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The Okanagan Valley normal fault from Penticton to Enderby
South-central British Columbia

by

Marc Bardoux

A thesis submitted to the Faculty of
Graduate Studies in partial fulfillment
of the requirements of the degree of
Doctor of Philosophy

Carleton University
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-FRONTISPIECE-

One of the best exposures of the Okanagan valley fault at Black Knight mountain east of Kelowna. This view looking north shows Eocene dacites overlying alluvial sedimentary rocks together stranded atop the shallow dipping structure underlain by retrograded mylonitic gneisses of the Okanagan complex. The fault surface is not exposed but several splay faults occur in the fields and along Mission Creek in the foreground.

ABSTRACT

The west flank of the southern Omineca Belt in south-central British Columbia is cut by the Okanagan Valley fault (OVF), a westward-dipping denudation surface bounded by a brittlely deformed hanging wall of Upper Crustal (layer) rocks and a ductilely strained footwall of mid-Crustal (layer) rocks mantled by a narrow syn-tectonic brittle zone. Fault segments of the OVF extend for 450 km from the NW flank of the Monashee Terrane in the north to the Colville complex in north-central Washington. Mylonites were observed mainly in the upper part of the gneissic footwall of the OVF and as far as 50 km east of Okanagan Lake. In most cases, the shear sense is clearly displayed, the hanging wall of the shear couple moved westward. Poorly constrained piercing points indicate a horizontal displacement of 50 km but indirect evidence suggest it may have reached 70 km indicating ~80% horizontal extension locally.

Pelites of unconstrained lithological affinity underlying the OVF preserve "peak" metamorphic assemblages of sillimanite+K-feldspar+biotite+almandine+quartz $\pm$ plagioclase $\pm$ muscovite coexistent with calcsilicate lenses containing the assemblage diopside+titanite+epidote+allanite+olivine. Pelitic rocks are locally retrograded to chlorite+muscovite+albite+epidote+calcite+quartz. Six samples tested by conventional thermobarometry provide "peak" T-P conditions averaging 750°C/600 MPa presumably related to Late Cretaceous regional D2 compressive event(s). More consistent retrograde T-P conditions of 600°C/450 MPa are related to Eocene decompression. The data obtained suggest that only a few equilibria are preserved in the shear zone environment, most garnets show flat compositional profiles attributed to high temperature homogenization whereas biotites show signs of re-equilibration. Contrasting mineral assemblages across the OVF suggest a minimum vertical offset of 5 km.

Plutons of the hanging wall preserve their Mesozoic crystallization ages whereas Ladybird-type anatectic melts and mylonitic gneisses in the footwall provided U-Pb, K-Ar and ⁴⁰Ar/³⁹Ar dates that tend to overlap between 52-51 Ma. The abrupt change in isotopic cooling dates across the Okanagan valley is directly related to lithospheric stretching and thermal cooling of the OVF. Hanging wall rocks remained at temperatures below 280°C during extension while zircons crystallized in footwall granitoids 56 Ma ago. Extensional fabrics may have started nucleating from granitic rocks as early as 56 Ma ago as magmas likely provided fluid rheology in the middle crust. The OVF may thus be contemporaneous to the Slocan Lake and the Columbia River normal faults. Ductile strain and high temperature gradients were maintained in the footwall of the OVF until 51 Ma, brittle deformation seems to have occurred mainly between 52-48 Ma but may have started as early as 59 Ma ago. Cooling rates of the exhumed footwall are estimated to be in the order of 65°C/Ma, exhumation rates were in the order of 2-6 mm/a.

High K calc-alkaline Eocene lavas formed widespread shield volcanoes across the southern Omineca Belt and are probably decompression melts cogenetic with several types of mid-crustal anatectic melts including Ladybird leucogranites and Coryell syenites. Altogether these rocks crystallized and cooled during footwall exhumation and cooling. Evidences of a parental magma of mantle origin for the Eocene lavas is lacking due to extremely large quantities of lower and mid-crustal contaminants. Eocene rocks near Kelowna are grouped in three sequences: (1) 54-52 Ma old andesitic to trachyandesitic Marron lavas; (2) Late middle Eocene alluvial sedimentary rocks of the White Lake Formation; (3) 46 Ma dacites, rhyolites and tuffs. Progressive fractional

crystallisation of orthopyroxene-olivine-magnetite in the intermediate lavas and amphibole-biotite-magnetite-zircon in the felsic magmas is indicated by trace element discrimination trends; it is proposed that the andesite to rhyolite magmatic suites are genetically related. Crosscutting relations between faults and subvolcanic and volcanic sequences, rotation of the sedimentary sections, and sedimentological evolution demonstrate that extension in the Kelowna area was active 53 Ma ago during the earliest eruptions and continued during the deposition of alluvial sequences. However, extension ended by 46 Ma since the last dacitic eruptions unconformably overlie and cut through the mylonitic footwall of the OVF. This does not, however, rule out younger brittle motion.

Relative to present horizontal, the magnetizations in the Marron lavas near Kelowna are widely scattered and after correction for bedding attitudes, there is no significant improvement. Some magnetizations were acquired before and others after tilting; that is, the horizontal plane at the time of magnetization sometimes did and sometimes did not coincide with the bedding plane. Partial unfolding experiments, carried out on both categories separately, yield a precision comparable to that expected for paleosecular variation, and a mean direction declination and inclination of 352° , 70° . The Marron is normally magnetized, lavas slightly older and others slightly younger are reversely magnetized and a transition from reversed to normal polarity occurs in basal beds of the Marron Formation. The overall mean direction of the Marron and stratigraphically adjacent units is 352° , 69° , yielding a paleopole at 85°N , 197°E in agreement with that for middle Eocene rocks of cratonic North America. Hence this part of the Quesnellia Terrane had reached its present position relative to North America by middle Eocene time, and there has been no significant rotation of it. In contrast, the direction after correction for bedding tilt implies a clockwise rotation of 28° which is incorrect because, the apparent rotation, is caused by wrongly assuming that the bedding plane always coincides with the paleohorizontal at the time magnetism is acquired. Post-tilting magnetization may have been caused by syn-extensional thermal processes. Further tests have been carried out on intrusive and metamorphic rocks collected by other workers in the Valhalla complex on the east flank of the Omineca Belt. These rocks are coeval with the Marron, and are located in both the hanging wall and footwall of the Slocan Lake normal fault, which dips 30° eastward. Paleodirections are very different from those at Kelowna. It is argued that this divergence is caused by tilting $37 \pm 10^\circ$ to the west antithetical to the Slocan Lake fault. Paleomagnetism thus provides a means by which tilts in metamorpho-plutonic complexes can be determined. Such tilts may have been responsible for some of the aberrant magnetizations observed in plutonic rocks of the Coast Plutonic complex west of the Omineca Belt.

Tectonism in south-central British Columbia has been significantly controlled by the convergence between North America and the Kula-Farallon plates. Northeast trending crustal thickening transferred around 59 Ma to east-west extension during a major plate reorganization as reduced rates of subduction progressively decreased compressional forces. This change in the balance of boundary forces was probably overtaken by a combination of "passive" and "active" mechanisms leading to a rapid transition to the state of higher gravitational energy and lithospheric stretching. At the end of compression the Omineca Belt and the adjacent Intermontane Belt formed elevated ranges. The combination of physical conditions such as: critical elevations; abundant anatectic melts at mid-crustal levels; mantle lithospheric

delamination; and boundary stress conditions caused by the oblique impingement of the Kula-Farallon ridge focused crustal extension mainly on the region of maximum crustal shortening and maximum elevation where anatectic melts were abundant. Simple calculations show that deviatoric stresses of as much as 100 MPa may have been generated by the combined effect of crustal thickening and lithospheric thinning.

The geophysical characteristics of the southern Omineca Belt are in many aspects similar to those of the Basin and Range Province. Both regions are characterized by a thin crust (~ 30 km) and low density upper mantle to a depth of 200 km, indicating that the lithosphere has been thinned to the base of the crust. Present-day heat flow data of 80 mWm^{-2} on the west flank of the Omineca Belt is also similar to that of the Basin and Range Province and could be reconciled with Eocene extension and is consistent with the magnitude of extension estimated with restored cross-sections in the order of 50%.

ACKNOWLEDGEMENTS

This work results from eight months of field study integrated with U-Pb, K-Ar and Ar-Ar geochronology carried out at the Geological Survey of Canada and Queen's University Geochronology Laboratories in Ottawa and Kingston, Ontario. The Paleomagnetic study made at the Pacific Geoscience Center is part of a research agreement to the Cordilleran Section of the GSC. Field work was carried out in solo in 1984 and 1986 but technical and financial field assistance with Don Mustard was provided by the GSC in 1985.

The project was partly funded by NSERC Operating Grant A2693 to R. L. Brown. I would like to acknowledge NSERC scholarships granted from 1984 to 1986 and a Geological Society of America research grant awarded in 1985. Travel funding was provided in 1985 and 1986 by the Graduate Students Association of Carleton University. Microprobe work was financed by a research grant of the Ministère de l'Energie et des Ressources du Québec awarded to Dr. Michel Gauthier in trade of a research assistantship in 1988.

I thank Richard Brown, Ted Irving, Dirk Tempelman-Kluit, Doug Archibald, Randy Parrish, Jean-Claude Mareschal, and Sharon Carr for their teaching, advice and discussions. I am grateful to the assistance of members of the Geochronology Laboratory of the Geological Survey of Canada in Ottawa and Doug Archibald at Queen's who provided mass spectrometry and data reduction. Thanks to Jane Wynne, Janet Baker, Brad Johnson, James Bourne, Serge Perreault and the above contributors for sharing their observations. Thanks also to staff members of the Département des Sciences de la Terre of l'Université du Québec à Montréal, especially Drs. Michel Gauthier, Michel Jébrak and Gilbert Prichonnet, who offered me teaching opportunities and research assistantships since 1988. Special thanks to Mrs. Michelle Laithier who has drawn certain figures.

This has been a challenging task delayed by contractual obligations. It would have been impossible to realize without the great patience and encouragement of my beloved Micheline and her dear children Eric and Karyne. I wish to dedicate years of struggle to them and to my unfortunate and courageous Carleton friend Jacques Bisson who committed suicide disappointed and confused while struggling to acquire a Master's degree in Geology at the age of Forty. Thanks to all for your stimulating moral support.

ORIGINAL CONTRIBUTION

During six months of field work in 1984 and 1985 I mapped two and one half 1:50,000 map sheets around Kelowna in the northern Okanagan region (82E/11, 82E/13, 82E/14) (Map 1). In 1986, I spent one month doing reconnaissance mapping in the southern Okanagan region and in the Okanagan Highlands of north-central Washington. I also spent one month working across the northern end of the Okanagan valley between Vernon and Sugar Lake on the west flank of the Pinnacles area.

Zircons, monazites, and thorites from 2 samples were dated utilizing the U-Pb procedures. Hornblende, biotite and muscovite from 19 samples were dated utilizing the K-Ar and/or the Ar-Ar methods. I performed all stages of rock crushing and grinding, mineral separation and mineral selection for all samples. Chemistry and mass spectrometry on zircons, monazites, thorites and K-Ar analyses on hornblendes and biotites were run by the staff of the GSC Geochronology Laboratory in Ottawa. Chemistry and mass spectrometry on hornblendes, biotites and muscovite for Ar-Ar analysis were performed by myself and D. A. Archibald at Queen's University in 1989, 1990 and 1991.

Thirty-six samples of Tertiary volcanic rocks were collected in the vicinity of Kelowna by myself and Don Francis. I performed the sample preparation but XRF analyses were run by S. T. Ahmedali, geochemist at McGill University.

Fifty-one sites were sampled for paleomagnetic work in 1985 and 1986. Preparation of the samples was done by Janet Baker then I spent a total of four months in 1986 and 1987 running demagnetization stages with the help of Jane Wynne and performing statistical treatments on programs available at the Pacific Geoscience Center in Sydney, British Columbia. A version of Chapter 7 has been published and partly written together with E. Irving. Contents of this chapter is totally scientific work performed by myself under the supervision of Ted Irving.

I spent ten days performing microprobe analysis at McGill University under the supervision of Mrs. Moyra McKimmon in 1988 and 1989. Serge Perreault taught me the methodology of the GEO-CALC program written by Berman, Brown and Perkins (1987).

Jean-Claude Mareschal provided two computer programs to treat heat flow data and model stress trajectories across south-central British Columbia.

I have benefitted from discussions with all colleagues cited above but I take full responsibility for the data and treatment of it in this thesis.

METHOD OF PRESENTATION

This thesis should represent 5 papers. A general introduction on the setting of the Okanagan valley is followed by; structural (Chapter 2), metamorphic (Chapter 3) and geochronological (Chapter 4) data in the northern Okanagan valley. Chapters 6 and 7 deal with the stratigraphy and paleomagnetism of Eocene lavas forming the Kelowna outlier. Chapter 8 reviews published geological and geophysical material including new data and speculates on constraints that may have led to Eocene extension in south-central British Columbia.

Due to the paper format certain chapters contain redundant text for which I apologize.

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Preamble

1. *Purpose of this study*

Mapping has been carried out in south-central British Columbia across the 119° meridian to decipher the tectonic setting of a major shear zone straddling the Okanagan valley. The work involved detailed structural and metamorphic analysis supplemented by sampling for geochemical, geochronological and paleomagnetic work.

Areas were selected at various structural levels across the shallow-dipping shear zone to document the progressive development of finite strain markers and examine indicators of the sense of shear (Chapter 2). Several geothermobarometric calculations were made on pelites within the shear zone to delineate the P-T history of the shearing event(s) (Chapter 3).

Isotopic dating with conventional methods (K-Ar, Ar-Ar, U-Pb) was carried out to constrain the age relationship between magmas stranded atop the shear zone and syn-kinematic metamorphic minerals (Chapter 4). The geochemical analysis of Eocene lavas contributes to our understanding of the evolution of magmatic sources and determines their mode of emplacement in the regional context of extension (Chapter 6). Paleomagnetic samples from these volcanic rocks were tested to decipher their pre- and post-tectonic positions relative to the North American craton and document their relative motion (Chapter 7).

The geological information obtained in the first chapters is integrated in a regional context and combined with available geophysical data from south-central British Columbia to illustrate certain geodynamic conditions that were speculatively prevailing in this region in the Early to Middle Eocene period (Chapter 8).

2. *Location and access*

Mapping in the northern Okanagan valley was first concentrated near Kelowna on two 1:50,000 map sheets (82 E/4 and 82 E/13 in the back pocket). Two smaller areas were also mapped north of Kelowna along the Coldstream valley and around the Trinity Hills near Vernon. The studied region spreads over a surface of 100 x 50 km starting at a distance of 70 km north of the U.S.A. - B.C. boundary (Fig. 1.1). Parts of the geological information collected near:

Kelowna have been used in the GSC open-file map no. 1969 (Tempelman-Kluit 1988).

Easiest access to Kelowna is by highway 3 from Hope connecting to highway 97 along the Okanagan valley. Bedrock exposures were found along highway roadcuts and canyon walls carved in the high plateaus on both sides of the valley. Spectacular sections of gneissic rocks occur along abandoned railway cuts on the south side of Mission Creek near Kelowna. High ridges bounding the valley are often capped by Cenozoic lavas and offer good exposures where recent logging cuts and roads were established. The main roads are in excellent repair and well traveled but branch roads are generally private but passable with the owner's authorization. Most logging roads have reasonable clearance and need usage of a four-wheel drive vehicle.

Maximum elevations reach 2.3 km (Big White Mountain) on the east side of the Okanagan valley and 2.6 km on the west side (Okanagan Range). The topographic slope is shallower on the east side down to elevations of 340m on the shore of Okanagan Lake which is about 650m deep (Mullins et al. 1990). The maximum relief in this area is therefore nearly 3 km and compares with those reliefs of the Coast Ranges and Death valley in southern California (Wernicke 1990).

3. *Historical overview*

Field investigations, in conjunction with geochronological analyses and seismic surveys, have led to the recognition of crustal-scale northerly-trending Eocene extensional faults and shear zones transecting the southern Omineca Belt in south-central British Columbia (Parrish et al. 1988; Brown et al. 1992). The present study examines the northern segment of one of these structures, the Okanagan valley fault (Parkinson 1985; Bardoux 1985; Tempelman-Kluit and Parkinson 1986) which links with the Eagle River fault further north (Journey and Brown 1986; Johnson and Brown 1989). Both fault segments exhumed the western flank of the Shuswap metamorphic complex (Fig. 1.1) and their geometry and tectonic history places constraints on cross-sections through the western boundary of the southern Omineca Belt.

The Shuswap complex consists of polydeformed rocks metamorphosed to the amphibolite facies exhumed as domal culminations by extensional faults. These faults created sharp metamorphic discontinuities for which various interpretations were given in the past. The Okanagan complex is a domal culmination underlying the Okanagan valley fault. Cairnes (1940)

recognized anatectic melts and proposed that the metamorphic assemblages and textures in the Okanogan complex were caused by the intrusion of Mesozoic batholiths. Waters and Krauskopf (1941) mapped an important mylonite zone along the Okanogan valley (described as a cataclasite) and concurred with Cairnes (1940) that gneissic layers were diapiric strain boundaries. Little (1961) mapped the metamorphic break crossing the Okanogan valley as a high-angle normal fault but the magnitude of this structure was minimized by Ross and coworkers (Ross 1972, 1974, 1981; Ross and Barnes 1972; Ryan 1973; Medford 1975; Ross and Christie 1969, 1979; Mathews 1981) who attempted to link structures and lithologies across the valley. Ross (1981) described mylonites within the Okanogan complex and suggested that noncoaxial strain markers indicated a northeast sense of shear making the shear zones eastward-verging thrusts.

The initial recognition of denudation faults must be credited to the late Richard Lee Armstrong. While *completing his graduate work in 1964*, Armstrong started accumulating a large geochronometric database guiding his analysis to what was to become a pioneering look at the relative age of metamorphic core complexes. While Cordilleran geologists spent most of their energy mapping compressional features (Misch 1960), Armstrong and Hills (1967) and Armstrong (1972) brought up several irrefutable arguments that became a break-through in the development of ideas on the existence of denudation structures juxtaposing young rocks on older polydeformed Paleozoic strata in the eastern Great Basin. For a decade Armstrong stood alone claiming that metamorphic domes of the Omineca Belt were exhumed by extensional processes.

The presence of low-angle normal faults at the southern end of the Okanogan valley had already been promoted by Snook in 1965 who proposed that this terrain boundary was subjected to rapid *extension* and *denudation* along *mylonitic zones* (described as a ductile strained rock). Snook (1965) identified *westerly directed kinematic indicators* and developed the concept of simultaneous broad arching, fracturing and brecciation of the strained boundaries. In essence, he only missed an isotopic database to demonstrate the precursory model of a Cenozoic "metamorphic core complex" as they were defined by Armstrong (1972). Snook should therefore be credited with the recognition of many of the physical aspects characterising such *...Fanes*.

In the early 1980's the concept of crustal-scale low-angle normal faulting gained much attention (Crittenden et al. 1980; Wernicke 1981) and field data accumulated in favor of the

existence of these features in the southern Omineca Belt (Read and Brown 1981; Harms 1982; McMechan 1983; Lane 1984; Parrish 1984; Parkinson 1985; Bardoux 1985; Monger 1986; Tempelman-Kluit and Parkinson 1986; Parrish et al. 1987a, 1988; Johnson 1988, 1989a; Coleman 1989; Moore 1989; Coleman and Parrish 1990; Carr 1990, 1991, 1992).

Uranium-lead zircon ages of syn-extensional granites (Solberg 1976; Carr et al. 1987; Parrish et al. 1988; Carr and Parkinson 1989; Carr 1990) have established the importance of Eocene histories of ductile shearing and subsequent brittle deformation in most extensional faults of south-central British Columbia. Potassium-argon mineral dates consistently record an early Tertiary thermal history in the high-grade rocks in the footwall of extensional faults (Baadsgaard et al. 1961; Ross 1974; Mathews 1981; Archibald et al. 1984), whereas the low-grade brittlely deformed hanging wall rocks preserve Mesozoic cooling ages (Parrish et al. 1988). Abundant volcanic flows are stranded atop the extensional faults and are also contemporaneous with the Eocene shearing event.

CHAPTER 1 Tectonic setting of the southern Omineca Belt.
Regional geology and possible stratigraphic correlations
of the northern Okanagan region

1. *Terrane accretion*

The Mesozoic - Early Tertiary tectonic evolution of the southern Omineca Belt relates to the collision between two outboard Superterrane (Intermontane and Insular) and the ancestral continental margin of western North America (Fig. 1.1). This setting was apparently analogous to the western Pacific comprising offshore arcs and marginal basins (Monger 1982).

The Intermontane Superterrane is made of Paleozoic to Triassic arc and oceanic terranes (*Eastern crust and basin*, *Quesnellia arc*, *Cache Creek crust and basin*, *Stikinia arc*) that coalesced before they accreted onto North America in the Jurassic (Monger et al. 1982). Following obduction, parts of the Intermontane Superterrane compressed and deformed the ancestral North American rise and slope sediments which were thickened and buried to depths of over 25 km (Journey 1986). By the Late Jurassic the internally deforming and thickening outer edge of the paleo continental margin together with the sutured Intermontane Superterrane were uplifted and transported eastward as a coherent crustal scale thrust sheet referred to as the Selkirk allochthon (Fig. 1.1; Read and Brown 1981). Large calc-alkaline batholiths intruded the southern Omineca Belt in the Jurassic (Armstrong 1988). Zircons in these plutons yield crystallisation ages of 160-165 Ma. This period of typical continental arc magmatism (Brandon et al. 1991) was followed by a period of tectonic and magmatic quiescence.

The Insular Superterrane was sutured westward of the Intermontane Superterrane in the Cretaceous (Fig. 1.1; Monger 1984) under transpressive conditions which led to large intracontinental strike-slip faulting (Gabrielse 1985; Price and Carmichael 1986). Crustal thickening culminated in the southern Omineca Belt in the Late Cretaceous to Paleocene (Carr 1990) creating a large plateau of unknown extent. This plateau provided abundant clastic sequences accumulated in the foreland foredeep (Price and Mountjoy 1970) and along the west coast of the Early Tertiary margin (Heller et al. 1987). Calc-alkaline plutons formed from mid-Cretaceous onward (110 - 115 Ma). These bodies mainly lie east of the mid-Jurassic batholiths

Figure 1.1:

Tectonic map of the southern Omineca Crystalline Belt, modified from Brown et al. (1986), Parrish et al. (1988), Johnson and Brown (1989). Insets locate the geologic belts of the Canadian Cordillera and the Intermontane (IM) and Insular superterrane (IS) modified from Monger et al. (1982). Other belts are (F) Foreland fold and thrusts (O) Omineca crystalline and (C) Coast Plutonic. The Omineca Belt is bounded on the east by the southern Rocky Mountain trench and on the west by the Okanagan Valley - Eagle River normal fault system (OVF/ERF) and terrane accretion boundary (TAB). Intermontane Superterrane lies west of TAB and is sutured westly to the Insular Superterrane via the Early Mesozoic subduction complex of the Bridge River and Cascadia Terranes (BRT). Late Mesozoic - Early Tertiary strike-slip faults overprint the suture zone. These are the Fraser - Straight Creek fault (FSF), Yalakom (YF), Hozomeen (HF) and Pasayten (PF) faults. Horizontal hatched pattern around Revelstoke locate the Shuswap metamorphic complex which dissects the TAB. Three crustal layers compose the Omineca Belt (Carr 1990). Basement rocks of the Lower Crustal layer (vertical lines) are exposed in the Monashee complex (MC) and in the Malton complex (M). The Middle Crustal layer (horizontal hatching) consists of metamorphic rocks characterized by the presence of abundant granitic sills and pegmatites and Early Tertiary K-Ar mica dates. Unpatterned area west of the Rocky Mountain Trench and rocks of accreted terranes (QT: Quesnellia; ET: Eastern) represent the Upper Crustal layer. Mesozoic - Paleocene thrust faults bounding or cutting crustal zones are the Monashee décollement (MD) and the Gwillim Creek shear zone (GCS). Thin black arrow along TAB indicates direction of emplacement of the Intermontane Superterrane in the Quesnel Lake area (Rees and Terri 1983). Amphibolite - grade culminations of the Middle Crustal layer occur between the Monashee and Malton complexes (Johnson and Brown 1989; Scammell 1990), south of the Monashee complex (Carr 1990) and further south in the Kettle - Grand Forks (K), Okanagan (OC), Valhalla (V), and Priest River (PR) complexes. These culminations are bounded by Eocene normal shears namely the Okanagan Valley - Eagle River (OVF/ERF), Greenwood (GWF), Granby (GF), Kettle (KF), Columbia (CF), Valkyr (VS), Slocan Lake (SLF), Newport (NF), and Purcell Trench (PTF) fault systems. Thick black arrows indicate stretching direction on the western and eastern flanks of the Omineca Belt (compiled from Chapter 2 and Carr et al. 1987).

and show REE distributions typical of a "within-plate" tectonic setting (Brandon et al. 1991). These rocks were emplaced continuously while crustal shortening and thickening culminated up to 60 Ma ago and some may have formed as anatectic melts (*see* section 3.2.2.1).

2. Regional geology

The Omineca Belt (Fig. 1.1) forms the hinterland of the Canadian Cordilleran orogen. It exhibits Proterozoic - Paleozoic sediments of the ancestral North American margin (Windermere Supergroup) resting on attenuated Archean crust (Armstrong 1983; Armstrong et al. 1991). The Omineca Belt also comprises Proterozoic - Early Jurassic arc and oceanic rocks of the Intermontane Superterrane imbricated into ancestral North American rocks between the Jurassic (Brown et al. 1986) and the Paleocene (Brown 1989; Brown and Carr 1990). It has been postulated that, by latest Paleocene, the southern Omineca Belt was underlain by thick imbricated basement layers creating a crustal welt approximately 60 km thick (Ranalli et al. 1989). Carr (1990) demonstrated that within 1-3 Ma of the cessation of compression the welt was denuded by moderate to low angle north-trending normal faults (Fig. 1.1). These faults have exhumed three different levels of the overthickened crust (Johnson 1989b; Carr 1990; Brown and Carr 1990) referred to as Basement, Middle Crustal and Upper Crustal layers. These layers record different deformation and cooling events and are separated from each other by regional-scale compressional and/or extensional shear zones.

Prior to extension the Selkirk allochthon was lying across the entire width of the Omineca Belt (Fig. 1.1). Eastward thrust Selkirk rocks and hosting early Proterozoic metasedimentary rocks of a shallow platformal environment unconformably overlying crystalline basement of North American affinity are exposed in the Monashee and Malton complexes (Fig. 1.1). Thrusting of the Selkirk allochthon was mainly accommodated on the westerly rooted Monashee décollement (Read and Brown 1981; Journeay 1986; Scammell 1986; Journeay and Brown 1986; Bosdachin 1989; Harrap 1990; Coleman 1989; Carr 1990; Brown et al. 1992). Rocks of the Selkirk allochthon reached upper amphibolite conditions upon their thrusting.

The Monashee décollement is a crustal-scale compressional shear zone of great areal extent that is part of a Mesozoic to Paleocene thrust system (Brown et al. 1986, 1992) that feeds

eastward into the sole thrust(s) of the Rocky Mountain Foreland Belt (Fig. 1.1; Price 1981; Journeay 1986; Parrish et al. 1987a; Journeay and Parrish 1989; Brown et al. 1992). Near Thor-Odin, the shear zone contains syn-kinematic leucogranites and pegmatites containing zircons that provided concordant ages of 62.1 ± 0.3 Ma and 59 ± 0.3 Ma (Carr 1990). Zircons in pegmatites that crosscut the décollement provided a concordant age of 58 ± 0.3 Ma (Carr 1990).

Rocks of the basal part of the Selkirk allochthon and of the North American basement are exposed in a domal culmination, the Shuswap metamorphic core complex (Fig. 1.1), which was exhumed in the Late Paleocene - Early Eocene by shallow-dipping normal faults. These faults have the particularity that they breakaway from where basement rocks are exhibited at the surface which is assumed to represent the crest of the ancestral crustal welt (*see* Brown et al. 1986). This is consistent with the fact that normal faults are outward-dipping from the basement exposure and juxtapose rocks of Upper Crustal levels against rocks of Middle and Basement crustal levels (Brown and Carr 1990; Carr 1992). According to Carr (1990) extensional structures are rare in the Basement layer but the Middle Crustal layer is characterized by penetrative early Cenozoic plastic strain (Parrish et al. 1988) and abundant intrusion of early Cenozoic leucogranites and early Cenozoic cooling history (Carr 1990). Rocks of Upper Crustal level systematically occur in the hanging walls of extensional faults and exhibit pre-Middle Jurassic deformation, Mesozoic plutonism and cooling dates showing that they were uplifted and maintained at high structural levels after the accretion of the Intermontane Superterrane before the Late Cretaceous (Read and Wheeler 1976; Brown and Tippet 1978; Parrish and Armstrong 1983; Parrish and Wheeler 1983). The Upper Crustal layer is also characterized by brittle extensional faulting. Seismic reflection data acquired by the Cordilleran Transect LITHOPROBE program demonstrated that compressional features of the southern Omineca belt extend presently to lower crustal levels. Reflectors related to extensional structures seem to flatten at mid-crustal levels except the Shuswap Lake fault that extends to the bottom of the lower crust (Brown et al. 1992; Cook et al. 1992).

The Okanagan valley exhibits the western boundary of the Middle Crustal layer underlying the Shuswap metamorphic complex comprising amphibolite grade gneisses interleaved with large granitoid masses altogether defining a metamorphic culmination referred to as the Okanagan complex (Fig. 1.2). This culmination was exhumed as a tectonic window through the greenschist-

grade assemblage of Upper Crustal level rocks related to the accreted Quesnellia Terrane (Fig. 1.1). Rock types of the Okanagan complex and Quesnellia are clearly displayed across the north-trending Okanagan depression underlain by the westerly-dipping Okanagan valley normal fault (Fig. 1.1; Parkinson 1985; Bardoux 1985; Tempelman-Kluit and Parkinson 1986). This fault connects northward with the Eagle River fault (Journey and Brown 1986; Johnson and Brown 1990) and southward with the Okan(o)gan fault (Snook 1965) which altogether form the longest structure of the vast array of normal faults developed in the southern Omineca Belt.

Estimated magnitudes of extension appear to decrease from 30% to 15% between the southern and northern limits of the Shuswap culmination (Brown and Journey 1987; Parrish et al. 1988). This may, in part, be related to a transfer of motion to major transcurrent faults bounding the extended domain. The Okanagan valley-Eagle River fault system is located *en échelon* between two large Cenozoic transcurrent faults, the Fraser-Straight Creek fault to the west and the Rocky Mountain Trench to the east (Fig. 1.1; Price and Carmichael 1986).

In summary, the southern Omineca Belt comprises antiformal metamorphic culminations exhumed by successive extensional structures leading to crustal scale denudation down to mid- to lower crustal levels. The Okanagan complex is a window into the western flank of the southern Omineca Belt. It comprises large Mesozoic and Cenozoic batholiths, eugeosynclinal Proterozoic and younger roof pendants, Early Proterozoic basement gneisses (Vaseaux Formation; Armstrong et al. 1991), and Cenozoic leucogranites which were probably syn-extensional anatectic melts. The Okanagan valley fault with its surface trace of nearly 450 km between Adams Lake in the north (Johnson and Brown 1990) and the Spokane valley (Snook 1965) in the south is comparable to the Wasatch fault zone which is the longest normal fault system of the Basin and Range Province in the northwestern United States (Machette et al. 1991).

3. *General geology of the northern Okanagan region*

Figure 1.2 is a geological compilation of the western boundary of the southern Omineca Belt. Assemblages of Quesnellia, referred to as the Upper Crustal layer, occur mainly west of the Okanagan valley whereas amphibolite-facies rocks and leucogranites of the Okanagan complex, the Middle Crustal layer, lie generally on the east side of the valley. Both assemblages

Figure 1.2: *Geological map of the Okanagan region. SB: Similkameen batholith; OB: Okanagan batholith; PB: Pennask batholith; OgC: Okanagan gneiss complex; C: Coryell intrusive suite; HR: Hunters range; T-O.: Thor-Odin. Early Tertiary normal faults are OVF: Okanagan Valley fault; GWF: Greenwood fault; GF: Granby fault; BF: Beaven fault. Stars locate new U-Pb samples near Kelowna and Mabel Lake and asterisks locate new ^{40}Ar - ^{39}Ar samples near Kelowna and Beaverdell (see Chapter 4). Map modified from Little (1961) and Parrish et al. (1988).*

	Miocene Plateau Basalts		Cretaceous Plutonic Complexes
	Middle Eocene syenite		Jurassic Plutonic Complexes
	Late Eocene dacites		Palaeozoic metasediments and metavolcanics
	Middle Eocene trachyandesites		Mylonitic gneiss
	Eocene granites		Denudation faults
			Normal faults

are cut or overlapped by Eocene and Miocene volcanosedimentary sequences (*see* Chapters 4, 6, 7). The following sections outline the stratigraphic sequences observed in both crustal layers mapped near Kelowna and Enderby in the northern Okanagan region.

3.1 Quesnellia Terrane near Kelowna (Upper Crustal layer)

The oldest rocks of Quesnellia Terrane near Kelowna comprise mainly mafic volcanic rocks interdigitated with siltstones and limestones containing Triassic conodonts (Tempelman-Kluit pers. comm. 1987). These stratified rocks occur mainly along Lambly Creek and north of Kelowna but they reappear in a tectonic klippe east of Kelowna (Map 1). Only one north-trending steep foliation is recorded in these rocks near Traders Cove and is interpreted to be axial planar to tight upright folds plunging shallowly to the north.

Large plutons intrude the Early Mesozoic strata and are deformed by discrete, annealed shear zones. The plutons are calc-alkaline hornblende-biotite granodiorites to diorites that are part of the Okanagan and Pennask batholiths (Daly 1905, 1912; Rice 1960; Peto 1973 a,b) which preserve mid-Jurassic Rb-Sr and K-Ar mineral dates (White et al. 1968; Roddick et al. 1972; Soregaroli and Christopher, unpublished; Medford 1975; Fox et al. 1976, 1977; Okulitch et al. 1975; Read and Okulitch 1977). These data indicate that rocks of the Upper Crustal layer were not reheated above 500°C ($530 \pm 40^\circ\text{C}$ mean closure temperature of Ar diffusion in hornblende; Harrison 1981; McDougall and Harrison 1988) since their emplacement. This also suggests that deformation in hosting Late Paleozoic-Mesozoic strata is older than mid-Jurassic which is consistent with the field observations whereby the plutons cut the axial planar fabric.

Leucocratic hornblende-biotite granodiorite bodies intrude the Okanagan and Pennask batholiths north of Summerland (Fig. 1.2; Peto 1973a). Their petrochemistry suggests that they bear no genetic affiliation to the evolving suite of the Okanagan batholith (Peto 1973a) which is also supported by younger K-Ar biotite ages ranging between 135-150 Ma (Peto 1973b). The latter may, however, represent a cooling age suggesting that the pluton and its hosting lithologies were not reheated above 280°C (mean closure temperature of Ar diffusion in biotite; McDougall and Harrison 1988).

Syenitic stocks of early Eocene age (approx. 52 Ma) have been found cutting through the

Okanagan batholith north and northwest of Summerland and along Okanagan Lake (Little 1961; Armstrong and Peto 1981; Ewing 1980, 1981a, b). Most of these Eocene syenites are grouped in the Coryell suite which forms large bodies in northeastern Washington and around Arrow Lake, 100 km east of the Okanagan valley (Little 1983; Fig. 1.2). Few Eocene stocks have been mapped in the northern Okanagan region. Stocks of feldspar porphyries occur in Whiteman Creek west of Vernon and near Terrace Mountain northwest of Kelowna (Fig. 1.2). The latter body provided a concordant Rb/Sr isochron date of 59 ± 2 Ma (Medford et al. 1983) and a K-Ar date of 47 Ma (Wanless et al. 1978). Medford et al. (1983) suggested that the discrepancy of nearly 12 Ma between Rb-Sr and K-Ar dates indicates a pattern of *slow cooling* following plutonic emplacement. Another Eocene stock, the Shingle Creek porphyry (Bostock 1941), underlies the White Lake basin (Church 1973) in the southern Okanagan region west of Penticton and provided a K-Ar date of 52.4 ± 2 Ma (Church 1979b).

Several outliers of Eocene lavas occur on the shoreline of Okanagan Lake (Fig. 1.2) where they are unconformably overlying Early Mesozoic strata and Mesozoic plutons. The Eocene lavas are mainly trachytes (Church 1973) overlain by thick alluvial sequences. Amphiboles and biotites in the lavas provided K-Ar dates ranging between 53-48 Ma (Church 1980a; see Chapters 4, 6). It has been proposed that these volcanic rocks are cogenetic to the Coryell suite (Church 1973; *but see* Chapter 6).

3.2 The Okanagan complex near Kelowna (Middle Crustal layer)

The east side of the Okanagan valley near Kelowna comprises mainly large granitoid bodies of variable compositions exhibiting a wide spectrum of ductile-to-brittle structures. In the less deformed zones it is clear that granitoids cross cut or were injected concordantly with the layering of the hosting paragneisses. Altogether these lithologies form the Okanagan complex.

Although they are generally poorly exposed, the paragneisses appear to form less than 20% of the Okanagan complex near Kelowna but their volume increases sporadically to 50% between large orthogneissic masses towards the northern and southern ends of the Okanagan valley. The largest volumes of paragneisses occur at Okanagan Falls (Fig. 1.2; Cairnes 1940; Bostock 1941; Little 1961; Christie 1973; Medford 1975) and south of Osoyoos near the

International Boundary (Fig. 1.2; Bostock 1941; Waters and Krauskopf 1941; Ryan 1973). Paragneisses also are more voluminous north of Kelowna where the Okanagan complex merges with the Shuswap complex near Enderby (Fig. 1.2; Jones 1959; Okulitch 1979; Johnson 1988).

