La abundance between 500 and 700, and lack negative Eu anomalies (Fig. 6). $(La/Yb)_n$ and $(La/Sm)_n$ ratios range from 50 to 80 and from 5 to 8, respectively. The absence of Eu anomalies implies that while ternary feldspar phenocryts are abundant in the rocks, their removal or that of a more calcic plagioclase during fractionation was not significant. The overall similarity of the REE patterns in the Yellow Lake lavas is consistent with their derivation from a common parental

magma and a cogenetic relationship through fractional crystallization.

Mantle-normalized trace-element abundance curves of the rocks of the Yellow Lake Member are distinctly enriched in large-ion-lithophile ele-
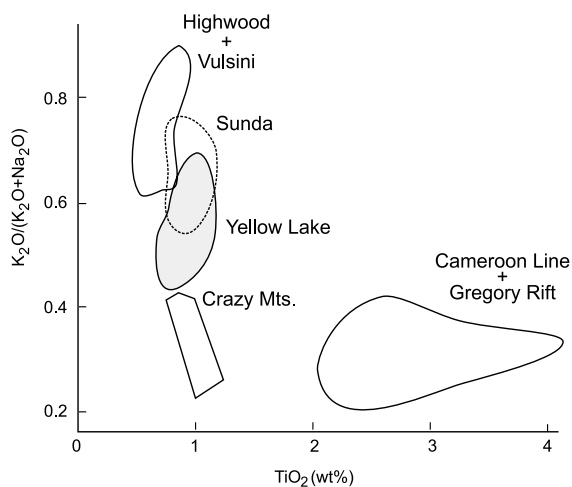
Fig. 4. $K_2O/(K_2O+Na_2O)$ vs. TiO_2 (wt%) diagram (modified from O'Brien et al., 1995) for the Yellow Lake volcanic rocks and selected alkaline rocks from Highwood Mountains and Crazy Mountains, Montana (O'Brien et al., 1995), Vulsini (central Italy; Rogers et al., 1985), Sunda arc (Indonesia; Stolz et al., 1988, 1990), Cameroon Line (west Africa; Fitton, 1987) and Gregory Rift (east Africa; Latin et al., 1993).
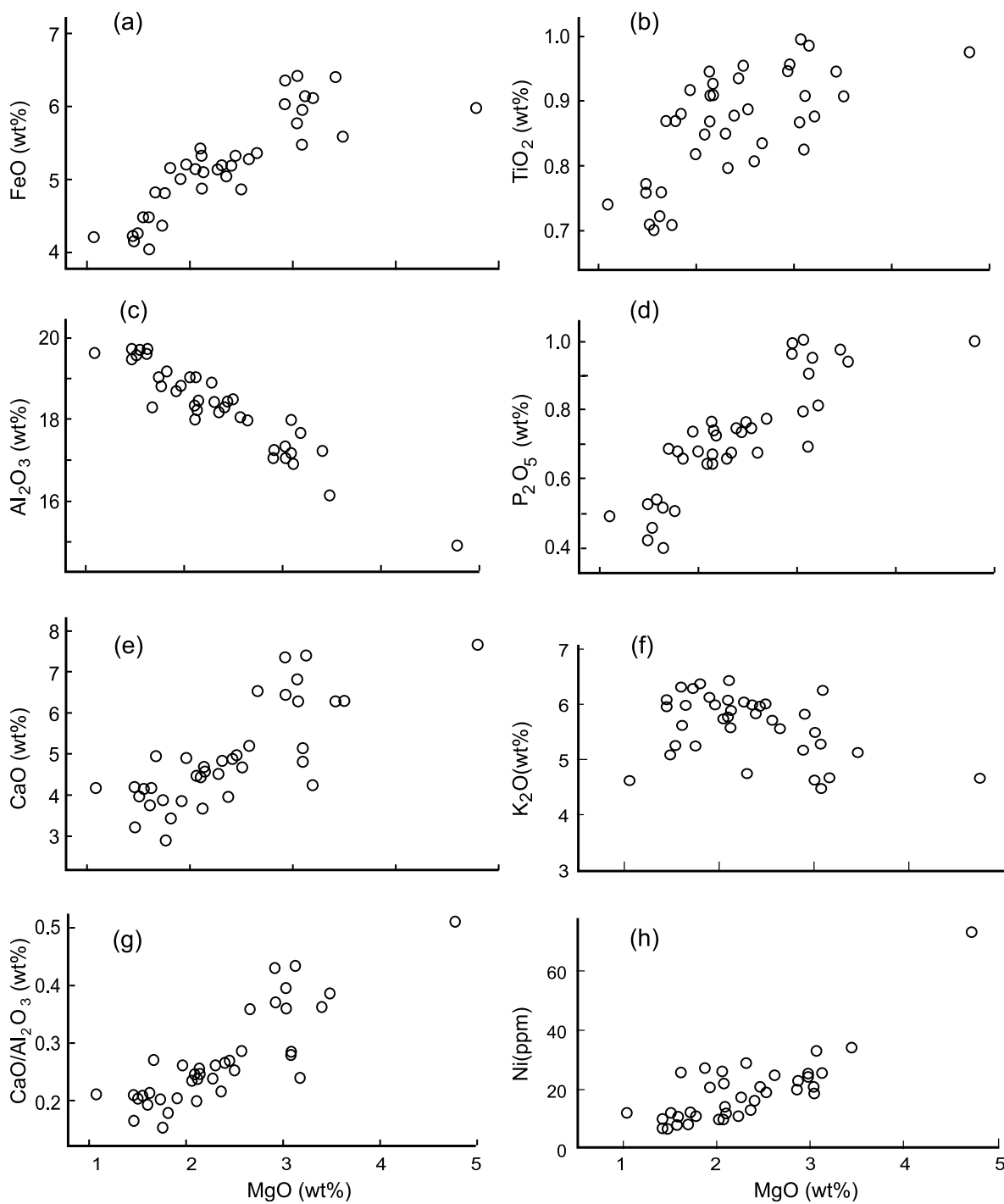
Fig. 5. Variations of FeO_{tot}, Al_2O_3, CaO, TiO_2, P_2O_5, CaO, K_2O (wt%), Ni (ppm) and CaO/Al_2O_3 vs. MgO (wt%) in the Yellow Lake volcanic rocks.
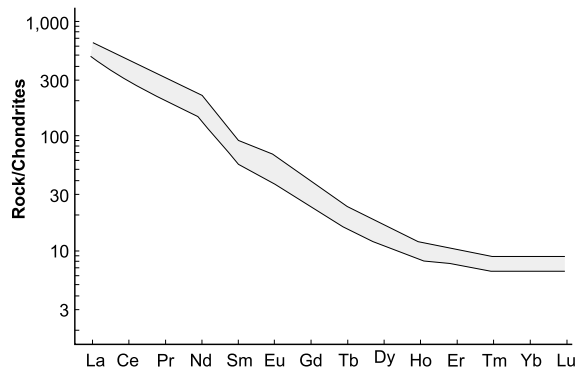
Fig. 6. Chondrite-normalized abundances of REE in the Yellow Lake volcanic rocks. Normalizing values after Sun and McDonough (1989).