3.2.1 *Paragneisses*

Most paragneissic units near Kelowna (Pgn on Map 1) are regionally consistent with those mapped in the southern Okanagan region (Christie 1973; Medford 1975) but stratigraphic tops are undetermined and lateral continuity is absent. Rare evidence suggests that paragneisses near Kelowna underwent at least two phases of isoclinal folding which is consistent with the structural history established in the southern Okanagan region (Christie 1973; Ryan 1973; *see* Chapter 2). The most distinctive units of the paragneissic assemblage are diopsidic marble, calcsilicate schists and amphibolites interlayered with semipelites. These form discontinuous lensoid bodies that are approximately 20m thick and 20m long in the direction of extensional transport (*see* Chapter 2) on either sides of Mission Creek (Map 1). These rocks are at upper amphibolite grade and may be tentatively correlated with subunits of the Vaseaux Formation as defined east of Okanagan Falls 40 km south of Kelowna (Fig. 1.2; Bostock 1941; Christie 1973). However, critical markers of the Vaseaux Formation such as quartzites and sillimanite-garnet-biotite-quartzofeldspathic paragneisses were not observed near Kelowna as they may be thinned by shearing and are probably hidden in treed areas.

Lithologies described in the Vaseaux Formation (Christie 1973) are repeated by sheared contacts transecting isoclinal fold limbs forming an estimated section 1200m thick. It is expected that intense shearing has transposed and rotated several units of the Vaseaux Formation into near parallelism (Christie 1973). Furthermore, these lithologies have been injected by abundant leucocratic granitic and pegmatitic material (Christie 1973).

Micaceous schists of the Vaseaux Formation near Okanagan Falls contain zircons with an upper intercept at 1899 ± 49 Ma (2) that probably indicate an Early Proterozoic source age related to cratonic North America (Armstrong et al. 1991) but may not be diagnostic of the age of the protolith. Nevertheless, because Sr, Pb and Nd isotopic data are systematically ranging between 1.9-2.2 Ga old it has been proposed (Armstrong et al. 1991) that paragneisses of the

Vaseaux Formation might represent pre-Purcell lithologies.

On the basis of lithological similarities, it appears that the Vaseaux Formation resembles the Fawn Lake assemblage exposed immediately above the Monashee décollement on the west flank of the Thor-Odin culmination east of Vernon (Fig. 1.2; Carr 1990). The Fawn Lake assemblage has been tentatively correlated with the semipelite-amphibolite (SPA) division which systematically surrounds the Monashee complex (Johnson 1988; Scammell 1991) and has been correlated with the Late Proterozoic Horsethief Creek Group (Windermere Supergroup; Poulton and Simony 1980; Raeside and Simony 1983; Pell and Simony 1984; Johnson 1988; Carr 1990; Scammell 1991). Correlating paragneisses of the Kelowna area with Windermere or Pre-Purcell stratigraphy is presently based on *ad hoc* assumptions and cannot be clearly established in the field because of poor exposures and extreme boudinage.

3.2.2 *Orthogneisses and leucogranites*

Eighty percent of the Okanagan complex exposed near Kelowna is made of quartzofeldspathic gneisses (Ogn on Map 1) which comprise variably strained leucocratic hornblende + biotite ± muscovite granite and pegmatite sheets that extend along several kilometers engulfing large paragneissic and amphibolite inliers (Am on Map 1). The plutonic rocks are largely undated. They clearly penetrate the paragneisses and become progressively concordant to the gneissic layering at the structural top of the Okanagan complex. Several pegmatitic bodies form long isoclinal sheets disrupted by penetrative extensional shear planes. The crosscutting relationships between paragneisses, pegmatites and granites and the apparent softening of the latter from field aspects suggest that these were injected during the progressive development of penetrative strain fabrics.

Several granitic bodies infolded with the paragneisses of the southern Okanagan complex contain several zircon populations which provide a record of concordant crystallization ages ranging between 210 - 53 Ma (Okulitch *in* Medford 1975; Parkinson 1985; *see* Chapter 4). Some of these igneous sheets exhibit as many as four non-coaxial phases of folding (Ryan 1973; Christie 1973; Medford 1975). Parkinson (1985) dated zircons from a pre-F2 hornblende-biotite granodiorite (Ryan 1973) cutting paragneisses related to the Late Permian Anarchist Group east

of Osoyoos (Fig. 1.2; Daly 1912; Little 1961). This granodiorite is 201.5 ± 2.2 Ma which nearly corresponds to the fossiliferous age of the hosting paragneisses. This data suggested that the major D2 structures in the southern Okanagan complex were post-200 Ma. Biotite-hornblende granodioritic sheets interpreted as syn-D2 intrusions (Christie 1973) are interlayered with paragneisses of the Vaseaux Formation east of Vaseaux Lake (Fig. 1.2). These rocks were considered to be pre-Jurassic (Christie 1973) before Parkinson (1985) found four magmatic zircon fractions that provided nearly concordant $^{206}\text{Pb}/^{238}\text{U}$ ages of 95.8 ± 1.2 Ma indicating that Christie's (1973) D2 event may be post- early Late Cretaceous (see Chapter 3).

The isotopic ages of various granitic sheets interlayered in the southernmost part of the Okanagan complex are apparently not always consistent with the deformation sequence observed in this small area. For example, the syn-D2 body at Vaseaux Lake described above is "younger" than Ryan's (1973) syn-D3 biotite quartz monzonite near Osoyoos (Fig. 1.2) which contains two nearly concordant zircon fractions dated at 160 ± 1.8 Ma and 161.4 ± 1.8 Ma with $^{207}\text{Pb}/^{206}\text{Pb}$ ages ranging between 173-178 Ma (Parkinson 1985). Certain samples analysed by Parkinson (1985) have probably suffered important Pb loss related to high strain conditions. Regionally, important metasomatic reactions from the Late Cretaceous onward have caused Pb loss or new crystal growth in high strain zones on the eastern flank of the Shuswap complex (Carr 1990).

Assuming that the geochronology data obtained by Parkinson (1985) is correct then it may be argued that the structural history established in the southern Okanagan region may be flawed (Christie 1973; Ryan 1973; Ross and Christie 1979). Nevertheless, the isotopic data obtained from diverse granitoid bodies of the Okanagan complex suggests that several of these rocks cooled below 750°C (closure temperature for Pb diffusion in zircon) in the Mesozoic.

Middle Crustal rocks of the Shuswap complex are invaded by large igneous bodies containing Early Eocene magmatic zircons and are referred to as the Ladybird suite (Carr et al. 1987; Parrish et al. 1988; Sevigny et al. 1989). These plutons are mainly calc-alkaline biotite granitoids comprising $\pm$ muscovite $\pm$ garnet $\pm$ sillimanite bearing quartz monzonite and granitic batholiths, stocks, sheets and pegmatites (Carr 1990). Field relationships, geochronological and geochemical data suggest that the Ladybird granitoids formed as anatectic melts during waning stages of compression in the Omineca Belt and were progressively strained under extensional

conditions (Carr 1990). Syn- to post-extensional Ladybird granitic melts span a period of approximately 10 Ma starting at 62.1 ± 0.3 Ma at the latitude of the study area (Parrish et al. 1988; Carr 1990). They get older to the north (Scammell 1991).

Besides Ladybird-type granites, syn- to post-extensional late Early Eocene syenitic stocks have also been found in the southern Okanagan complex. The undeformed Allendale syenite located east of Okanagan Falls cuts the gneissic foliation and provided a K-Ar age of 53 Ma (Church et al. 1983). Nearby, a mylonitized rhomb porphyry, concordant with the gneissic layering provided two K-Ar ages of 45.3 Ma (biotite) and 48.3 Ma (hornblende) (Ross 1974). This rock also provided two nearly concordant fractions of magmatic zircons dated at 51 ± 1 Ma (Parkinson 1985) interpreted as a magmatic age. The extensional fabrics overprinting this rock suggest that part of the ductile shearing was accommodated in the gneisses of the Okanagan complex in the middle Eocene though it may have started earlier.

In northeast Washington, biotite-granites of the Keller Butte suite form a large part of the Okanagan complex. Because they yield Early Eocene K-Ar dates these bodies have been correlated with the Ladybird suite (Holder and McCarley Holder 1988). The geochemistry of the Keller Butte plutons indicates that indeed they formed as intracontinental melts but no absolute age is available for them. It is noticeable that these rocks are geochemically indistinguishable from isotopically dated Mid-Cretaceous granites throughout the southern Omineca Belt (Brandon et al. 1991) therefore the Keller Butte suite might as well be mid-Cretaceous.

Because there are several granitoids containing Eocene zircons throughout the southern Omineca Belt (Parkinson 1985; Parrish et al. 1988; Carr 1990, 1992; *see* Chapter 4) it is tempting to link most of these rocks with the Ladybird suite (*see* Plate 1 of Parrish et al. 1988). However, the largest plutonic bodies southeast of Kelowna are feldspar-phyrlic hornblende-biotite granodiorites which resemble in composition the 165 Ma old Okanagan batholith (Roddick et al. 1972; Peto 1973) and the 77 Ma old Whatshan batholith (Carr 1990; 1991) and contain Mesozoic crystalline zircons (Parkinson 1985).

The contact relationships between most igneous sheets and paragneisses in the Okanagan complex are generally knife sharp at the bottom and grade to narrow zones of migmatite at the upper contact. In certain cases the transition from paragneisses to strained granitoids is so

gradational that a contact cannot be distinguished and regardless of the size of the igneous body there is never a recognizable thermal aureole at their contact. Using mappable metamorphic isograds Ryan (1973) demonstrated that large leucocratic biotite quartz monzonite bodies in the southernmost Okanagan region are spatially confined to paragneisses metamorphosed to upper amphibolite grade and are overprinted by extensional ductile fabrics. Because they lie mainly in upper amphibolite rocks with diffused contacts it appears that many igneous sheets throughout the Okanagan complex are probably anatectic. These rocks may have formed by similar processes but diachronously from the typical Ladybird suite (Carr 1990; 1992).

With a systematic geochronometric database, Carr (1990) has demonstrated that Ladybird leucogranites formed during late compressional events. Similarly mid-Jurassic plutons, such as MBX32 and MBX39 (Fig. 1.3), formed while the Intermontane Superterrane accreted onto North America (Brown et al. 1986; Armstrong 1988) and may also be anatectic melts. Based on field relationships, Carr (1990) suggested that Eocene melts facilitated the development of extensional shear structures across the southern Omineca Belt. If this causality between anatexis and extension holds, given proper tectonic and thermal conditions (England and Thompson 1986), then it could be expected that extensional structures in the southern Omineca Belt may have existed in the mid-Jurassic as there were large volumes of melt formed. There is no evidence yet for pre-Eocene extension in south-central British Columbia.

3.2.2.1 Geochemistry of some Okanagan granitoids

Undated granitoids lying in the Upper and Middle Crustal layers in the northern Okanagan valley have been compared with dated bodies to test their bulk compositional variations with time. Upper crustal bodies MBX32 and MBX39 provided respectively hornblende and biotite K-Ar ages of 158 ± 2 Ma (*see* Chapter 4). A concordant zircon fraction in sample MXE19 and two thorite fractions in sample MX261 in the Middle Crustal layer are 56 ± 0.3 Ma and 51 ± 0.5 Ma respectively (*see* Chapter 4). Two other biotite leucogranites interlayered in the Okanagan complex near Kelowna (samples MX27, MX43) are undated. Regardless of their crustal layer these bodies define a linear compositional trend that crosses the calc-alkaline/tholeiitic field on the SiO_2 versus $\text{FeO}^*/\text{FeO} + \text{MgO}$ diagram (Fig. 1.3). Though this diagram cannot serve as a

source discriminant, it is noticeable that the trend followed by the Okanagan granitoids closely matches the range of compositions of the Whatshan batholith (W on Fig. 1.3) that provided a 77 Ma U-Pb age (Carr 1990). The Whatshan batholith occurs in the footwall of the breakaway of the Okanagan valley fault, the Beaven normal fault (Cook et al. 1992), 60 km east of Kelowna (Carr 1990, 1991). Both Eocene bodies MXE19 and MX261 (Figs. 1.2, 1.3) are interpreted as Ladybird S-type granites (*see* Chapters 2, 4; Carr 1990, 1992) and plot as the most evolved members of the trend together with the most evolved member of the Whatshan suite (Brandon et al. 1991) which may be therefore another Ladybird rock.

Rare-earth elements from the granitoids reported on Figure 1.3 were analysed and plotted in ORG normalized spidergrams (68-74% SiO₂; Fig. 1.4; Pearce et al. 1984). These rocks are indistinguishable from one another and they all show enrichments in Rb, Th, Ta, and Ce that are typical of calc-alkaline melts produced in a "within-plate" setting (Pearce et al. 1984). These profiles overlap generally with those of mid-Cretaceous plutons intruding the southern Omineca Belt (Brandon et al. 1991). Thus, undated leucogranites of the Okanagan region may have the characteristics of both Mesozoic and/or Cenozoic granitoids. Therefore, without isotopic age it is impossible to establish which granitoids in this region are time correlative. It would be dangerous to accept geochemical similarity as evidence for contemporaneity.

Figure 1.4 outlines an interesting feature of the Okanagan granitoids that may explain both their range in composition and the variety of U-Pb zircon ages recovered from them (Parkinson 1985). Six amphibolitic rocks hosting the granitoids are illustrated on the REE profiles. These amphibolites contain at least 80% hornblende with severely uranitized pyroxene cores. These rocks form lensoid bodies intimately related to paragneisses in the field. Their protolith is unknown but they may represent ancient calc-alkaline diabase dikes. The composition of the amphibolites generally overlaps those of the granitoids. The latter are slightly enriched in the most incompatible elements (Rb, Th, Ta, and Ce) probably due to selective crystallization of biotite, hornblende and K-feldspar. Late alteration to chlorite and sericite has probably reduced the Y and Yb contents of K-silicate minerals in both amphibolites and granitoids (Pearce et al. 1984). The chemical similarities with lithotypes of the Okanagan complex may suggest that leucogranites have inherited their compositions directly from their hosting lithologies consistent

Figure 1.3: *FeO/FeO + MgO versus SiO₂ diagram (after Miyashiro 1974) showing that both Mesozoic and Cenozoic granitoids of the northern Okanagan area (dots) exhibit compositional trends that are similar to those of the Late Cretaceous Wathshan batholith (Carr 1990; 1991) located east of the Okanagan valley (data from Brandon et al. 1991). Eocene magmas of the Okanagan suite plot as the most evolved of the compositional trend and are interpreted as anatectic melts from field relationships. The range in silica contents for different lithologies - QD-D: Quartz diorite to diorite; QMD: quartz monzodiorite; GD: granodiorite; Gr: granite. See Figure 1.2 for sample locations. Sample E19 is 56 Ma (U-Pb, zircon), sample 261 is 52 Ma (U-Pb, zircon), sample MBX32 is 160 Ma (⁴⁰Ar/³⁹Ar, biotite), and sample MBX39 is 159 Ma (⁴⁰Ar/³⁹Ar, biotite).*

Figure 1.4: *ORG-normalized rare earth element spidergrams of seven plutonic bodies underlying the Upper Crustal layer and the Middle Crustal layer in the northern Okanagan region. Compositional trends are typical of within plate granitoids (Pearce et al. 1984) formed in attenuated continental lithosphere or anatectic melts with crustal contaminants. These profiles resemble those of mid-Cretaceous plutons intruding the southern Omineca Belt (Brandon et al. 1991). Note that REE abundances of amphibolitic rocks of the Middle Crustal layer overlap the granitoids compositions and further indicates that they are the likely source of the anatectic melts.*

with the field relationships and the idea that leucogranites are anatectic melts.

It may be proposed that compositional variations amongst leucogranites of the northern Okanagan region, including Ladybird magmas, reflect the composition of hosting lithologies which are mainly amphibolitic (metavolcanic rocks) and/or metasedimentary rocks. When tested by field relationships it appears that amphibole-bearing granitoids occur mainly where mid-crustal lithologies are dominantly amphibolitic, as in the Okanagan region, whereas S-type granitoids, such as the typical Ladybird suite, are hosted in metasedimentary rocks (Carr 1990, 1992). Perhaps, differing granitoid facies interlayered in the Okanagan complex and their gradual changes in composition from amphibole-biotite to biotite $\pm$ muscovite granitoids resulted from extended reaction histories as invoked by Speer (1987). Textural relationships in thin sections demonstrate that biotite pseudomorphs generally replace euhedral hornblende in the Okanagan granitoids. It is envisioned that magmas interlayered in the northern Okanagan complex began crystallizing as *in situ* melts from Liquid = Hornblende reaction to a Liquid + Hornblende = Biotite reaction. Such changes could be brought about by movement of magma, adding or removing melt phases of residual liquid phases, or an influx of water from the country rocks under decreasing temperature conditions and increasing fO_2 (Speer 1987).

3.3 Quesnellia terrane in the Enderby region (Upper Crustal layer)

The Enderby region is located in the northernmost part of the Okanagan valley (Fig. 1.2). Low- to medium-grade metavolcanic and metasedimentary rocks are exposed in a large klippe underlying the Trinity Hills valley east of Enderby (Fig. 1.2; see Chapter 2). Similar rocks are exposed west of Vernon and comprise upper Paleozoic (Nicola Group) to Jurassic strata (Jones 1959; Okulitch 1979) of the Quesnellia Terrane cut by Middle Jurassic granodiorite and unconformably overlain by Tertiary volcanic and sedimentary rocks. Altogether these rocks resemble those lying west of Kelowna and relate to the Upper Crustal layer. These rocks exhibit only one steep foliation that is axial planar to NW plunging tight folds and is locally overprinted by metamorphic minerals in the contact aureole of the Enderby stock (see Chapter 3). Altogether these rocks are juxtaposed against amphibolite-grade gneisses, referred to as the northernmost part of the Okanagan complex, along the nearly flat lying Okanagan valley fault (Fig. 1.2).

3.4 The Okanagan complex in the Enderby region (Middle Crustal layer)

Lithologies mapped as part of the Okanagan complex in this region are mainly metasedimentary rocks comprising rusty-coloured, flaggy-weathering migmatitic sillimanite-garnet quartzo-feldspathic paragneiss, calcsilicate schist, and amphibolite gneiss. Extensive quartzite layers occur southwest of Mabel Lake (Fig. 1.2). Stratigraphic correlations between high-grade rocks of the Enderby region and other regions can not be clearly established for none of the lithologies have been systematically mapped. Nevertheless the area is located only 25 km west of the Thor-Odin culmination where stratigraphic sequences have been more clearly observed (Carr 1990). Tentative correlations may be proposed between the high-grade lithologies of the Enderby-Mable Lake region and the upper member of the Mount Fosthall composite unit also referred to as the Gold Range assemblage on the southern flank of Thor-Odin (Fig. 1.2; Carr 1990). The Gold Range assemblage is correlated with Paleozoic to Mesozoic strata of the Kootenay Arc (Höy 1977; Klepacki 1985; Carr 1990). In the Thor-Odin area the Gold Range assemblage is sheared over the Fawn Lake assemblage via the unannealed Slate Mountain shear zone which is interpreted as an easterly rooted Eocene extension fault as it cuts down section in the direction of transport (Carr 1990). By comparison with the stratigraphic sequence of the Thor-Odin-Pinnacles area (Carr 1990) it may be proposed that the Slate Mountain shear zone may extend between the Vernon and Kelowna areas as paragneisses in the latter region are tentatively correlated with the Fawn Lake assemblage (see section 3.2). This suggests that the Slate Mountain shear zone may root to the south and to the east (Carr 1990).

Paragneisses of the Trinity Hills region could also be correlative with migmatic gneisses underlying the Hunters Range north of Enderby (Fig. 1.2; Johnson 1988; 1989b). A thick sheet of mylonitized leucocratic granite associated with the Ladybird suite has been found interdigitated with the Trinity Hills paragneisses near Mabel Lake (site E19 on Fig. 1.2). The granitic sheet contains Eocene zircons (see Chapter 4) and indicates that Early Tertiary shearing has affected nearly simultaneously the paragneisses of the Kelowna and Enderby regions.

Mathews (1981) outlined a sharp thermal discontinuity in the Enderby region when he found Early Cretaceous K-Ar dates (130-150 Ma) in the low grade rocks of the Nicola Group and Eocene K-Ar dates (40-60 Ma) in the underlying high-grade gneisses. The time of cooling of

high-grade rocks of the northernmost Okanagan region is clearly coeval with that recorded in the gneisses in the southern Okanagan region (see Chapter 4) and supports the interpretation of a major denudation surface carrying low-grade over high-grade rocks on a low- to moderately-dipping fault surface herein related to the Okanagan valley fault system. This surface would have formed under thermal conditions exceeding 500°C and decompressed rapidly. Unfortunately direct field evidence of the fault surface is lacking near Enderby.

3.5 The boundary zone between the Okanagan Complex and Quesnellia Terrane

Rocks at the contact of the Upper Crustal (base of the Quesnellia assemblage) and the Middle Crustal layers (uppermost Okanagan complex) are pervasively metasomatized to lower greenschist assemblages. This zone varies in thickness but never exceeds 100 m and it is systematically accompanied by intense cataclasis (see Chapter 2). K-Ar cooling dates analysed throughout the gneissic layers underlying the cataclastic zone range between 63.5 Ma and 48.4 Ma on hornblende fractions and between 56.4 Ma and 47.8 Ma on biotite in the Kelowna region. Fission-track ages from apatite and titanite extracted from the same structural levels near Kelowna range between 61.9 Ma and 35.6 Ma (Tables A, B of Appendix 1). The following chapter describes the features related to extensional faulting in the Kelowna and Enderby regions.

4. Summary

The Okanagan complex is a window through the west flank of the Omineca Belt and exhibits rocks of middle crustal level on the east side of the Okanagan valley. In contrast rocks on the west side of the valley are of upper crustal level comprising Permo-Triassic lithologies that are part of Quesnellia terrane which was accreted onto the north American margin in the Mesozoic. These rocks of two contrasting metamorphic grades are juxtaposed across a major denudation surface, the Okanagan valley fault. Hanging wall low-grade rocks are referred to as the Upper Crustal layer and are cut by Mesozoic undeformed granites altogether capped, locally and unconformably, by Eocene lavas and sedimentary rocks. Rocks of the footwall of the Okanagan valley fault define the Middle Crustal layer which comprises paragneisses, amphibolites, orthogneisses and leucogranites of various compositions. Footwall lithologies

contrast along strike and are stratigraphically difficult to correlate on a regional scale. Paragneisses and amphibolites in the Kelowna area are systematically interdigitated and are tentatively correlated with the Vaseaux Formation (Bostock 1941). These fragmentary assemblages also resemble the description of the Late Proterozoic (SPA) Fawn Lake assemblage described in the hanging wall of the Monashee décollement near Thor-Odin (Carr 1990). Rocks of the Vaseaux Formation may, however, be as old as pre-Purcell stratigraphy (Late Archean-Early Proterozoic; Armstrong et al. 1991). Paragneisses in the Enderby region north of Vernon are essentially metasedimentary rocks tentatively correlated with the Early Paleozoic Gold Range assemblage that structurally overlies the Fawn Lake assemblage in the Thor-Odin area (Carr 1990). If correlations are valid then paragneisses of the Kelowna area appear to be older than those exposed near Enderby.

Orthogneisses and leucogranitic rocks are interlayered with paragneisses in the Okanagan complex. Large volumes of these bodies occur near Kelowna and extend in the southern Okanagan complex where they are transposed parallel to the gneissic layering of the complex. These bodies contain Mesozoic-Cenozoic mineral dates (Parkinson 1985; see Chapter 4) and their modal compositions define a linear trend transecting the calc-alkaline-tholeiitic fields. End-members of these trends are undersaturated Jurassic (calc-alkaline) and saturated Eocene (tholeiitic) and seem to have evolved similarly to the Whatshan granitic suite (Brandon et al. 1991). REE profiles of the Mesozoic to Cenozoic Okanagan granitoids are typical of a "within-plate" setting interpreted as *in situ* melts formed by different degrees of melting of hosting mid-crustal lithologies generating either hornblende / hornblende-biotite / biotite / biotite-muscovite granitoids. These compositional changes could have been brought about by different country rocks compositions and adding or removing melt phases of residual liquids, or water input from country rocks under decreasing temperature conditions and increasing fO_2 (Speer 1987).

CHAPTER 2 Structural features along the northern Okanagan valley

1. *Introduction*

This chapter considers the finite products of progressive extensional strain on the western flank of the Omineca Belt in the hinterland of the Canadian Cordillera. Field mapping with joint programs of isotopic dating across the Okanagan valley have led to the discovery of a major extensional Eocene low-angle shear zone straddling the valley and extending 450 km on a northerly trend (Fig. 2.1; Parkinson 1985; Bardoux 1985; Tempelman-Kluit and Parkinson 1986; Brown and Journeay 1987; Parrish et al. 1988; Johnson 1988; but *see* Okulitch 1987). The following outlines the kinematic relationships and documents the variable intensity of strain fabrics through about 1 km of vertical section within ductilely sheared gneisses of the Okanagan complex. A particular emphasis is given to strain fabrics near the uppermost portion of the strain zone herein referred to as the mylonitic front of the Okanagan valley fault. It will be demonstrated that strain fabrics nucleated from voluminous Eocene *in situ* melts that coalesced within paragneisses of the Okanagan complex. Strain-insensitive and strain-sensitive fabrics (Hanmer and Passchier 1991) in these rocks indicate that they have undergone noncoaxial progressive deformation in a simple shear flow type. The top of the mylonitic front is overprinted by cataclastic structures thus the Okanagan valley fault is a plastic-brittle normal fault. Like most other normal faults across the southern Omineca Belt, the Okanagan valley fault system created major metamorphic offsets by the juxtaposition of greenschist-grade rocks (Quesnellia) maintained under shallow crustal rheological conditions in the hanging wall against amphibolite-grade gneisses (Okanagan complex) subjected to mid-crustal rheological conditions. Extensive Eocene lavas unconformably overlie the Upper Crustal layer and small klippen of this layer are stranded on top of the Middle Crustal layer east of Kelowna.

2. *Deformation history in the Middle Crustal layer (Okanagan complex)*

Rocks of the Okanagan complex are characterized by well developed parallel shear surfaces related to extensional events which have overprinted pre-existing structures related to compressional events preserved only in paragneisses.

Figure 2.1:

Geological and tectonic compilation map of the Okanagan valley area extending to northern Washington. OB: Okanagan Batholith; SB: Similkameen Batholith; PB: Pennask Batholith; OgC: Okanagan gneiss complex; C: Coryell plutonic suite. HR: Hunters Range; T-O: Thor-Odin culmination. Tertiary extension faults OVF: Okanagan valley fault (Bardoux 1985); GWF: Greenwood fault system (Little 1983); GF: Granby fault (Parrish et al. 1988); BF: Beaven fault, the breakaway of the Okanagan valley fault (Carr 1990; Cook et al. 1992); CF: Cherryville fault (Carr 1990). Star symbols locate new U-Pb sites (samples E19, 261) in the northern Okanagan region. Asterisks locate sites for new $^{40}\text{Ar}/^{39}\text{Ar}$ dates (samples 78, 48) and K/Ar dates (samples 32, 39). Sections A-A' and B-B' indicate sections illustrated in Figures 2.4 and 2.5.

2.1 Pre-mylonitization

Paragneisses of the Middle Crustal layer were subjected to polyphase folding before they were invaded by voluminous igneous bodies and submitted to bulk simple shear generating mylonites and isoclinal folds. Pre-mylonitic folds are preserved in semipelites interleaved with calcsilicates on the south side of Mission Creek east of Kelowna (Map 1). These rare structures are extremely incoherent in terms of geometry and attitude but they preserve evidence that the protoliths of the metasedimentary rocks have undergone at least two phases of coaxial isoclinal folding. The resulting geometry is a northerly-trending hook-shaped interference pattern presumably formed under compressional conditions as indicated by larger and more continuous folds of similar aspect in high-grade metasedimentary and metavolcanic rocks in the southern Okanagan complex (Christie 1973; Ross and Christie 1979; Ross 1981).

Large volumes of leucosomes, megacrystic granite, and pegmatites invaded paragneisses in the Kelowna area. Many of these granitoids are in gradational contact with hosting paragneisses and suggest they formed by *in situ* melting processes (Ryan 1973; Christie 1973). Knife sharp planar contacts between strongly foliated leucogneisses and paragneisses are interpreted as extensional shear planes. Non-foliated granites and pegmatites mainly lie on the eastern flank of the Okanagan complex but micaceous laminae develop progressively to protomylonitic and ultramylonitic fabrics towards the western flank of the complex.

2.2 Non-coaxial extensional flow structures in the Okanagan complex

Extensional structures in the region generally exhibit westerly-dipping shear planes, strain-sensitive fabrics and pervasive mineral extension lineations (Map 1). The latter tracks a shear direction consistently oriented N285E on individual shear planes (*see* Fig. 2.4). The total finite strain increases to high but unestimated magnitudes at the mylonitic front straddling the boundary between the Middle and Upper Crustal layers (Fig. 2.1).

2.2.1 Kinematic framework

The structural features described in the following sections suggest that the bulk boundary conditions of most extensional structures developed at mid-crustal level under progressive two-

dimensional non-coaxial flow by bulk simple shear. Although the lower boundary of the shear zone is poorly exposed, shear features were observed on a maximum width of 1 km and it appears that the rotational component of the flow was accommodated mainly along discrete shear planes. The kinematic frame of reference of the shear features recorded in the Okanagan complex can be described as a top-to-the-west sense of shear with respect to the shear plane dipping gently to the west as reflected by the systematic orientation of shear-sense indicators. In this context the maximum instantaneous stretching and the direction of maximum finite extension of the strain ellipsoid plunges approximately 15° to the east. In this context rocks of the Upper Crustal layer, referred to as hanging wall, were dropped westly over the Middle Crustal layer referred to as the footwall. A summary of all structural features is given in Figure 2.2.

2.2.2 *Shear-sense indicators*

Strain features commonly nucleate in homogeneous isotropic porphyritic granitic sheets which contain elongated anisotropic paragneissic and amphibolitic rafts. Complex shape fabrics (C/S, asymmetrical extensional shear bands), pressure shadows, and quarter structures (Hanmer and Passchier 1991) are commonly outlined in the polymineralic aggregates of the granitic bodies. Implicit is the progressive involvement of various generations of granitic sheets in the incremental strain gradient towards the top of the Okanagan complex partly demonstrated by granitic sheets being involved in various shear related asymmetrical folds. Many bodies are rotated progressively and amplified into single sheath folds in shear parallel layers indicating high finite strain conditions at the top of the complex. Sheath folds are often cut by shear planes creating transposed limbs and rootless fold hinges. Similar structures observed in the southern Okanagan complex were erroneously interpreted as compressional features (Christie 1973; Ross and Christie 1978). The axial plane of the tightest folds outlined by leucogneisses has consistently rotated anticlockwise looking north with respect to the axial plane of more open folds outlined by pegmatites indicating that the upper member of the shear couple moved to the west.

Strain-insensitive quartz shape fabrics were observed in semipelitic and quartzite layers hosting the granitic sheets. Fabrics in both rock types are unannealed and systematically indicate a sinistral sense of shear suggesting they were involved in the extensional processes.

Figure 2.2:

Summary table of the extensional fabric distribution in the upper 1000m of the northern Okanagan complex. Most extensional features occur in a narrow Zone A approximately 200m thick mantling the complex. Brittle features are concentrated in the uppermost 50-100m of Zone A referred to as Zone A₁. Depth boundaries between each zone may fluctuate locally.

Well-developed extension lineations and shear planes had been mapped by Krauskopf in 1941. These lineations are defined by stretched minerals (quartz, feldspar, hornblende, biotite) on shear planes throughout different structural levels of the complex. The persistence and the consistent trend of this lineation along the entire length of the Okana(o)gan complexes (Krauskopf 1941; Snook 1965; Ryan 1973; Christie 1973) is assumed to represent the transport direction of the Okanagan valley fault. Thin sections cut perpendicular to the lineation show irregular mineral shapes and anastomosing schistositys forming ridges and grooves suggesting that the deformation mechanism in these rocks was a non-ideal two-dimensional simple shear strain.

It is proposed that the total finite strain is significant in the extensional shear fabrics and that all material lines tend to rotate towards the flow plane. Therefore the shear plane is assumed to be subparallel to the biotite and amphibole foliations in orthogneisses and the micaceous foliations in paragneissic rafts (Hanmer and Passchier 1991).

2.2.3 *Plastic-brittle behavior*

Extensional ductile fabrics are overprinted by cataclastic fabrics which increase in density toward the structural top of the Okanagan complex (Fig. 2.2). Thus the Okanagan valley fault is referred to as a plastic-brittle structure. Rocks of the Upper Crustal layer are systematically underlain by the cataclastic zone. These hanging wall sequences are cut by discrete listric normal faults synthetic and antithetic with the sense of shear recorded in the footwall mylonites.

Extensional fabrics and preexistent metamorphic isograd surfaces are only partly affected by late upright cylindrical folds (Ryan 1973; Christie 1973; Ross 1981) which produced the large-scale undulations of the Okanagan valley fault trace (Fig. 2.1). These open structures plunge at about 15° to the west-northwest parallel to the stretching direction. No penetrative fabric is associated with these folds except for some widely spaced N-S and E-W fracture cleavages filled with chlorite, epidote, and calcite along which centimetre-scale displacements were measured.

3. *Extensional strain in the footwall*

Many intrusive bodies in the Okanagan complex are probably related to the Eocene Ladybird suite (Carr et al. 1987; Parrish et al. 1988; Carr 1990). These rocks postdate the

regional compressional deformation affecting the paragneisses and therefore record different finite strain.

3.1 Strain nucleation

Progressive extensional deformation is exhibited by granitic sheets depending on their spatial and temporal relationship to the Okanagan valley fault system. These markers are used as a qualitative gauge of Eocene strain though most granitic sheets have yet to be dated. Granitic sheets are strongly deformed on the west side of the Okanagan complex whereas granites of possibly the same age are mainly undeformed on the southeast side of Little White mountain (Fig. 2.3; Map 1). The lower limit of the Eocene ductile strain gradient can be projected at the top of the Big White mountain granite where discrete ultramylonite bands are observed at an elevation of 2.3 km (*see* Chapter 5). However, because a steep eastward-dipping brittle normal fault with an unknown displacement is expected on the west side of Big White mountain (Fig. 2.3; *see* Chapter 4) it is not valid to project the base of the Okanagan fault directly from the top of Big White mountain to the Okanagan Lakeshore (Fig. 2.5).

3.2 Strain variation

The variation of strain throughout the Okanagan complex can be described in terms of three categories shown as zones A, B, and C on Figures 2.2 and 2.3. These zones are defined by the incremental appearance of unannealed mylonitic textures. Zone A fabrics are mid- to lower amphibolite mylonites or protomylonites that form the high strain domain of the Okanagan valley fault. Unannealed Zone B amphibolite-facies protomylonitic fabrics nucleate from granites exhumed in the central part of the Okanagan complex. Zone C fabrics are confined to granoditic and pegmatitic bodies on the eastern side of the complex and are outlined by the development of a very weak sub-horizontal foliation. Figure 2.2 presents a regional compilation of the data for the kinematic elements observed in the northern Okanagan complex. Most of these fabrics have been confirmed in thin section and were systematically recognized from Penticton in the south to Enderby in the north.

Figure 2.3:

Structural compilation map along the trace of the Okanagan valley fault showing the distribution of westerly-directed kinematic indicators in the upper mylonitic zone of the detachment surface. Fabric categories observed in the footwall of the Okanagan valley fault refer to the relative distribution illustrated in Figure 2.2. These are: A: unannealed mid- to lower amphibolite-greenschist facies mylonites; B: weak to moderate foliations and/or protomylonitic fabrics; C: undeformed igneous textures and quartz ribbons (myrmekitic and zoned feldspars). Two quartz-oligoclase porphyries at Shingle Creek west of Penticton and Beaverdell, used as probable markers across the Okanagan valley fault (see Chapter 5) are outlined by black patches. Structural symbols are shear foliation (angle of dip given) and lineation in west-directed mylonite.

3.2.1 Zone A

Mylonitic fabrics of Zone A occur on the shoreline of Okanagan Lake, on the ridges of the highest mountains, underneath or near tectonic klippen, and immediately beneath Eocene volcanic outliers (Fig. 2.3). Granitic rocks and pegmatites are ultramylonitic over a maximum thickness of 500m with well developed shear planes containing a west-northwest-plunging stretching lineation (Fig. 2.4). Strain-sensitive fabrics (C/S planes), rotated feldspar and hornblende porphyroclasts, asymmetrical extensional shear bands, sheath folds and asymmetric folds throughout the zone indicate consistently that the upper member of the shear couple moved to the west (Berthé et al. 1979; Platt and Vissers 1980; Brown and Murphy 1981; Simpson and Schmidt 1983; Lister and Snoke 1984; Platt 1984; Hanmer and Passchier 1991).

Mineral assemblages in granitic sheets consist of plagioclase, quartz, K-feldspar, hornblende, and biotite with accessory sphene, apatite, zircon, thorite, opaques and epidote probably of retrograde metamorphic origin. Feldspars are commonly myrmekitic. Mafic silicates are often retrograded to chlorite and plagioclase is generally sericitized. Paragneisses in this zone exhibit thick bands of segregated felsic (feldspar, quartz) and mafic (pargasitic hornblende, biotite) minerals interleaved with centimetre-scale quartzo-feldspathic leucosomes (Robin 1979). Hornblende amphibolites, diopsidic marbles and calcsilicates are lens shaped megaboudins interlayered with fibrolite-garnet bearing semipelites. These assemblages indicate that the metasedimentary rocks were at amphibolite grade prior to or during extension (*see* Chapter 3).