ments (LILE), including Ba (500–900× primitive mantle), Th and light REE (LREE), relative to the heavy REE (HREE) and HFSE accompanied by distinct negative anomalies for Nb and Ti (Fig. 7). The anomalies are present despite the high elemental abundance of HFSE in the rocks. For example, the abundance of Nb varies between 83 and 130 ppm (120–180× primitive mantle). They differ from those of intraplate potassic mafic rocks including kimberlites and OIB (Sun and McDonough, 1989), which are not depleted in Nb (Fig. 7). The patterns of the Yellow Lake rocks are typical of subduction-related mafic rocks or continental intraplate basalts that have been significantly contaminated or derived from a lithospheric mantle modified by an ancient subduction process (e.g. Kempton et al., 1991). They resemble the profiles of other subduction-related potassic lavas, including the Eocene calc–alkaline volcanic rocks of the Challis–Kamloops belt (e.g. Dostal et al., 1998) and the potassic and ultrapotassic rocks from the Montana alkaline province (O'Brien et al., 1995) and SE Spain (Contini et al., 1993). In common with rocks from the Montana province (O'Brien et al., 1991, 1995), the trace-element patterns of the Yellow Lake volcanics also show small depletions in Zr and Hf (Fig. 7).

7. Isotope geochemistry

The initial $^{87}Sr/^{86}Sr$ ratios for Yellow Lake samples are elevated and vary in a narrow range of 0.70645–0.70649 (Table 3). Their values are higher than those of many calc–alkaline rocks from Eocene successions well north of the study area, with typical upper limits of ca. 0.705 (e.g. Ewing, 1981b; Armstrong, 1988) including high-K calc–alkaline and shoshonitic suites from Buck Creek (e.g. Dostal et al., 1998, 2001). The Yellow Lake rocks are located east (on the North American side) of the $^{87}Sr/^{86}Sr = 0.704$ isopleth (Fig. 1) of Ghosh (1995) and Armstrong (1988) and overlie Precambrian crust, as indicated by the Lithoprobe data (Clowes et al., 1998). The 0.704 isopleth was predominantly based on Mesozoic magmatic rocks, suggesting that the mantle is characteristically radiogenic in this part of the Cordillera. The initial $^{143}Nd/^{144}Nd$ ratios of the rocks (0.512262–0.512353; $\varepsilon_{Nd} = -6.1$ to -4.3) require a large time-integrated LREE enrichment.
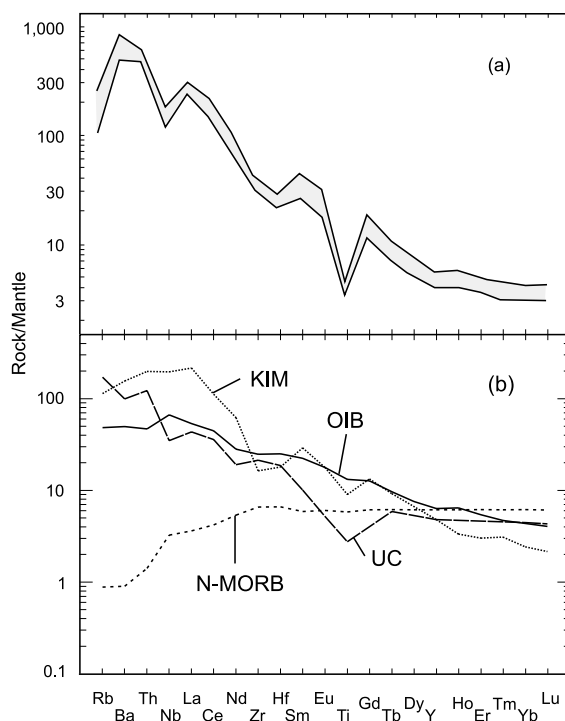

Fig. 7. Mantle-normalized abundances of incompatible trace elements in (a) the Yellow Lake volcanic rocks, (b) N-MORB (N-type MORB; Sun and McDonough, 1989), OIB (Sun and McDonough, 1989), KIM (kimberlite; Mitchell, 1986), and UC (upper crust; Taylor and McLennan, 1985). Normalizing values after Sun and McDonough (1989).

The values differ significantly from other rocks of the Eocene magmatic belt in British Columbia (Fig. 8), although they plot on an extension of the mantle array towards the EM II mantle reservoir of Zindler and Hart (1986). The values are comparable to the Sr and Nd isotopic data, reported by Ghosh (1995), for regional samples including the Marron Formation (analcime phonolite: 0.706439 and −7.6) and Coryell intrusions (0.706723–0.707480 and −7.5 to −9.3, respectively).

On the initial ε_{Nd} versus $^{87}Sr/^{86}Sr$ diagram (Fig. 8), the rocks fall in a restricted field near that of the leucitites of central Italy (Fig. 8), and plot between the fields of the Eocene high-K calc–alkaline lavas of the Challis–Kamloops belt, and of the Eocene potassic rocks of the Highwood Mountains and Bearpaw Mountains of Montana, which were emplaced through the Archean/Proterozoic crust. There is no obvious correlation be-

Fig. 9. Variations of Ba/Nb vs. ε_{Nd} values in the Yellow Lake volcanic rocks.

tween major elements and the isotopic compositions suggesting decoupling of isotopes and major elements. However, there is a correlation between the LILE/HFSE ratios such as Ba/Nb and La/Nb and the Nd isotopic ratios (Fig. 9).

8. Geochronology

Four mica K–Ar ages (Table 4) on the Penticton Group are reported here and all agree within analytical error. The first biotite is from a rhyolite porphyry (52.4 ± 1.8 Ma) at the base of the Penticton Group (Kettle River Formation) and the second one is from the White Lake Formation volcanic unit in the upper part of the Penticton Group (52.5 ± 1.8 Ma). The other two biotite/phlogopite samples are from volcanic rocks of the Marron Formation, occurring stratigraphically between these two units. The data confirm Early to Middle Eocene ages for the Penticton Group and imply the group was emplaced during a short time interval. These ages coincide with other available isotopic ages for the Penticton Group (54–49 Ma; Hunt and Roddick, 1989; Bardoux, 1993) as well as for other magmatic suites within the Challis–Kamloops volcanic belt in southern British Columbia (54–48 Ma; e.g. Hunt and Roddick, 1987; Read, 1993, 2000; Hickson et al., 1991; Breitsprecher, 2002).