Strained leucogneisses in Zone A contain nearly 20% porphyroclasts that are mainly ovoid to tabular K-feldspar (Plates 2.1, 2.2). Shear sense was ascertained in at least fifty localities from field and thin section observations of porphyroclasts of which the best examples are reported in Figure 2.3 (*see also* Map 1) and illustrated in Plate 2.2. The best preserved rotated porphyroclasts are those surrounded by a fine-grained groundmass. These are rare due to the coarseness of the plutonic protoliths and their original equidimensional grain geometries. As a result mylonitic leucogneisses exhibit a polymineralic coarse-grained, heterogeneous, competent groundmass. Direct interaction of this groundmass with the porphyroclasts (Plate 2.2) has caused a certain amount of reorientation leading to a variety of shear-sense interpretations.

Plate 2.1: *Oblong myrmekitic feldspar observed in the uppermost portions of the detachment surface. The quartz-oligoclase intergrowths are strain related as they occur along grain boundaries that face the maximum finite shortening direction at an angle of 10° with the well developed S foliation. Sample RLB3 collected along highway 97, immediately north of Swan Lake, 5 km north of Vernon.*

Plate 2.2:

S-C tectonites in the Okanagan valley fault showing unambiguous westward shear sense in all cases. Mylonites in b, and f formed in orthogneissic protoliths showing variable porphyroclast asymmetries due to polycrystalline interactions but developed good C, S fabrics. Mylonites formed in quartz-biotite-garnet assemblages a, c, d, and e mainly show the development of asymmetric mica "fish", winged porphyroclasts, and well developed C, S, and C' fabrics (Berthé et al. 1979). Sample locations (see Fig. 2.1) are: a (OF1) from the Okanagan Falls area; b, c (MX173, MX61) from the Kelowna area; d, f (MXE42, MXE19) from the Enderby area; and e (RLB3) from the Vernon area. West is to the right in a, b, and c and to the left in d, e, and f. Mineral abbreviations are: K: K-feldspar; Mu: muscovite; Pl: plagioclase; Q: quartz; Si: sillimanite.

Most of the hornblende and approximately 50% of the feldspar porphyroclasts are pulled-apart in this zone. Gashes and microfractures are filled with unannealed quartz and are generally oriented perpendicular to the long dimension of the porphyroclasts, that is perpendicular to the maximum instantaneous stretching axis. Unfractured porphyroclasts are often asymmetric with their long dimension tilted east in the direction of maximum stretching. In the uppermost 100m of Zone A porphyroclasts of K-feldspar ($\text{Ab}_{10}\text{An}_{90}\text{Or}_{100}$) are reduced in size and exhibit elongated ovoid shapes and become more myrmekitic when in contact with plagioclase (Plate 2.1). Myrmekitic intergrowths are observed mainly along those faces parallel to the long dimension of the grain or at a large angle to the shear surfaces outlined by micaceous trails. Although myrmekites occur in less deformed igneous bodies lying in the deeper zones B and C, the quartz-oligoclase intergrowths in leucogneisses of Zone A are undoubtedly strain related (Hanmer 1982; Simpson 1985). They occur along grain boundaries perpendicular to the maximum finite shortening direction which is nearly coplanar with the well developed flattening foliation in these rocks (Plate 2.2). In most cases, the edges of the myrmekite intergrowths face into the host porphyroclast and are angular, suggesting that some recovery took place (Simpson 1985). Myrmekitic intergrowths also occur within shear planes. The intergrowths have planar boundaries and relate to a mechanism of both diffusional creep exsolution and replacement of K-feldspar by plagioclase (Phillips 1980).

Asymmetric hornblende and feldspar porphyroclasts in Zone A developed simple triangular pressure shadows of polycrystalline fine grained recrystallized feldspar and quartz. These appendages formed in the extensional quadrants of the bulk flow (Simpson 1985; Hanmer and Passchier 1991). They are not sigmoidal and their long dimensions lie clockwise looking north with respect to the mylonitic foliation indicating that the upper member of the shear couple moved westward. The indistinct outlines of the pressure shadows and their variable composition from the rigid object suggests that they might be metasomatic in origin and formed by mass transfer and diffusion processes (Robin 1979; Hanmer and Passchier 1991).

Several feldspar porphyroclasts are also fragmented into pieces that have undergone a certain amount of domino-type rotation and translation. Rotations are generally clockwise to the east looking north which is an antithetic movement to the bulk shear plane probably caused by

local non-ideal shear behaviour (Hanmer and Passchier 1991). At depths of approximately 300m below the upper surface of the Okanagan complex, porphyroclasts start to be dynamically recrystallized and cataclastic elements are rare.

The interpretation of the kinematic significance of asymmetric rigid inclusions relates to their rotation rate which depends upon the shape and the instantaneous orientation of the inclusions with respect to the strain field under a given flow type (Ghosh and Ramberg, 1976; *see* Hanmer and Passchier 1991). In progressive simple shear all inclusions should rotate with the same sense as the bulk flow. Most porphyroclasts considered above are unwinged (Hanmer and Passchier 1991) and have aspect ratios in the order of 2:1. It is most likely that these inclusions were oriented randomly in the protolith therefore making empirical predictions of the kinematic significance from their present asymmetries is a hazardous undertaking.

C/S mylonites (Berthé et al. 1979; Lister and Snoke 1984) are frequently observed throughout Zone A (Plate 2.2). It is expected that S (flattening) - and C (shear) -surfaces formed concurrently during progressive deformation. In thin section, C-surfaces are thin layers of recrystallized polymineralic aggregates or trails of mica in quartzites. S-surfaces are defined by the mineral-shaped preferred orientation of the old grains (generally micas) formed by dynamic recrystallization between C-surfaces. Micas are also frequently aligned in these orientations. The average angle between C- and S- surfaces in Zone A ranges between 20° and 35° indicating a moderate amount of strain (Berthé et al. 1979). Sinistral discontinuities across C- surfaces are locally recorded by grain-size displacements of feldspar and micas.

Asymmetrical extensional shear bands (Hanmer and Passchier 1991; Platt and Vissers 1980) developed mainly in anisotropic amphibolite-rich layers at angles between 15° and 25° to the principal shear surface (Dennis and Secor 1987). These centimetric shear bands are defined by the alignment of small biotite plates (Plate 2.2a, d, e). Only one set of these structures was observed and confirms that the deformation mechanism in these rocks was non-coaxial (Platt and Vissers 1980) and unidirectional. The sense of slip is synthetic to the slip direction of the whole shear zone as indicated by offset markers and drag effects across the shear bands. Locally, asymmetrical extension shear bands have created the asymmetric boudinage of layers showing no compositional contrasts. The motion of the boudins is systematically down-to-the-west and

does not exceed several centimetres. Shear band surfaces are curved at both ends of the boudins merging into the principle shear surface transferring motion parallel to the shear surface. Asymmetric extensional shear bands are abundant in the upper 200 m of Zone A and are consistent with an increase of the finite strain gradient towards the ultramylonitic front at the top of Zone A. Extensional shear bands develop where grain-sizes are most reduced due to strain softening conditions. Mesoscopic melanocratic gneissic bands developed pinch-and-swell structures (Hanmer and Passchier 1991) that rotated antithetically with respect to the vorticity of the bulk flow recorded by other strain features. These back-rotated elements are compatible with a top-to-the-west sense of shear.

Mylonitized metaquartzites found near Okanagan Falls and east of Vernon (Fig. 2.1) are projected into Zone A. In thin section these rocks exhibit flattened ovoid K-feldspar and garnet porphyroclasts (Ross 1981) in a matrix of quartz ribbons showing strong preferred fabrics (Wilson 1975; Boullier and Bouchez 1978) and muscovite fish (Plate 2.2a). Ribbons are generally thinner than 2mm and form extensive series of equidistant laminae. In thin section they typically contain small subgrains forming mosaics with strong preferred orientations interpreted as strain-insensitive fabrics (Hanmer and Passchier 1991). Elongated grains can have length to breadth ratio of up to 40:1. Quartz and muscovite crystals are oriented with their long dimensions systematically dipping east at approximately 25° with respect to the shallow westerly-dipping mylonitic foliation decorated with trails of fine muscovite. No recrystallization texture was recognized at the mica boundaries (Lister and Snoke 1984). Quartzites exhibit a second type of grain boundaries oriented $70-90^{\circ}$ to the shear surfaces which may reflect grain-boundary migration recrystallization along misorientations such as boundary slides and microshears in the host grain (Lister and Dornsiepen 1980). Analysing the deformation mechanisms and slip directions in the ribbons was not undertaken as other types of strain fabrics found in these rocks give an unequivocal sense of shear. Nevertheless, the two-dimensional geometry in the ribbons suggests a large amount of strain (Berthé et al. 1979). Orientations of the ribbons is sub-parallel to S-surfaces consistent with a top-to-the-west sense of shear (Plate 2.2).

Berthé et al. (1979) and Cobbold and Quinquis (1980) suggested that, at advanced stages of deformation in a shear zone, asymmetric folds should develop; thus the unique plane of

anisotropy in layered rocks would be deformed. Such folds have been observed within 200 m of the upper limit of Zone A near Kloo Creek east of Kelowna (Map 1). The asymmetry of those centimetric-scale, reclined to isoclinal, intrafolial folds is compatible with a top-to-the west sense of shear. Many non-cylindrical, sheathlike folds concurrent with C- and S- surfaces were also mapped throughout Zone A. Fold axes of these structures are curved within the axial planes though they are mostly colinear and parallel to the pervasive mineral lineation. Several of these folds developed by passive amplification of initial deflections or by homogeneous folding of mafic layers that were initially oblique to the shearing plane (Cobbold and Quinquis 1980). Progressive stages of folding amplification have been observed throughout the thickness of Zone A. The amplitude of a few centimetres and the hinge span of several metres of the elliptical sheath folds suggests that shear strains in these rocks reached a minimum value of $\gamma \sim 10$ (Cobbold and Quinquis 1980). Sheath folds of larger amplitudes may repeat the gneissic layering throughout Zone A but the lack of large exposures precluded their recognition.

The timing of fabric development in Zone A are constrained by a syn-kinematic leucocratic granitic sheet of the Ladybird suite in the Mabel Lake area (MXE19 on Fig. 2.1; see Chapter 4). This mylonitized body of biotite-muscovite granite contains zircons with a crystallization age of 56.4 Ma indicating that ductile shearing at the top of the Okanagan valley fault may have been active in the Early Eocene. Syn-kinematic muscovite in the Mabel Lake granite (Plate 2.2f) provided an $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age of 49.2 ± 0.4 Ma suggesting that rocks of Zone A cooled below 300°C at this time (see Chapter 4).

3.2.2 Sub-Zone A_c

High strain ductile fabrics of Zone A are overprinted in the upper 100m by syn-kinematic chlorite-epidote-sericite bearing greenschist-facies cataclastic fabrics defining sub-Zone A_c. Cataclasites and related brittle fabrics are typically observed in two physiographical contexts; (1) near the projected trace of the main structural break which is generally at high elevations or along Okanagan Lake; (2) around isolated Eocene volcanic klippen east of Kelowna and Enderby.

Shallow dipping fractures wrapping around brecciated mylonite fragments are conspicuous along the trace of the Okanagan valley fault and suggest a protracted superposition of brittle over

ductile fabrics (Plate 2.3). The brittle zone is systematically accompanied by pervasive saussuritization in the groundmass and in veins. Some veins are also filled with framboidal pyrite and gypsum typical of hydrothermal deposits. Fracture planes exhibit sigmoidal geometries suggesting motion was still underway on the main slip surface after their formation. These are increasingly abundant within 10 m of the principal fault surface.

Decametric lenses of cataclastic ultramylonite occur at the top of Zone A₁. These are surrounded by centimetre-thick clay zones bounded by polished surfaces containing multidirectional fault striae. Although motions across the gouge zones are difficult to assess it is proposed that these structures are the major slip surfaces. Two main types of clay zones were observed and consist of centimeter-thick flat lying gouge surfaces cutting steeper, thinner, and less extensive argillaceous gouge zones. Flat lying black gouge zones extend down section to nearly 100 m below the main structural break of the Okanagan valley fault at the foot of Black Knight Mountain east of Kelowna (*see* Frontispiece). Cross-cutting relationships between several fault sets at this locality suggest that the brittle elements splayed vertically and horizontally (and diachronously?) into several anastomosing branches. Although it was impossible to clearly establish the relative timing between cataclastic branches and the principal slip surface it is suggested from their arcuate shape that the brittle elements were created at steeper angles and subjected to tilting and rotation on later throughgoing structures. Steep structures may have been splays of an earlier shearing episode later superimposed and transected by throughgoing shallow dipping fault surfaces during post peak shear resistance. The brittle event thus may have started as individual dislocated tracks creeping and merging gradually to form large denudation surfaces.

The intensity of granulation increases to the top of Zone A₁ where grains are reduced to micron sizes. Such granulation is a function of increasing confining pressure and increasing displacement along the shear fractures. In a few cases, grains and brecciated fragments have been clearly rotated in a manner that agrees with the ductile sense of shear.

The presence of clays in the Okanagan fault zone must have reduced the coefficient of internal friction (μ) to about 0.1 (Wang and Mao 1979). A strict functional relation does not exist between the mean thickness of a gouge zone and the shear displacement parallel to this zone. Although experimental studies may not apply to natural examples it may be proposed that the

Plate 2.3: Initial stages of brecciation of a mylonite in sub-Zone A, from the bottom of the uppermost 100 meters of the Okanagan complex showing the granulation caused by the superposition of brittle fractures onto older ductile fabrics during protracted deformational sequences. Sample MX63 underlying the Joe Rich Klippe located 30 km east of Kelowna (see Fig. 2.2).

Plate 2.4: Development of normal faults in the uppermost portions of the Okanagan complex illustrating the effects of brittle failure following ductile shear along the Okanagan Valley fault. Fracture and fault densities increase to nearly 200m⁻¹ in this zone. Sample MX133 along Mission Creek.

minimal displacement across the brittle layer is at least an order of magnitude greater than the thickness of the gouge zone (100m; Engelder 1974). Thus the minimal motion associated to the sub-horizontal brittle surfaces is at least 1 km.

Small scale longitudinal and extensional fractures generally filled with cryptocrystalline chlorite, epidote, quartz or pyrite extend down section from Zone A₁ to Zone A₂. Offsets on these fractures are on the order of a few centimetres (Plate 2.4). Most fractures are generally steep, have centimetric spacing and strike N-S perpendicular to the stretching direction. Dip motions on the fractures are predominantly down to the west, but reverse senses were also observed locally suggesting that contraction also took place locally and could accommodate "room" problems. The total westward drop accumulated on these microfaults relative to the footwall is estimated at a minimum of 500m. A second set of steep fractures trends generally E-W. They are rarer and more widely spaced than the N-S trending fractures. Their displacements are exclusively dip slip and on the order of a few centimetres either to the north or south. The total drop on these structures has not been estimated.

Brittle structures are found to the bottom of Zone A but beyond a depth of approximately 300m they consist mainly of sporadic extension fractures filled with quartz $\pm$ hornblende suggesting greater thermal trapping conditions at depth.

Argillaceous gouge of three cataclastic fault splays provided whole-rock $^{40}\text{Ar}/^{39}\text{Ar}$ ages ranging between 59.7 ± 1 Ma and 48.5 ± 1.4 Ma (*see* Chapter 4). These dates are concordant plateaus but they give a rather widespread range which is consistent with the idea that fault splays are diachronous. However, some contamination may be suspected in the oldest sample (*see* Chapter 4). Thus it appears that cataclastic fabrics in Zone A₁ quenched Ar diffusion only 2 Ma after muscovites cooled below 300°C in Zone A₂.

3.2.3 Zone B

Weakly deformed granitoid rocks are confined to structural levels between 500 and 1000m below the uppermost surface of the Okanagan valley fault. Rocks of this level have been eroded by deep canyons surrounding Mission Creek in the central part of the Okanagan complex (Figs. 2.1, 2.2). These structures represent distributed strain in the footwall of the Okanagan fault.

This zone consists mainly of granitic rocks overprinted by mesoscopic shear planes comprising stretching lineations and unannealed protomylonitic fabrics visible mainly in thin section. The mineral assemblage consists of plagioclase, quartz, K-feldspar, biotite, hornblende and accessory zircon and apatite. Thin leucogneissic bands are often pulled apart creating thin boudins approximately 25m long. Due to this very large aspect ratio the boudins have not generated back-rotated structures (Hannier 1986) and pull-aparts are undeformed and generally filled with annealed quartz $\pm$ hornblende.

The mylonitic foliation disappears progressively downwards but the east-west lineation stays well developed. Recrystallized feldspars in this zone preserved very few asymmetries. Quartz ribbons consist of several rows of primary grains containing deformation bands and serrated boundaries along which subgrains are aligned.

A weakly foliated biotite granite sheet at the top of Zone B southeast of Kelowna (MX261 on Fig. 2.1) contains thorite crystals which provided concordant ages of 51.3 Ma that is a minimum age of crystallization (*see* Chapter 4). This could suggest that ductile deformation in the footwall of the Okanagan valley fault lasted until the late Early Eocene and propagated down section from Zone A as mylonitic bodies in the upper zone are apparently 5 Ma older. Alternatively this age could reflect diachronous plutonic events which would not place constraints on the onset of deformation in Zone B.

3.2.4 Zone C

Primary igneous textures are preserved in the large intrusions lying on the east side of the Okanagan complex in the Kelowna region (Figs. 2.2, 2.3). Mineral assemblages are the same as in Zone B but feldspar are nearly euhedral and plagioclase crystals are often zoned. Rocks of this zone underlie the highest peaks of the Okanagan Range. Locally a faint biotite lineation develops together with discrete 10-50cm wide westerly-dipping and top-to-the-west directed shear zones synthetic with the Okanagan valley fault system. Quartz ribbons in Zone C consist of individual grains with coarse rectangular geometries and straight boundaries perpendicular to the ribbon walls. Rare subgrains in these ribbons may have formed under post-tectonic static thermal conditions causing recrystallization.

Granitoids in Zone C contain small rafts of amphibolite-grade paragneisses which exhibit two foliations that may be correlative with compressional events (Ryan 1973; Christie 1973). Thus it can be proposed that because the igneous bodies are not penetratively deformed they post-date the compressional event(s) and lie below the zone of major extensional strain.

No U-Pb ages were obtained from the granitoids in Zone C but some may be petrographically correlative with the Eocene Ladybird suite and therefore link with the various granitic sheets and leucogneisses found in the higher strain zones A and B.

The affiliation between magmatic bodies in each structural zone is indirectly demonstrated in the field by the absence of sharp intrusive contacts between plutonic phases. It appears that most granitic bodies grade into orthogneisses which, in turn, cut polydeformed paragneissic layers. A granite-orthogneiss affiliation is supported by petrographic resemblances between bodies of either strain zones. For example, two of the highest peaks near Kelowna, Little White mountain and Big White mountain (Fig. 2.1; Map 1), are underlain by intrusions containing 10 to 20% K-feldspar phenocrysts of various sizes. The Little White biotite granite lies at the limit between Zones B and C. The unfoliated porphyritic granite grades progressively to the west to a foliated granite and further west to a layered porphyroclastic orthogneiss in Zone A. Similarly the biotite granite underlying Big White mountain is cut at its summit (2300m) by several discrete mylonitic and lineated bands containing westerly-directed kinematic indicators (*see* Chapter 5). These shear zones are the most eastward found near Kelowna and display all the petrographic aspects of progressive strain whereby textural changes are easily identified.

The boundaries between porphyritic plutons, porphyroclastic gneiss and complexly swirled and banded leucogneiss are somewhat arbitrary. However, the transition from granite to gneiss is frequently marked by the presence of pegmatites and partial migmatization invading paragneisses. The pegmatites, which rarely exceed a few metres in length and 30 cm in width, have a quartz monzonite composition interpreted as an igneous differentiate of the large plutonic masses. The acquisition of foliations at the granite-gneiss transition zone coincides with the local occurrence of muscovite-bearing pegmatites mainly observed at the limit between Zone C and Zone B east of Kelowna. The pegmatites have diffused contacts in the granitic bodies, their grain sizes are reduced by a factor of one half towards the layered gneiss and this reduction is

accompanied by the appearance of a faint foliation defined by muscovite trails. The persistence of muscovite at the transition between Zone C and Zone B implies that fluids may have led to strain softening processes facilitating strain. Melt-enhanced deformation is expected to provide an active mechanism for heat transfer by fluid circulation and may have accelerated strain by melt lubrication (Hollister and Crawford 1986). Deep seismic surveys in the extended regions of the southwestern United States exhibit highly reflective layers interpreted as melt layers bounded by ductile shear zones at present mid-crustal depths (McCarthy and Thompson 1988).

3.3 Regional distribution of fabric elements in the Kelowna region

The mylonitic shear foliation and its accompanying stretching lineation in the vicinity of Kelowna are shown on the equal area plots of Figure 2.4. Lineation plunges WNW towards a mean direction of 285° with a plunge of 15° on westerly-dipping foliation and ESE towards 105° with the same plunge on easterly-dipping shear surfaces. This lineation is interpreted as the principal finite direction of shear. It is noticeable that the average contoured foliation, estimated from nearly 300 measurements, dips approximately 15° to the northeast (Fig. 2.4).

Contouring the shear foliation and stretching lineation data of the Kelowna area by domains reveals some important variations. In order to account for this phenomenon the region has been divided into domains which show the distribution of the structural partitions (Fig. 2.4b). For this purpose, the fabrics surrounding Mission Creek, east of Kelowna, were regrouped in five structural and physiographic domains. Each domain comprises a minimum of 30 measurements which define contoured maxima that differ from one domain to the other (Fig. 2.4b). Foliations of domain I, in the southwest, predominantly dip 20° north. Foliations of domain II, south of Mission Creek, dip shallowly at 10° to the northeast. Domain III, farther east, comprises foliations which systematically dip 10° to the west-northwest. The foliation in domain IV, north of Mission Creek, dips 20° east. Excluding domain V, the foliation attitudes surrounding Mission Creek in the footwall of the Okanagan valley fault define a broad open antiform and a broad open synform with north-south trending axes. Kinematic indicators on either flanks of these structures indicate a top-to-the-west shear sense parallel to the lineation trend. This implies that the shear zone that generated the mylonitic fabrics could have been syn-kinematic and that arching

Figure 2.4: *Stereographic projections, lower hemisphere equal-area of structural features in the Okanagan complex near Kelowna. a. the regional distribution of foliation poles (contour interval = 3% starting at 3%) and mineral stretching lineations (contour interval = 3% starting at 3%); b. poles to foliation subdivided in five domains which outline a domal geometry between domains I, II, III, and IV in the Okanagan complex. Data of domain V are erratic and show the effects of bulk rotation caused by late brittle reactivation in the highest structural level Zone A, of the Okanagan valley fault (contour interval = 5% starting at 5%).*

POLES TO SHEAR FOLIATIONS

STRETCHING LINEATIONS

STRUCTURAL DOMAINS

was a later feature accounting for the slight dispersion of the lineations outlined in Figure 2.4a.

The orientations of the shear foliations in domain V, located beneath the Eocene outlier of Black Knight Mountain (*see* Frontispiece), are erratic. This is related to the fact that this area represents the highest level of the strain zone (Zone A₁) where cataclastic fabrics and shallow splay faults have dismembered the earlier plastic fabrics. Despite the cataclastic features, the distribution of poles to foliation planes in domain V are not random but rather concentrate in the SW quadrant exhibiting steeper dips than the "regional" defined in the other domains. This concentration of poles suggests that tilting on individual fault surfaces in the cataclastic zone was nearly systematic in a northeastern direction. If such direction corresponds to a motion vector this one trends nearly 60° south of the ductile stretching vector and would suggest that brittle stretching occurred in a direction diagonal to the ductile stretching. This lateral motion may have occurred either concurrently or after the ductile extensional processes. Lateral stretching during extension has been documented in regions where pre-rifting geometries are easily reconstructed by global plate dynamics such as in the Gulf of Aden (Courtilot and Vink 1983).

4. *Deformation history of the Upper Crustal layer (Quesnellia Terrane)*

Rocks related to the Upper Crustal layer comprise greenschist-grade Late Paleozoic metasedimentary and metavolcanic rocks intruded by Mesozoic batholiths. Altogether these rocks are capped by Eocene lavas and alluvial sedimentary rocks. The finite displacement on the Okanagan valley fault brought the present part of the Upper Crustal layer on top of the Middle Crustal layer. Protracted denudation of the Okanagan fault has caused megascopic exhumation and arching of both layers. Erosion since the Late Middle Eocene has dissected the crustal layers leaving isolated klippen and cutting down section to a few kilometers into the Upper Crustal layer. Detached klippen lay on the east side of the Okanagan valley and may be indirectly linked to more continuous sections occurring on the west side of the valley.

4.1 Detached klippen

Two wedge-shaped klippen underlain by a shallow-dipping fault surface occur north of Mission Creek, 30 km east of Kelowna and are referred to as the Mission Creek klippen (Figs.

2.1, 2.2; Map 1). The underlying faults are interpreted as remnants of the normal fault system because they are characterized by thick chloritized and cataclastic mylonites typical of Zone A, including abundant angular fragments derived from overlying Eocene strata. The fault surfaces dip about 20° to the south into the mylonitic gneisses of Zone A making a shallow angle with the bedding or overlying Eocene argillites. The nearly parallel attitude between bedding in the sedimentary strata and the slip surface suggests either that the strata were deposited near the end of the extensional episode or that the motion on the fault plane was mainly non-rotational. Referring to relative motions expected in a ramp/flat geometry, hanging wall rotations would occur on the ramp segments of listric faults whereas flat segments may be non-rotational (Wernicke and Burchfiel 1982). Non-rotational extension of the hanging wall of the Okanagan valley fault near Kelowna is consistent with the paleomagnetic data obtained in these rocks (Bardoux and Irving 1989; *see* Chapter 7). Elsewhere rotational motion is clearly documented in the White Lake basin (Church 1973) on a southern segment of the Okanagan valley fault. Based on seismic profiles (Cook et al. 1992) the Okanagan valley fault seems to be made of several ramp-flat segments (Wernicke and Burchfiel 1982). It is therefore consistent that ramp segments have recorded rotations whereas flat surfaces would only record translations of strata along the direction of transport.

Another detached block referred to as the Joe Rich klippe (Fig. 2.3) straddles the south side of Mission Creek right across from the Mission Creek klippen. The stratigraphic elements preserved in this large klippe comprise a basal section of brecciated metavolcanic rocks of the Nicola Group and limestone breccia. The latter may be correlative to limestone breccias occurring at the base of the Brooklyn Formation which contains Ladinian age faunas. The breccias are thought to have formed by submarine destruction of a carbonate platform in the Paleozoic (Fyles pers. comm., 1990). The nearest *in situ* limestone breccias occur in the Greenwood area (Fig. 2.1; Little 1983) that is 110 km to the southeast of the Joe Rich klippe. Assuming these are the nearest source it seems improbable that the Joe Rich klippe was displaced 110 km in a NNW direction (*see* Chapter 5). Hence it is proposed that the source of the Paleozoic strata lies amongst undivided Triassic strata mapped in the direction of transport of the Okanagan valley fault, that is 60 km east of the Joe Rich klippe at the headwaters of Rendell

Creek (Little 1957). The total displacement of the Joe Rich klippe would then be tentatively estimated at 60 km. Alternatively it may be proposed that the source of the limestone breccias has been eroded off the eastern flank of the Okanagan complex in the Tertiary. Brecciated limestones of the Joe Rich klippe are unconformably overlain and cut by Eocene trachytes capped by alluvial sedimentary rocks and dacitic lavas (see Chapter 6) which can be partly correlated with lithologies exposed west of Kelowna (see section 4.2). Feeders to the trachytes are nowhere found near the Joe Rich klippe whereas feeders to the younger dacites cut the mylonitic gneisses of the Okanagan complex nearby the klippe. Biotites in the dacite were dated at 46-48 Ma by K-Ar methods (see Chapter 4) and indicate that the Joe Rich klippe was probably locked to mid-crustal rocks by the time dacites were extruded.

The Joe Rich klippe lies in a valley and is slightly tilted southwards (Fig. 2.3). It is surrounded by saussuritized and unannealed mylonitic gneisses related to Zones A and A₁, preserving west-over-east sense of shear. This indicates that the underlying slip surface is part of the Okanagan fault system which therefore extends 30 km east of the Okanagan valley.

The estimated offsets of the Mission Creek klippen seem to be smaller than that of the Joe Rich klippe and may reach only a few hundred meters. These 'klippen are composed strictly of alluvial sediments capped by rhyodacitic lava flows similar to those found in the upper parts of the Joe Rich klippe. It is proposed that the faults underlying the Mission Creek klippen lie at higher structural levels than the fault surface underlying the Joe Rich klippe. Normal faults of the Mission Creek klippe may thus form an imbricate system (Wernicke and Burchfiel 1982) cutting down section southward towards the Joe Rich klippe. Similar imbricate fault systems have been observed at smaller scales at the uppermost level of the footwall of Okanagan valley fault. The presence of anastomosing imbricate fault systems indicates that the Okanagan valley fault has propagated along several sub-parallel slip surfaces creating a *chaos* structure (Wernicke and Burchfiel 1982).

The shallow-dipping fault segments underlying the Mission Creek klippen trend E-W nearly parallel to the main stretching direction. They may have been formed as minor transfer faults (Lister et al. 1986; Gibbs 1987) accommodating differential movement within the hanging wall of the Okanagan valley fault. It is suggested that transfer faults also underlie the Coldstream

valley and the Trinity Hills valley east and southeast of Vernon (Fig. 2.1).

4.2 Brittle structures in the Kelowna outlier

The lithologies occurring west of Kelowna lie in the hanging wall of the Okanagan valley fault and are referred to as the Kelowna outlier. The bottom of the outlier exhibits Late Paleozoic rocks correlated with the Apex Mountain Group and the Nicola Group (Tempelman-Kluit 1988). These rocks show one well developed schistosity that is axial planar to tight upright to northeastwardly-reclined folds. Mesozoic granodiorites cut the Paleozoic sequence, contain locally a weak schistosity, and have been metasomatized. A decametric shear zone was found in a granitoid window underlying Eocene lavas near Lambly Creek (Map 1). Fabrics in this subvertical shear zone are annealed suggesting they are unrelated to footwall extensional fabrics (Zone A) and must be related to an older (?Mesozoic) tectonic event. Eocene rocks unconformably overlying older lithologies, show absolutely no ductile fabrics.

The rocks of the Kelowna outlier are cut by discrete brittle structures trending generally N-S. Steep normal and reverse faults are concentrated mainly at the boundary of the outlier near cataclastic Zone A_c in the upper part of footwall mylonites. Hanging wall fractures and faults, have dip-slip offsets smaller than a few hundred meters. South of Westbank, pre- to early-Eocene left-lateral faults have brecciated Mesozoic granitoids and appear to control the deposition of pyroclastic flow deposits at the base of the Eocene volcanic edifice (*see* Chapter 6; Map 1). Reactivation of these faults prior to the deposition of the main volcanic flows is suggested from the cataclastic aspect of the pyroclastic fragments.

Steep normal faults cut through the Kelowna outlier similarly to those faults defining graben systems in the Eocene White Lake basin located 40 km to the south near Penticton (cross-sections A-A' and B-B' on Figs. 2.1, 2.5, 2.6). The latter is cut by a dozen normal faults grouped in two sets (Church 1973) that are presumably listric and converge at depth rooting into the surface of the Okanagan valley fault (Tempelman-Kluit and Parkinson 1986). The easterly faults in the White Lake basin are westward-dipping creating domino-style rotations (Fig. 2.5; Wernicke and Burchfiel 1982) causing easterly tilts of Eocene volcanic blocks. Westerly blocks are tilted westward on easterly-dipping normal faults which merge into westerly rooted faults.

The fault geometry underlying the White Lake basin probably reflects the presence of a local ramp in the denudation surface (McClay and Ellis 1987; Fig. 2.5, *modified from* Church (1973) and Tempelman-Kluit and Parkinson (1986)). The tilted lava flows are in turn overlain by syn-tectonic alluvial sedimentary rocks which show increasing dip with age as a result of progressive collapse of the east half of the graben. The sedimentary rocks are tilted antithetically to the transport direction recorded in the underlying mylonitic gneisses (Parkinson 1985; *see* preceding section 3). Large brecciated slide blocks derived from nearby hanging wall and footwall lithologies have been gravitationally transported in this unstable tectonic environment and deposited onto the tilted Eocene sequence (Church 1973; Tempelman-Kluit and Parkinson 1986).

Several normal faults extending for 10 km along N-NE trends cut the Kelowna outlier (cross-section A-A' on Fig. 2.1 and Fig. 2.6). Two of these faults dip easterly, three dip westerly, whereas several others have unknown attitudes. It appears that the number of hanging wall faults in the Kelowna outlier is about half the fault density cutting the White Lake basin which may indicate that the footwall is flatter (McClay and Ellis 1987) in the Kelowna region. Sedimentary rocks lying at the top of the Kelowna outlier have probably been tectonically tilted to the east and northeast between 10° and 75° (Fig. 2.4). Rotations around vertical axes in the underlying Eocene lavas are undetectable by paleomagnetic techniques. Small scale rotations around horizontal axes in individual fault blocks may have occurred though they seem to have been partly obliterated by thermal overprints during extensional processes (*see* Chapter 7).

Two Eocene outliers occurring near Enderby in the northern Okanagan valley (Fig. 2.3) are cut by only two normal faults (Mathews 1981) that are interpreted to flatten along the Okanagan valley fault in the subsurface. The first fault marks the eastern limit of the Enderby volcanic block (ca. 48 Ma old; Ewing 1981) which is tilted 15° to the east against retrograded and brecciated high-grade mylonitic gneisses whereas the second fault cuts through the western contact of the Trinity Hills lavas east of Enderby causing no apparent rotation and only minor offsets (Mathews 1981).

The tectonic elements found in the Eocene outliers and in the detached klippen along the northern Okanagan valley suggest that rocks of the Upper Crustal layer have not slipped passively onto the gouge zone mantling the Middle Crustal layer. Generally, the bedding attitudes in each

half-graben lying in the hanging wall of the Okanagan valley fault are opposite to the sense of rotation expected on most normal faults mapped (Wernicke and Burchfiel 1982). This may be partly due to internal deformation mechanisms associated with a change in slip vectors between the brittle upper crust and ductile transition zone during the protracted evolution of the fault zone (Jackson and McKenzie 1983). Many of these structures may have formed as antithetic faults (Jackson and McKenzie 1983). Perhaps the original attitudes of the lava flows and alluvial fans in these Eocene basins were not horizontal before steep normal faults were created (*see* Chapter 7). Because most of these faults are parallel they are expected to have formed during the same episode (Jackson and McKenzie 1983). Their slightly concave surface traces to the east suggest that they may be listric at depth extending to the base of the Upper Crustal layer. However, some of these structures may cut beyond the main detachment surface of the Okanagan valley fault into the Middle Crustal layer. Indeed there are a few examples of such faults cutting into Zone A of the Okanagan complex near Kloo Creek southeast of Kelowna (Plate 2.4; Map 1).

Hydrothermal activity along the hanging wall faults is suggested by the local occurrence of extension veins filled with quartz-chlorite-calcite-epidote. Davis and Lewis (1984) found drastically decreasing heat flow values across the highly fractured rocks of Zone A, near Penticton (Fig. 2.1). This indicates that cataclasites are still flooded with cold meteoritic ground water. This borehole information also confirms that Zone A, of the brittle edge of the Okanagan valley fault projects in the subsurface.

Syn-extensional sedimentation is recorded in nearly 1.2 km thick sequences of alluvial sediments capping the hanging wall lavas of the Kelowna outlier (Fig. 2.1). Basal conglomerates contain large rounded boulders derived mainly from hanging wall sources. There is a gradual appearance of very few mylonitic boulders most certainly derived from eastern sources in the upper 100m of the sedimentary sequence (*see* Chapter 6). This evidence suggests that mylonitic footwall gneisses were exposed at the surface near the end of the depositional cycle. Pollen of late Middle Eocene age was found near outcrops containing the mylonitic clasts in the vicinity of Kalamo Park west of Kelowna, and gives an approximate time at which the Okanagan valley fault breached through the surface (*see* Chapter 6; Map 1).

Figure 2.5: *Section B-B' located on compilation map of Figure 2.1. The White Lake basin, an Eocene outlier, is cut by listric normal faults defining a graben structure (redrawn from the compilation of Tempelman-Kluit and Parkinson 1986). The geometrical distribution of the normal faults possibly reflects the presence of a ramp flat in the surface of the Okanagan valley fault (see McClay and Ellis 1987).*

Figure 2.6: *Section A-A' located on the compilation map of Figure 2.1. Structural features surrounding the Okanagan valley fault near Kelowna. The shallow westerly-dipping Okanagan valley fault is underlain by mid-crustal gneisses overprinted by extensional ductile fabrics for approximately 1 km down section. These fabrics overprint Eocene granitoids and are in turn overprinted by cataclastic structures mantling the uppermost 100 meters of the sheared gneisses. Hanging wall rocks are detached from mid-crustal rocks and do not show any evidence of unannealed ductile deformation. They are cut by steep normal faults that may become listric down section and probably merge with the Okanagan valley fault. The surface of the Okanagan valley fault is cut by 48 Ma old dacite plugs feeding the lavas of the Joe Rich klippe. The elevation of the Big White shear zone is indicated on the right scale and represents the base of the Okanagan valley fault and should project 20 km to the east of section A-A'.*

5. Discussion

Protomylonites in Zone B formed under non-coaxial conditions and simultaneously developed shear planes and planes of preferred mineral orientations expected to have been initially oriented at 45° from one another. As deformation intensified in Zone A, grain size reduction and recrystallization were accompanied by a reduction in the angle between the flattening surface S and the principle shear surface C to 15° . Locally the S- surface has rotated into near parallelism due to very high strain conditions (Berthé et al. 1979). The density of C- surfaces increases steadily towards the top of the Okanagan complex which may result from the creation of new shear planes. On the other hand S- surfaces may have been activated where their angle with the C- surfaces is decreased to less than 10° .