Fig. 8. ε_{Nd} vs. $^{87}Sr/^{86}Sr_i$ for the Yellow Lake volcanic rocks (YL). Initial isotope ratios were calculated at an age of 50 Ma. Shown for comparison are the compositions of Eocene alkaline rocks of the Highwood Mountains (HM) and Bearpaw Mountains (BM) from the Montana alkaline province and alkaline rocks of central Italy (Roccamonfina [RC] and Vulsini [VU]) (modified after O'Brien et al., 1995, Dudás and Harlan, 1999 and Buhlmann et al., 2000), Eocene high K-calc–alkaline lavas of the Challis–Kamloops belt from the Buck Creek complex (BC), central British Columbia (Dostal et al., 2001) as well as the main oceanic mantle reservoirs of Zindler and Hart (1986). DM, depleted mantle; BE, bulk silicate Earth; EM I and EM II, enriched mantle; HIMU, mantle with high U/Pb ratio. The Mantle Array of Zindler and Hart (1986) is defined by oceanic basalts.
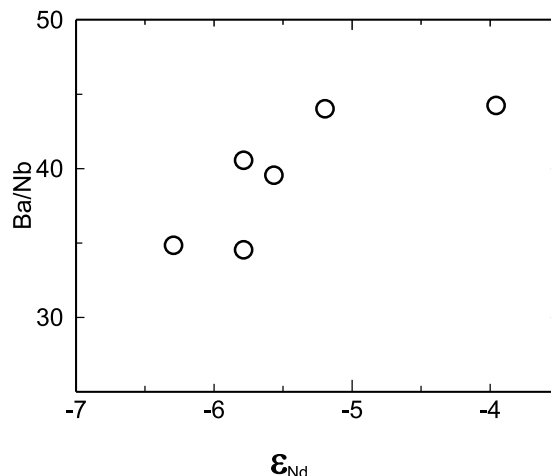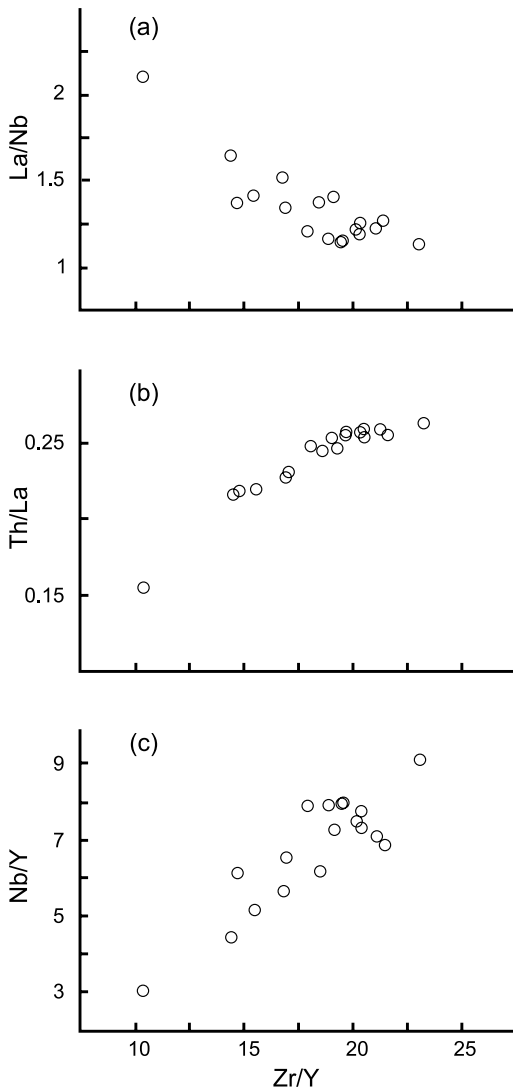
Fig. 10. Variations of La/Nb, Nb/Y and Th/La vs. Zr/Y in the Yellow Lake volcanic rocks.

9. Petrogenesis

The rocks of the Yellow Lake suite are characterized by high contents of total alkalis (>9 wt%), particularly K_2O (>4.5 wt%). Their mantle-normalized trace-element patterns (Fig. 7) are distinctly enriched in LILE including Ba and Th, and in LREE relative to HREE and HFSE, accompanied by negative anomalies for Nb and Ti. The radiogenic isotope characteristics of these

rocks (Fig. 8) include elevated but uniform initial radiogenic $^{87}Sr/^{86}Sr$ (~ 0.7065) ratios and the distinctive non-radiogenic ε_{Nd} (~ -7.6 to -4.3) values which are closer to the EM II mantle component than other rocks of the Challis–Kamloops belt, specifically those of Buck Creek volcanic complex (Fig. 8), which are much closer to DM and HIMU components. Rocks with geochemical characteristics similar to the Yellow Lake suite have been described from back-arc, rifted arc and intraplate settings, particularly those associated with continental lithosphere (Mitchell, 1996).

The origin of the potassic, particularly leucite- and/or analcime-bearing alkaline rocks remains controversial (Edgar and Mitchell, 1997; Conticelli and Peccerillo, 1992; Shaw, 1996). The mantle-normalized patterns of some of these rocks, including the Yellow Lake tephriphonolites, are significantly different from those of magmas which are assumed to be generated within the asthenosphere, namely MORB and OIB (Fig. 7), which peak at Nb and have no negative Ti anomaly (Sun and McDonough, 1989). The parental magmas of these potassic rocks cannot be generated by partial melting of primitive upper mantle.
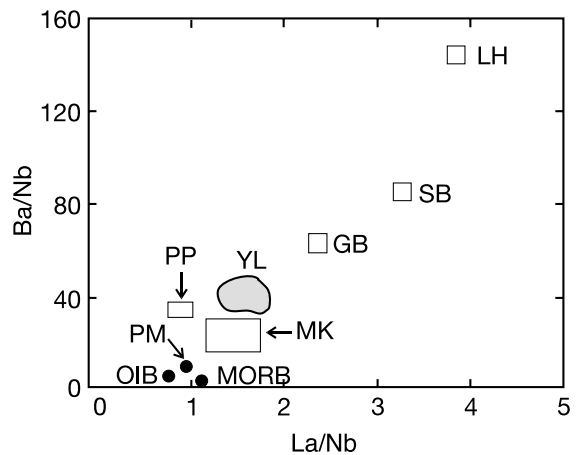

Fig. 11. Ba/Nb vs. La/Nb plot for the Yellow Lake volcanic rocks (YL) and some potassic lithospheric mantle melts (modified from Miller et al., 1999). MORB (N-type MORB), OIB and PM (primitive mantle) are after Sun and McDonough (1989); PP (phlogopite peridotite) after Erlank et al. (1987), phlogopite lamproites (GB, Gaussberg; LH, Leucite Hills; SB, Smoky Butte) are after Mitchell and Bergman (1991) and micaceous kimberlites (MK) after Mitchell (1986).
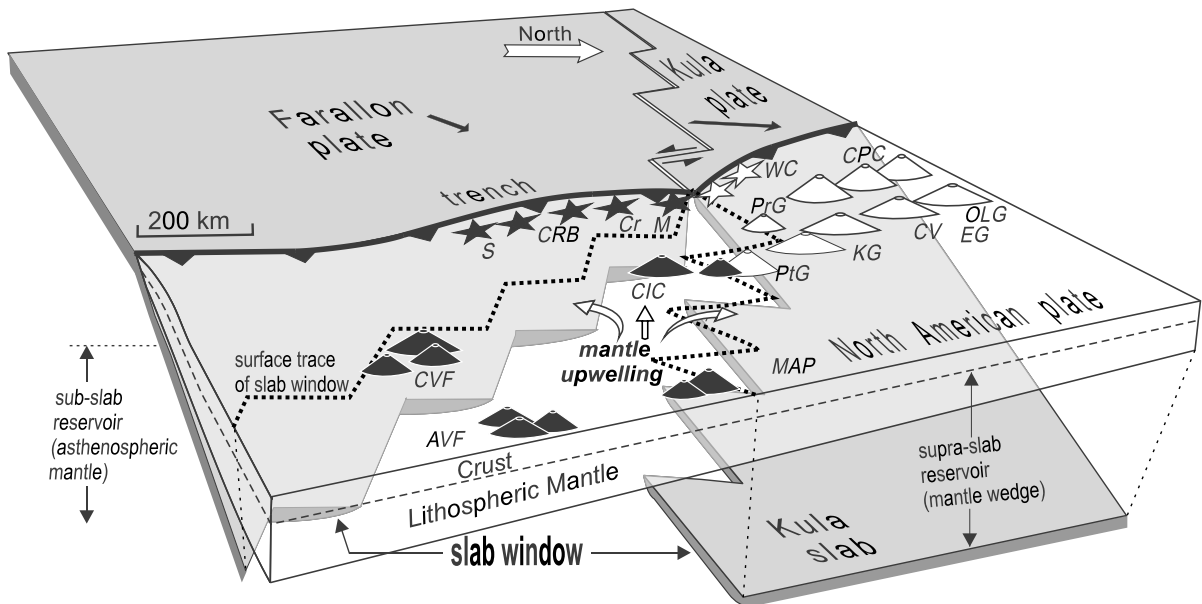
Fig. 12. Cartoon showing the proposed model of the mantle environment beneath the southern Cordillera in the Middle Eocene. The Farallon plate is shown to dip more gently than the Kula plate; Farallon and Kula plate motions, relative to a fixed North America, are 12.6 and 18.3 cm/yr, respectively (shown graphically with black arrows). Locations of Middle Eocene magmatism are shown schematically as volcanoes and stars although in some regions the igneous activity is recorded partly or wholly by intrusions. White volcanoes represent igneous complexes with arc character derived mainly from the mantle wedge above the Early Cenozoic subduction zone. Black volcanoes represent complexes derived largely from Precambrian mantle lithosphere accessed mainly in the slab window. Black stars indicate oceanic centers of Siletzia derived from the mantle and lacking significant arc character. White stars are felsic fore-arc centers on Vancouver Island. CPC, Coast Plutonic Complex; CRB, Coast Range Basalts; S, Siletz Basalts; Cr, Crescent Basalts; M, Metchosin Basalts; WC, Walker Creek Intrusions; CV, Clisbako Volcanics; OLG/EG, Ootsa Lake, Buck Creek and Endako Groups; KG, Kamloops Group; PtG, Penticton Group; PrG, Princeton Group; CIC, Colville Igneous Complex; CVF, Challis Volcanic Field; MAP, Montana Alkaline Province (including Highwood and Bearpaw Mountains); AVF, Absaroka Volcanic Field (Wyoming). Surface trace on the North American plate of the slab window is shown by bold dotted lines (modified from Breitsprecher et al., 2003).

Petrogenetic models which are relevant to the origin of silica-undersaturated alkaline rocks such as those of the Yellow Lake suite require that: (1) they contain a significant crustal contamination; or (2) they were derived from a source enriched in LREE and depleted in HREE and HFSE relative to primitive mantle.

9.1. Crustal contamination

The fact that the Yellow Lake rocks were erupted through continental crust raises the possibility that crustal contamination produced some of the isotopic and trace-element signatures. High concentrations of incompatible trace elements in these rocks, significantly higher than those in average upper crust or typical upper crustal melts (Taylor and McLennan, 1985) or an average Precambrian crust of the Canadian shield (Shaw et al., 1976), cannot be generated only by progressive or wholesale assimilation of typical middle to upper crustal material (Fig. 7). Because the lavas are enriched in LREE and depleted in HREE relative to continental crust and MORB and/or OIB, crustal contamination would lead to a decrease of LREE and an increase of HREE in the magma, leading to a flatter slope of the REE patterns. Assimilation of dry lower crustal granulites (like the Middle Proterozoic Monashee orthogneiss; Parrish et al., 1988) typically depleted in LILE

could hardly contribute a high abundance of the LILE in the tephriphonolites. Thus, contamination alone of OIB- or MORB-type magma by crustal material could not have produced the tephriphonolites. Minor assimilation of crustal rocks may well be present and would lower the LILE and LREE contents of the rocks therefore being difficult to resolve. The low contents of SiO_2 (generally below 53%) in the potassic silica-undersaturated mafic rocks, however, limits the amount of crustal material that could have been assimilated.