Mechanical processes associated with high-strain extensional faults depend on the modal mineralogy, the heterogeneity of the rock and the temperature and pressure conditions prevailing during deformation. The behavior of high-strain zones is thus controlled by a relative depth factor at which the structures develop and propagate. Accordingly shear zones are classified as either ductile or brittle if they form either below or above a rheological boundary termed the brittle-ductile transition. Under average geothermal conditions ($30^\circ\text{C}/\text{km}$) the brittle-ductile transition occurs at a depth of approximately 15 km (Sibson 1977) which is the theoretical base of the greenschist grade of metamorphism (Scholz 1989) and the lower boundary of the upper crust. Certain greenschist assemblages may be superimposed on high-grade rocks by fluid circulation parallel to permeable fault surfaces (Kerrick 1988). The Okanagan valley fault exhibits brittle-ductile fabrics. The presence of ductile fabrics overprinting amphibolite-facies minerals in the footwall suggest that they formed under amphibolite-grade conditions at depths of approximately 15 km assuming a geothermal gradient of $30^\circ\text{C}/\text{km}$. Brittle fabrics are clearly accompanied by greenschist-grade assemblages and formed at shallow crustal levels (*see* Chapter 3). The relative timing between these features has been tentatively established (*see* Chapter 4) and suggest that both ductilely and brittlely deformed rocks were created nearly simultaneously.

The present-day attitude of the Okanagan valley fault suggests it is a low-angle surface (15°) at shallow depths. Seismic reflectors associated with the Okanagan valley fault west of Vernon (Cook et al. 1992) indicate that the fault surface becomes concave upwards with three

ramps/flat segments and flattens to present day mid-crustal levels. This geometry resembles that of most major detachment faults observed in the Basin and Range Province (Wernicke 1981, 1985). Whether the Okanagan valley fault was initiated as a planar or a listric structure in both the brittle and ductile seismic zones (Jackson and McKenzie 1983) is critical in estimating the magnitude of extension (Wernicke and Burchfiel 1982) when critical offset markers are missing (see Chapter 5). Of course, the present day fault orientation has been affected by substantial differential exhumation and trap-door rotation about the vertical in both the footwall and hanging wall blocks. Nevertheless it seems probable that the present day attitude of the Okanagan valley fault represents an exhumed Eocene flat. It is noticeable that syn-extensional sediments in the hanging wall near Kelowna have not been substantially rotated ($20-35^\circ$) into the fault surface. If we assume large magnitudes of extension and large displacement across the Okanagan valley fault one would expect that the hanging wall sedimentary basins would have been rotated to steeper angles (Wernicke and Burchfiel 1982) if the fault surface was rotational. Graphically estimated magnitudes of extension assuming rotational fault conditions, comparing the dip of bedding against an angle between bedding and fault of 35° , reach 100% with a planar surface geometry but it decrease to approximately 60% using a listric geometry. Because the Okanagan valley fault seems to have been non-rotational the amount of extension reached from its present-day configuration therefore approximates 100%.

The current debate on the dynamics of extensional faulting is far from creating a consensus. Recently Carr (1990) documented that extensional dislocations nucleated at mid-crustal depths from anatectic melts on the east flank of the Okanagan complex near Thor-Odin. This study indicates that extensional shear fabrics along the Okanagan valley fault appear to have formed progressively downwards possibly nucleating from leucocratic anatectic melts in a relatively short time span. The superposition of brittle structures onto ductile fabrics are consistent with a progressive exhumation of the footwall of the Okanagan valley fault.

It has been suggested for some time that extensional faults are produced by reactivation of compressional structures (Bally et al. 1966; Smythe et al. 1982) and vice versa normal faults may be reactivated as thrusts (Jackson 1980; Stoneley 1982) and there is no theoretical limit on the number of times the direction of motion on a fault can reverse. Ivins et al. (1990) have

demonstrated that preexistent crustal inhomogeneities, such as abandoned thrusts, may be reactivated as extensional structures under high P_{max} reducing the effective strength of the failing material. Fluid rheology may have been provided by coalescent anatectic magmas formed in the Okanagan complex. There is no direct evidence of ancient compressional shear zones in the Okanagan complex near Kelowna but structures mapped by previous workers in the southern Okanagan complex (Christie 1973; Ryan 1973) may be compressional in origin.

6. Conclusions

The sharp tectonic and metamorphic contrast across the Okanagan valley is controlled by a major extensional structure named the Okanagan valley fault. Footwall gneisses of the Okanagan complex exhibit typical features of protomylonites and mylonites. A stretching lineation is defined by the long axes of hornblende crystals and feldspar aggregates, pressure shadows, ribbon-shaped quartz, linear biotite streaks, elongate feldspar porphyroclasts and sheath fold axes aligned with an average trend of N285W. This lineation is pervasive throughout the Okanagan complex on an estimated structural thickness of approximately 1 km and as far as 50 km east of Kelowna.

Extensional strain in the Okanagan complex is partitioned in three zones representative of a progressive strain gradient. Deepest rocks of Zone C are mainly highly recrystallized plutonic rocks passing upward into protomylonitic leucogneisses of Zone B with synkinematic biotite and hornblende. Plagioclase and hornblende porphyroclasts are reduced in size in Zone A and are progressively chloritized and sericitized while K-feldspar has developed myrmekite microtextures and has been partly recrystallized to muscovite. Ultramylonites in Zone A systematically exhibit a WNW sense of non-ideal progressive simple shear and may have been formed as early as 56 Ma ago. The top of Zone A is a mylonite front with superimposed brecciation referred to as sub-Zone A₁ which formed most certainly 50 Ma ago and possibly as early as 59 Ma ago.

Rocks in the hanging wall of the Okanagan valley fault have been affected by extensional brittle structures. The displacement of the Joe Rich klippe east of Kelowna may have been as large as 60 km.

1. *Introduction*

Metamorphism associated with crustal extension has received little attention but where data exist, as in the Aegean region, it appears that present day extension follows compressional regimes in which metamorphic assemblages record prograde decompression followed by retrograde decompression (Buick and Holland 1989). This text describes the structural and metamorphic evolution of pelitic schists in the Okanagan complex (Christie 1973; Ryan 1973) which lies several kilometers above the projected surface of the principal zone of compressional shear in the hinterland of the southern Canadian Cordillera, the Monashee décollement (Brown et al. 1986; Brown et al. 1992; Cook et al. 1992). Rocks of the Okanagan complex were exhumed by the Okanagan valley normal fault in the Early Eocene (*see* Chapter 4) soon after compression ceased on the Monashee décollement (Carr 1990). Metapelitic textural and compositional evolution together with certain P-T estimates are consistent with decompression during heating followed by rapid cooling and decompression of the Okanagan complex during extension. It is proposed that the Barrovian metamorphism in schists of the Okanagan complex was relatively stable during the transfer from a compressional to an extensional tectonic regime under non-coaxial strain conditions at mid-lower crustal levels. However, results presented herein suggest that mineral homogenization has probably been enhanced by strain and cannot be considered as a purely temperature dependent diffusion process.

2. *Regional metamorphism*

The Okanagan complex straddles the western boundary of the southern Omineca Belt. It is a tectonic window through the southwestern flank of the Shuswap metamorphic complex a domal culmination comprising middle and lower crustal rocks that were juxtaposed by thrusting along the northeasterly directed Monashee décollement (Brown et al. 1986).

Lower crustal rocks underlying the Monashee décollement define the Monashee terrane (Ranalli et al. 1989). These rocks were affected by three contractional phases (Journeay 1986;

Read and Klepacki 1981). (1) Pre- peak of metamorphism E verging D1 folds are overprinted by; (2) syn-peak of metamorphism D2, NE to N verging, tight to isoclinal folds that are in part synchronous with the earlier movement on the Monashee décollement and; (3) NE verging D3 folds postdating the peak of metamorphism and the shearing on the Monashee décollement.

Middle crustal rocks overlying the Monashee décollement are exposed in the Thor-Odin culmination. They exhibit six phases of deformation (Carr 1990), three of which are major contractional phases: (1) rare D1 isoclinal folds trending N with a NE vergence similarly to; (2) tight to isoclinal D2 folds with shallow plunging axes trending WNW formed during peak metamorphism reaching upper amphibolite grade; (3) upright to reclined D3 folds nearly coaxial with D2 structures formed near the end of peak metamorphic conditions. Mineral dating of granitic bodies affected by D2 structures suggests that peak metamorphic conditions in the Thor-Odin area lasted until the Late-Cretaceous-Paleocene (Carr 1990, 1992) and correspond to the final event of plastic strain accommodated by the Monashee décollement (Coleman 1989, 1990). Compressional structures in Thor-Odin are affected by open folds. They are also affected by east- and west-dipping extensional shear structures such as the west-dipping Beaven fault interpreted as the breakaway of the Okanagan valley fault (Carr and Brown 1990; Cook et al. 1992).

The Okanagan complex lies 50 km south-southwest of the Thor-Odin culmination. Stratigraphic correlations may exist between these regions (Jones 1959; Solberg 1976; *see* Chapter 1) but they probably represent different mid-crustal levels (Cook et al. 1992). Paragneisses of the Okanagan complex were affected by three major phases of folding overprinted by open folds and brittle structures (Ryan 1973, Christie 1973, Medford 1976, Solberg 1976). These structures comprise: (1) rootless isoclinal D1 fold hinges of various sizes with N trending axes; (2) syn-peak of metamorphism isoclinal D2 folds, mainly rootless with WNW trending axes; (3) post- peak of metamorphism D3 open to recumbent folds with an average trend nearly coaxial with D2 folds. Many granitic bodies, described as migmatitic leucogneisses (Ryan 1973; Medford 1976), have been injected during D2 and/or D3 throughout the Okanagan complex.

Structures and metamorphic histories affecting the Okanagan complex and the Thor-Odin culmination need not have been formed simultaneously but it appears that metasedimentary rocks of both regions were affected by as many as and similar phases of deformation. "Peak"

metamorphism seems to have occurred during D2 in both regions while subhorizontal isoclinal folds and mineral stretching lineations were forming along WNW trends. Both regions exhibit D3 folds coaxial with D2 axes, furthermore, D2 and D3 events of both regions are marked by the injection of abundant leucogranites (Medford 1976; Carr 1990; *see* Chapter 1).

D2 structures (foliations and isoclinal folds) in the Okanagan complex may be Eocene and extensional. A monzonitic body affected by D2 foliations near Vernon (Fig. 3.1) provided a Sr/Rb whole-rock age of 42 ± 10 Ma that was interpreted as a thermal reset date (Solberg 1976) but it may, alternatively, indicate the cooling age of the D2 event. Within error, this whole-rock date is consistent with a magmatic zircon age of 51 ± 1 Ma obtained from a porphyry affected by extensional shear fabrics near Okanagan Falls (Parkinson 1985). It is noticeable that extensional fabrics affect mainly granitic bodies. Incidentally these fabrics are sub-parallel to D2 foliations described as compressional features in paragneisses hosting the granitoids (Ryan 1973; Christie 1973; Medford 1976; Solberg 1976). Perhaps extensional fabrics have to be distinguished from compressional fabrics on the basis of the lithotypes they are affecting. Consequently the age of compressional D2 fabrics in the Okanagan complex is unknown.

Regionally across the southern Omineca Belt D2 structures are described as compressional features formed in the Late Cretaceous-Early Tertiary under "peak" metamorphic conditions (Carr 1990; Scammell 1991). Carr (1990) demonstrated that D2 structures at the base of the Thor-Odin culmination, immediately above the Monashee décollement, ceased to form 58 Ma ago. Extensional fabrics on the west flank of Thor-Odin, in the Okanagan complex, may have formed as early as 56 Ma ago. This is inferred from the age of a concordant zircon fraction found in a mylonitized leucogranite underlying the Trinity Hills (*see* Chapter 4) 40 km west of Thor-Odin. Based on their present structural position it is expected that D2 compressional structures in Thor-Odin formed at deeper levels than similar structures in the Okanagan complex (Brown et al. 1992; Cook et al. 1992). Perhaps rocks of the Okanagan complex have been affected by different contractional features thus metamorphism may have been diachronous during the D2 event. It is tentatively proposed that syn-D2 peak metamorphic assemblages preserved in the Okanagan complex may have been formed under contractional conditions but were affected by extensional fabrics immediately after the "peak" of metamorphism leading to rapid quenching of the

metamorphic assemblages under rapid exhumation rates. Rapid exhumation of the Okanagan complex is consistent with the fact that high-grade minerals are stretched along the direction of extensional transport. These minerals are merely overprinted by lower-grade phases (kyanite, andalusite, staurolite, cordierite).

Paleozoic rocks included in the Upper Crustal layer exhibit regionally as many as three phases of deformation (Carr 1990) but those lying west of the Okanagan complex in the hanging wall of the Okanagan valley fault, were regionally affected by two foliations of which D1 is dominant. These rocks are part of the allochthonous Quesnellia Terrane and locally reached lower amphibolite grade during the formation of WNW trending D1 folds. The latter were overprinted by post-peak of metamorphism E trending angular D2 folds west of Vernon (Solberg 1976). A K-Ar date of 178 ± 6 Ma provided by amphiboles parallel to the S1 foliation led Solberg (1976) to suggest that peak metamorphic conditions in the Upper Crustal layer were reached in the Jurassic. This is consistent with a 159.9 ± 0.7 Ma age obtained from an undeformed biotite granodiorite intruding Paleozoic rocks and cutting D1 structures near Kelowna (*see* Chapter 4). In the Eocene, rocks of the Upper Crustal layer have been faulted several kilometers down along the Okanagan valley fault (*see* Chapter 5). They were reheated to approximately 100°C by syn-extensional metasomatic fluids as indicated by Eocene apatite fission track dates in Jurassic plutons (Medford 1975).

3. *Metamorphic contrasts across the Okanagan valley*

Several attempts were made in the past to correlate structural and metamorphic histories across the Okanagan valley (Okulitch 1969; Ryan 1973; Solberg 1976; Ross 1981) on the basis of similar orientations and geometries. However recent mapping has revealed that both sides of the valley represent different crustal blocks that were deformed and metamorphosed diachronously (Parkinson 1985; *see* Chapter 4). On the basis of regional correlations we propose that D2 structures affecting metapelites of the Okanagan complex are Late-Cretaceous-Early Tertiary whereas D1 and locally D2 affecting Paleozoic strata of the Upper Crustal layer are pre- to syn-Late Jurassic (Ryan 1973; Solberg 1976; Fig. 3.1).

Figure 3.1: *Sites of thermobarometric samples collected nearly at the same structural level in the footwall of the Okanagan valley fault. E: Enderby; TV: Trinity valley; HR: Hunters Range; SF: Shuswap Falls; V: Vernon; K: Kelowna; O: Osoyoos; OF: Okanagan Falls; T: Tonasket. Metamorphic isograds are only loosely constrained and partly compiled from (Jones 1959; Ryan 1973; Okulitch 1979; Carr 1990). Staurolite-out and muscovite-out isograds separate metamorphic zones which contain: St: staurolite-muscovite; S-M: sillimanite-muscovite; S-K: sillimanite-K-feldspar-melt. Diagonally hatched pattern indicates Upper Crustal level rocks which are generally at greenschist grade but reaches lower amphibolite grade near Osoyoos (O; Okulitch 1969) and Vernon (V; Solberg 1976).*

3.1 Petrography of the Upper Crustal layer

Prior to extension, Paleozoic rocks of the Upper Crustal layer were involved in the obduction of the Intermontane Superterrane onto the paleocontinental margin of North America in the Late Jurassic (Monger et al. 1982; Brown et al. 1986). Intermontane lithologies straddling the northern Okanagan region include two Late Paleozoic metavolcanic series correlated with the Harper Ranch Group (Monger 1982) near Enderby and with the Nicola Group west of Kelowna (Bardoux 1986). These rocks are interdigitated with graphitic schists, pelites, and carbonates of uncertain stratigraphic correlation (Little 1961; Tempelman-Kluit 1988).

Syn- to post-D1 assemblages consist, in volcanic rocks, of biotite+actinolite+chlorite+plagioclase+epidote whereas metapelites interdigitated with the volcanic rocks contain biotite (partly retrograded to chlorite)+garnet+quartz+plagioclase.

An undeformed biotite-magnetite leucogranite body has developed a narrow thermal contact aureole in hosting Paleozoic lithologies southeast of Enderby (Jones 1959; Okulitch 1979; Fig. 3.1). Metasedimentary and metavolcanic rocks in the aureole contain respectively the assemblages biotite+garnet+muscovite+chlorite+plagioclase and plagioclase+epidote+actinolite which indicate that the intrusion was emplaced at 200-300MPa pressure (7-10 km) after M1 metamorphism. The granitic body contains rafts of sillimanite-bearing paragneiss which probably belong to the Middle Crustal layer underlying the Okanagan valley fault. The age of the intrusion is unknown but it is petrographically comparable to the Eocene Ladybird suite as described by Carr et al. (1987) and Parrish et al. (1988). This body may be a residual phase of mid-crustal anatexis melting.

3.2 Petrography of the Middle Crustal layer

Rocks of the Okanagan complex represent the Middle Crustal layer. It comprises two alternating units of: (1) sillimanite and/or staurolite-bearing pelites and semipelites interlayered with quartzites, diopside-calcic calcisilicate and amphibolite; and (2) foliated leucosomes grading progressively eastward to large undeformed bodies of porphyritic hornblende-biotite±muscovite granites. Staurolite schists are mainly exposed in the northernmost Okanagan complex east and west of Trinity valley (Jones 1959) and in the southernmost Okanagan complex near the U.S.-

Canada border (Fig. 3.1; Ryan 1973). Sillimanite-bearing schists occur mainly in the central Okanagan complex (Christie 1973; Ryan 1973; Medford 1976) but they have also been mapped on all sides of the Trinity valley (Johnson 1939b; Fig. 3.1). Migmatitic leucogneisses containing pelite, marble and amphibolite rafts near Kelowna, Okanagan Falls and Osoyoos are tentatively interpreted as core zones of the Okanagan complex.

The absence of contact metamorphism between plutonic bodies and hosting sillimanite-bearing paragneisses (Christie 1973; Ryan 1973) suggests they are *in situ* melts or else they intruded hot rocks. An anatectic origin is consistent with the gradual appearance of leucogranites across the staurolite-out isograd (Fig. 3.1; Ryan 1973) and with the abundance of leucosomes in zones of highest metamorphic grade throughout the Okana(o)gan complex(es) near Tonasket (Holder and McCarley-Holder 1988), Osoyoos (Ryan 1973), Penticton (Christie 1973) and Kelowna. Regional metamorphism seems inseparable from syn-kinematic leucocratic anatexis and plutonism and associated potash metasomatism which has given rise to orthoclase porphyroblasts.

Non-coaxial strain fabrics are preserved in the upper 500m of the Middle Crustal layer whereas annealed textures occur at deeper levels. Perhaps the absence of extensional fabrics at greater depths suggests that extensional deformation outlasted metamorphism at the structural top of the Middle Crustal layer but maybe extensional faulting was discrete and did not penetratively deform the entire layer.

We sampled pelitic and Al-poor semipelitic schists (Table 3.1) which are generally partitioned in two petrographic domains created by metamorphic segregations (Robin 1979; Vernon 1988). Layers classified as domain M are richer in mica (biotite $\pm$ muscovite) than layers in domain Q which are mainly composed of quartzo-feldspathic leucosomes (migmatite sweats) containing garnet porphyroblasts. Textures in Q domains are attributed to a mechanism of dislocation flow (Vernon 1988), garnets are often ovoid and asymmetric becoming increasingly serrated in those samples of the northernmost Okanagan region (45, E4). Pressure shadows of the garnet porphyroblasts are composed of annealed quartz, feldspar, and biotite and are probably synkinematic (Plate 3.1a, b; Hanmer and Passchier 1991). Layers of domain M are often intergrown with fibrolite aggregates (Plate 3.1) and represent zones of high non-coaxial strain. Fibrolite needles observed in the field are generally elongated parallel to the pervasive stretching

2

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Plate 3.1:

Photomicrographs of mineral textures from mylonitized metapelitic assemblages in the Okanagan complex illustrating Q and M domains (as described by Vernon 1988). Domain Q consist of quartzo-feldspathic leucosomes. Domain M contains mainly micas intergrown with fibrolite aggregates. (a) Sample E4 showing porphyroblastic sillimanite and embayed garnets. Garnet fragments (upper right) under extensional strain conditions became passive rigid bodies rotating in the matrix. A rootless micrometric fold hinge is observed in the upper left corner. This fold hinge is interpreted as an ancient contractional feature overprinted by C and S fabrics. Biotite lies mainly in pressure shadows and as inclusions in garnet (lower left and upper right). Ovoid garnet in lower right exhibits an inclusion-rich core (biotite + quartz) and an inclusion-free rim. (b) Sample 47A, intensely fractured garnet porphyroblast with only few quartz inclusions. Biotite inclusions along rims could be an unused reactant overgrown as garnet grew or else it overgrows garnet. Pressure shadows comprise mainly biotite with minor amounts of annealed quartz and plagioclase. Porphyroblast shares contact with a K-feldspar porphyroblast in upper half. (c) Sample 45, biotite + garnet assemblage affected by extensional shear fabrics C, S, C' indicating upper-member-to-the-west sense of shear. Garnet porphyroblast to the lower left is pulled apart along inclusion-rich trails oriented perpendicular to the shear foliation in the direction of transport during extension. Abundant micrometric inclusions are mainly quartz and minor hematite as shown in enlargement (d) where the straight alignment of inclusions is interpreted to be a pre-extensional foliation. Mineral abbreviations: Bi: biotite; Gt: garnet; Si: sillimanite; Pl: plagioclase; Qtz: quartz; Hm: hematite; K: K-feldspar.

direction overprinted on shear foliations throughout the Okanagan complex (*see* Chapter 2). Garnets tend to fracture perpendicular to the stretching direction (Plate 3.1b, d). Samples from Kelowna (5, 7) exhibit only a weak foliation whereas those of Okanagan Falls (OF1), Vernon (45, 47) and Enderby (E4) exhibit unannealed textures indicating top-to-the-west sense of shear. Because extensional features are pervasive in the samples collected ancient foliations can only be inferred locally and indirectly. Two cases are illustrated in Plate 3.1; a micrometric fold hinge in sample E4 (Plate 3.1a) interpreted as an ancient structural feature similarly to abundant quartz inclusions defining linear trails oriented perpendicular to the extensional shear foliation in sample 45 (Plate 3.1d). These features may relate to a contractional event.

The highest grade assemblage mapped in the northern Okanagan complex occurs in sample E4 at the southern end of the Hunters Range east of Enderby (Fig. 3.1). It consists of abundant folia of fibrolite overgrown by idiomorphic sillimanite anastomosing around intrafolial zones composed of plagioclase + garnet (with staurolite inclusions) + quartz + biotite + K-feldspar + graphite. Idiomorphic sillimanite randomly overgrows the fibrolite aggregates suggesting that they recrystallized from the aggregates under reducing non-coaxial strain and prograde conditions (Vernon 1988). The Enderby schists are interdigitated with diopsidic marbles and amphibolites that are tentatively correlated with similar lithotypes associated with the Late Proterozoic Horsethief Creek Group in the Hunters Range north of Enderby (Johnson 1988; *see* Chapter 1).

4. *Analytical procedure*

Six pelitic and semipelitic samples were collected from unconstrained stratigraphic units at nearly the same structural level in the strained zone of the Okanagan complex. These are from the northern Okanagan complex near Vernon (samples 45, 47), Enderby (E4), Kelowna (5, 7), and Okanagan Falls (OF1). Four of these samples exhibit unannealed extensional shear fabrics (E4, 45, 47, OF1) and preserved stable mineral assemblages though they are located in the high strain Zone A underlying the Okanagan valley fault (Fig. 3.1; *see* Chapter 2). The samples from Kelowna (5, 7) exhibit annealed shear fabric textures preserving by quartz ribbons and asymmetric feldspar porphyroblasts and represent slightly deeper structural levels.

A minimum of two domains of biotite-garnet-plagioclase-K-feldspar sharing mutual

contacts were probed in all six samples including two duplicates (5 and 47) to check for data coherency. Each domain analyzed is <2 mm in radius and contains at least one grain of each mineral phase. Domains in each sample are separated by at least 5 mm. Porphyroblast transects mainly are parallel to the long dimension of each grain and were made with spot-analysis every 10-30 μ m to test for retrogressive dissolution and strain-induced chemical disequilibrium in pressure shadows (Duebendorfer and Frost 1988). Mineral assemblages analysed are given in Table 3.1 and follow a north-south distribution spread over 100 km longitudinally (Fig. 3.1). Several grains of each mineral were analysed several times with a Cameca Camebax WDS Electron microprobe at the Department of Geological Sciences at McGill University. The operating conditions were as follows: accelerating voltage, 15 kv; sample current, 10 nano-amperes; beam diameter, 2 μ m; spectra collecting time, 20 s. The standards used for each element are: Na, albite; Mg, olivine; Al, Si, K, orthoclase; Ca, Fe, andradite; Ti, Mn; MnTiO₂. The treatment of the raw data was done using ZAF correction procedures established by Cameca. Representative analyses of garnet cores and rims, biotite cores and rims, and plagioclase cores and rims are given for each sample in Tables 3.2 to 3.7. Complete rendition of the microprobe data is given in Appendix 2. Details of the mineral chemistry of each mineral phase are discussed in section 7.

TABLE 3.1. Mineral assemblages in the pelitic rocks of the northern Okanagan complex

Sple no.	Text. (*)	Loc. (#, @)	Phases									
			Ga[Chl]	Si[Chl]	St[Chl]	Mu	Bt[Chl]	Kf[Ser]	Pl[Ser]	Qz	Ilm	Rt
E4	(a)	En/p	X	X	X _w	--	X	--	--	X	X	X
45	(u)	En/p	X	--	X	X	X	--	--	X	X	X
47	(u)	Vn/p	X	X	--	--	X	X	X	X	--	--
5	(a)	Kl/sp	X	X	--	--	X	X	X	X	--	--
7	(a)	Kl/sp	X	X	--	--	X	X	X	X	--	--
OF1	(u)	OF/p	X	X	--	X	X	X	X	X	X	--

(*) Observed textures; u: unannealed extensional; a: annealed.

(#) Localities (north to south); En: Enderby; Vn: Vernon; Kl: Kelowna; OF: Okanagan Falls.

(@) Rock type; p: pelite; sp: semipelite.

[] Retrograde mineral phases; Chl: chlorite; Ser: sericite.

_w Inclusion only

5. *Metamorphic reactions in the Middle Crustal Layer*

The regional distribution of peak metamorphic zones and isograds along the Okanagan valley is shown in Figure 3.1. Systematic variation of mineral assemblages are rare suggesting that, in general, the Middle Crustal layer is at the same metamorphic grade. A sharp metamorphic break trending east-west straddles the southern end of the Hunters Range stepping into the Trinity valley. The latter exhibits greenschist-grade metavolcanic and metasedimentary rocks correlated with the Mount Ida Group (Fig. 3.1; Jones 1959; Okulitch and Cameron 1976). At this locality high-grade rocks of the Hunters Range provided Eocene K-Ar cooling ages in sharp contrast with Late Jurassic K-Ar cooling ages preserved in the greenschist-grade rocks underlying the Trinity valley (Mathews 1981).

Porphyroblastic staurolite+garnet schists are exposed along the western slope of Silver Star mountain east of Vernon and between Trinity Creek and Shuswap Falls. These two localities bound the western and eastern sides of greenschist grade rocks underlying the Trinity valley (Fig. 3.1). The staurolite schists are linked with the Okanagan complex because they exhibit mylonitic strain fabrics indicating top-to-the-west sense of shear (*see* Chapter 2; Plate 3.1c) and for several other reasons explained below.

Rocks exposed near Shuswap Falls extend continuously northwards to Mabel Lake via the eastern slope of the Trinity Hills (Fig. 3.1). They link with the high-grade rocks of the Hunters Range where westerly-directed extensional fabrics overprint older ?compressional fabrics (Johnson 1988, 1989b). Slightly tilted Eocene sedimentary rocks and lavas cap the low-grade rocks of the Trinity Hills which are underlain by gneisses of the Shuswap Falls area. The gneisses are interlayered with an Eocene granitic body also affected by westerly-directed mylonitic fabrics (Plate 2.2f; Fig. 2.2; *see* Chapter 4). The extensional shear fabrics in the Shuswap Falls rocks seem to post-date high-grade metamorphism and are consistent with the petrographic textures observed in the southern Okanagan complex and on the west flank of Thor-Odin east of Mable Lake (Fig. 3.1; Carr 1990).

Rocks of the west flank of Silver Star mountain comprise three mineral assemblages increasing in grade from east to west towards Swan Lake. These rocks cross the staurolite-out isograd and the muscovite-out isograd separating the sillimanite-muscovite and sillimanite-K-

feldspar zones (Fig. 3.1). Staurolite-garnet-biotite-muscovite schists near the summit of Silver Star mountain, are severely retrograded near the contact with greenschist-grade graphitic schists of the Trinity valley. Further west near Vernon sillimanite-K-feldspar pelites are overprinted by top-to-the-west ultramylonitic fabrics (see Plate 2.1) suggesting that these rocks have suffered high strain in the Early Eocene after high-grade metamorphic conditions were reached.

Metamorphic gradients in the Shuswap Falls and Silver Star mountain regions are similarly increasing but in opposite directions that is away from the greenschist rocks of the Trinity valley. Rocks of both regions are also overprinted by co-directional extensional features whereas rocks underlying the Trinity valley are unaffected by these structures. This suggests that the low-grade rocks underlying the Trinity valley form a thin klippe related to the Upper Crustal layer. This klippe probably rests tectonically on amphibolite-facies rocks of the Middle Crustal layer similarly to the situation prevailing further south (Fig. 3.1). Though structural features and metamorphic events may be regionally diachronous it is proposed that peak metamorphic assemblages of the Silver Star and Shuswap falls gneisses are M2 assemblages formed during regional compression events ending in the Early Tertiary (Brown et al. 1986; Carr 1990). Because extensional shear fabrics are overprinting these non-retrograded high-grade assemblages the latter may be compressive M2 assemblages. These textural relationships could suggest that compression transferred simultaneously to an extension regime in the Early Tertiary at this latitude of the southern Omineca Belt (Carr 1990). Peak assemblages of sillimanite-K-feldspar have probably been quenched by rapid exhumation because they are nowhere overprinted by lower-pressure and lower-temperature kyanite $\pm$ andalusite assemblages. Greenschist-grade phases do occur in narrow zones mantling the Middle Crustal layer but these assemblages were formed along permeable brittle zones by fluids flushed along the Okanagan valley fault system under seismic pumping conditions (Sibson 1977).

5.1 Peak mineral assemblages

One could write several reactions involving the phases most commonly observed in the pelitic rocks analysed; (1) staurolite + quartz + biotite + rutile = garnet + sillimanite + K-feldspar + ilmenite + H₂O. Garnets generally contain quartz and biotite inclusions but muscovite

and sillimanite inclusions were found in OF1 and suggest the reaction whereby; (2) muscovite+biotite+quartz=garnet+sillimanite+K-feldspar+H₂O. Muscovite is absent in samples 5, 7, 47, and E4 and may be related to breakdown reactions such as; (3) muscovite+quartz=sillimanite+K-feldspar+H₂O or; (4) muscovite+quartz=sillimanite+anatectic melt (Kerrick 1972) or in accordance with the absence of feldspars in certain sillimanite schists E4 and staurolite schist 45; (5) staurolite+muscovite=garnet+biotite+quartz. Three pelitic assemblages OF1, 45, E4 contain accessory minerals ilmenite and/or rutile occurring as idioblastic grains in the matrix or as inclusions in garnets and/or biotite. Apatite, tourmaline, zoned zircon and discontinuous folia of graphite are also present in most pelites.

Calcsilicates interleaved with the semipelites of the Kelowna area (5, 7) typically contain amphibolite-grade assemblages of diopside+titanite+scapolite+epidote+allanite±olivine±spinel. The latter two phases are severely serpentinized. Orthogneisses interfoliated with these rocks are predominantly composed of quartz+plagioclase+K-feldspar+hornblende+biotite±chlorite (retrograde) with accessory minerals such as titanite, zircon, monazite and thorite.

5.2 Retrograde assemblages

Retrogression has affected most peak assemblages observed in the Okanagan complex (Table 3.1). Sillimanite needles, K-feldspar, garnet and staurolite porphyroblasts at the highest structural levels of the complex have been pulled apart and retrograded to chlorite+muscovite (Plates 3.1d, 3.2). In many cases there is obvious retrogression of biotite reacting to form chlorite±K-feldspar with exsolutions of ilmenite and/or rutile.

Pervasive growth of hydrated minerals occurs in tension fractures in the cataclastic Zone A, mantling the Okanagan complex (see Chapter 2). This assemblage consists mainly of epidote+sericite+calcite+pyrite. Retrograde chlorite crystallizes generally from biotite and hornblende in orthogneisses cutting the pelitic rocks whereas feldspars in the same rocks are pervasively replaced by white mica (sericite). Amphibolitic lenses near the cataclastic Zone A, often contain clinozoisite indicating that these rocks were severely hydrated.

The breakdown of the high grade phases can occur according to several hydration reactions; (6) garnet+K-feldspar+sillimanite+H₂O=muscovite+biotite+quartz; (7) plagioclase+

Plate 3.2: *Pulled apart and retrograded sillimanite. Retrograde assemblage is chlorite + muscovite.*

K-feldspar + H₂O = muscovite + albite + zoisite + quartz; (8) sillimanite + biotite + quartz + H₂O = muscovite + chlorite + ilmenite (Hollocker 1987); (9) staurolite + biotite + quartz + H₂O = muscovite + chlorite + ilmenite (Hollocker 1987); (10) garnet + biotite + H₂O = chlorite + muscovite + O₂. Plagioclase may have reacted with garnet and biotite to form chlorite and excess Ca may be removed from reaction (10) and redeposited as calcite or epidote so that a more general reaction would be; (11) garnet + biotite + plagioclase + H₂O = chlorite + muscovite + albite + epidote + quartz + calcite. If hornblende participated as a retrograded phase with the preceding reactants, a general reaction would be; (12) hornblende + biotite + plagioclase + garnet + ilmenite + H₂O = chlorite + actinolite + albite + epidote + titanite + quartz + K⁺ or; (13) hornblende + plagioclase + biotite + ilmenite + H₂O + CO₂ = chlorite + epidote + titanite + calcite + quartz.

All retrograde reactions probably involved undetermined volumes of H₂O and CO₂ (Kerrick and Hyndman 1986). These fluids may have been derived from deep crustal sources by prograde metamorphism and dehydration of metasedimentary rocks following crustal thickening (Smith et al. 1991; see Chapter 4). Mantle degassing processes (Gough 1986; Beaudoin et al. 1991) and subduction processes may also have provided fluids to the system.

6. *Mineral chemistry in the pelitic assemblages*

Mineral compositions showing best data quality and considered representative of average modes in the pelitic samples tested are given in Tables 3.2 to 3.7. Biotites are more abundant than garnets in each sample and are probably residual phases of peak metamorphic conditions. Biotite composition (Table 3.2) varies considerably between sites and depending on whether it is in the matrix, sharing contacts with garnet, and/or included within garnet. Compositional variations amongst biotites represent either exsolution of ilmenite, exsolution of rutile, or Fe-Mg exchange with surrounding ferromagnesian minerals. Regional variations likely represent bulk compositional variations (Fig. 3.2a). Matrix biotites exhibit Fe/(Fe + Mg) ratios decreasing towards the Kelowna region and increasing progressively towards the Hunters Range and Okanagan Falls (samples E4 and OF1; Figs. 3.1, 3.2a). Matrix biotites are more Ti rich in the Hunters Range area (sample E4; Table 3.2) but biotite inclusions in garnet show the widest variety in Ti content. These inclusions have generally lower Fe/(Fe + Mg) than matrix biotite

suggesting that Fe-Mg exchange with garnet has occurred during retrogression. Ti content in biotite is expected to increase with metamorphic grade (Spear et al. 1990). Though biotite inclusions in garnets have high Ti contents they all have undergone retrograde Fe-Mg exchange evidenced by diffusion halos in garnet adjacent to the inclusions (Plate 3.3b). This retrogression may be minor relative to rim retrogression. Biotite inclusions within quartz grains themselves included in the core of garnet in sample 47 may record a composition on the prograde P-T path.

Except for sample 45, garnet cores and matrix biotites, show systematic variations in Fe-Mg contents (Fig. 3.2a). Both mineral phases exhibit the smallest Mg/Mg+Fe ratios in samples from both ends of the sampling region (OF1, E4). The most Mg rich biotites and garnets are those of sample 47. These features probably relate to primary bulk compositional variations which reflects, for example, in X_{Mg} that increases progressively from 0.60 to 0.81 northward (Table 3.3; Fig. 3.2b). Most garnets exhibit slightly concave upward almandine growth profiles. Samples 5, 7, 45 and E4 have homogeneous cores and rims that are compositionally zoned (Fig. 3.2b; Plate 3.3c). The concave upward profile of sample 47 may indicate this sample has suffered more intense retrogression than the other samples. Only in sample OF1 and in the local vicinity of biotite inclusions are the garnet cores zoned in Fe and Mg (Fig. 3.2b; Plate 3.3c). The slight compositional changes along garnet rims regardless of the mineral phase in contact and regardless of the trend of the transect suggests that even those rim faces parallel to the maximum instantaneous stretching axis have been homogenized. However, the compositional variations in sample OF1 which exhibits mylonitic non-annealed textures may reflect strain-induced processes (Duebendorfer and Frost 1988). The profile of sample OF1 is symmetric and concave downward with significant chemical growth zoning outlined by two stages of Fe loss that are homogenized by Mg and Mn enrichments at the garnet rim and near muscovite and biotite inclusions (Plate 3.3a). The preservation of this growth zoning may suggest that the southern part of the Okanagan complex did not spend as much time under homogenization conditions or did not attain the homogenization temperatures of the northern Okanagan complex. Conversely, these data might reflect faster exhumation rates in the southern Okanagan complex.

Average matrix plagioclase compositions in the pelites range around An₂₅ but it increases to An₄₀ in samples 5 and 7 (Table 3.4). These local fluctuations may also reflect

Table 3.2. Representative biotite compositions in pelites of the northern Okanagan complex
(m: matrix phase; r: rim composition; i: inclusion phase)

SPL	OF1m	OF1r	5Am	5Ar	5Ai	5Bm	5Br	5Bi	7m	7r	7i	47Am	47Ar	47Ai	47Bm	47Br	47Bi	45m	45r	E4m	E4r	E4i	
FeO	21.62	19.47	20.17	18.11	18.03	15.81	18.13	18.19	17.62	18.69	17.74	18.80	17.34	18.70	15.44	17.42	18.24	17.07	18.35	20.74	20.48	20.91	19.96
MnO	0.53	0.50	0.56	0.12	0.18	0.12	0.31	0.25	0.13	0.21	0.16	0.30	0.09	0.17	0.05	0.11	0.12	0.07	0.04	0.10	0.09	0.01	0.06
Na ₂ O	0.12	0.13	0.10	0.16	0.17	0.28	0.12	0.19	0.31	0.19	0.27	0.17	0.29	0.13	0.28	0.21	0.22	0.35	0.34	0.26	0.30	0.31	0.20
Al ₂ O ₃	18.57	19.27	18.75	19.43	19.33	18.82	19.61	19.58	18.92	19.26	19.67	19.53	18.60	18.81	19.16	18.37	18.49	18.23	20.30	19.86	18.74	20.16	19.64
SiO ₂	35.96	35.43	35.83	35.93	35.78	35.56	35.63	35.68	36.77	35.37	36.03	35.15	36.31	35.92	35.64	35.70	36.29	36.14	36.38	35.32	35.48	35.28	35.93
MgO	7.99	8.49	8.15	9.98	10.28	11.03	9.82	10.06	9.44	9.06	9.44	8.86	10.56	10.06	12.18	11.11	10.62	11.32	9.20	9.13	7.88	8.92	8.77
K ₂ O	10.36	10.26	10.24	10.04	9.97	9.39	10.00	9.98	9.48	10.04	10.18	9.92	9.68	10.34	9.85	10.04	9.98	9.78	9.69	9.68	9.58	8.78	8.48
CaO	0.02	0.02	0.02	0.03	0.04	0.03	0.03	0.04	0.04	0.04	0.04	0.04	0.04	0.04	0.04	0.03	0.04	0.01	0.01	0.01	0.02	0.06	
TiO ₂	2.68	2.70	1.56	2.76	2.30	4.18	2.32	2.34	3.66	2.95	2.85	2.82	2.78	2.97	3.24	3.04	2.77	3.40	1.89	1.16	3.90	2.03	3.76
OH	3.97	3.95	3.90	4.00	3.98	4.00	3.98	3.99	4.06	3.95	4.00	3.94	4.01	4.01	4.02	3.98	4.01	4.01	4.01	3.95	3.96	3.98	3.95
TTL	101.97	100.23	99.26	100.55	100.08	99.23	99.86	100.29	101.40	99.73	100.36	99.49	99.75	101.11	99.90	99.99	100.79	100.39	100.20	100.19	100.41	100.46	100.17

Cations per 22 oxygens

Fe	2.73	2.46	2.59	2.26	2.27	1.98	2.28	2.28	2.17	2.37	2.22	2.39	2.17	2.34	1.92	2.19	2.28	2.13	2.30	2.63	2.59	2.64	2.49
Mn	0.07	0.06	0.07	0.02	0.02	0.01	0.04	0.03	0.02	0.03	0.02	0.04	0.01	0.02	0.01	0.01	0.02	0.01	0.01	0.01	0.11	0.11	0.01
Na	0.03	0.04	0.03	0.05	0.05	0.08	0.03	0.05	0.09	0.06	0.08	0.05	0.09	0.04	0.08	0.06	0.06	0.10	0.10	0.08	0.09	0.09	0.09
Al	3.30	3.44	3.39	3.43	3.42	3.33	3.48	3.47	3.29	3.44	3.47	3.50	3.29	3.32	3.36	3.25	3.26	3.20	3.58	3.55	3.34	3.58	3.36
Al ^{iv}	0.72	0.82	0.90	0.80	0.80	0.66	0.85	0.82	0.71	0.80	0.87	0.84	0.74	0.69	0.66	0.62	0.68	0.59	1.02	0.91	0.72	0.90	0.75
Si	5.42	5.37	5.50	5.38	5.38	5.33	5.37	5.36	5.42	5.36	5.39	5.34	5.45	5.37	5.30	5.37	5.42	5.39	5.44	5.36	5.37	5.31	5.39
Mg	1.80	1.92	1.86	2.23	2.30	2.46	2.20	2.25	2.29	2.05	2.11	2.01	2.36	2.24	2.70	2.49	2.36	2.52	2.05	2.07	1.78	2.00	1.97
K	1.99	1.98	2.00	1.92	1.91	1.79	1.92	1.91	1.78	1.94	1.94	1.92	1.85	1.97	1.87	1.93	1.90	1.86	1.84	1.87	1.85	1.69	1.62
Ti	0.30	0.31	0.18	0.31	0.26	0.47	0.26	0.26	0.41	0.34	0.32	0.32	0.31	0.33	0.36	0.34	0.31	0.38	0.21	0.13	0.44	0.23	0.42
OH	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
TTL	17.64	17.61	17.64	17.58	17.65	17.47	17.60	17.63	17.47	17.58	17.56	17.57	17.56	17.64	17.62	17.65	17.62	17.60	17.53	17.76	17.48	17.55	17.37

N _a	0.486	0.442	0.462	0.403	0.401	0.355	0.405	0.404	0.401	0.424	0.401	0.427	0.385	0.416	0.340	0.385	0.404	0.375	0.411	0.457	0.459	0.458	0.442
N ₂	0.301	0.431	0.412	0.464	0.502	0.552	0.391	0.368	0.400	0.367	0.381	0.356	0.422	0.398	0.478	0.441	0.415	0.448	0.357	0.360	0.316	0.347	0.344
N ₃	0.983	0.880	0.985	0.976	0.97	0.953	0.960	0.960	0.960	0.970	0.960	0.959	0.953	0.983	0.959	0.971	0.970	0.949	0.947	0.959	0.954	0.947	0.947
N ₄	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000

N ₂	0.486	0.442	0.462	0.403	0.401	0.355	0.405	0.404	0.401	0.424	0.401	0.427	0.385	0.416	0.340	0.388	0.404	0.375	0.411	0.457	0.459	0.458	0.442
N ₂	0.301	0.431	0.412	0.404	0.502	0.552	0.391	0.396	0.400	0.367	0.351	0.359	0.422	0.399	0.478	0.441	0.418	0.448	0.357	0.360	0.316	0.347	0.345
N ₂	0.983	0.980	0.985	0.976	0.971	0.953	0.960	0.960	0.960	0.970	0.960	0.975	0.975	0.980	0.959	0.970	0.970	0.949	0.947	0.959	0.954	0.947	0.947
N ₂	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000	1.000

Table 3.5. Representative muscovite compositions in pelites of the northern Okanagan complex (cc: core of crystal; tr: rim of crystal; it: inclusion in garnet)

	SPLE	OFcc	OFtr	OFit	45cc	45tr	45it
FeO	1.57	1.47	1.58	1.15	1.23	1.37	1.37
Na ₂ O	0.55	0.35	0.56	0.88	0.79	0.97	0.97
Al ₂ O ₃	34.51	34.09	33.76	35.09	35.39	36.35	36.35
SiO ₂	46.23	46.21	46.18	46.24	46.93	46.43	46.43
MgO	0.77	0.79	0.84	0.74	0.72	0.46	0.46
K ₂ O	11.10	10.91	11.08	10.05	10.34	10.22	10.22
TiO ₂	0.84	0.63	0.59	0.59	0.36	0.45	0.45
OH	4.49	4.45	4.44	4.49	4.53	4.55	4.55
TTL	100.07	99.00	99.02	99.27	100.32	100.80	100.80

Cations per 24 oxygens

Fe	0.18	0.17	0.18	0.13	0.14	0.15	0.15
Na	0.14	0.09	0.15	0.23	0.20	0.22	0.22
Al	5.43	5.40	5.37	5.52	5.51	5.64	5.64
Al ^{iv}	3.60	3.60	3.59	3.70	3.71	3.75	3.75
Si	6.17	6.22	6.23	6.18	6.20	6.11	6.11
Mg	0.15	0.16	0.17	0.15	0.14	0.09	0.09
K	1.89	1.87	1.91	1.77	1.74	1.71	1.71
Ti	0.08	0.06	0.06	0.06	0.04	0.05	0.05
OH	2.00	2.00	2.00	2.00	2.00	2.00	2.00
TTL	16.05	15.99	16.05	15.97	15.97	15.96	15.96

Table 3.4. Representative plagioclase compositions in pelites of the northern Okanagan complex (cc: core of crystal; tr: rim of crystal)

	SPLE	OFcc	OFtr	5A _{cc}	5A _{tr}	5B _{cc}	5B _{tr}	7A _{cc}	7A _{tr}	47A _{cc}	47B _{cc}	47B _{tr}
FeO	0.37	---	---	0.29	---	0.04	---	---	---	0.17	0.1	0.16
Na ₂ O	8.22	8.15	6.78	6.61	6.70	6.31	8.30	8.10	7.63	7.65	7.56	7.91
Al ₂ O ₃	24.02	24.28	25.65	26.20	25.62	26.68	23.20	23.79	24.87	24.39	24.87	24.63
SiO ₂	62.16	61.25	57.89	58.01	58.57	57.14	61.78	61.53	60.51	60.39	60.83	60.47
K ₂ O	0.22	0.28	0.33	0.19	0.39	0.24	0.48	0.32	0.30	0.20	0.02	0.26
CaO	5.64	5.62	5.03	8.63	7.78	8.89	5.16	5.95	6.46	6.48	6.60	6.22
TTL	100.71	99.65	98.77	99.95	99.15	99.77	99.96	99.76	99.82	99.29	100.31	99.67

Cations per 8 oxygens

Fe	0.01	---	---	0.01	---	0.01	---	---	0.02	0.01	---	---
Na	0.70	0.70	0.60	0.57	0.55	0.55	0.72	0.70	0.66	0.66	0.65	0.68
Si	2.74	2.73	2.62	2.60	2.64	2.57	2.77	2.74	2.70	2.70	2.70	2.70
Al	1.25	1.27	1.37	1.38	1.36	1.42	1.23	1.25	1.31	1.29	1.30	1.29
K	0.01	0.02	0.02	0.01	0.02	0.01	0.03	0.02	0.02	0.01	0.02	0.01
Ca	0.27	0.27	0.41	0.41	0.38	0.42	0.25	0.28	0.31	0.31	0.31	0.30
TTL	4.95	4.95	5.00	4.99	4.98	5.00	4.96	4.96	4.99	4.99	4.99	5.00

Na	71.61	71.20	59.30	57.40	59.52	55.47	72.39	69.84	69.95	67.35	66.26	65.69
Na ^{iv}	1.50	1.60	1.90	1.10	2.29	1.37	2.76	1.80	1.73	1.14	1.79	1.46
Na ^{vi}	27.20	27.20	35.80	41.50	38.10	43.16	24.85	28.36	31.32	31.51	31.95	29.85

Table 3.6. Representative staurolite compositions in pelites of the northern Okanagan complex.

SPL	45-1	45-2	E4-2	E4-3
FeO	13.17	12.74	13.25	13.75
MnO	0.11	0.15	0.09	0.11
Al ₂ O ₃	53.98	53.65	54.37	54.16
SiO ₂	27.99	28.20	26.95	27.11
MgO	1.44	1.53	1.76	1.75
K ₂ O	0.02	0.01	0.04	---
TiO ₂	0.52	0.38	1.01	1.08
OH	2.16	2.15	2.15	2.16
TTL	99.39	98.83	99.72	100.15

Cations per 24 oxygens

Fe	1.53	1.49	1.54	1.60
Mn	0.01	0.02	0.01	0.01
Al	8.84	8.82	8.90	8.85
Si	3.89	3.93	3.74	3.80
Mg	0.30	0.32	0.36	0.36
Ti	0.05	0.04	0.10	0.11
OH	1.00	1.00	1.00	1.00
TTL	15.63	15.62	15.70	15.70

X _{Fe}	0.831	0.816	0.810	0.817
X _{Mn}	0.167	0.167	0.157	0.180
a _{Fe}	0.477	0.443	0.431	0.445

Table 3.7. Representative limonite-rutile compositions in pelites of the northern Okanagan complex.

SPL	OF1	OF3	45-1	45-2	E4-3	E4-2
FeO	33.58	31.95	46.22	45.38	43.96	43.96
TiO ₂	51.55	47.39	51.41	50.95	50.67	50.67
MnO	12.09	10.12	0.45	0.43	1.96	1.96
SiO ₂	0.07	3.83	---	0.07	0.45	0.45
Al ₂ O ₃	0.05	2.56	---	---	0.39	0.39
MgO	0.01	1.14	---	---	0.08	0.08
TTL	97.35	97.03	98.08	97.77	97.52	97.52

Cations per 6 oxygens

Fe	1.35	1.33	1.98	2.00	1.89	1.89
Ti	2.00	1.77	1.99	1.98	1.96	1.96
Mn	0.53	0.43	0.02	0.01	0.08	0.08
Si	---	0.19	---	0.01	0.02	0.02
Al	---	0.15	---	---	0.02	0.02
Mg	---	0.08	---	---	---	---
TTL	3.99	3.96	4.00	4.00	3.99	3.99

X _{Fe}	99.97	95.42	100.00	100.00	99.68	99.68
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variations in bulk compositions (Appendix 2). There are no plagioclase inclusions in the garnets. Calcium zoning in garnets sharing a contact with matrix plagioclase has not been observed (Fig. 3.3). Perhaps a certain amount of Na has diffused out of the plagioclase leading to a relative increase of Ca in the plagioclase. Because this phenomenon occurs only near garnet contacts it can be alternatively proposed that these variations may reflect local chemical equilibrium.

Figure 3.3 illustrates a south to north distribution of the average compositional variations of Fe, Mg, Mn, Ca, Na and K in weight percent losses or gains of rim compositions relative to core compositions of garnets, biotite, and plagioclase sharing mutual contacts. This diagram summarizes the chemical data for each sample analysed (Appendix 2). It includes duplicates of samples 5 and 47 which show small coefficients of variation except maybe for Ca in plagioclase and Mn in garnet of sample 47B. Although many element variations from core to rim are small (within detection limits) most phases appear to remain stoichiometric. For example, Fe variations between cores and rims of biotites throughout the Okanagan complex seem to be balanced by equivalent losses or gains of Mg. Similarly gains or losses of Ca in most plagioclase are nearly compensated by gains or losses of Na.

There are only few examples of an equilibrated cationic exchange between minerals of each sample analysed. For example, Fe is systematically lost in biotites and garnets of sample OF1 whereas Fe and Mg are gained in both biotites and garnets of samples 47 and 45. Garnets and biotites in sample E4 appear to be homogeneous. Altogether this data suggest that cationic exchanges have not been equilibrated systematically at the end of crystallisation indicating that many minerals remained stoichiometric but most assemblages are non-equilibrated. Compositional variations are considered significant but inconsistent because they occur in most samples and were duplicated and triplicated amongst each sample. These behaviours are probably caused by conditions of re-equilibration which complicates the application of thermobarometry. Nevertheless the chemical data may be rationalized by suggesting that Fe and Mn are generally enriched and Mg is slightly depleted at garnet rims and added proportionally in biotites. Ca in garnets does not vary much though it is clearly gained along most plagioclase rims which remained stoichiometric.

Figure 3.2: (a) *Mg/Mg+Fe ratios of garnets and biotites in metapelites of the Okanagan complex. Solid dots and solid lines illustrate core and matrix compositions whereas circles and hatched lines indicate rim-rim pairing compositions. Garnet core and biotite matrix composition reflect bulk composition variations amongst the samples tested. Rim compositions indicate that the majority of the assemblages have exchanged either Mg and/or Fe. Garnets in samples 47 and E4 appear to be completely homogenized. Biotites sharing contacts with garnets tend to cluster to average Mg/Mg+Fe ratios of .30 and .35 suggesting the system has tentatively tried to homogenize in a bimodal fashion during decompression. (b) Almandine chemical zoning profiles along six core-to-rim traverses show a progressive almandine enrichment from the southern Okanagan pelites to the northern Okanagan pelites consistent with bulk compositional variations. Except for sample OF1 and in the local vicinity of biotite inclusions all garnets exhibit cores that are homogeneous and zoned rims, attributed to early high-grade homogenization overprinted by retrograde continuous reactions with matrix minerals biotite and/or muscovite inclusions. FeO is in weight percent.*

Plate 3.3:

Compositional profiles across garnets in the Okanagan complex as located in Figure 3.1. (a) Center left to right transect in an oblong anisotropic garnet of sample OF1 containing numerous large inclusions of biotite and muscovite and exhibits two stages of almandine depletion from core to rim indicating prograde crystallization conditions (Tracy 1982). If a garnet grows under extensional conditions its core will be modified subsequent to peak conditions and continuous zoning with increasing almandine contents would be expected. The rim of sample OF1 starts off 1 mm inside the grain from a low almandine composition suggesting it recrystallized under retrograde (?extensional) conditions and was progressively reheated to be enriched in almandine contents near the end of its crystallisation. Heating for the second-stage growth may have been provided by in situ melts such as those interleaved within the Okanagan complex. Preserving such a profile requires very rapid exhumation of the hosting lithologies consistent with the presence of a major extensional fault. Rim temperatures obtained with the PTX program (see specified conditions in text) are 575°C and may be considered as syn-extensional conditions. (b) Sample 5 is an annealed metapelite containing garnets with a homogeneous core and slightly zoned rims typical of high-grade garnets (Tracy 1982). The center left to right transect shows that all components vary only at the rim of the ovoid garnet, but that it is mainly pyrope and spessartine which vary the most. Almandine and pyrope vary adjacent to biotite inclusions and indicate retrograde conditions. (c) Top left to bottom right transect; although they have been severely strained pulled apart garnets in sample 45 (see also Plate 3.1d) exhibit homogeneous compositions with slight variations in almandine and pyrope adjacent to matrix mineral typical of retrograde conditions. Samples 5 and 45 exhibit modified primary compositions typical of high-temperature homogenization conditions. The residual growth profiles indicate prograde conditions due to diffusional processes at temperatures reaching at least 600-650°C.

OF1

5

45

Figure 3.3: *South to north distribution of element variations (Fe, Mg, K, Mn, Ca, Na in weight percent) between cores and rims of biotite, garnet, and plagioclase crystals in each metapelite analysed for geothermobarometry. Plagioclase is absent in samples 45 and E4. Element gains relative to core compositions are indicated by hatched pattern whereas losses are blackened. Note that most phases appear to remain stoichiometric. Gains of one element in one phase is balanced by losses in one or more phases; for exemple losses or gains of Mg in biotites of all samples is generally balanced by equivalent gains or losses of Mg in garnets in corresponding samples. However, there is a net gain of Fe in rims of biotites and garnets lying in the northernmost Okanagan region. Mn in garnets has severely fluctuated in many samples which indicates that the exchange equilibria with coexistent biotites is probably not equilibrated.*

7. Compositional trends

Grossular zoning in garnet is mainly pressure-dependent when developed in the presence of aluminosilicates+plagioclase+quartz; almandine-pyrope zoning is essentially temperature-dependent. Compositional trends have been evaluated from all garnets analysed to verify the homogeneity of the Ca-Mg-Fe-Mn exchange processes (Martignole and Nantel 1982; Berman 1990) and evaluate the range of bulk compositions before testing various thermometric calibrations.

Two main types of garnet compositional trends are observed (Fig. 3.4): samples OF1, 5, 7, and E4 indicate conditions of *cooling-unloading* (trend 2); sample 45, from Shuswap Falls, exhibits an *isobaric-heating* trend (trend 1 reversed). The latter sample, a staurolite-garnet schist, probably recorded metamorphic conditions of an older prograde metamorphic event that has been isolated from the more common unloading conditions recorded by the other samples. Local metamorphic perturbations may be partly supported by the presence of transitional conditions affecting garnets of sample 47 located southwest of sample 45 (Fig. 3.1). Assuming garnets were formed under synchronous thermal conditions, these data together with textural evidences suggest that certain parts of the Okanagan complex have been kept hot while others were decompressing and cooling. In the context of superposed compressional events followed by rapid exhumation these rocks need not to have all achieved similar P-T conditions.

Porphyroblast assemblages of staurolite-garnet-rutile and sillimanite-garnet-rutile of samples 45 and E4 are progressively wrapped by biotite-muscovite folia rich in fibrolite and ilmenite. The implications of these reaction textures in the TiKFMASH system, assuming water saturation and $X_{Mn}^{Al} > X_{Mn}^{Fe} > X_{Mn}^{Ca}$ and $X_{Fe}^{Al} > X_{Fe}^{Ca} > X_{Fe}^{Mn}$, constrains the P-T path of these samples to a trajectory of heating and decompression (Buick and Holland 1989). This prograde reaction is consistent with growth profiles and compositional trends mentioned previously (Fig. 3.3). Knowing that these rocks are affected by extensional fabrics this indicates that they reached their peak conditions during or just before the extensional event.

Figure 3.4: *Compositional trends in metapelites of the Okanagan complex. Large symbols are garnet core data whereas dots are rim data. Excepting sample 47 the other data sets suggest that the gneisses of the Okanagan complex have undergone a major cooling-unloading history. Lower left inset illustrates compositional trends and associated P-T conditions from Martignole and Nantel (1982). Trend 1 ==> isobaric - cooling; Trend 2 ==> cooling - unloading; Trend 3 ==> isothermal - unloading. The number of domains tested per site are given in parenthesis.*

8. *P-T estimates*

In principle, thermobarometry is simplistic when one considers the lack of experimental data at appropriate compositions and P, T, X conditions of interest and a poor understanding of non-ideal mixing processes. In practice, it is as complicated to determine which measured mineral compositions should be used for equilibrium calculations especially when dealing with sheared rocks which exhibit prograde and retrograde reactions.

We make the very general assumption that all garnets tested are of the same generation and may be regionally referred to as M2 porphyroblasts, that is they are expected to have formed during the last regional compressive event affecting the southern Omineca Belt (Carr 1990). This assumption is partly constrained by the paragenetic sequences observed in thin section and from field observations. All garnets are single porphyroblastic phases. Abundant biotite inclusions within garnet and in the matrix suggest biotite was present throughout the history of the rocks. Garnets may have been present at this time, or formed shortly thereafter, as there are sillimanite and staurolite pseudomorphs and inclusions in several garnets. Muscovite is present as inclusions only in sample OF1 and was probably present in the early assemblage together with staurolite. Though they are affected by extensional shear fabrics and homogenized, in the southern Okanagan complex, fabrics such as aligned inclusions perpendicular to extensional shear planes and sillimanite folded by minor folds (Plate 3.2) suggest that these assemblages formed prior to extension in a compressional environment. These metamorphic histories are consistent with those established by Christie (1973) and Ryan (1973) in the southern Okanagan complex. However, the age of the pre-extensional metamorphic event is not demonstrable because no mineral crystallization dates have been obtained directly from these rocks. The only indirect evidence that exists in this region is provided by a 56 Ma concordant zircon fraction obtained from a migmatitic granite north of sample 45. This migmatite is similar in composition to several others found throughout the Okanagan complex.

8.1 Procedures

Applying thermobarometry to the Okanagan pelites requires some discussion of mineral compositional variability and zoning. The homogenized cores of the garnets suggest these were

formed under high thermal metamorphism (Tracy 1982). Compositionally zoned rims, which occur in most samples, indicate continuous reaction with matrix minerals due to retrograde Fe-Mg exchange. The garnet rim composition may therefore be used with the mineral sharing a mutual contact to infer a point on the cooling path. The composition of monophase biotite inclusions and adjacent garnet are used to infer cooling temperatures. Choosing a biotite composition representative of the prograde metamorphic conditions is hazardous. High-grade biotites were not convincingly found except for a few Ti-rich inclusions. Nevertheless, because biotites are much more abundant than garnets it can be expected that they did not homogenize systematically with garnets and therefore are likely to preserve "peak" compositions. Because there are no plagioclase inclusions in garnet only the rim of matrix plagioclase (Table 3.4) and garnet rim compositions are used to infer a pressure along the retrograde P-T path.

The chemical data presented in Tables 3.2 to 3.7 and Appendix 2 was used to estimate temperatures using experimental and empirical calibrations of the Fe-Mg exchange between garnet and biotite phases (Ferry and Spear 1978). The data for all samples were tested with ideal-mixing models (Thompson 1976; Ferry and Spear 1978; Chippera and Perkins 1988). Non-ideal mixing models to account for the effects of Ca, Mn, Ti and Al³⁺ components (Hodges and Spear 1982; Perchuk and Lavrent'eva 1983; Dymek 1983; Indares and Martignole 1985) were also tested because Ca and Mn in garnet and Al³⁺ in biotite reduces the calculated temperatures while Ti causes the opposite effect. Most garnets show Ca and Mn variations that may affect the thermometric estimates (Table 3.3; Fig. 3.3). Their Fe/(Fe+Mg) ratio of 0.7 is nearly comparable with that used by Indares and Martignole (1985) which provides the most consistent temperature data (Table 3.8). Pressures were evaluated in those samples using the discontinuous reaction plagioclase = garnet + sillimanite + quartz (GASP; Ghent 1976) experimentally calibrated by Newton and Haselton (1981) and revised by Koziol and Newton (1988) using the Goldsmith (1980) plagioclase stability curve.

P-T grids were also generated in the TiKFMASH system from the bulk composition sillimanite-muscovite-garnet-ilmenite-rutile $\pm$ staurolite + H₂O pelites using a 1990 version of the PTX program from the GEO-CALC software written by Berman et al. (1987) following thermodynamic data sets derived and compiled by Berman (1988; it may now be obsolete). The

PTX program calculated all possible equilibrium reactions implied by a given assemblage of products and reactants from measured mineral compositions. It locates stable and metastable equilibria by calculation rather than by experimental plots. The crossing of multiple reaction curves (metastable) at a single intercept for a given assemblage on the diagrams produced by PTX defines a specific P-T value. Curves falling off a single intersection or widespread in P-T space are regarded as being in disequilibrium (Fig. 3.5). Stable and metastable reaction curves created with PTX used the ideal mixing for biotite, Berman's (1990) mixing properties for quaternary Ca-Mg-Fe-Mn garnet, and Furhnam and Lindsley's (1988) activity model for anorthite. As expected few of the reactions tested were in total chemical equilibrium and it appears that K-feldspar is always unstable not surprisingly for upper amphibolite-facies pelites (Essene 1989). The pressures obtained with the GASP were compared with PTX and compared with pressure conditions in those samples lacking plagioclase using garnet + rutile + Al₂SiO₅ + ilmenite + quartz (GRAIL; Bohlen et al. 1983) and garnet + biotite ± staurolite ± muscovite reactions.

8.2 Results

The average data obtained for each calibration and with PTX are summarized in Table 3.8 with the corresponding compositions used in the calculations. Maximum temperature and pressure of metamorphism estimated with the different calibrations for the six samples tested range between 690-825°C and 5-8.4 kbar (500-840MPa). These data may reflect "peak" metamorphic conditions reached at the end of the compressional event and are consistent with the presence of anatectic melts and homogeneous garnet profiles typically assigned to temperatures above 650°C (Yardley 1977; Anderson and Olympio 1977; Woodsworth 1977; Tracy 1982; Selverstone and Chamberlain 1990). This suggests that rocks of the Okanagan complex spent a sufficiently long time at elevated temperatures to allow at least partial homogenization. Retrograde temperatures and pressure range between 550-690°C and 4.5 kbar (450MPa) in most samples.

Three of the best results obtained with PTX are shown in Figure 3.5. Precise P-T intersections were obtained with different mineral pairing modes, in several cases the garnet-core/biotite-core pair (5cc in Fig. 3.5) is at equilibrium whereas in other cases the garnet-rim/biotite-rim pairing produces a plausible P-T intersection (47rr and 45rr in Fig. 3.5). Although

Table 3.8

Compositional parameters of coexisting biotite-garnet and plagioclase with temperatures calculated using four garnet-biotite equilibria (Thompson 1976; Hodges and Spear 1982; Perchuck and Lavrent'eva 1983; Indares and Martignole 1985) and pressures (kbar) calculated from the anorthite-grossular-Al₂SiO₅-quartz equilibrium (Newton and Haselton 1981). Also included are temperature and pressure estimated with the PTX program (Berman 1989). Mineral pairing modes are; CC: garnet core with matrix biotite core; RR: garnet rim with matrix biotite rim; i: biotite inclusion and adjacent garnet.

Sample	Biotite				Garnet				Muscovite				Plagioclase				Grt-Bt				F	
	AlFe	Mg	Al	Al ^{IV}	AlFe	Mg	Al ^{IV}	Al ^{VI}	Al ^{IV}	Al ^{VI}	Al ^{IV}	Al ^{VI}	Al ^{IV}	Al ^{VI}	Al ^{IV}	Al ^{VI}	Al ^{IV}	Al ^{VI}				
51cc	0.485	0.391	0.983	0.055	0.141	0.704	0.132	0.120	0.044	0.070	0.897	0.931	0.716	0.013	0.270	803	753	688	6.4	490/30	5.5/0.5	
51rr	0.442	0.431	0.983	0.087	0.157	0.600	0.097	0.249	0.032	0.050	0.904	0.954	0.712	0.016	0.272	635	633	599	5.4	575/25	4.5/0.5	
51i	0.462	0.412	0.985	0.032	0.150	0.617	0.087	0.236	0.050	0.071	0.899	0.927	0.714	0.014	0.271	638	636	601	5.2	575/25	4.5/0.5	
51cc	0.403	0.494	0.976	0.056	0.142	0.669	0.175	0.098	0.058	-----	-----	-----	0.593	0.019	0.388	775	713	673	6.5	635/35	5.0/0.5	
51rr	0.401	0.502	0.970	0.046	0.142	0.655	0.149	0.137	0.058	-----	-----	-----	0.574	0.011	0.415	684	663	623	4.6	625/15	6.0/0.5	
51i	0.355	0.552	0.953	0.084	0.118	0.693	0.145	0.104	0.055	-----	-----	-----	-----	-----	-----	660	560	642	610	4.3	550/05	4.5/0.5
53cc	0.405	0.391	0.960	0.046	0.151	0.644	0.173	0.126	0.037	-----	-----	-----	0.595	0.023	0.382	792	755	746	684	6.2	780/10	5.3/0.3
53rr	0.404	0.390	0.969	0.046	0.145	0.668	0.149	0.127	0.056	-----	-----	-----	0.555	0.014	0.432	696	656	661	623	4.0	710/25	5.5/0.3
53i	0.401	0.409	0.991	0.073	0.127	0.673	0.145	0.126	0.028	-----	-----	-----	-----	-----	-----	637	570	631	603	4.1	660/05	5.0/0.2
71cc	0.424	0.367	0.970	0.061	0.143	0.590	0.163	0.133	0.032	-----	-----	-----	0.724	0.028	0.249	799	740	754	687	5.7	750/50	6.3/0.2
71rr	0.401	0.381	0.960	0.058	0.157	0.690	0.135	0.143	0.032	-----	-----	-----	0.698	0.018	0.284	779	702	737	678	6.4	660/40	4.5/0.5
471cc	0.388	0.422	0.953	0.056	0.132	0.659	0.205	0.080	0.052	-----	-----	-----	0.670	0.017	0.313	812	768	759	692	6.8	740/50	7.5/0.2
471rr	0.416	0.399	0.980	0.059	0.123	0.676	0.211	0.059	0.055	-----	-----	-----	0.674	0.011	0.315	-----	-----	-----	-----	-----	840/20	7.2/0.3
471i	0.340	0.478	0.959	0.064	0.117	0.656	0.208	0.080	0.044	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	700/20	4.8/0.3
471cc	0.388	0.441	0.970	0.060	0.110	0.666	0.194	0.092	0.047	-----	-----	-----	0.663	0.018	0.320	-----	-----	-----	-----	-----	725/10	6.5/0.5
471rr	0.404	0.418	0.970	0.055	0.120	0.715	0.189	0.057	0.040	-----	-----	-----	0.687	0.015	0.299	-----	-----	-----	-----	-----	750/05	6.5/0.2
471i	0.378	0.448	0.949	0.068	0.105	0.676	0.221	0.061	0.038	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	780/05	6.0/0.1
457cc	0.411	0.367	0.947	0.038	0.183	0.780	0.103	0.046	0.054	0.118	0.918	0.881	-----	-----	-----	-----	-----	-----	-----	-----	560/40	5.6/0.8
457rr	0.457	0.360	0.959	0.023	0.158	0.800	0.118	0.021	0.058	0.102	0.922	0.897	-----	-----	-----	-----	-----	-----	-----	-----	665/03	6.5/0.1
842cc	0.459	0.316	0.954	0.078	0.128	0.782	0.159	0.027	0.032	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	825/05	8.4/0.2
842rr	0.458	0.347	0.947	0.040	0.156	0.783	0.160	0.025	0.035	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	780/05	8.4/0.2
842i	0.442	0.348	0.947	0.075	0.133	0.810	0.132	0.036	0.032	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	-----	700/05	7.5/0.2

Figure 3.5: *GEO-CALC P-T diagrams (plotted with both stable and metastable reactions; Berman et al. 1987) showing three examples of assemblages in equilibria with reliable intersections. Sample 5 from the Kelowna region shows cores of garnet, biotite, and plagioclase. Sample 47rr, from the Vernon area, shows garnet, biotite, plagioclase, rutile, ilmenite reaction curves for these grains using rims of grains sharing mutual contacts. Sample 45rr, from the Shuswap Falls area (north of Vernon), shows P-T conditions using rims of garnet-biotite-staurolite-plagioclase sharing mutual contacts. Reaction numbers 1) $MU + GR + ALM = ANN + 3 AN$; 2) $3 SI + 3 RT + 2 GR + ALM = 6 AN + 3 ILM$; 3) $6 RT + GR + 2 SI = 3 AN + 6 ILM + 3 aQZ$; 4) $PY + MU + GR = 3 AN + PHL$; 5) $2 SI + aQZ = 3 AN$; 6) $3 SI + 3 RT + 2 GR + ALM = 6 AN + 3 ILM$; 7) $25 aQZ + 6 ST = 12 W + 46 SI + 8 ALM$; 8) $17 ALM + 25 MU + 6 ST = 12 W + 96 SI + 25 ANN$; 9) $8 MU + 17 QZ + 6 ST = 12 W + 62 SI + 8 ANN$; 10) $8 PHL + 25 aQZ + 6 ST = 12 W + 46 SI + 8 PY + 8 ANN$; 11) $12 W + 31 PY + 23 MU + 8 ANN = 31 PHL + 48 aQZ + 6 ST$; 12) $12 W + 23 PY + 8 ALM = 23 PHL + 48 aQZ + 6 ST$. Activity models used are: BI ideal; GA (Berman 1990); PLAG (Furhman and Lindsley 1988); $ST a_{Si}^{ST} = X_{Si}^4$*

the assemblages compared are assumed to be in equilibrium the data indicate that core-core pairing yield lower temperatures and pressures of 635°C/4.5 kbar (450MPa) than rim-rim pairing modes 675-750°C/6 kbar (600MPa) of samples 45 and 47. The latter confirms the prograde conditions outlined by compositional trends but it clearly illustrates the difficulty in interpreting the proper pairing mode to distinguish prograde from retrograde conditions in rocks that have been thoroughly homogenized. Alternatively these results indicate that chemical equilibrium has been modified more severely in the southern Okanagan complex.

9. *Validity of the P-T data and P-T-t evolution*

As pointed out above, P-T paths determined between and within sites present an incoherent set of thermobarometric conditions which are not always compatible with mineral parageneses. For example, temperatures exceeding 750°C are clearly overestimated because none of the rocks examined exhibit granulite-facies assemblages. This again demonstrates the difficulty of pairing phases re-equilibrated while undergoing high strain or the non preservation of equilibrium compositions. Prograde pressure estimates range from 5.2 to 8.4 kbar (520 - 840 MPa) but most equilibria give pressures of 5.5 kbar (550 MPa) comparing garnet-core with matrix biotite-core compositions. Retrograde pressure estimates vary from 4.1 to 7.5 kbar (410 - 750 MPa) with a mean value of 4.8 kbar (480 MPa) comparing garnet-rim and biotite-rim compositions. Corrections for compressibility and expansion for GASP barometry may shift the calculated equilibrium by as much as 1kbar (100 MPa) when temperatures exceed 500°C (Essene 1989) thus the peak pressure attained by the pelites may have averaged 6.5 kbar (650 MPa) while retrograde pressures reached perhaps 5 kbar (500 MPa).