CaO shows a weak correlation with ε_{Nd} values, where three samples with high CaO have higher ε_{Nd} values than the low CaO samples, which have more negative ε_{Nd} values (Tables 2 and 3). With this exception, however, there is a lack of correlation of isotopic compositions, particularly $^{87}Sr/^{86}Sr$, with major elements, namely SiO_2 and MgO contents and other indices of fractionation in the Yellow Lake rocks. A lack of such correlation further argues against a simple assimilation–fractional crystallization model for the lavas (DePaolo, 1981). Small degrees of partial melting can probably produce melts with very high concentrations of incompatible elements and the retention of garnet in the restite can lead to enhanced LREE compared with HREE. Since these melts contain high concentrations of these elements, they can have a significant effect on the relatively low concentrations of incompatible elements in a mantle melt while leaving the major element composition virtually untouched.

However, relatively uniform trace-element and isotopic compositions of these rocks and their similarities to the well-documented suites of the Montana alkaline province and elsewhere (e.g. Shaw, 1996; O'Brien et al., 1995) suggest that the distinctive geochemical characteristics (the highly fractionated REE patterns, high LILE/HFSE ratios and decoupling of major element chemistry from radiogenic isotopic ratios) reflect their mantle source and its mineralogy. As the low MgO and Ni contents and sparse olivine, among other features, indicate that the rocks underwent significant differentiation, resulting in modified geochemical parameters. The apparent trend towards at least some degree of differentiation, combined with low SiO_2 contents of the Yellow Lake lavas, places severe constraints on the relative percentage of crustal material which might have been assimilated into the magma.

9.2. Source enrichment

Elongated trends on graphs of ratios of trace elements such as Th/La vs. Zr/Y, Nb/Y vs. Zr/Y and La/Nb vs. Zr/Y (Fig. 10) cannot be generated by variable degrees of melting or fractional crystallization but imply source heterogeneity or source mixing. Current models for the genesis of silica-undersaturated potassic melts such as the Yellow Lake suite typically invoke the presence of a veined mantle source (e.g. Foley, 1992). The vein assemblages are thought to consist of hydrous and anhydrous phases enriched in incompatible elements, such as phlogopite, K-amphibole and apatite.

The lavas display negative Nb and Ti anomalies and LILE enrichment relative to HREE (Fig. 7), suggesting that a subduction process produced the enrichment of the mantle source for the Yellow Lake rocks. Significant amounts of LILE and volatiles, in such an enriched mantle source, possibly concentrated in phlogopite. The elevated concentrations of Ba and Sr in the Yellow Lake rocks are consistent with the presence of phlogopite and clinopyroxene in their source and the relatively high La/Yb ratios indicate that garnet may have been an important phase in the residuum. Experimental studies suggest that an assemblage rich in phlogopite, K–Ti amphibole, clinopyroxene and apatite could be the source of ultrapotassic rocks (Sato et al., 1997; Edgar and Mitchell, 1997). Furthermore, these minerals are stable throughout the subcontinental lithospheric mantle.

Miller et al. (1999) argued that the low Ti/Y (200–300) and negative ε_{Nd} values in some ultrapotassic rocks are a feature of recycled continental material subducted back into the upper mantle. The Yellow Lake rocks also have these characteristics, including low Ti/Y. On the La/Nb vs. Ba/Nb graph (Fig. 11), the Yellow Lake rocks plot on a trend defined by potassic rocks from the Montana province and by other rocks

that are derived from subcontinental lithospheric mantle, such as lamproites and kimberlites (Mitchell, 1996; Mitchell and Bergman, 1991).

10. Petrotectonics

The melting of phlogopite-rich mantle enriched by Kula–Farallon subduction processes can explain the major and trace-element characteristics of the mafic potassic rocks of the Yellow Lake suite, but it cannot readily account for the unusual isotopic compositions of these rocks. In particular, non-radiogenic neodymium isotopic compositions require an ancient, LREE-enriched component. A radiogenic strontium isotopic composition implies either time-integrated, long-term enrichment of Rb in the source or a primary high initial Sr isotopic ratio which is inherent to this part of the mantle. In particular, the high Sr abundances (1600–6500 ppm), low Rb/Sr ratios (0.01–0.1) and elevated but restricted Sr isotopic ratios ($\sim$0.7065) in the Yellow Lake rocks could be either due to an unusual source enrichment by Cenozoic subduction processes or, alternatively and more likely, they could reflect Precambrian lithosphere. The Yellow Lake suite is geographically located east of the $^{87}Sr/^{86}Sr = 0.704$ isopleth of Ghosh (1995), which delimits lower (primitive/oceanic) values to the west and higher (evolved/continental) values to the east (Fig. 1). That the Sr isotopes of the Yellow Lake rocks fit into regional variations suggests they are related to crustal lithosphere rather than to the recent history of subduction. Elsewhere, isotopic compositions, characterized by high $^{87}Sr/^{86}Sr$ ratios and low ε_{Nd} values, have been attributed to derivation from ancient metasomatized lithospheric mantle sources (Menzies, 1989). Additionally, to produce the observed Nd isotopic compositions, the mantle sources (the subcontinental lithospheric mantle) had to be isolated from convecting asthenospheric mantle since the Proterozoic. From Lithoprobe interpretations (Clowes et al., 1998), this part of the Challis–Kamloops volcanic belt in southern British Columbia is underlain by the westernmost edge of Precambrian cratonic North America.