With the exception of sample 45, all temperature and pressure estimates are consistent with the mineral assemblages above the muscovite-out isograd and consequently may represent prograde "peak" temperatures of 750°C. Retrograde temperatures of 550°C on most rim and inclusion assemblages indicate that pelites were rapidly cooled afterwards and crystallized chlorite. Pressure estimates of footwall rocks are within analytical uncertainties but temperature ranges are extremely variable and only those data obtained from assemblages preserving chemical zoning are regarded as reliable. The temperatures of 750°C obtained with the Indares and

Martignole calibration (1985) brings the pelites in the field of *in situ* partial melting if X_{H_2O} was sufficiently high. This is consistent with field observations whereby leucosomes and granitoid bodies are systematically interleaved with sillimanite-bearing pelites.

Certain variations in P-T values may be caused by disequilibria between the Fe-Mg solid-solutions and deficient Ca exchanges between garnet and plagioclase. This phenomenon is most certainly due to homogenization and differential re-equilibration upon cooling and exhumation of the strained pelites. Nevertheless, it is proposed that the pelites contain mineral phases which were quenched at different stages of the extensional P-T path.

It is proposed that the matrix biotite-garnet core pairing mode provides minimum pressure estimates for the early segment of the P-T-t trajectory at the end of the compression period established regionally between 70 Ma and 60 Ma (Carr 1990; Scammell 1991; point a on Fig. 3.6). Anatectic melts in the Okanagan complex were cooling 56 Ma ago probably near the time of "peak" thermal conditions (*see* Chapter 4; point b on Fig. 3.6). It is proposed that the rim-rim pairing mode in the pelites provides minimum pressure and temperature estimates for the onset of extension on the P-T-t trajectory circa 52 Ma (*see* Chapter 4; point c on Fig. 3.6). Because no andalusite or kyanite are present in the retrograde assemblages it appears that rocks of the Okanagan complex were exhumed too rapidly preventing the crystallization of such phases. The extensional P-T trajectory is partly constrained at lower P-T by the cooling of retrograde minerals in the cataclastic zone of the Okanagan valley fault (point d on Fig. 3.6; *see* Chapter 4) and the cooling of granitoids interdigitated in the Okanagan complex (point e; Medford 1975).

10. *P-T conditions of the Upper Crustal layer*

Peak assemblages in the Upper Crustal layer within the map area never exceeded lower amphibolite grade. At the southwestern end of the Okanagan valley, Okulitch (1969) estimated crystallization conditions with the calcite-dolomite geothermobarometer and suggested that upper crustal rocks reached peak conditions of the order of 450°C and 1 to 3 kbar (100 to 300 MPa). This data is consistent with P-T values estimated at 550°C/3kbar (Solberg 1976) in the northern Okanagan region west of Vernon.

Figure 3.6: *The P-T-t history for sillimanite-bearing pelites from the Okanagan complex lying in the hanging wall of the Monashee décollement (Carr 1990). Thrusting on the Monashee décollement stopped 59 Ma ago (Coleman 1990; Carr 1990) but extensional features started forming soon after peak metamorphic conditions between 70 and 60 Ma ago (point a extrapolated from Scammell 1991) in the hanging wall of the Monashee décollement. Several studies demonstrate that crustal extension formed as a continuum from compressional conditions throughout the southern Omineca Belt (Carr 1990, 1992; Scammell 1991). These extensional features appear to have been activated after the culmination of diachronous M2 compressional metamorphic conditions (somewhere in the high P and T field). Geochronological data from rocks of the Middle Crustal layer; a: U-Pb monazite and xenotime from late-compressional leucogranites in the northern part of the Shuswap metamorphic complex (Scammell 1991), b: U-Pb zircon in syn-extensional Ladybird-type leucogranite at Mabel Lake (sample E19 Chapter 4) emplaced in undated sillimanite-grade migmatitic schists similar to those of sample E4 from which P-T data range around 750°C and 650 MPa (Table 3.8), c: $^{40}\text{Ar}/^{39}\text{Ar}$ biotite cooling age in amphibolites of the Kelowna area (Chapter 4), d: Apatite fission-track cooling dates in amphibolites of the Kelowna area (Medford 1975). Estimations of peak-M2 metamorphic conditions affecting metapelites of the Okanagan valley samples are shown with uncertainty boxes centered at 725°C/620 MPa (core-core data) and 650°C/500 MPa (rim-rim data) respectively.*

P-T conditions across the Okanagan valley fault are difficult to compare because they are temporally distinct. Hanging wall metamorphism is probably Jurassic whereas footwall metamorphism is probably Late Cretaceous-Early Tertiary. However, it seems inevitable that rocks in the hanging wall were kept under conditions of greenschist grade or less during the extensional event in the Early Eocene. It is proposed that the post-extensional metamorphic break across the Okanagan valley fault was of the order of 300°C and 3-4 kbar (300-400 MPa). This confirms that the cataclastic zone straddling the Middle Crustal layer and the Upper Crustal layer along the Okanagan valley is a fundamental crustal dislocation that may have generated a minimal vertical offset of at least 10 km assuming normal geothermal gradients.

11. *Conclusions*

The results of this study indicate that P-T paths within high strain zones can be documented but would require considerable detailed microprobe analysis of metamorphic assemblages within and outside the high strain zone. Although care has been taken to analyse mineral phases in detail it cannot be determined with certainty "peak" metamorphic conditions that affected the pelitic rocks of the Okanagan complex. This is mainly because it has not been possible to recover biotite phases that have convincingly avoided being re-equilibrated during syn-kinematic retrograde processes. Nevertheless, most P-T data obtained by combining garnet core compositions with certain biotite compositions are generally in the order of 700°C and 6 kbar (600 MPa) which is consistent with: the homogeneous profiles of garnets; most mineral assemblages and textures observed; and the presence of abundant anatectic melts throughout the Okanagan complex mainly in zones exceeding the muscovite-out isograd. Sillimanite + muscovite and staurolite + garnet + muscovite assemblages contain only thin quartzo-feldspathic selvages and appear to be truncated by the Okanagan valley fault. Rocks containing sillimanite + muscovite and staurolite-bearing assemblages provided T-P data in the order of 600°C/5 kbar (500 MPa). As expected, retrograde P-T data and garnet compositional trends obtained throughout the northern Okanagan complex indicate that high-grade rocks underwent mainly cooling-unloading conditions at a minimum T-P of 550°C/4.5 kbar (450 MPa).

Based on regional assumptions the timing of "peak" metamorphic conditions in the

Okanagan complex is associated with the final stages of Late Cretaceous-Early Tertiary D2 compression documented in the southern Omineca Belt suggesting that melting occurred in these rocks during the compressional event. Retrograde data are related to extension and might be used to support the idea of rapid exhumation of the Okanagan complex in the Eocene as most assemblages quenched nearly at the same P-T of 500°C/4 kbar (400 MPa). Integrating temporal constraints it may be proposed that "peak" temperature and pressure conditions in the Okanagan complex indicate that at the transition from compressional to extensional regimes rocks were undergoing prograde decompression locally but they underwent retrograde decompression rapidly from 750°C to 550°C and from 6.5 to 4.5 kbar (650 to 450 MPa) between 56 and 50 Ma.

Because they overprint the metamorphic assemblages of the Okanagan complex it is suggested that extensional fabrics outlasted "peak" metamorphic conditions and led to the rapid exhumation of these rocks quenching cationic diffusion and precluding the crystallization of aluminosilicates such as andalusite and kyanite. Perhaps the preservation of kyanite±andalusite on the west flank of Thor-Odin (Carr 1990) indicates that exhumation of the Okanagan complex was more rapid than the Middle Crustal layer at Thor-Odin.

Field evidence shows that rocks mantling the Okanagan complex were rendered permeable by cataclasis and affected by abundant fluids percolating at temperatures < 350°C creating greenschist-grade assemblages confined to a narrow tectonic zone straddling the trace of the Okanagan valley fault. Retrograde conditions affecting rocks underlying the cataclastic zone may be related to intracrystalline cationic diffusion.

1. *Introduction*

This investigation complements a significant background of geochronometric data acquired from the Okanagan region over 20 years. Detailed accounts of these data can be found in Roddick et al. (1972), Ross (1974), Medford (1975), Mathews (1981), Parkinson (1985), Parrish et al. (1988), and Armstrong et al. (1991). Most of the earlier workers had observed that the Okanagan valley was the locus of a sharp break in K-Ar and apatite fission-track ages and proposed that rocks of the Okanagan complex had been subjected to a major thermal resetting event in the Eocene (Ross 1974; Medford 1975, 1978; Mathews 1981). Recent workers, however, suggested that Eocene ages reflect the cooling related to the exhumation of footwall rocks underneath the Okanagan valley fault. Large displacement faulting had been challenged by Ross and Christie (1979) who felt that they had demonstrated continuity of lithology and structure across the Okanagan valley. This led Medford (1975) to suggest that large heating cells were confined to the Okanagan complex on the east side of the valley. This interpretation followed his demonstration that apatite fission track dates (retention temperatures below 75°C) jumped from 35 Ma in the Okanagan complex to 70 Ma west of Okanagan Lake (*see* Appendix 1, Table B).

A fundamental breakthrough in the understanding of the geochronometric evolution of the Okanagan complex was developed by Parkinson (1985) who redated by the U-Pb method a strained porphyritic sill which had previously released Eocene K-Ar dates (Ross 1974). This rock exhibits ductile extensional shear fabrics and contains several zircon fractions overlapping concordia at 51 ± 1 Ma suggesting that it was deformed at mid-crustal level in the Middle Eocene. Parkinson (1985) investigated many other leucogneisses and leucogranites in the Okanagan complex and found concordant zircon crystallization ages ranging between 200 and 95 Ma which indicates that not all granitoids in the Okanagan complex are Eocene (*see* Fig. 1 of Parrish et al. 1988) and/or that certain zircon fractions were inherited from older protoliths. Parkinson's work demonstrated that extensional shearing occurred in the Okanagan complex in the Eocene and confirmed the presence of a major denudation fault mantling the complex (Parrish et al. 1985) now documented by regional mapping (Bardoux 1985; Tempelman-Kluit 1988; Johnson 1988).

The geochronological database constraining the history of deformation and exhumation of the Okanagan valley fault required considerable refinement. In this perspective various mineral phases were separated from geological units in the hanging wall and footwall of the northern segment of the Okanagan valley fault (*see* Chapter 2). Samples were collected near Okanagan Falls, Kelowna, and Enderby (Fig. 4.1) and minerals were dated by U-Pb, K-Ar, and $^{40}\text{Ar}/^{39}\text{Ar}$ isotopic methods and agree with the regional geochronometric framework established by previous workers. The data constrain the temporal evolution of the Okanagan valley fault between 56 and 48 Ma (Fig. 4.1) and indicate that Ladybird type granitic bodies found in most metamorphic domes of the southern Omineca Belt (Carr 1990) also occur in the Okanagan complex.

2. *Geological setting of the Okanagan valley*

A north - south-trending belt of amphibolite-grade schists and paragneisses underlies the Okanagan complex in the footwall of the Okanagan valley fault (Fig. 4.1). These rocks exhibit polyphase compressional features and are cut by voluminous granitic bodies of possible anatectic origin. Granitic rocks are particularly affected by penetrative non-coaxial brittle-ductile extensional fabrics indicating top-to-the-west sense of shear and defining the Okanagan valley fault (*see* Chapter 2). Rocks on the west side of the Okanagan complex define the hanging wall of this shear system and consist of a mixed sequence of low-grade polydeformed Paleozoic miogeoclinal rocks cut by undeformed Mesozoic plutons. Both of these units are unconformably capped by calc-alkaline lavas and clastic sedimentary rocks of Eocene age that are locally affected by normal faults (*see* Chapter 6). The boundary between hanging wall and footwall rocks is marked by the presence of a thick chloritized breccia mantling the Okanagan complex and is herein referred to as the cataclastic zone of the Okanagan valley fault. Several dacite dikes intrude the Okanagan complex and the cataclastic zone east of Kelowna.

3. *Sample distribution*

Twenty-three samples were collected from metamorphic, plutonic and volcanic rocks in order to decipher a certain part of the deformation history of the northern segment of the Okanagan valley fault. The petrographic descriptions and coordinates of these samples are given

in Appendix 3. Figure 4.1 shows a compilation of the isotopic data from this study and includes some of the published K-Ar and apatite fission track dates obtained by Medford (1975).

Two granitic bodies transposed parallel to the gneissic layering of the Okanagan complex were sampled for U-Pb dating (Fig. 4.1; Map 1). Sample MX261 was collected near Hydraulic Lake east of Kelowna in a sheeted body of fine-grained biotite-microgranite approximately 10 m thick. This magma seems to have been injected concordantly between amphibolitic paragneisses and diopsidic marbles. Sample MX19E is another sheet-like body of biotite-muscovite granite collected near Mabel Lake east of Enderby at the northern end of the Okanagan valley (Fig. 4.1). Strain fabrics in MX19E and MX261 (*see* Plate 2.2f) indicate top-to-the west sense of shear. They are more intense (mylonitic) in MX19E which lies immediately underneath Eocene alluvial sedimentary rocks and trachyandesites of the Trinity Hills (Mathews 1981).

Six samples of mylonitic gneisses of the Okanagan complex and two samples of granitic bodies intruding these rocks together with three volcanic units from the hanging wall of the Okanagan valley fault were analysed by K-Ar methods. Seven samples of amphibolitic gneiss of the Okanagan complex were analysed by $^{40}\text{Ar}/^{39}\text{Ar}$ step heating and total fusion methods. Three of these samples MX229, MX49, MX48 were collected from west to east at equidistant intervals parallel to the pervasive extensional stretching direction and across half the thickness (500m) of the ductile strain zone developed in the footwall of the Okanagan valley fault (*see* Chapter 2). This sampling was planned to decipher differential cooling histories at different levels in the footwall. Sample MX300 is a strained hornblende diorite collected east of Skaha Lake near Penticton (Fig. 4.1). This unit lies at the same structural level as the Kelowna sample MX229 and was collected to test the lateral extension and the systematic variation of the cooling pattern in the footwall of the Okanagan valley fault. Four samples of brecciated mylonite MX03, MX10, MX21 and MX78 were collected to establish the cooling history of the cataclastic zone.

4. *Analytical techniques, procedures and results*

Mineral separates were obtained by standard crushing, grinding, Roger's table, heavy liquid, and Frantz magnetic separation techniques carried out on 20-40 kg rock samples.

Figure 4.1: *Geological map and sites of isotopic samples collected in the northern Okanagan valley. Circle symbols for U-Pb sites (samples MX19E, MX261), Zr: zircon; Mz: monazite; T: thorite. Open squares for $^{40}\text{Ar}/^{39}\text{Ar}$ dates (samples MX78, MX48), H: hornblende; B: biotite. Filled squares for K-Ar dates (samples MBX32, MBX39). Cross symbols correspond to published K-Ar and apatite fission-track dates (compiled from Medford 1975; Wanless et al. 1978; Stevens et al. 1982; see Appendix 1 Table A).*

4.1 U-Pb dating

Samples MX261 and MX19E were pulverized and preliminary mineral separation was carried out at Carleton University using a Frantz isodynamic separator. Samples were then passed through pure and slightly diluted methylene-iodide heavy liquids (Specific Gravity = 3.325) to obtain a heavy minerals separate. Zircons, monazites, and thorites were separated again using another Frantz isodynamic separator at the Geochronology Laboratory of the Geological Survey of Canada in Ottawa where all phases were seived in alcohol.

With the exception of one monazite fraction all mineral phases selected are non-magnetic. Thorites of MX261 and one fraction of monazite in MX19E were abraded using the method of Krogh (1982). All mineral phases were separated into small populations of 10-50 clear inclusion-free grains. Four fractions of euhedral zircons were found in sample MX19E together with one fraction of subhedral monazite (Table 4.1). Sample MX261 contained only one fraction of small euhedral zircons together with two abundant fractions of euhedral thorites (Fig. 4.2, Table 4.1).

Multiple crystals of zircon, monazite and thorite were leached for 30 minutes in warm 3N HNO₃ before dissolution. These minerals were spiked with a mixed ²³⁴U-²³⁵U-²³³U-²⁰⁵Pb tracer prior to dissolution (Parrish and Krogh 1987). Zircons were dissolved in 48% HF for 30 to 60 hours at 230-240°C whereas monazites were dissolved in 12N HCl at 150°C for 60 hours. Thorites did not dissolve in HF at 240°C after 90 hours but 3.1N HCl was added and heated to 230°C for 14 hours. U and Pb were separated using ion exchange column chemistry. Chemical procedures follow the methods of Parrish et al. (1987). Isotopic compositions of Pb and U were measured on a Finnigan-MAT 261 multicollector mass spectrometer (Roddick et al. 1987). Procedural U blanks were mostly less than 3 pg (9 pg for Sample MX261 fraction A), and Pb blanks range from 11 to 18 pg (Table 4.1). Corrections for common Pb were done assuming model Pb compositions derived from the growth curve of Stacey and Kramers (1975) using a ca. 55 Ma model Pb. Decay constants are those recommended by Steiger and Jäger (1977).

Errors in U/Pb ratios refer to one standard error of the mean (Table 4.1), whereas error ellipses (Fig. 4.2) are two standard errors of the mean. Pb-U errors have been estimated by direct propagation of errors as proposed by Roddick (1987). An analysis is considered concordant if the error ellipse overlaps concordia. The mean intersection of the error ellipse with concordia is

thus interpreted as the age of the rock. Due to excess ^{206}Pb from ^{230}Th decay (Shärer 1984) in monazite and thorite, U-Pb ages used are the $^{207}\text{Pb}/^{235}\text{U}$ as described by Parrish (1990).

4.2 Age interpretations

Samples MX261 and MX19E are described with their analytical data in Table 4.1. Concordia diagrams are presented in Figure 4.2 on which zircons are labelled with letters, monazites with numbers, and thorites with capital roman numbers. Zircons are generally of very high clarity and quality. They predominantly occur as colourless faceted euhedral crystals in the form of tetragonal prisms with a length to width ratio of 2:1 to 4:1. These morphological characteristics are typical of zircons of Ladybird-type granitoids and regarded as being indicative of a magmatic origin (Carr 1990). The clearest and most inclusion free and crack-free crystals were chosen for multiple crystal analysis.

Monazites in MX19E are yellow and occur in fragments or as subhedral crystals, interpreted to represent an igneous morphology. Two thorite fractions were found in sample MX261 and occur as euhedral tabular green crystals with a length to width ratio of 2:1 to 4:1 also interpreted as igneous morphologies.

4.2.1 Sample MX261

This rock lies in structural Zone B (see Chapter 2) and consists of a protomylonitic biotite (chloritized) microgranite. The only zircon fraction (A) recovered in this rock yields discordant data with a $^{207}\text{Pb}/^{235}\text{U}$ age of 52.0 ± 0.2 Ma and has either experienced Pb loss or has inherited Pb (Table 4.1; Fig. 4.2a). Two thorite fractions (I, II) yield concordant ratios with very small error ellipses plotting separately on the concordia curve with $^{207}\text{Pb}/^{235}\text{U}$ ages of 50.7 ± 0.1 Ma and 51.2 ± 0.1 Ma respectively. These data do not allow a precise age interpretation to be made, except that the igneous age is older than 51.3 Ma. Both thorite fractions have high concentrations of U of at least 135,000 ppm (Table 4.1) and perhaps this has lowered the resistance of the crystals to Pb loss. The slight differences in age between the two fractions may be related to progressive crystallization at temperatures below 750°C (the closure temperature of Pb in thorite) and results from high-temperature Pb loss related to intense shearing conditions (Carr 1992).

Table 4.1. U-Pb analytical data

<p>MX261: Strained granulite bled (chert) microgranite interbedded with paragneiss along the abandoned CP railway 5 km west of Hydraulic Lake. This unit is overprinted by westerly directed non-coaxial strain fabrics. The chemical characteristics of this unit are common to several plutonic bodies of the Okanagan complex and may be considered as a late aplitic phase of the "Ladybird" suite as defined by Carr et al. 1987, Parrish et al. 1988, and Carr 1990.</p>													
Mineral Fraction / size/abraded	wt. (mg)	U (ppm)	Pb ¹ (ppm)	$^{206}\text{Pb}/^{204}\text{Pb}$	Pb ² (ppm)	$^{206}\text{Pb}/^{204}\text{Pb}$	Pb ³ (ppm)	$^{206}\text{Pb}/^{204}\text{Pb}$	Pb ⁴ (%)	$^{206}\text{Pb}/^{238}\text{U}$	$^{207}\text{Pb}/^{235}\text{U}$	Corr. $^{207}\text{Pb}/^{206}\text{Pb}$	$^{207}\text{Pb}/^{235}\text{U}$ (Ma)
Ac/-105/no	0.028	1092	8.7	1338	14	10.1	10.1	.00797[.27]	.05257[.36]	.05257[.36]	.04778[.25]	88.9 ± 6.0	52.0 ± 0.2
I/-105/yes	0.027	150500	2312	227560	1308	53.5	53.5	.00789[.10]	.051187[.12]	.051187[.12]	.04701[.04]	50.0 ± 1.8	50.7 ± 0.1
II/+105/yes	0.018	135110	2111	15043	811	53.9	53.9	.00797[.13]	.05168[.15]	.05168[.15]	.04701[.04]	49.8 ± 2.0	51.2 ± 0.1
<p>MX192: Muscovite-biotite-bearing fine grained mylonitic granite showing diagnostic kinematic fabrics suggesting top-to-the-west sense of shear. Flagstone porphyroblasts are asymmetric. This rock is also qualified as a "Ladybird-type" granite (Carr et al. 1987; Parrish et al. 1988; Carr 1989, 1990).</p>													
Ac/+149/no	0.065	1702	13.8	393	79	2.4	2.4	.00877[.19]	.057129[.51]	.057129[.51]	.04722[.40]	60.5 ± 19.0	56.4 ± 0.6
Cz/-149+105/no	0.162	1933	14.9	350	218	2.8	2.8	.00832[.25]	.05490[.57]	.05490[.57]	.04766[.44]	92.6 ± 20.5	54.3 ± 0.6
Ep/-149+105/no	0.043	4132	31.4	2763	71	2.6	2.6	.00821[.13]	.05354[.17]	.05354[.17]	.04728[.09]	63.5 ± 4.3	53.0 ± 0.2
Fz/-105+74/no	0.090	1483	12.2	4800	59	3.1	3.1	.00878[.11]	.06152[.15]	.06152[.15]	.05079[.08]	231.3 ± 3.6	60.6 ± 0.2
I/+149/no	0.088	3197	169.9	1754	122	85.6	85.6	.00844[.15]	.053847[.21]	.053847[.21]	.04622[.12]	9.4 ± 5.9	53.3 ± 0.2
2/+149/yes	0.261	3485	167.6	1424	387	84.1	84.1	.00841[.32]	.05359[.35]	.05359[.35]	.04621[.12]	8.8 ± 5.9	53.0 ± 0.4

1: Letters are zircon fractions (c: clear colourless; b: clear brown; cb: cloudy brown); roman numerals are green thorite fractions; numerals are yellow monazite fractions.

2: Radiogenic Pb corrected for spike, fractionation, blank and common Pb.

3: Measured ratio, corrected only for spike and fractionation.

4: Total common Pb in analysis corrected for fractionation and spike.

5: Corrected for blank Pb and U, common Pb; uncertainties in brackets are 1 standard error of the mean in percent.

6: Corrected for blank and common Pb; uncertainties in brackets are 1 standard error of the mean in Ma.

7: Correlation coefficient of errors in $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{235}\text{U}$

Figure 4.2: *U-Pb concordia diagrams for samples MX261, MX19E located on Figure 4.1. Sample MX261 from the Kelowna area contains two concordant thorite fractions (I, II) plotting at 51.4 ± 0.1 Ma but their U content suggest these crystals have suffered important Pb loss. Similarly the only zircon fraction (A) recovered in this sample exhibits either an inherited Pb component or Pb loss. The age of this rock is probably older than 51 Ma. Sample MX19E is a mylonitized biotite-muscovite leucogranite underlying Trinity Hills. It contains one concordant zircon fraction (A) at 56.2 ± 0.2 Ma interpreted as the age of crystallization. Three other discordant zircon fractions may line up on a discordia chord with a lower intercept at 52.5 ± 0.5 Ma and an upper intercept at 1850 Ma. However, the scattered fractions (B, C, E) may be related to variation in age of inheritance and variable Pb loss systematics prior to or during growth of magmatic crystals. This sample probably contains two populations of zircons. Monazites in MX19E show signs of excess ^{206}Pb formed from the decay of initial disequilibrium amounts of ^{230}Th . All error ellipses are shown at the 95% confidence level (isotopic data provided by R. R. Parrish, Geological Survey of Canada, Ottawa).*

4.2.2 Sample MX19E

This sheeted body consists of mylonitized biotite-muscovite±garnet granite lying in structural Zone A as indicated by irrefutable strain markers suggesting top-to-the-west sense of shear (*see* Chapter 2). Four zircon fractions and one monazite fraction were dated. The largest fraction of clear colourless zircons (A) provided concordant data at 56.2 ± 0.2 Ma that are interpreted to be the crystallization age of the granite. Medium sized clear colourless zircons (fraction C) provided slightly discordant data with a $^{207}\text{Pb}/^{235}\text{U}$ age of 54.3 ± 0.6 Ma and another set of medium sized brown zircons (fraction E) provided a nearly concordant $^{207}\text{Pb}/^{235}\text{U}$ age of 53.0 ± 0.2 Ma which may represent another crystallization age if corrected for thorium disequilibrium (*see* Carr 1990 and Parrish 1990). The smallest sized cloudy brown zircons (fraction F) provided highly discordant data showing signs of loss or inherited Pb with a $^{207}\text{Pb}/^{235}\text{U}$ age of 60.6 ± 0.2 Ma. A regression line was tentatively calculated using zircon fractions C, E and F (MSWD = 1) and gives an upper intercept of 1850 ± 50 Ma falling in the range of inheritance of many other zircons in Mesozoic (Sevigny et al. 1989; Scammell 1991) and Cenozoic (Parrish et al. 1988; Carr 1990) granitoids of the southern Omineca Belt. The lower intercept of the discordia chord is at 52.5 Ma. Within error it coincides with the interpreted age of fraction E. The coexistence of the concordant fraction A and the trend defined by fractions C, E, and F may indicate that two populations of magmatic and xenocrystic zircons were analysed in this sample.

Nonmagnetic and magnetic grains of a single monazite fraction (1 and 2) provided overlapping $^{207}\text{Pb}/^{235}\text{U}$ ages of 53.2 ± 0.2 Ma and 53.0 ± 0.4 Ma respectively. Both analyses are "reversely" discordant (Parrish 1990) but were not corrected for Th (Sharer 1984; Parrish 1990). The ages of the monazite within error coincide with the age of the lower intercept of the zircon discordia chord of 52.5 Ma and coincide also with the age of the nearly concordant zircon fraction E of 53 Ma. This indicates that the monazites may be xenocrystic.

4.3 K-Ar dating

Samples were collected from both metamorphic and volcanic rocks lying across the Okanagan valley near Kelowna. Biotite and hornblende were leached in 2N HNO₃ for 20 minutes followed by rinsing in pure H₂O. K and Ar concentrations were determined at the Geochronology

Laboratory of the Geological Survey of Canada by standard atomic absorption techniques (Parrish and Roddick 1985) and isotope dilution mass spectrometry utilizing an ^{39}Ar spike. K analysis was made by atomic absorption spectrometry, and ^{40}Ar by vacuum fusion and isotope dilution mass spectrometry. Cooling ages determined are compiled in Table 4.2 and were corrected for atmospheric Ar present.

4.4 Age Interpretations

All K-Ar samples are located on Figure 4.1 and described in Appendix 3. Table 4.2 presents a summary of the analytical data for each sample analysed. The analyses are classified as the Upper Crustal layer and the Middle Crustal layer referring to the hanging wall or footwall of the Okanagan valley fault respectively. Cataclasites separate these two crustal layers.

Table 4.2. K-Ar analytical data

Sample no.	material ¹	Wt%K	Rad.Ar ²	%Atm. Ar	Age	Lat.	Long.
=====							
<i>Undeformed granites (Upper Crustal layer)</i>							
=====							
MBX32	Hb	0.765±0.49	0.488±0.49%	25.0	157 ± 2	49°56'35"	119°36'55"
MBX39	Bi	7.742±0.48	4.999±0.61%	2.1	159 ± 2	49°37'52"	119°08'30"
=====							
<i>Okanagan gneiss complex (Middle Crustal layer)</i>							
=====							
MX5	Hb	0.779±0.49	0.158±0.98%	22.1	51 ± 1	49°46'08"	119°07'07"
MX5	Bi	7.636±0.67	1.482±0.62%	4.3	49 ± 1	49°46'08"	119°07'07"
MX48	Hb	1.065±0.52	0.185±1.20%	74.7	44 ± 1	49°48'15"	119°14'50"
MX48	Bi	7.522±0.56	1.412±0.67%	18.0	48 ± 1	49°48'15"	119°14'50"
MX229	Hb	0.993±0.56	0.201±0.67%	58.2	51 ± 1	49°46'35"	119°34'15"
=====							
<i>Cataclasite (top of the Okanagan complex)</i>							
=====							
MX133	Hb	1.184±0.98	0.258±0.68%	14.3	55 ± 1	49°49'35"	119°21'52"
=====							
<i>Volcanic rocks (overlying the Upper Crustal layer)</i>							
=====							
MBX1	Bi	6.929±0.76	1.378±0.61%	5.0	50 ± 1	49°48'15"	119°44'35"
MBX2	Bi	6.685±1.71	1.414±1.20%	5.5	54 ± 2	49°54'30"	119°36'56"
MBX4	Bi	7.564±0.40	1.449±0.61%	2.2	49 ± 1	49°42'35"	119°05'50"

¹: Hb: hornblende; Bi: biotite

²: in 10⁴ ccg⁻¹

4.4.1 *Upper Crustal layer*

Plutonic bodies mapped in the Upper Crustal layer preserve Jurassic K-Ar dates interpreted as crystallization ages if extremely rapid quenching occurred. MBX32 is a large body of hornblende-biotite granodiorite exposed in an erosional window through the Westbank Eocene outlier west of Kelowna (Fig. 4.1). Hornblende in this sample contains 25% atmospheric Ar and provided an age of 157.2 ± 2.4 Ma. MBX39 is an undeformed biotite-hornblende granodiorite lying on the east flank of the Okanagan complex near Carmi (Fig. 4.1; Map 1). Biotites in this sample contain only 2% atmospheric Ar and are dated 158.9 ± 2.4 Ma. Because of its structural position MBX39 was first interpreted as a lithology underlying the Okanagan valley fault because it projects structurally below and east of the Eocene outlier of Nipple Mountain. However, Soregaroli (unpublished data 1979) reports Eocene K-Ar ages from an undeformed porphyritic biotite-muscovite granitoid (muscovite = 57.3 ± 2.0 Ma; biotite = 57.8 ± 1.9 Ma) in the immediate footwall of the Nipple Mountain outlier only 10 km southwest of MBX39 (Fig. 4.1; Map 1). If MBX39 lies underneath the strain zone of the Okanagan valley fault it becomes difficult to explain how rocks that are structurally below this zone could have stayed at temperatures below 280°C (the closure temperature for Ar diffusion in biotite). Perhaps the Eocene thermal event so common in the Okanagan complex was focused parallel to the strain zone of the Okanagan valley fault. Alternatively, MBX39 may be part of a klippe composed of rocks of the Upper Crustal layer underlain by an eastward-dipping normal fault. Evidence for this structure has only been identified on air photographs outlined by a significant lineament along the northern branch of the West Kettle River. The lineament extends northwards to Cherryville east of Vernon (Fig. 4.1; Map 1). The presence of this structure would be consistent with the occurrence of several shallow dipping brittle faults underlying isolated klippen composed mainly of Jurassic plutons on the northeastern flank of the Okanagan complex west of the Mount Fosthall-Cherryville region (Carr 1990, 1991). It is proposed that the West Kettle River lineament underlies what is referred to as the Carmi klippe.

The preservation of Jurassic K-Ar mineral dates reveals that the intrusives and their hosting lithologies forming the Upper Crustal layer have cooled below 300°C and stayed below 300°C since the Late Jurassic. Within the analytical precision, these results fall in a widespread

age range of K-Ar mineral dates (155-165 Ma) obtained from the Similkameen and Okanagan batholiths in the southwestern part of the Okanagan valley (Roddick et al. 1972; Peto 1973; Medford 1975). These K-Ar ages are close to the probable age of a major metamorphic event recorded along the eastern flank of the southern Omineca Belt (Archibald et al. 1983) and coincide within error with the only age data obtained by Solberg (1976) west of Vernon in the northern Okanagan valley.

4.4.2 *Middle Crustal layer*

High-grade rocks affected by extensional shear fabrics in the Middle Crustal layer preserve K-Ar mineral dates ranging between 51 ± 1 Ma and 48 ± 1 Ma. For example, biotites in sample MBX5 are only 2 Ma younger than cogenetic hornblende which ranges around 50 ± 1 Ma. The amphibolite sample MX229 provided a hornblende fraction containing as much as 58% atmospheric Ar that are dated 51 ± 1 Ma. A reverse discordance of K-Ar ages between hornblende-biotite pairs is recorded in sample MX48 an amphibolitic gneiss collected near Hydraulic lake east of Kelowna (Fig. 4.1; Table 4.2). Hornblendes in MX48 are dated 44 ± 1 Ma whereas cogenetic biotites yield an age of 48 ± 1 Ma. The relatively young age for the hornblende fraction is probably related to excess atmospheric Ar (Table 4.2) that may have been adsorbed during metamorphic or igneous processes. Biotite thus provides a more realistic cooling age as it contains only 18% atmospheric Ar. This situation has been investigated with the $^{40}\text{Ar}/^{39}\text{Ar}$ method discussed in section 4.5.

4.4.3 *Cataclasite*

The Okanagan valley fault is mantled by brecciated mylonite and chloritic-sericitic gouge material. Sample MX133 is a brecciated mylonitic hornblende gneiss lying along Mission Creek, 8 km southeast of Kelowna (Fig. 4.1). Hornblende in this sample contains 14% atmospheric Ar and provided an age of 55.2 ± 1.3 Ma which may be suspicious (*see* section 4.6.3).

4.4.4 Eocene lavas

Rocks of the Kelowna volcanic outlier unconformably overlie the Upper Crustal layer west of Kelowna (Fig. 4.1). The outlier comprises nearly 2 km of intermediate volcanic flows capped by 1.2 km of alluvial sedimentary rocks which were rotated into the plane of the Okanagan valley fault after their deposition (*see* Chapter 6). The trachyandesite sample MBX2 is part of the Marron Formation (Church 1973, 1981b) and lies near the top of the outlier. Biotites in this sample contain only 5% atmospheric Ar and provided an age of 53.6 ± 2.2 Ma (Table 4.2). A quartz-orthoclase porphyritic laccolith MBX1 intruding the base of the trachyandesitic flows west of Westbank (Fig. 4.1) provided a biotite age of 50.5 ± 1 Ma (Table 4.2). The K-Ar ages obtained from MBX1 and MBX2 likely indicate their age of crystallization for there is no detectable evidence of K-metasomatism in these rocks (*see* Chapter 6).

Several dacitic dikes cut through mylonitic gneisses of the Middle Crustal layer southeast of Hydraulic lake (Fig. 4.1). One of these dikes MBX4 (Table 4.2) contains fresh biotite which provided an age of 48.6 ± 0.7 Ma and confirms the field observations that it feeds the youngest dacitic flows in this region (*see* Chapter 6). It is proposed that this date brackets the last increment of ductile shearing on the Okanagan valley fault as MBX4 is undeformed.

4.5 ⁴⁰Ar/³⁹Ar dating

Mineral separation was carried out at Carleton University and ⁴⁰Ar/³⁹Ar analyses were run by myself and Dr. D. J. Archibald at the Queen's University Geochronology Laboratory. Each sample was pulverized to -60/+200 mesh size, using a disk mill. Iron fillings were first separated with magnets enveloped in cotton cloths. A Frantz isodynamic separator was used to extract the ferromagnesian minerals and then the samples were washed in water and cleaned with acetone. Magnetic separation was most efficient while setting the isodynamic separator at variable side tilts (5° to 28°) and magnetic fields of .25 to .59 Å. Samples were 80% pure and were put in an ultrasonic bath for 10 minutes to separate feldspar aggregates from ferromagnesian minerals. The separates were then sized in two fractions of -100/+140 microns and -140/+200 microns. Visual examination indicated that hornblende and biotite separates were 99% pure. Further purification was obtained by hand-picking the best crystals to ensure a 99.9% purity.

Fifteen samples and four flux monitors were irradiated with fast neutrons in position 5C of the McMaster Nuclear Reactor in Hamilton, Ontario, for 1.375 days. The monitors were distributed throughout the irradiation container, and J-values (the irradiation parameter) for individual samples were determined by interpolation (Table 4.3).

The bakeable, UHV, stainless-steel, Ar-extraction system is operated on-line with a modified, A.E.I. MS-10 mass spectrometer run in a static mode. Total-fusion analyses were done in resistively-heated niobium crucibles in a quartz tube. Radioactive gas was concentrated through four high-vacuum valves separating two charcoal collectors before the purified gas is admitted to the mass spectrometer. Isotopic analysis commenced immediately after the purified gas entered the spectrometer. Measured mass-spectrometric ratios were extrapolated to zero-time, corrected to an $^{40}\text{Ar}/^{39}\text{Ar}$ atmospheric ratio of 295.5, and corrected to neutron induced ^{40}Ar from potassium, and ^{39}Ar and ^{38}Ar from calcium. Dates and errors were calculated using formulae of Dalrymple et al. (1981) and the decay constants recommended by Steiger and Jäger (1977).