The question arises whether the subduction imprint of trace-element distribution results from a relatively young enrichment by Kula–Farallon subduction or represents lithospheric mantle previously enriched above some prior subduction zone. Trace elements together with the isotope ratios can distinguish between these two options. If the high LILE/HFSE ratios are inherited from a mantle source enriched by metasomatic processes during Cenozoic subduction, then these ratios should not correlate with evolved isotope ratios because the subduction-related metasomatism would have been too recent to have produced isotopic enrichment (Kempton et al., 1991). However, the Ba/Nb ratios correlate with Nd isotopic depletion (correlation coefficient: 0.80; Fig. 9) implying that arc-imprint on the source is not the result of Cenozoic subduction, but of some older subduction event. The correlation of these LILE/HFSE ratios with the isotopes also suggests that the relative depletions of HFSE reflect long-lived enrichment of this Precambrian lithospheric mantle.

Depleted mantle Nd model ages (T_{DM}) for these rocks are within the narrow range of 890–950 Ma (Table 3); it can be speculated that the age refers to ancient subduction-related metasomatism of the mantle beneath this part of southern British Columbia. Compared to the Yellow Lake lavas, the rocks from the Montana alkaline province (Fig. 8) have even more radiogenic $^{87}Sr/^{86}Sr$ and more non-radiogenic ε_{Nd} values. It is probable that this component under the Wyoming craton was more enriched or was in isolation from convecting asthenospheric mantle longer (exceeding 1 Ga; O'Brien et al., 1995) than under southern British Columbia.

11. Tectonic setting

Geochemical evidence indicates that the alkaline rocks were derived from an enriched lithospheric mantle that was isolated since the Precambrian. However, the details of the tectonic processes and heat responsible for the origin of the magmas have not been firmly established. In particular, the importance of subducted slab ge-

ometries and their impact on mantle dynamics has not been fully explored.

The Eocene volcanic field in southern British Columbia developed during crustal extension in a convergent margin setting in which the Kula and Farallon plates were subducting beneath western North America. Magma genesis was broadly synchronous with core-complex formation in south central British Columbia (e.g. Tempelman-Kluit and Parkinson, 1986; Parrish et al., 1988). Thus, the origin of the Penticton volcanic suite could be related to rifting at a consuming plate margin. However, despite the apparent suitability of an extending-arc model as an explanation for the foregoing geochemical and structural observations (e.g. Ewing, 1980), a broader view of the Eocene igneous field in western North America raises some additional issues. Firstly, magmatic trends are not limited to an east–west increase in alkalinity. Instead, the Eocene igneous field displays a progressive change in composition from arc to back-arc-like rocks in central British Columbia (Smith, 1986; Dostal et al., 1998) to mixed arc and non-arc rocks in southern British Columbia (Ewing, 1981b; Smith, 1986; Church, 1985), to rocks with predominantly non-arc character in the northwestern United States (Norman and Mertzman, 1991; Dudás, 1991; O'Brien et al., 1991; Morris et al., 2000; Feeley et al., 2002). Secondly, the geometry of the subducting slabs beneath western North America in the Late Cretaceous and Early Tertiary may have been significantly different from the persistent model of a single, subhorizontal Farallon slab (e.g. Dickinson and Snyder, 1979). As indicated by Engebretson et al. (1985), both the Kula and Farallon plates were subducting beneath western North America during this interval, and the location of the intersection between the Kula–Farallon ridge and North American trench is imperfectly constrained. Consequently, attempts to relate Eocene magmatism to simple subduction beneath southern British Columbia and the northwestern United States or solely to extension at a consuming plate margin may be inappropriate.

The southward change from arc to non-arc character is effectively explained by the existence of a slab window beneath the northwestern United States (Fig. 12; Thorkelson and Taylor, 1989). A slab window, which is a gap between two downgoing slabs (Dickinson and Snyder, 1979), formed as the Kula–Farallon ridge was subducted beneath North America from Late Cretaceous to Eocene time, ca. 83–45 Ma (as inferred by Engebretson et al., 1985).

During this interval, the Kula and Farallon plates as well as their subducted parts diverged from one another. The subducted slabs were surrounded by hot mantle peridotite and any magma generated between the slabs would have risen upward, past the trailing slab edges, and into the overriding lithospheric plate (cf. Dickinson and Snyder, 1979; Thorkelson, 1996). Although the location of the Kula–Farallon slab window in Late Cretaceous to Paleocene time is controversial (Thorkelson and Taylor, 1989; Haeussler et al., 1995), its position has been inferred from fore-arc magmatism beneath Washington (Babcock et al., 1992) or southern British Columbia (Groome, 2000) in Middle Eocene time when the Penticton Group was extruded. A southern British Columbia to northwestern United States position for the window has been inferred from geochemical evidence in the back-arc (Breitsprecher et al., 2003). Such a position is compatible with all published tectonic models, particularly if restoration of northerly displaced fore-arc terranes is considered (Bol et al., 1992; Haeussler et al., 1995).

The slab-window model for Eocene evolution of southern British Columbia to the northwestern United States is depicted in Fig. 12 as a cartoon block diagram, which portrays the Challis–Kamloops volcanic belt successions in the context of the model. The Kamloops and Princeton Groups, with their predominant arc character, were erupted above the subducting Kula plate (Fig. 12). The alkaline members of the Penticton Group volcanic rocks, in particular the Yellow Lake Member and the associated Coryell syenite intrusions, formed above or just beyond the southern edge of the Kula plate, adjacent to the slab window. Farther to the south in the northwestern United States, Colville igneous complex, Challis volcanic field and successions comprising the Montana alkaline province (Highwood and Crazy volcanic fields) developed entirely above the

Kula–Farallon slab window, where supra-slab, metasomatized mantle lithosphere became underlain by or was partly replaced by upwelling asthenospheric mantle from the sub-slab reservoir (cf. Thorkelson and Taylor, 1989; Johnson and Thorkelson, 1997). Upwelling of the hotter, drier asthenospheric mantle to the base of the North American lithosphere beneath the northwestern United States and southeastern British Columbia may have been the thermal trigger necessary to induce melting of the fertile metasomatized lithospheric mantle. As noted by Dudás (1991) in his analysis of Eocene magmatism in the Montana alkaline province, upwelling of hot asthenosphere is preferable to the flat-slab subduction-driven model with its consequent mantle cooling.

The key element that the slab-window setting provides for the Yellow Lake volcanics is a heat source. Northwest of the study area, the intervening Kula slab is a heat sink and melting is restricted to more ordinary metasomatized supra-slab mantle lithosphere. To the southeast, the region of old metasomatized Precambrian lithospheric mantle overlies the extensive area exposed to the heat source provided by asthenosphere upwelling through the slab window. The asthenosphere is an effective heat source and triggers melting in the overlying lithospheric mantle. Geochemical evidence for the involvement of MORB-type asthenosphere is provided by Pb isotope values from the Highwood Mountains in Montana (O'Brien et al., 1995). Because of the unusual composition of the alkaline rocks like those at Yellow Lake, it can be inferred that most of the melting for their derivation was from metasomatized Precambrian mantle lithosphere. Coeval, more typical calc–alkaline volcanics to the north in central British Columbia were derived from adjacent but less enriched supra-slab mantle.

Contemporaneous with slab-window magmatism, diverging stress fields in the region were produced by opposing tractions on the North American plate by subduction of the Kula plate (north-northeast directed) and the Farallon plate (east-northeast directed), as indicated by Engebretson et al. (1985). This effect, along with thermal weakening from the uprise of asthenosphere and derivative magmas, facilitated the collapse of the Ju-rassic to Paleogene orogen, and helped to generate the massive Eocene core complexes and abundant grabens in southern British Columbia and the northwestern United States (Ewing, 1980; Parrish et al., 1988). Thus, a slab window in the mantle and extensional tectonism in the crust combined to produce the special conditions under which the Penticton Group and related successions with anomalous alkaline compositions were generated.

12. Conclusions

In southeastern British Columbia in the vicinity of the Canada–USA border, the Challis–Kamloops volcanic belt contains a suite of potassic silica-undersaturated mafic rocks (~ 53–52 Ma old). They are typically porphyritic, with phenocrysts of rhomb-shaped ternary feldspar and clinopyroxene, and minor phlogopite and analcime. These rocks form the western edge of the Montana alkaline province (Larsen, 1940; Chadwick, 1985). Geochemically, the rocks possess subduction characteristics, including a HFSE depletion and a LILE enrichment, but have unusual isotopic compositions compared to most other rocks within the Challis–Kamloops volcanic belt. Negative ε_{Nd} values of the Yellow Lake suite require a Precambrian subcontinental lithospheric mantle source. Elevated but relatively constant Sr isotopic ratios (~ 0.7065) are typical of their geographic position in the Cordillera irrespective of age of magmatism and probably reflect a persistent Precambrian lithospheric mantle source rich in Sr with low Rb/Sr ratios. The similarity of the Sr isotopic ratios to those of older magmatic rocks in the region also implies that the mantle lithosphere beneath southeastern British Columbia has been in place and not tectonically moved since Mesozoic terrane accretion and probably since the Late Precambrian.

The rocks of the Yellow Lake Member overlie Precambrian crust and were derived at least in part from ancient metasomatized subcontinental lithospheric mantle. The geographic position of the Penticton Group and regional geochemical variations between these lavas and those from

the southeastern region of the Challis–Kamloops belt suggest a unique and ancient mantle lithosphere was responsible for the genesis of these rocks. It is proposed that the alkaline rocks of the Challis–Kamloops belt are related to a slab-window environment. In particular, the Yellow Lake rocks were formed above the southern edge of the Kula plate, adjacent to the slab window, whereas the Montana alkaline province situated well to the southeast was formed above the Kula–Farallon slab window.

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