4.6 Age Interpretations

$^{40}\text{Ar}/^{39}\text{Ar}$ age spectra from 10 samples collected in the Okanagan region are illustrated on Figure 4.3 and the analytical data is summarized in Table 4.3. Detailed descriptions and complete analytical data of the samples are given in Appendix 3. Samples tested include multigrain fractions of hornblende, biotite, muscovite and sericitic fault gouge that represent distinctive levels of the structural zones underlying the Okanagan valley fault (*see* Chapter 2). The plutonic rock MBX32, gneissic samples MX48, MX229, and the cataclasite sample MX133 were analysed by the K-Ar method (*see* sections 4.4.1, 2, 3) and redated with the $^{40}\text{Ar}/^{39}\text{Ar}$ method on similar minerals. Sample MX19E analysed with the U-Pb method on zircon was dated by $^{40}\text{Ar}/^{39}\text{Ar}$ on muscovite. Single total fusion analyses of samples MX49, MXE42, MXE100, MX300, MXOF1 and duplicate analyses of samples MX03 and MX21 were obtained and listed in Table 4.4.

Table 4.3. Summary of the $^{40}\text{Ar}/^{39}\text{Ar}$ step-heating data.

Sample Number ¹	Mat. ¹	Grain Size	Mass (mg)	J-value $\times 10^2$	K Wt%	Int. Age ² Ma $\pm 2\sigma$	Plateau Age ³ Ma $\pm 2\sigma$, ($\%^{39}\text{Ar}$)
=====							
<i>Okanagan Batholith</i> (Upper Crustal layer)							
=====							
MBX32 ²	Bi	>60	280	0.4927	6.80	159.90 $\pm$ 0.70	160.40 $\pm$ 0.60(83)
=====							
<i>Okanagan gneiss complex</i> (Middle Crustal layer)							
=====							
MX19E'	Mu	+60	122	0.6885	6.10	48.92 $\pm$ 0.85	48.69 $\pm$ 0.62(52)
MX19E'	Mu	-60	294	0.6920	6.40	49.17 $\pm$ 0.60	49.17 $\pm$ 0.39(92)
MX48 ²	Hb	>60	280	0.4920	0.91	51.50 $\pm$ 0.90	50.40 $\pm$ 0.90(77)
MX48 ²	Bi	>60	280	0.4910	7.30	48.90 $\pm$ 0.50	48.70 $\pm$ 0.40(98)
MX49 ²	Hb	>60	212	0.3157	0.73	48.70 $\pm$ 4.40	48.50 $\pm$ 2.40(92)
MX229 ²	Hb	>60	67	0.3173	0.86	56.40 $\pm$ 5.10	47.30 $\pm$ 3.70(90)
MX229 ²	Hb	>60	98	0.4880	1.00	55.20 $\pm$ 5.70	51.60 $\pm$ 4.60(94)
MX300 ²	Hb	>60	126	0.3187	1.30	49.10 $\pm$ 2.30	48.70 $\pm$ 1.00(77)
=====							
<i>Cataclasite</i>							
=====							
MX03 ²	Mu	+40	138	0.6880	7.10	50.57 $\pm$ 0.82	50.50 $\pm$ 0.60(92)
MX10	WR	60/80	254	0.6873	1.80	48.54 $\pm$ 3.05	48.53 $\pm$ 1.42(48)
MX78	WR	-60	225	0.4885	0.80	59.10 $\pm$ 2.80	59.70 $\pm$ 1.00(94)
MX78	WR	80/100	351	0.3278	0.82	58.83 $\pm$ 1.80	50.51 $\pm$ 3.25(7)
MX133 ²	Hb	>60	251	0.6848	0.85	53.67 $\pm$ 2.29	52.09 $\pm$ 1.75(98)

¹: Analysed by two methods; *: $^{40}\text{Ar}/^{39}\text{Ar}$ and K-Ar; *: $^{40}\text{Ar}/^{39}\text{Ar}$ and U-Pb; °: $^{40}\text{Ar}/^{39}\text{Ar}$ and total-fusion

²: Bi: biotite; Hb: hornblende; Mu: muscovite; WR: whole rock

³: Includes more than 90% of the incremental steps

⁴: On average includes 60% of the incremental steps

Table 4.4. Total-fusion analytical data.

Sample Number	Run Number	Material	Grain Size	C/P	Mass mg	J-Value	40/39	36/39	37/39	Vol. 39Ar E-8 cm ³	%40Ar Rad.	Age t _{1/2}	+/-	Error 2 sigma
MX03	D-615	Muscovite	40/60	47/48-11	6	4.8750E-03	13.534	0.0254	0.022	0.574	44.42	52.12	+/-	4.05
MX-21	D-614	Muscovite	80/100	47/48-13	13	4.8600E-03	9.431	0.0122	0.005	2.458	61.52	50.17	+/-	0.42
MX-300	D-613	Hornblende	60/80	47/48-12	107	4.8650E-03	9.937	0.0125	3.168	0.958	65.14	56.05	+/-	2.14
MX49	D-612	Hornblende	60/80	47/48-7	100	4.9000E-03	14.591	0.0269	6.155	0.485	48.59	61.85	+/-	3.22
MX49	D-611	Biotite	40/60	47/48-8	51	4.8950E-03	6.628	0.0035	0.992	5.244	84.06	48.54	+/-	0.62
MX21	D-690	Muscovite	>60	76/35	36	6.9150E-03	5.831	0.0059	0.000	13.277	69.77	50.05	+/-	0.20
MXE100	D-691	Muscovite	>60	76/37	38	6.9050E-03	8.043	0.0126	0.000	13.807	53.41	52.74	+/-	0.28
MXE-42	D-692	Muscovite	>60	76/32	13	6.9300E-03	6.229	0.0088	0.034	4.769	67.37	51.73	+/-	0.48
MXOF01	D-693	Muscovite	<60	76/35	12	6.9100E-03	7.618	0.0119	0.000	3.544	53.47	50.07	+/-	0.91
MXOF01	D-694	Muscovite	>60	76/38	5	6.8925E-03	7.783	0.0127	0.000	1.411	49.29	47.02	+/-	11.47
MX03	D-695	Muscovite	>60	76/33	5	6.9250E-03	6.845	0.0088	0.474	1.790	61.96	52.24	+/-	2.09

Figure 4.3: *⁴⁰Ar/³⁹Ar incremental-release age spectra of hornblende, biotite, and muscovite concentrates and whole-rock samples collected in the footwall and hanging wall of the Okanagan valley fault. Uncertainties in age (2 σ) are indicated by vertical width of bars. Integrated and plateau ages (% of Ar released in parentheses) are listed on each of the spectra. Map patterns as in Figure 4.1 and analytical data are presented in Table 4.3. (Footwall data - Middle Crustal layer) Hornblende concentrates from amphibolite samples MX49.1, MX229, and MX133 provided variably discordant ⁴⁰Ar/³⁹Ar release spectra. Samples MX48 and MX300 provided a nearly concordant release pattern with plateau ages of 50.4 ± 0.9 Ma and 48.7 ± 1.0 Ma, respectively. MX300 also provided a total-fusion age of 56.05 ± 2.14 Ma (see Table 4.4). Biotite from sample MX48 from which hornblende was analysed provided a concordant plateau age of 48.7 ± 0.4 Ma. Two muscovite concentrations of sample MX19E provided internally concordant release spectra with plateau ages of 49.2 ± 0.4 Ma and 48.7 ± 0.6 Ma, respectively. Muscovite concentrates from a cataclasite at the top of the Okanagan valley fault underlying the Joe Rich klippe provided an internally concordant spectra of 50.5 ± 0.6 Ma. Wholerock samples of two other cataclasites exhibit contrasting age spectra. Sample MX10 contains a significant proportion of white mica and chlorite and provided markedly discordant, saddle-shaped release spectra which record a plateau age of 48.5 ± 1.4 Ma. Sample MX78 provided an extremely concordant release spectra which define a plateau age of 59.7 ± 1.0 Ma. This data is ambiguous for it suggests that brittle rocks of the Okanagan valley fault were cooling prior to ductile fabrics (see text). (Hanging wall data - Upper Crustal layer) Biotite concentrates in sample MBX32 provided internally concordant release spectra which define a plateau age of 160.4 ± 0.6 Ma consistent with a hornblende K-Ar age of 157.2 ± 2.4 Ma (Table 4.3). The data shows that hanging wall rocks were not reheated above 280°C since the Jurassic whereas footwall rocks were cooled from 550°C to 250°C between 51 Ma and 48 Ma indicating extremely rapid exhumation rates.*

4.6.1 *Granitoid of the Upper Crustal layer*

Sample MBX32, a granodiorite underlying the Kelowna volcanic outlier near Westbank (Fig. 4.3), yields a concordant biotite plateau age of 160.4 ± 0.6 Ma that is slightly older than a K-Ar hornblende date of 157.2 ± 2.4 Ma. This data confirms the Late Jurassic age of cooling of this pluton and indicates that major post-late-Jurassic thermal events have not affected the hanging wall of the Okanagan valley fault. The Jurassic age falls in the range of a regional igneous event recorded in south-central British Columbia (Archibald et al. 1983; Armstrong 1988).

4.6.2 *Amphibolites of the Middle Crustal layer*

Samples of high-grade gneisses of the Okanagan complex yield a variety of plateau-shaped age spectra that are generally indicative of normal closure of Ar diffusion in hornblende and do not show evidence of a reheating event (Fig. 4.3; McDougall and Harrison 1988). Hornblende age spectra generally exhibit anomalous steps starting around 50% cumulative ^{39}Ar which may indicate that amphiboles have lost a certain quantity of Ar. This pattern has been found systematically in most amphiboles dated from the southern Omineca Belt (D. A. Archibald, pers. comm., 1991). Within error limits, minimum ages of the hornblende spectra of the Kelowna region range between 46 Ma to 56 Ma. Integrated ages for a biotite fraction in sample MX48 yield a plateau age of 48.7 ± 0.4 Ma that is 1.5 Ma younger than its cogenetic hornblende fraction. This data allows us to discard the anomalous K-Ar hornblende age of 44 Ma (see section 4.4.2). Sample MX49 lies immediately west of sample MX48 (Fig. 4.1) and provided a plateau age of 48.5 ± 2.4 Ma on hornblende. Total-fusion data from the same sample yield an age of 61.9 ± 3.2 Ma on hornblende and a 48.5 ± 0.6 Ma age on biotite (Table 4.4). The older total-fusion hornblende age may be caused by the presence of excess Ar similarly as for hornblende in sample MX300 which yields a total-fusion age of 56.1 ± 2.1 Ma.

Muscovite occurs within extensional flattening and shear surfaces in sample MX19E which provided a zircon fraction with a crystallization age of 56 Ma (see section 4.2.2). Two muscovite fractions exhibit concordant plateaus at 48.7 ± 0.6 Ma and 49.2 ± 0.4 Ma. Lying at the same structural level but nearly 130 km south of sample MX19E, sample MXOF1 from the Okanagan Falls region (Fig. 4.3) is a metasedimentary rock in which muscovite defines flattening and shear

planes created under extensional non-coaxial strain conditions (see Fig. 2.2c). These muscovites provide a total-fusion age of 50.1 ± 1 Ma (Table 4.4) which may be considered coeval to the muscovites in sample MX19E.

4.6.3 *Cataclasites*

Cataclastic rocks immediately underlying the footwall of the Okanagan valley fault exhibit various age spectra. Sample MX133 yielded a hornblende plateau age of 52.1 ± 1.8 Ma with relatively large errors and nearly overlaps the 55 ± 1 Ma age obtained from the same mineral separate by the K-Ar method (Tables 4.2, 4.3; Fig. 4.3). Muscovite of sample MX03 provided a concordant plateau age of 50.5 ± 0.6 Ma (Table 4.3; Fig. 4.3) which overlaps within error with total-fusion ages of 52 ± 3 Ma provided by two other muscovite fractions (Table 4.4). These data suggest that total-fusion ages represent maximum cooling ages and/or are affected by the smaller grain size of certain total-fusion samples (McDougall and Harrison 1988). MX21 is another cataclastic mylonite underlying the Joe Rich klippe east of Kelowna (see Chapter 2). Two muscovite fractions from this sample were analysed 2 years apart and provided the same total-fusion age of 50.1 ± 0.2 Ma (Table 4.4) which is younger than the total-fusion age of MX03 cited above. Because samples MX03 and MX21 occupy the same structural position, and are only 4 km apart, it may be proposed that they reflect slightly different cooling ages.

Two whole rock samples of sericitic fault gouge underlying the Eocene dacitic outlier of Black Knight Mountain near Kelowna (see Frontispiece) reveal some intriguing information (Tables 4.3, 4.4). Sample MX10 was collected in a cataclastic zone immediately beneath the Eocene volcanic outlier in one of the best exposed sections of the upper part of the Okanagan valley fault. Sample MX78 was collected on a splay fault approximately 100 m below MX10. The latter yields a complex spectrum with an average plateau age of 48.5 ± 3.1 Ma whilst MX78 provided a much older but concordant plateau age of 59.7 ± 1.0 Ma (Fig. 4.3). The cooling history of these samples is difficult to assess in the present tectonic framework. A solution may arise from their Ar systematics as indicated by their distinctive spectra (Fig. 4.3). Heating steps of MX10 are grouped in three segments breaking at 21% and 53% ^{39}Ar released which define a saddle-shaped pattern with relatively large errors. The first segment of this pattern shows

important loss of ^{39}Ar whereas the second and third segments provide average plateaus at 52 Ma and 48 Ma. In contrast, the plateaus of sample MX78 are extremely concordant with very small errors for more than 95% of the purified gas released. The irregular spectrum of MX10 may be explained by poor Ar retention of variable grain sized sericite. Variable grain sizes may trap or release various quantities of Ar causing a certain amount of recoiling (McDougall and Harrison 1988). There may be reasons to reject the data of sample MX78 simply because it provides a suspicious age amongst most cataclastic samples, MX03, MX21, and MX10 which provide muscovite ages ranging around 52-50 Ma. Perhaps MX78 has been subjected to severe leaching of K leaving excessive Ar concentrations as commonly recorded in brittle fault zones (Kerrick 1988). These effects might have been obliterated by heating the sample too rapidly during the first stages of the step-heating procedures so the sample was analysed a second time and provided a concave age spectra. This data shows that the sample contains small amounts of excess Ar in the low temperature range though it reaches an integrated age of 58.8 ± 1.8 Ma (Table 4.3) which could still indicate the age of cooling of this rock.

5. Discussion

The isotopic data acquired from the northern Okanagan region can be treated to decipher: the crystallization age of anatectic melts in the Okanagan complex; the age of ductile and brittle deformation of the Okanagan valley fault; the cooling rate recorded in the footwall and the hanging wall of the Okanagan valley fault; and the age of crystallization of intermediate volcanic flows in the hanging wall of the Okanagan valley fault.

5.1 Origin of the leucogranites

Several leucogranite bodies underlying the metamorphic culminations of the southern Omineca Belt were formed as anatectic melts (see Chapter 1). Many of these bodies crystallized in the Early Tertiary and created the Ladybird suite (Carr et al. 1987; Parrish et al. 1988; Carr 1989, 1990, 1992). Because of their mineralogy and field relationships with their hosting lithologies; the presence of zircons inherited from Proterozoic sources; and the presence of zircons with middle Eocene crystallization ages, samples MX19E and MX261 can be correlated

with intrusions classified as late to post-kinematic Ladybird suite in the Valhalla complex (Parrish et al. 1988). Carr (1990) observed that Ladybird intrusions are generally absent in lower crustal rocks underlying the Monashee décollement in Thor-Odin but are abundant in mid-crustal rocks. Assuming this observation applies latitudinally across the southern Omineca Belt, the presence of Ladybird-type granitoids is consistent with the interpretation that the Okanagan complex is a middle crustal layer.

U-Pb zircon systematics in both samples MX19E and MX261 have probably been modified by high-temperature or fluid-enhanced diffusion during intense shearing along the Okanagan valley fault. This is noticeable because most zircon fractions exhibit Pb loss. It may be expected that the monazite and thorite systematics in these samples reflect also Pb loss processes enhanced by strain mechanisms.

Granitic melts can be formed as *in situ* metamorphic segregations derived from quartzofeldspathic rocks under conditions of regional metamorphism (Brown and Fyfe 1970; Wyllie 1977; Robin 1979; Whitney 1988). Increasing temperature, decreasing pressure or increasing P_{H_2O} can trigger melting in the crust (Wyllie 1977). Based on the tectonic history of the southern Omineca Belt (Brown et al. 1986; Journeay 1986; Scammell 1986) documented by an excellent geochronometric database it appears that melts of the Ladybird suite, as well as many other (older) leucogranites in this region, were formed while "peak" metamorphic conditions were reached during at least two compressional events. These events led to major crustal thickening and at least two diachronous periods of metamorphism starting around 136 Ma (Scammell 1991). According to other workers the last episode of peak metamorphism affecting the southern Omineca Belt was diachronous between the Late Cretaceous and Late Paleocene (Sevigny et al. 1989; Carr 1990; Scammell 1991) when final thrusting along the Monashee décollement occurred (Coleman 1989; Carr 1990; 1992) and converted simultaneously to extensional conditions.

5.2 Relationship of intrusion to extension along the Okanagan valley fault

A concordant zircon fraction in MX19E suggests that extensional shear features overprinting this rock and associated with the Okanagan valley fault may have been formed as early as 56 Ma. This age corresponds to the late kinematic phase of the Ladybird suite (Parrish

et al. 1988; Carr 1990) intruding the Valkyr shear zone which exhibits top-to-the-east sense of shear in Valhalla complex (Carr et al. 1987; Parrish et al. 1988). Although older phases of the Ladybird suite may occur in the Okanagan complex the present U-Pb data argues in favor of a temporal distinction between extensional shearing on the Valkyr shear zone (Parrish et al. 1988) and top-to-the-west shearing on the Okanagan valley fault. However, it appears that the Okanagan valley fault may have been activated at the same time as two other easterly-dipping and easterly-verging structures straddling the eastern boundary of the southern Omineca Belt, that is the Columbia River fault (Lane 1984) and the Slocan Lake fault. The latter cuts the Valkyr shear zone (Parrish et al. 1988). This suggests that both flanks of the Omineca Belt were denuded simultaneously partly because melt enhanced softening caused by Ladybird-type granites in the middle crust probably played the same role in the nucleation of the Okanagan valley fault as it did for easterly-dipping shear zones (Carr 1990).

5.3 Cooling history of the Okanagan complex

The zircon age in sample MX19E indicates that this granitoid and possibly its hosting high-grade metamorphic rocks from which it was segregated started cooling below 750°C (closure temperature for Pb in zircon; Parrish and Roddick 1985) at 56 Ma and cooled below 650°C (closure temperature for Pb in monazite; Parrish and Roddick 1985) at 53 Ma. There is little systematic variation in the K-Ar ages in the migmatitic gneisses of the Okanagan complex thus contouring these data does not outline any cooling pattern (Archibald et al. 1984) which is consistent with the idea that the gneisses underwent rapid exhumation. Some of the lithologies dated by the K-Ar method may also have been jostled by late brittle faults. In general terms, the K-Ar ages on hornblende and biotite in the gneisses of the Kelowna region are consistent with the K-Ar ages provided by similar minerals in the southern Okanagan region (Fig. 4.1; Appendix 1). These data are consistent with the regional pattern of Eocene cooling dates documented in south-central British Columbia (Baadsgaard et al. 1961; Armstrong 1974, 1982, 1983, 1985; Ross 1974; Medford 1975; Miller and Engels 1975; Fox et al. 1977; Parrish 1979; Mathews 1981, 1983; Okulitch 1984; Price et al. 1981; Price 1985; Archibald et al. 1984; Parkinson 1985; Parrish et al. 1988; Carr 1990). The frequency histogram of Figure 4.4 shows the distribution

Figure 4.4: *Frequency histogram showing the distribution of 100 K-Ar dates in south-central British Columbia (compiled from Armstrong 1974, 1982, 1983, 1985; Ross 1974; Medford 1975; Miller and Engels 1975; Fox et al. 1977; Mathews 1981, 1983; Parrish 1979; Okulitch 1984; Price 1985; Price et al. 1981; Archibald et al. 1984; Parkinson 1985; and this study). Range of ages is between 40 and 68 Ma with more than 50% concentrated in the 48-54 Ma bracket interpreted as the time spanned to close most of the K-Ar system for phases with decreasing closing temperature.*

of 100 K-Ar dates (this study included) ranging from 68 to 40 Ma. At least 60% of them fall in the 54-48 Ma range which is interpreted as the time during which most of the Ar system closed sequentially in hornblende, muscovite, and biotite in the southern Omineca Belt.

Except for sample MX229 which shows evidence of excess Ar, samples MX48, MX49, MX300 (Table 4.3) provided $^{40}\text{Ar}/^{39}\text{Ar}$ step-heating and total-fusion data overlapping with the K-Ar data of the same or similar rock types. With better precision the $^{40}\text{Ar}/^{39}\text{Ar}$ data indicates that amphiboles cooled below 550°C (the closure temperature of Ar in hornblende; Harrison 1981) between 52 and 48.5 Ma. Extremely consistent $^{40}\text{Ar}/^{39}\text{Ar}$ muscovite ages of 50 Ma indicate that the Okanagan complex cooled simultaneously below 350°C (closure temperature of Ar in muscovite; McDougall and Harrison 1988) over a surface of approximately 4500 km^2 from Enderby to Okanagan Falls (Fig. 4.1). Ductile shearing in the footwall of the Okanagan valley fault thus seems to have lasted approximately 7 Ma ending around 49 Ma when rocks of the footwall cooled below 350°C entering the crustal domain of brittle deformation assuming normal exhumation conditions provided by tectonic denudation and erosion.

In theory it may be envisaged that extensional faults nucleate from shear zones developed in the middle crust and propagate diachronously in the upper crust while the lower crust is progressively exhumed (Wernicke 1985). Cooling ages in the brittle zone would therefore be expected to be younger than cooling ages in the ductile zone and would be consistent with the observation that fragments in the brittle zone of the Okanagan valley fault are dominantly ultramylonites. Few cooling ages in the fault blocks presently juxtaposed may relate to the time when the fault started breaking away. Cooling events related to the period of first break-up have a good chance to be preserved if the brittle zone is cooled by cold fluids or simply because these rocks never get buried deeply. Conversely, footwall rocks are maintained hot until they reach the closure temperatures of the minerals analysed. Hence these rocks provide cooling ages related to final stages of exhumation. The data obtained from samples in the ductile zone of the Okanagan valley fault suggest these rocks cooled below 550°C 52 Ma ago. Most samples of the cataclastic zone mantling the Okanagan complex suggest that brittle deformation occurred below 250°C (closure temperature for Ar in mica; McDougall and Harrison 1988) around 50 Ma. An exceptional $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age of 59 Ma was obtained from sample MX78, a splay fault in the

cataclastic zone, and a 59.9 Ma K-Ar date was obtained from a gneissic rock in the southern Okanagan region (Medford 1975; sample 3-10; *see* Appendix 1). Assuming they do not contain excess Ar, these rocks may preserve an early stage of the extension history. If this data is valid it would indicate that the Okanagan valley fault has been activated at the same time as the Valkyr shear zone in the Valhalla complex (Carr et al. 1987; Parrish et al. 1988) suggesting that the Okanagan fault, like easterly-dipping extensional faults, formed during the last moments of thrusting on the Monashee décollement (Carr 1990).

The cooling rates estimated from diverse geochronological methods in the Okanagan complex exceed $65^{\circ}\text{C Ma}^{-1}$ which is three times greater than the Alpine cooling rate (Clark and Jager 1969). Given a hypothetical geothermal gradient of $40^{\circ}\text{C km}^{-1}$ realistic in a context of extensional failure and an erosion rate of 2 mm a^{-1} (Heller et al. 1987) exhumation rates during extension between 53 Ma and 49 Ma would have been in the order of 4 mm a^{-1} (Fig. 4.5) which is slightly more than the current uplift rate of this area (Holdhal, pers. comm., 1986). Apatite fission track dates obtained by Medford (1975) in the Okanagan complex yielded an average age of 45 Ma (track retention is essentially complete in apatite at temperatures below 105°C) suggesting that cooling and exhumation rates in the Okanagan complex were decreasing slowly between 49 and 45 Ma to perhaps $1\text{-}2 \text{ mm a}^{-1}$ nearly in equilibrium with erosion rates.

Jurassic $^{40}\text{Ar}/^{39}\text{Ar}$ biotite dates clearly indicate that rocks of the Upper Crustal layer have not been heated above 280°C during the cooling event affecting the Middle Crustal layer. Apatite fission-track dates in Mesozoic granitoids of the Upper Crustal layer yielded a mean age of 54 Ma (Medford 1975) indicating that these rocks were heated to approximately 100°C in the Middle Eocene (Fig. 4.5). Low temperature heat sources may have been provided by syn-extensional volcanic lavas and/or from mid-crustal fluids percolating from the lower and middle crust.

When we examine the data from the northern Okanagan region it is not obvious that geochronology has helped to decipher with consistency the temporal evolution of certain thermal events related to the Okanagan valley fault. The data reveals no definite cooling history affecting the rocks of the Okanagan complex (Fig. 4.5). It looks as if these rocks cooled from 650°C to 250°C in less than 2 Ma when the isotopic data is considered altogether within error (Fig. 4.5). The cooling path indicated by these data is probably oversimplified and indicates the effects of

Figure 4.5: *Cooling curve of the hanging wall (hatched trend) and footwall (shaded trend) of the Okanagan valley fault. Sizes of boxes reflect mineral ages summarized from Tables 4.1, 4.2, 4.3, 4.4 and closure temperatures are discussed in text. Apatite fission-track data for both hanging wall and footwall compiled from Medford (1978). Anatectic melts are assumed to have formed under peak metamorphic conditions reaching at least 700°C (see Chapter 3) at the end of compression probably in the Early Tertiary. The age of peak metamorphism has been extrapolated from the studies of Carr (1990) and Scammell (1991). The melts are overprinted by extensional shear fabrics which may have started forming as soon as the melts formed. The thermal conditions and the time of igneous crystallization therefore may represent the time that rocks in the footwall of the Okanagan valley fault were at 700°C.*

complicated systematics caused by massive elemental diffusion (Pb loss, inherited Ar etc...) leading to a certain homogenization of the isotopic systems. Intensely deformed rocks (mylonites and cataclasites) contain crystal defects which may modify isotopic diffusion mechanisms and can reset most isotopic ratios. Isotope diffusion may be particularly modified while crystals are dislocated at high-temperatures or affected by intense metasomatism (Kerrick 1988).

5.4 Timing of volcanism versus cooling of the Okanagan complex

The Kelowna volcanic outlier overlying the Upper Crustal layer (see Chapters 6, 7) contains a trachyandesite unit MBX2 that was cooling below 280°C (the closure temperature of Ar in biotite; McDougall and Harrison 1988) 54 Ma ago. A porphyritic body MBX1 intruding the base of the Kelowna outlier was also cooling below 280°C, 51 Ma ago. The different biotite ages of these units indicate that the dates obtained are crystallisation ages.

The data suggest that mafic lavas overlying the Upper Crustal layer cooled below 280°C simultaneously as metamorphic rocks of the Middle Crustal layer cooled from 550°C to 350°C. The age of crystallization of the trachyandesite unit MBX2 (54 Ma) is nearly coeval with that of the granitic body MX19E (56 Ma). The age of crystallization of the porphyritic laccolith MBX1 is coeval to the crystallization age provided by zircons (51 ± 1 Ma) in a mylonitized porphyritic sill intruding the southern Okanagan complex (Parkinson 1985; Parrish et al. 1988). Though the units cited above were probably never forming intrusive-extrusive pairs their respective crystallization ages indicate that volcanism in the hanging wall of the Okanagan valley fault occurred while anatectic melts were crystallizing and cooling in the footwall of the Okanagan valley fault. The time correlation established between hanging wall volcanic rocks and footwall gneisses therefore indicates a close spatial and temporal association between the onset of large-magnitude extension and voluminous calc-alkaline magmatism throughout south-central British Columbia. It is proposed that trachyandesitic lavas were extruded slightly before the main break-up of the Okanagan valley fault at the earliest stages of extension. The cooling age provided by apatite fission-track dates in the granitoids underlying the Eocene lavas (Medford 1975) confirms that these rocks were colder than 105°C at 54 Ma. This suggests that the oldest lithologies in the hanging wall of the Okanagan valley fault were probably reheated by extruding lavas and partly

by hot footwall rocks as the hanging wall was rapidly juxtaposed against the rapidly unroofing Middle Crustal layer.

It appears that a general decrease in extensional strain rates is associated with a decrease in volcanic activity after 48 Ma. A dacitic dike MBX4 that cuts mylonitic gneisses of the Okanagan complex provided a biotite age of 46 Ma and clearly indicates the end of ductile shearing in the Okanagan region as the dike is undeformed and feeds volcanic flows lying unconformably or in brittle contact with the Okanagan complex (see Chapter 6).

6. Conclusions

Eocene leucogranites of the Ladybird suite form an unknown volume of the Okanagan complex and indicate that the latter is part of the Middle Crustal layer overlying the Monashee décollement. Ladybird-type melts invade undated polydeformed high-grade paragneisses assumed to have been metamorphosed to peak conditions in the Late Cretaceous - Early Tertiary during a regional compressional event. Two sheeted bodies of leucogranite 70 km apart lie within the ductile zone of the Okanagan valley fault system and exhibit extensional shear fabrics. The presence of one fraction of concordant zircon in one of these bodies suggests that shearing may have started as early as 56 Ma ago. Other zircon, monazite, and thorite fractions in these rocks provide U-Pb ages of 51 Ma and are thought to have lost Pb at high temperatures as a result of fluid or high-temperature diffusion as both samples were submitted to intense deformation.

The dynamics and the tectonothermal evolution of the Okanagan valley fault resembles and seems to have been contemporaneous with shearing along the easterly dipping Columbia River and Slocan Lake extensional faults indicating that both flanks of the southern Omineca Belt have been denuded simultaneously circa 55 Ma.

A kinematic link between extensional faulting and cooling of the Okanagan complex is provided by the K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages of several mineral phases. Though they overlap within error around 50 Ma, gneissic minerals indicate that the Okanagan complex has been exhumed rapidly by ductile shearing along the Okanagan valley fault at an average rate of 4 mma^{-1} between 54 Ma and 48 Ma while cooling from 650°C to 250°C reducing to 2 mma^{-1} between 50 and 45 Ma while cooling from 250°C to 75°C assuming geothermal gradients of 40°C/km⁻¹.

Rocks in the brittle zone mantling the Okanagan complex yield cooling ages generally overlapping those of the ductile zone suggesting that both deformation mechanisms occurred nearly simultaneously. This is contradictory with the idea that ductile deformation precedes brittle deformation as the footwall of a denudation fault is progressively unroofed from deep to shallow crustal levels. Perhaps hydrothermal alteration has caused severe depletions in K feldspars and other K-bearing minerals leading to the re-equilibration of the K-Ar isotopic systems.

Timing constraints are placed on the cooling of the Middle Crustal layer and intermediate volcanism onto the Upper Crustal layer. The ages of crystallization of Eocene lavas are in perfect concordance within error of most K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages recorded by the metamorphic minerals of the Middle Crustal layer suggesting that both phenomena occurred simultaneously between 54 Ma and 50 Ma. Hanging wall Paleozoic and Jurassic rocks of the Upper Crustal layer remained cooler than 280°C but the extrusion of intermediate lavas approximately 54 Ma ago and the intrusion of porphyries 51 Ma ago probably reheated the hosting lithologies to approximately 100°C. It is clear that volcanism occurred while rocks of the Okanagan complex were exhumed and cooled suggesting that magmatism has played an active role in the extensional regime of the Okanagan valley fault.

1. *Introduction*

Estimating magnitudes of displacement on shallow dipping normal faults is a difficult task when there are no reliable markers across the principal fault surface which has been partly eroded and arched after major exhumation of the footwall rocks. Estimates of the magnitude of displacement along the Okanagan valley fault rely on indirect evidence leading to large uncertainties. We will try to constrain minimum and maximum values of displacement given the structural, metamorphic and geochronological information presented in Chapters 2, 3, and 4.

2. *Timing, exhumation rates and magnitude of displacement*

Rapid exhumation has led to rapid cooling of the Okanagan complex, the footwall of the Okanagan valley fault (*see* Fig. 4.5). Cooling of anatectic melts and their hosting sillimanite-grade gneisses in the footwall started at temperatures exceeding 500°C and perhaps as high as 750°C (*see* Chapter 3). This was followed by the creation of extensional ductile fabrics between ~56 Ma and 50 Ma ago (*see* Chapter 4). Rapidly exhumed footwall rocks passed the 500°C isotherm between 54 and 52 Ma and the 350°C isotherm between 52 and 50 Ma. By 48 Ma footwall rocks had cooled below 250°C and were continuously but more slowly exhumed after 48 Ma as they cooled below 100°C (Medford 1975; *see* Chapter 4). Brittle fabrics mantling the footwall rocks were likely formed below 350°C between 59 Ma and 48 Ma.

Given a period of denudation of approximately 10 Ma and a steady isothermal condition, the exhumation rate of the footwall of the Okanagan valley fault is estimated to be between 2 and 6 mma^{-1} (*see* Fig. 4.5). These values compare to the Holocene slip rates recorded on the Wasatch Fault zone, the longest normal fault zone (450 km) in the mid-Tertiary extensional system developed in the northwestern United States (Machette et al. 1991).

Unequivocal piercing points across the Okanagan valley fault have not been found in the field. The only valuable reference are two peculiar porphyritic plugs (Fig. 5.1) composed of large euhedral quartz and centimetric rhombohedral (twinned) orthoclase crystals trapped in an aphanitic felsic matrix. One of these bodies is exposed at Shingle Creek, 3.2 km southwest of

Penticton (Bostock 1941; Little 1961; Church 1973) in the hanging wall of the Okanagan valley fault. The second porphyry occurs 5.6 km south of Beaverdell (Little 1961; Fig. 5.1) and may be interpreted as a footwall unit because it seems to intrude the Okanagan complex. However, this body lies on the east side of the West Kettle River lineament (Fig. 5.1) and projects structurally above a Jurassic pluton (MBX39 in Chapter 4) itself part of the Carmi Klippe (*see* Chapter 4, p. 103). Thus we are probably not dealing with real hanging wall/footwall cutoffs.

The Shingle Creek porphyry has been dated at 51.2 ± 1.2 Ma (Soregaroli unpublished data) by K-Ar on biotite but the Beaverdell porphyry is undated. Both porphyries are undeformed and intrude Eocene trachytic flows of the Marron Formation (Church 1973; *see* Chapter 6). They are separated by 45 km along an azimuth trend of N285W that is exactly parallel to the extensional transport direction found in the gneisses of the Okanagan complex (*see* Chapter 2). Though the structural setting of the Beaverdell porphyry is unclear, it is tentatively proposed that it was juxtaposed to the Shingle Creek porphyry or that both porphyries were part of the same intrusive system before they were separated by the Okanagan valley fault. The present horizontal distance between the porphyries is therefore assumed to represent an apparent offset of the Okanagan valley fault and may only be an increment of the total displacement as both plugs may have been part of the same hanging wall block. Because there are no other unequivocal piercing points across the Okanagan valley fault that would relate to the period of ductile deformation it is proposed that 45 km of horizontal motion is a minimum value (Bardoux 1986).

Finding a good fit between hanging wall and footwall lithologies across the Okanagan valley fault is extremely difficult and may suggest indirectly that displacements were large. Accommodating 45 km of displacement forces the Okanagan batholith on the west side of the Okanagan valley to overlap the voluminous granitoids of the Okanagan complex (Fig. 5.1). Knowing that the Okanagan batholith is mainly Mesozoic (Roddick et al. 1972), the reconstruction would suggest that the granitoids underlying the Okanagan complex were in large part Mesozoic and may have been the rooting system of the Okanagan batholith. This interpretation is consistent with the discovery of Mesozoic zircon dates in granites of the Okanagan complex (Parkinson 1985) and would suggest, but does not prove, that Eocene

Figure 5.1: *Geological map of the Okanagan region indicating the location of porphyry bodies used as markers across the Okanagan valley fault (black spots south of Beaverdell and Penticton separated by arrow striking parallel to the mineral stretching direction in the Okanagan complex [OgC]). Also shown are sites of isotopic samples: star symbols for U-Pb sites (samples MX19E, MX261); asterisks for $^{40}\text{Ar}/^{39}\text{Ar}$ dates (samples MX78, MX48) and K-Ar dates (samples 32, 39).*

Miocene Plateau Basalts	Cretaceous Plutonic Complexes
Middle Eocene syenite	Jurassic Plutonic Complexes
Late Eocene dacites	Paleozoic metasediments and metavolcanics
Middle Eocene trachyandesites	Mylonitic gneiss
Eocene granodiorites	Denudation faults
	Normal faults

Figure 5.2: *Regional elevation maps of the northern Okanagan valley and the Slocan valley with cross-section of average elevation data. The topographic pattern of these extended regions differs though they both are slightly asymmetric on the trough side with a breakaway side high. The more asymmetric profile of the Okanagan valley suggest it has been affected by an extensional structure of greater magnitude of displacement than that underlying the Slocan valley (Wernicke 1990) which has an estimated displacement of 15-20 km.*

granitoids (Parkinson 1985; see Chapter 4) form a small volume of the Okanagan complex.

3. *Palinspastic restoration of the Okanagan valley fault*

The Okanagan valley is a prominent physiographic feature of the Canadian Cordillera (Mullins et al. 1990). Its physiographic relief reaches 2.3 km (Fig. 5.2a) and compares to the relief of severely extended regions of the Sierra Nevada and Death Valley in southern California (Wernicke 1990). The subdued asymmetric trough of the Okanagan valley contrasts with the more abrupt profile of the symmetric trough defined by the eastward-dipping Slocan valley fault on which 15-20 km of displacement has been proposed (Fig. 5.2b; Carr et al. 1987). According to Wernicke (1990), areas that have undergone the greatest extension should have a thinner crust and be topographically lower than surrounding unextended areas. Thus a genetic relationship may be established between large-scale tectonic denudation and the aspects of topographic depressions. In this perspective the Okanagan valley being deeper than the Slocan valley and asymmetric suggests that it is underlain by a fault that has more than 20 km of displacement.

The magnitude of down-dip displacement along the Okanagan valley fault can be tested by palinspastic restorations (Fig. 5.3). A first restored section [1] is drawn based on the poorly constrained horizontal offset of 45 km between porphyritic bodies cited in the preceeding section; a second restored section [2] is drawn based on equally poor estimates of a 90 km offset between Marron lavas and one of their probable parental magmas the Coryell syenites exposed mainly east of Beaverdell (Fig. 5.1; Tempelman-Kluit and Parkinson 1986; see Chapter 6). We assume that the geometry of the Okanagan valley fault is, as drawn from the LITHOPROBE seismic data (Cook et al. 1992), a shallowly dipping surface with at least three ramp/flat segments and we contend with Carr (1990) that the breakaway zone of the Okanagan valley fault is the cataclastic Beaven fault exposed 60 km east of Vernon (Fig. 5.1). The effects of small extensional structures such as the Cherryville fault (Carr 1990) are neglected.

Model 1 with a displacement of 45 km (Fig. 5.3) restores the hanging wall of the Okanagan valley fault to a pre-extensional position carrying its breakaway zone nearly 15 km above the present-day surface that is immediately above the breakaway zone of the Columbia River fault which recorded top-to-the-east sense of shear and formed simultaneously with the

Okanagan valley fault 56 Ma ago (Parrish et al. 1988; Lane et al. 1989; *see* Chapter 4).

Model 2 accommodates a displacement of 90 km (Fig. 5.3) carrying the hanging wall of the Okanagan valley fault approximately 25 km above the present-day surface and eastward of the Columbia River fault. In this situation the breakaway zone of the Okanagan valley fault lies above the Kootenay Arc that is east of the breakaway zone of the Slocan Lake fault another major extensional structure with top-to-the-east sense of shear (Carr 1987; Carr et al. 1987).

The first model has some validity because westward directed extensional shear indicators would lie systematically west of eastward directed extensional shear features which altogether would define a symmetric and nearly simultaneous spreading system (*see* Chapter 4). The second model is improbable because the hanging wall of the Okanagan valley fault crosses the Slocan Lake fault which is a slightly older (Parrish et al. 1988) or contemporaneous feature (*see* Chapter 4). In order to keep westward directed shear sense indicators west of the Slocan Lake fault the maximum down-dip displacement allowable on the Okanagan valley fault is 70 km (Fig. 5.3).

We may consider another aspect of both models of Figure 5.3. Restored sections assume that the pre-extensional lower crust was at least as thick as shown on the LITHOPROBE profile (Cook et al. 1992). This interpretation leads to the proposition that the pre-extensional crust of the southern Omineca Belt was at least 47 km thick and could have been as thick as 60 km. It can be demonstrated that at least 5 km of upper crustal rocks have been eroded from the southern Omineca Belt since the Eocene (Heller et al. 1987; Mack and Jerzykiewicz 1988). We may indirectly test both restorations by verifying the metamorphic assemblages preserved at a fixed level in the footwall of the Okanagan valley fault prior to extension assuming normal orogenic burial, rapid exhumation and normal geothermal gradients. The Okanagan complex, from which P-T data have been estimated (*see* Chapter 3), is chosen as a fixed reference level (Fig. 5.3). If the Okanagan valley fault had 90 km of motion then rocks of the Okanagan complex would have been buried down to at least 23 km reaching pre-extensional peak metamorphic conditions in excess of 700 MPa/700°C. In the case that motion on the Okanagan valley fault was 45 km the Okanagan complex would have been buried down to 15 km generating peak P-T conditions in the order of 500 MPa/500°C. The peak temperatures obtained from pelitic rocks in the Okanagan complex cannot resolve this problem because they range too broadly between 550-750°C due to

Figure 5.3:

Palinspastic reconstruction of the Okanagan valley fault (OVF). Restored section 1 assumes that the hanging wall has been displaced by 45 km. This would suggest that the breakaway, the Beaven Fault (Carr 1990; Cook et al. 1992) would have been near the position of the breakaway of the east-dipping Columbia River normal fault (CRF; Lane 1984; Lane et al. 1989). Restored section 2 assumes a displacement of the hanging wall of 90 km (Tempelman-Kluit and Parkinson 1986) and carries the breakaway east of the east-dipping Slocan Lake fault (SLF; Carr 1985; Carr et al. 1987). Because west-dipping extensional structures are expected to lie systematically west of east-dipping structures section 2 is an unlikely situation. This is consistent with the maximum P-T conditions recorded in the Okanagan complex which do not exceed 756°C and 600 MPa (see Chapter 3). The maximum displacement that can be accommodated with this scenario is 70 km.

reequilibration processes (*see* Chapter 3) but closer constraints on pressures suggest that rocks of the Okanagan complex reached peak pressures of 600 MPa. This data is compatible with a displacement on the Okanagan valley fault closer to 70 km. This value will be used to calculate the overall extension across the southern Omineca Belt in Chapter 8.

4. *Conclusions*

Because unequivocal piercing points are lacking across the cataclastic front of the Okanagan valley fault palinspastic restorations have been tentatively used to estimate the most probable hypothesis of the displacement of the Okanagan valley fault which cannot exceed 70 km and would be consistent with the average P-T data recorded in the Okanagan complex that is 650°C/600MPa. A displacement greater than 70 km would be inconsistent with the fact that extensional structures with top-to-the-west sense of shear indicators should lie systematically west of extensional structures with top-to-the-east sense of shear indicators forming together a symmetric and nearly contemporaneous spreading system above the apex of the compressive crustal welt underlying the Shuswap metamorphic complex.

CHAPTER 6 Stratigraphy, petrology and tectonic setting of the Eocene magmatic and sedimentary rocks of the Kelowna outlier

1. *Introduction*

The Early Tertiary tectonic history of south-central British Columbia is marked by the transition from a period of compression characterized by crustal shortening followed by Eocene extension (Brown and Carr 1990). The Eocene intermediate lavas of the Kelowna area were extruded mainly during and after the extension episode which also corresponds to a period of regional reorientation of the subduction regime of the Farallon-Kula ridge system (Tabor et al. 1984; Engebretson et al. 1985; Lonsdale 1988; Thorkelson and Taylor 1989).

The Kelowna lavas make up one of eight Eocene outliers stranded on top of the Okanagan valley extensional fault. The Kelowna lavas are underlain by cataclasites and mid-crustal rocks that provided whole-rock, hornblende, muscovite, and biotite K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ cooling dates ranging between 54 and 48 Ma (*see* Chapter 4). K-Ar ages provided herein suggest that the Kelowna lavas crystallized between 54 and 51 Ma. This data indicates that volcanism in the hanging wall of the Okanagan valley fault was coeval with the ductile shearing and cooling of the mid-crustal rocks underlying the fault. The presence of cataclasites underneath most of the Kelowna outliers indicates that they have moved along brittle surfaces after their extrusion.

This chapter summarizes the petrologic, stratigraphic, and structural data that pertains to the onset of Eocene volcanism in the Okanagan region. It is proposed that volcanism was directly related to partial melting processes occurring simultaneously at mid-crustal depths. Rare-earth elements have not been analysed. We emphasize on major and trace elements Zr, Nb, Ti, Y considered relatively immobile during secondary alteration processes (Pearce and Norry 1979; Menzies and Seyfried 1979) to understand the evolution of the Kelowna volcanic field. A second objective concerns the relationship between Eocene sedimentological processes and the dynamics of the Okanagan valley fault. Rotation in individual fault blocks within the Kelowna outlier has not been sufficient to be detected by paleomagnetic data (*see* Chapter 7) but field observations clearly indicate that block faulting related to extension has outlasted sedimentation.

2. *Regional setting of the Okanagan valley*

Eocene intermediate lavas crop out in a broad northwest-trending swath of discontinuous exposures extending between Kamloops in south-central British Columbia to Republic in northeast Washington (Muessig 1967; Church 1973, 1978, 1979a,b,c, 1981, 1982b; Pearson and Obradovich 1977; Ewing 1981; Thorkelson 1989). Most lava sequences range in thickness from 500m to 2 km and overlap high-grade metamorphic rocks that were exhumed by extensional faults in the Eocene. Eocene lavas commonly evolved from andesitic to felsic compositions and are interdigitated with thick alluvial sedimentary sequences (Church 1973; Ewing 1981). It has been recognized that altogether the Eocene rocks were deposited in association with coeval strike-slip faulting (Ewing 1981; Taylor et al. 1988; Thorkelson 1989) and extensional faulting (McMechan 1983; Bardoux 1985; Tempelman-Kluit and Parkinson 1986; Parrish et al. 1988).

The Kelowna area exhibits three volcanic outliers of Eocene lavas (Fig. 6.1) and is one of very few regions in south-central British Columbia where Miocene to Recent Columbia Plateau type basalts are preserved (Church 1982a). Only the Eocene volcanic suite bears a temporal and structural relationship with the Okanagan valley fault but Bardoux and Irving (1989) demonstrated that these units are autochthonous with respect to North America (*see* Chapter 7).

Besides Eocene lavas the hanging wall of the Okanagan valley fault comprise an assemblage of Late Paleozoic low-grade rocks of Quesnellia arc terrane (Monger et al. 1982). These rocks are of the Intermontane Superterrane that accreted onto North America in the Jurassic (Monger et al. 1982) and was invaded by calc-alkaline batholiths. The high-grade gneisses underlying the Okanagan valley fault are part of the Omineca Belt (Brown et al. 1986) and are characteristically overprinted by mylonitic shear fabrics dipping at approximately 15° to the west (*see* Chapter 2). Mylonitic fabrics overprint older compressional structures and they are systematically brecciated at the contact with rocks of the Intermontane Superterrane and/or Eocene volcanic units (Fig. 6.1).

The Okanagan valley fault is the largest of an array of easterly and westerly dipping normal faults that have denuded both flanks of the Monashee metamorphic complex, a major domal culmination exposed north of Kelowna (Brown and Journeay 1987; Johnson 1989b; Brown et al. 1992). The southern segment of the Okanagan valley fault straddles the west flank of the

Figure 6.1: *Geological map of the Kelowna area. Rocks in the hanging wall of the Okanagan valley fault comprise: (1) Miocene to Pleistocene plateau basalts; (2) Rhyodacitic to rhyolitic lavas and tuffs; (3) Alluvial sedimentary rocks correlated with the White Lake Formation (Church 1973); (4) Trachytic lavas correlated with the Nimpit Lake member of the Marron Formation (Church 1973); (5) Trachyandesitic lavas correlated with the Kitley Lake member of the Marron Formation (Church 1973); (6) Andesitic lavas; (7) the Trépanier dacite; (8) the Trépanier porphyry; (*) mudflow deposits; (9,10) Mesozoic biotite-granodiorite intruding Late Paleozoic metasediments. Rocks in the footwall of the Okanagan valley fault comprise: (A) ?Eocene hornblende-biotite granite and Mesozoic biotite-hornblende granodiorite; (B) mylonitic gneisses; (C) rhyolitic plugs.*

119°30'

119°00'

50°

Fig.2

"Colville Batholith" (Waters and Krauskopf 1941) composed of variably strained plutons which provided Early Tertiary cooling and crystallization ages (Holder and McCarley-Holder 1988). Further south the fault disappears beneath the thick sequence of Miocene Columbia River Plateau basalts (Mangan et al. 1986) which probably masks Eocene extensional features that may have served as feeders to the Miocene lavas.

3. *Stratigraphy of the Kelowna outlier*

Because they are cut by shallow to moderately dipping northerly trending brittle faults (Fig. 6.1), a composite section of Eocene rocks has been established near Kelowna and will be collectively referred to as the Kelowna outlier (Figs. 6.1, 6.2). The oldest Eocene unit exposed near Trepanier consists of basal mudflows preserved in three decametric graben structures bounded by right-lateral and normal faults which cut into Mesozoic granodiorites. The clasts in the mudflows are rounded and a few centimeters in diameter and essentially are granitic and low-grade metavolcanic pebbles derived from the nearby basement lithologies. It is noticeable that the clasts are systematically fractured or split in angular fragments suggesting that a certain tectonic activity preceeded the main volcanic event. Because of their stratigraphic position in the Eocene sequence, the mudflows may be tentatively correlated with the Tranquille Formation which underlies the Kamloops volcanic suite north of Kelowna (Ewing 1981).

Approximately 200m of thinly laminated dacitic flows, referred to as the Trepanier dacite, overlie the basal mudflows (Figs. 6.1, 6.2). The Trepanier dacite is only exposed in a lenticular body tapering eastward near Westbank (Figs. 6.1, 6.3) and extending westward above another lenticular body of quartz-oligoclase porphyry referred to as the Trepanier porphyry. Several dikes of plagioclase + augite phyric andesites and trachyandesites cut the Trepanier dacite along highway 97. These dykes feed approximately 1500m of thick shield-type andesitic-trachyandesitic flows overlying the Trepanier dacite to the south and pre-Tertiary rocks to the north (Fig. 6.1). These lavas are correlated with the Kitley Lake Member (Church 1973, 1979b, 1980a) and grade up-section into hematitic scoriaceous trachytes correlated with the Nimpit Lake Member (Church 1973, 1979b, 1980a). The latter two units are part of the Marron Formation as defined in the type-sections of the Penticton outlier (Church 1973) and also correlate with the Sanpoil lavas in

Figure 6.2: *Schematic stratigraphic section of the Kelowna outlier (Church 1980a, 1982b; this study). Notice that the basal mudflow unit is confined to small graben and systematic fracturing of the mudflow fragments suggest syn-tectonic deposition. The Black Knight sill complex is interpreted to have intruded into unconsolidated clastic strata correlated with the White Lake Formation (Church 1978) creating "invasive flow" structures (Byerly and Swanson 1978). The top of the Black Knight sill complex is composed of a dome shaped dacitic body. Trachyandesites of the Kitley Lake Member, the Tréponier orthoclase-quartz porphyry and dacitic dykes feeding the Joe Rich dacite are dated 53.6 ± 2.2 Ma, 50.5 ± 1 Ma, and 48.6 ± 0.7 Ma respectively by K-Ar method on biotite (see Chapter 4). Miocene to Pleistocene plateau basalts are locally underlain by fluvial uranium-rich sedimentary rocks (Boyle 1982).*

NE Washington (Muessig 1967; Pearson and Obradovich 1977). Altogether the Marron and Sanpoil lavas form the most continuous and most widespread lithotype of the Eocene volcanic field in south-central British Columbia and NE Washington (op. cit.).

The Marron lavas near Kelowna are overlain by several coarsening upward alluvial sequences which occupy the same stratigraphic position, and therefore are correlated with, the White Lake Formation exposed in the Penticton outlier (Church 1973).

Dacitic and rhyodacitic lavas are only exposed on the east side of the Okanagan valley underlying Mount Dillworth, Black Knight mountain and the Joe Rich klippe (Figs. 6.1, 6.2; Map 1). These felsic flows possess a distinctive reversed paleomagnetic polarity and are tentatively correlated with the Marama Formation (Church 1973; *see* Chapter 7). The Joe Rich klippe at the eastern limit of the Kelowna map sheet (Fig. 6.1; Map 1) exhibits at least 400m of felsic lavas comprising laminated and columnar jointed flows interdigitated with thin sub-horizontal tuffaceous beds. These rocks lie in observed disconformity above a wedged sequence of clast supported conglomerates. These conglomerates are correlated with the White Lake Formation because they overlie an attenuated sequence (200m) of red-weathering trachyandesites correlated with the Kitley Lake Member. The latter lies, also in observed disconformity, above limestone breccias correlated with the Middle Triassic Brooklyn Formation (Little 1983; *see* Chapter 2). The breccias form the basal unit of the Joe Rich klippe and rest in fault contact with high-grade mylonitic gneisses of the Okanagan complex. From their stratigraphic position in the Joe Rich klippe the rhyodacites are interpreted as the youngest unit of the Kelowna outlier. This stratigraphic relationship is consistent with the presence of an important swarm of dacitic plugs (MBX4 in Table 6.1) cutting across mylonitic rocks associated with the Okanagan valley fault. Most of these plugs occur in the neighbourhood of the felsic lavas capping the Joe Rich klippe suggesting that these flows have been little displaced from their feeding source (Fig. 6.1). However, two felsic outliers lying north of the Joe Rich klippe are underlain by cataclastic gneisses of the Okanagan complex which indicates that felsic lavas have been tectonically displaced. Perhaps the absence of mafic dikes cutting the gneisses of the Okanagan complex in the Kelowna map sheet indicates that the volcanic outliers exposed near and west of Kelowna (made mainly of trachyandesites) have been displaced by a distance exceeding the width of the

Kelowna map sheet, that is more than 40 km (Fig. 6.1).

Eocene syenites and diorites collectively known as the Coryell suite (Little 1961) have been interpreted to be the intrusive equivalents to the Marron lavas (Church 1973). No plutons of this paragenesis occur in the Kelowna map sheet (Fig. 6.1). Further east on the Kettle River (east half) map sheet (Little 1957) the nearest syenitic body from Kelowna is located 10 km east of Big White mountain. Assuming that the Big White syenite represents the root of the Kelowna outlier which remains a matter of conjecture (Tempelman-Kluit and Parkinson 1986; Bardoux 1986; *see also* Okulitch 1987; *see* Chapter 5) this would suggest that a minimum post-Marron displacement on the Okanagan valley fault was in the range of 60 km.

4. *Geochronology*

Biotites from three lava types of the Kelowna outlier have been dated using conventional K-Ar techniques to establish a stratigraphic sequence and tight constraints on timing between extensional faulting and the evolution of the Eocene lithotypes. Samples are located in Figure 6.1 and described in Appendix 3. The data are summarized in Table 6.1 are of good quality as shown by the small percentages of atmospheric Ar.

TABLE 6.1. K-AR analytical data from the Kelowna lavas

UNIT ^a	Sample	Min.	%Atm. Ar	Ar% K $\pm 1\sigma$	Rd.Ar $\pm 1\sigma$	Age $\pm 2\sigma$
Kitley Lake Member	MBX2	bi	5.5	6.7 ± 1.7	141.4 ± 1.2	53.6 ± 2.2
Qz-Feld. porphyry ^b	MBX1	bi	5.0	6.9 ± 0.8	137.8 ± 0.6	50.5 ± 1.0
Dacite ^c	MBX4	bi	2.2	7.6 ± 0.4	144.9 ± 0.6	48.6 ± 0.7

^a: Sample localities; ^b: Mount Law, ^c: Trepanier, ^d: Joe Rich Creek

The oldest rocks of the Kelowna outlier are the trachyandesites of the Kitley Lake Member which provided an age of 53.6 ± 2.2 Ma on fresh biotite interpreted as an age of crystallization. This duplicates a K-Ar date (53 ± 2 Ma; Church 1979b) obtained on biotite of a similar unit

collected north of Trepanier (Fig. 6.1). The Trepanier quartz-oligoclase porphyry (Fig. 6.1) provided a K-Ar date of 50.5 ± 1.0 Ma on fresh biotite and also duplicates the age obtained by Church (1979b) on altered biotites from the same unit. Because the porphyry underlies the trachyandesites of the Kitley Lake Member, it is interpreted to be a laccolith of post-Marron age and may be considered as coeval, within error limits, to the Shingle Creek porphyry, exposed near Penticton, in which biotites provided a K-Ar age of 52 ± 2 Ma (Church 1979b). Oligoclase porphyries correlated with the Coryell suite in the Granby Creek area, 60 km southeast of Kelowna, were dated at 51.1 ± 0.5 Ma by U-Pb on zircon (Carr and Parkinson 1989). Perhaps the porphyritic bodies straddling the Okanagan valley, such as MBX1, can be related to the Coryell suite and may represent shallow apophyses. It is noticeable then that these rocks intruded 2-3 Ma after the trachytic lavas of the Marron Formation were cooled below 280°C (the closure temperature for Ar in biotite; McDougall and Harrison 1988). The younger age of the Trepanier porphyry is somewhat inconsistent with the parentage established between Marron and Coryell rocks (Church 1973). However, it is consistent with the fact that it is characterized by a reversed magnetic polarity as opposed to a normal polarity for the Marron lavas (*see* Chapter 7).

A dike of rhyodacitic composition similar to those surrounding the Joe Rich klippe yields a K-Ar age on biotite of 48.6 ± 0.7 Ma which is consistent with an uncorrected 46 ± 2 Ma age provided by biotites of a rhyodacitic lava flow sampled on top of the Joe Rich klippe (Mathews unpublished data 1964). The crystallization age of this dike confirms the stratigraphic relationships established in the field and suggests that the felsic lavas of the Kelowna suite are nearly coeval with the estimated age of extrusion of most lavas of the Kamloops Group that are exposed 100 km north of Kelowna (Ewing 1981a). Proposing that the rhyodacitic lavas of the Joe Rich area are younger than the rest of the Kelowna lavas is consistent with the fact that they yield a reversed magnetic polarity (*see* Chapter 7).

5. *The alluvial sequences*

The sedimentary rocks correlated with the White Lake Formation (Church 1973) include poorly indurated, well rounded and well bedded polymictic conglomerate and sandstone strata exposed in three localities near Kelowna. The Westbank basin comprises an estimated section

Figure 6.3: *Map of the Westbank basin and schematic section of the sedimentary sequence. Site numbers indicate type-localities of members of the composite section discussed in the text and positioned on the right margin of the section. Scale bars on the schematic stratigraphic column are at intervals of 100m.*

- Mylonitic gneiss
- Eocene Alluvial sediments
- Eocene Trachyandesites
- Eocene Rhyodacites
- Typical sections
- Okanagan Valley detachment

of 1.2 km of thinly laminated to massive clastic strata spread over an area of about 50 km². Similar strata occur locally along Mission Creek and in the Joe Rich klippe (Fig. 6.1).

5.1 Stratigraphy

Five subdivisions were established in the stratigraphic section of the Westbank basin (Fig. 6.3). Exposed at the junction of highway 97 and Glenrose road south of Westbank (site 1, Fig. 6.3), the lower 200m of the section is dominated by crudely stratified conglomerate and volcanic debris flow with clasts up to 30cm in diameter. These rocks are interbedded with and fine upward to mudstone and thin coal seams containing fern-like leaf fossils and needle-bearing branch fossils. The conglomerates represent proximal stream flow, and some debris-flow units (Rust and Koster 1984). The only paleocurrent direction noted in this part of the section is easterly.

The second unit in the composite section is exposed in Powers Creek (site 2) and consists of approximately 200m of sandy conglomerate, sandstone and mudstone. These sedimentary rocks represent stream flow, debris flow and channel deposits of the middle and outer parts of an alluvial fan. Some conglomerates are matrix rich and suggest debris-flow deposition while small fining upward sequences of sandstone represent channel-fill or flood deposits. The succeeding 300m are dominated by stratified sandy conglomerate well exposed south of Mount Boucherie (site 3). These sediments represent stream flow and debris flow deposits of the proximal parts of an alluvial fan. Clast size in these conglomerates is bimodal ranging either in the 1 cm or 5 cm sizes. East of Mount Boucherie another fossiliferous sandstone-mudstone sequence occurs (site 4) that is interpreted as a younger unit representing channel-fill or flood deposits that gradually coarsen upward into a sandy conglomerate at Kalamoir Park (site 5). The Kalamoir Park unit is the youngest exposed sedimentary section representing the top of the Westbank basin. It is composed of matrix supported debris-flows that contain boulders nearly one meter in diameter, by far the largest in the clastic sequence.

Partial exposures of sedimentary strata along Mission Creek and in the Joe Rich klippe preclude their correlation with the subdivisions established in the Westbank basin.

5.2 Sedimentological environment

Sedimentary sequences similar to the White Lake Formation are documented throughout the Eocene successions of south-central British Columbia (Church 1973, 1979a; Ewing 1981a). Although compositions vary between sites in this vast region it is proposed that the sedimentary sequences originate from a system of braided-stream deposits interdigitated with debris flow which formed thick alluvial fans and subaqueous deltas bounding relatively high relief. This period of sedimentation occurred during a magmatic lull as there are no volcanic flows interdigitated with the sedimentary rocks.

The systematic variation of clast types and their relative proportions in the Westbank basin are guides of the erosional process. Debris derived from the Marron Formation predominates at the bottom of the section whereas Mesozoic granitoid boulders and Paleozoic greenstone clasts are abundant slightly above the lower contact. All these clasts are likely derived from rocks lying west of the basin. Their large sizes and angularity indicate that the sources were proximal and that erosion was probably rapid suggesting that the volcanic edifice and the basement underneath the Westbank basin were at higher relative elevations than presently. Clasts of Paleozoic and Mesozoic rocks become increasingly abundant higher in the section indicating the volcanic edifice was rapidly eroded. Above the sandy middle subdivision, conglomerates become gradually coarser grained and contain up to 95% undeformed plutonic clasts (site 3, Fig. 6.3). Some of these clasts are foliated and may be derived from strained granitoids underlying the Okanagan complex (Fig. 6.1) suggesting simultaneous erosion of both hanging wall and footwall of the Okanagan valley fault. The time of deposition of the upper part of the Westbank basin is constrained by palynological assemblages which yield an age no older than Late Middle Eocene (sites 3 and 4 in Fig. 6.3; McIntyre pers. comm. 1986). Strained granitic clasts and chloritized mylonitic clasts become increasingly more abundant, though still extremely rare (5-10%), in the upper subdivision at Kalamo Park (site 5, Fig. 6.3). This indicates a gradual change and the addition of the mylonitic gneisses underlying the Okanagan valley fault as a new source area.

Very few large clasts of retrograded breccia or cataclastic mylonite were found in the Westbank basin. This could be partly due to the high friability of such clasts that would rapidly be fragmented to clay sizes upon transport and easily evacuated along the drainage pattern and

carried out of the source area. The few breccia clasts seen around Mount Boucherie (site 4 in Fig. 6.3) suggest the beginning of the unroofing of the retrograded cataclastic zone mantling the Okanagan valley fault near the end of the sedimentary cycle.

Six paleocurrent directions based on cross-bedding were measured in sandstone beds of the lower and middle divisions of the Westbank basin. These are interpreted to have been dominantly to the northeast throughout most of the section suggesting that the Intermontane Belt may have been high during this period. Paleocurrent directions seem to have changed gradually towards a southwesterly direction as the Okanagan complex breached the surface.

5.3 Paleodepth of the Kelowna outlier

Vitrinite reflectances obtained from coal seams in the Westbank basin are $> 1.0\%$ near the surface trace of the Okanagan valley fault decreasing rapidly westward and away from the fault trace (Mathews and Bustin 1985). These data have been interpreted to be the result of abnormally high thermal conditions created by hydrothermal fluids circulating along the permeable cataclastic zone of the Okanagan valley fault. It has also been proposed that the vitrinite maturity was caused by surface coal fires or by magmatic reheating related to Neogene volcanism that spread over paleodepressions in the Kelowna area (Mathews 1987). Alternatively it may be proposed that under conditions of normal burial, the reflectance values in the White Lake Formation would indicate an Eocene paleotemperature of nearly 100°C equivalent of 3.5 km of overburden when a continuous compactional geothermal gradient is assumed (Karweil 1956; Dow and O'Connor 1982). These thermal conditions are not improbable as the Eocene lavas underlying the sedimentary rocks contain laumontite crystals which indicate a certain degree of post-depositional low-grade metamorphism. It is, however, impossible to estimate the nature of the overburden which has been eroded during the last 45 Ma.

5.4 Syn-extensional sedimentation

Strata in the Westbank basin and those in the Joe Rich klippe generally dip 20° eastward. Alluvial fans are rarely flat lying but it is suggested that the Eocene beds have rotated down to the east from a sub-horizontal position into the plane of the Okanagan valley fault during

Figure 6.4: *Sketch looking south showing steeply, westerly inclined and folded sandstone and conglomerates of the White Lake Formation along Mission Creek. Deformation was probably caused by the magmatic injection of the Black Knight dacitic sill complex (see Figure 6.2) lying in the treed area immediately west of this site. Sandstone dikes were injected perpendicular to bedding during and after deformation (not shown). View looking south, trees in background are approximately 15 m high.*

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extension (Fig. 6.3). Strata of the Mission Creek section are folded and dip at steep angles to the west opposite to the Westbank strata (Fig. 6.4). These features together with syn-sedimentary deformation structures (slump folds and sandstone dikes) may suggest that syn-depositional tectonic activity affected these rocks. Alternatively, it is proposed that thick sill-like bodies lying east of the Mission Creek strata jostled them while they were still unconsolidated creating so-called "invasive flow" structures (Byerly and Swanson 1978). This scenario would imply that some volcanism occurred after or during lacustrine sedimentation which is consistent with the presence of younger felsic flows east of Kelowna and in the Joe Rich klippe.

The conglomerates of the Westbank basin show an inverse stratigraphy of clast types starting at the bottom with hanging wall derived boulders gradually decreasing in percentage up section and being replaced, in part, by footwall derived clasts at the top of the section. Postulated unconformities within the Eocene section and the repetition of coarsening-upward sequences in the conglomerates suggest a pulsating history of faulting and deposition. Sedimentary transport during the denudation process seems to have been in the direction of tectonic transport only in the upper parts of the sedimentary section.

Sedimentary facies and distinctive clasts within the Westbank conglomerates can put some constraints on the amount of displacement along the Okanagan valley fault during the latest phase of movement. Most fan facies suggest that clasts are deposited within 2 or 3 km of the fan apex. The radius of a fan is generally less than 5 km which is consistent with the clast size distribution (Rust and Koster 1984). Clasts of hornblende-granodiorite in the middle part of the Westbank basin near Mount Boucherie are now at least 20 km west of their postulated footwall source at Little White mountain. Allowing for a few kilometers of sedimentary transport, these relations broadly suggest that the amount of transport during later displacement of the Okanagan valley fault was in the order of 10 to 15 km. The rareness of mylonitic clasts in the Eocene strata also suggests that the mylonitic footwall of the Okanagan valley fault breached the surface at a very late time of the extensional history.

Similar syn-extensional sedimentary sequences and evolutions have been documented for metamorphic core complexes in the western U. S. (Spencer and Reynolds 1986; Miller and John 1988) where mylonitic and cataclastic clasts derived from detachment faults are better preserved.

5.5 Sedimentary output versus extension modes and rates

Crystallization ages of trachytic lavas underlying (54 Ma) and dacitic lavas overlying (48 Ma) the Eocene sedimentary rocks near Kelowna suggest that 1.2 km of compacted strata were deposited at a depositional rate in the order of 2 m'a⁻¹ that is typical of regions with high relief.

The volume of sedimentary rocks preserved in the alluvial sequences correlatable with the White Lake Formation throughout south-central British Columbia is estimated at 500 km³. This represents approximately 5% of the volume of rocks involved in the extensional zone comprised between the Shuswap metamorphic complex in the north (Brown et al. 1986) and the Colville complex in the south (Holder and McCarley-Holder 1988). Large volumes of syn-extensional detritus have been carried away from the extended area towards the Cordilleran foreland (Mack and Jerzikiewicz 1989) and along the west coast of Oregon, Washington, British Columbia, and further into the Pacific. The latter regions exhibit voluminous Eocene continental and marine deposits, up to 7 km thick, derived from metamorphic source terranes (Johnson 1984; Heller et al. 1987; Tabor et al. 1984; Underwood and Bachman 1986; Taylor et al. 1988). An estimate of 7 km of Eocene strata in Washington is probably conservative as it does not account for compaction and omits post-Eocene erosion, which redeposited Eocene detritus in younger basins.

It appears that during early stages of extension the southern Omineca Belt was high since Early Eocene sedimentary rocks have been deposited outside this region and perhaps outside the Intermontane Belt as well. This suggests that the Omineca and Intermontane Belts may have formed a high plateau region in the Early Eocene. The collapse of the plateau seems to have occurred by denudation and erosion. Coarser sedimentary rocks were collected in individual basins next to extensional faults as the plateau was progressively returning to a lower physiography. The paleocurrent data in the Early Tertiary sequences in south-central British Columbia are insufficient to demonstrate drainage reversals between pre-extensional and syn-extensional physiographies. It is expected that if a pre-extensional "plateau" existed there would have been a drainage divide from which paleocurrents would have drained in opposite directions away from the high region. Conversely if the "plateau" collapsed paleocurrent directions would be expected to have been in random orientations or converging toward physiographic depressions.

6. *Setting of the Eocene lavas*

Compositions of Eocene lavas closely resemble those of intracontinental calc-alkaline suites of northern Chile (Francis et al. 1984; Hildreth and Moorbath 1988) and Greece (Mann 1983). Though they may be back-arc related (Thorkelson 1989) elemental ratios exhibit mainly large crustal contributions to the magma genesis.

Eocene lavas in south-central British Columbia and north-central Washington are found discontinuously across a 300 km wide area located nearly 500 km inboard of the present continental margin. Eocene lavas were emplaced within the Omineca Belt while it was denuded by large-magnitude extension that culminated approximately 52 Ma ago (see Chapter 4). This syn-extensional magmatism lacks primitive basalt compositions. It is distinct from the basaltic and/or bimodal series formed during late stages of extension in the Basin and Range Province (Christiansen and Lipman 1972; Suneson and Luchitta 1983) as well as the alkalic, silica-undersaturated suites of typical continental rifts (Gass 1970; Neumann and Ramberg 1978).

6.1 *Petrography of the Kelowna lavas*

Eocene volcanism in the Kelowna region exhibits some interesting time-composition-volume trends. Early, small-volume flows are mainly felsic. Subsequently more voluminous intermediate flows of the Marron Formation (Church 1973) were extruded but final eruptions became felsic again. Of the total volume of Eocene lavas exposed near Kelowna about 10% are andesites, 70% are trachyandesites, 15% are felsites and 5% are quartz-oligoclase porphyries (Fig. 6.1). Volcanic compositions range from basaltic andesite with 54% SiO₂, to rhyolite with 73% SiO₂, with a few compositional gaps at the 65 and 70% SiO₂ (Fig. 6.5).

Despite the indication of low to medium temperature hydrous alteration (δO^{18} values per mil) along the Okanagan valley (Nesbitt and Muehlenbachs 1989) the textures seen in thin section are relatively unaltered and preserve their trachytic aspects. Most mafic flows contain up to 20% univertual phenocrysts and/or clots of diopside $\pm$ hypersthene $\pm$ plagioclase which are generally zoned from An₈₀ to An₂₀ at their rims. Olivine pseudomorphs were observed in certain trachyandesites but they have been completely serpentinized while pyroxenes are intensely uralitized and carbonatized confirming that a certain degree of deuteric alteration affected these

Figure 6.5: *Top, Na₂O + K₂O/SiO₂ (wt%) variation diagram. KELOWNA SUITE: (□) intermediate lavas; (○) dacitic flows; (△) rhyolitic lavas and porphyries; KAMLOOPS SUITE: (■) intermediate lavas. (○) Miocene and Pliocene Plateau Basalts. Bottom, sample locations of Tertiary lavas in the Kelowna area. Lower Plate units refer to rocks underlying the Okanagan valley fault, unit 11: highly strained paragneisses and leucogneisses; 10: granitic bodies of unconstrained age. Upper Plate units refer to the hanging wall of the Okanagan valley fault. Units 9: Late Paleozoic metavolcanic and metasedimentary rocks of the Nicola Group; 8: Mesozoic plutons; 7: Trépanier dacite; 6: andesites; 5: trachyandesites of the Kitley Lake Member; 4: trachytes of the Nimpit Lake Member; 3: alluvial sedimentary rocks of the White Lake Formation; 2: dacites to rhyolites; 1: Miocene to Pliocene Plateau Basalts.*

rocks. Feldspar, biotite, and hornblende are xenocrystic phases in the felsic lavas which also contain up to 5% subhedral to partly resorbed quartz xenocrysts in a vitrophyric groundmass. Accessory minerals vary with the lava composition and consist respectively of apatite in trachytic rocks, magnetite and ilmenite in dacitic flows and titanite and zircon in porphyries.

The textures of xenocrysts in most of the acid lavas and especially the corrugated aspect of biotites and hornblendes indicate that they may have been inherited from metamorphic rocks such as those of the Okanagan complex (Fig. 6.1). The paragenetic sequence of the felsic lavas indicate these may be residual liquids derived from *in situ* melts such as the biotite-hornblende granitic bodies of the Ladybird suite interlayered in the Okanagan complex (Carr 1990; *see* Chapters 1, 4). The local appearance of K-feldspar in certain lavas (Church 1973) may further indicate variable source compositions, which is also consistent with an anatectic magma source.

6.2 Whole-rock geochemistry of the Eocene lavas

A total of 27 samples of the Kelowna and Kamloops suites were analysed for major, minor and trace elements (Rb, Th, Sr, Nb, Y, Zr) by XRF at McGill University and are listed in Tables 6.2 and 6.3 and located on Figure 6.5. REE were omitted but concentrations of the less mobile elements such as Nb, Zr, Ti, and Y shed some light on the genesis of the Eocene lavas (Winchester and Floyd 1977; Menzies and Seyfried 1979; Pearce 1982).

6.2.1 Major elements

The Kelowna andesite-dacite-rhyodacite suite forms coherent, roughly linear arrays from 53 to 73% SiO₂ (5.48 - 0.38% MgO; Tables 6.2, 6.3). Although mafic phases tend to plot near the tholeiitic/calc-alkaline boundary the felsic lavas plot in the calc-alkaline field on a SiO₂ vs FeO*/MgO diagram and straddle the alkaline/sub-alkaline division on the Na₂O+K₂O vs SiO₂ diagram (Fig. 6.5). Rocks of the Kelowna suite have the same general characteristics as those of the Kamloops suite though the latter are slightly more alkaline (Table 6.3; Fig. 6.5). The Kelowna andesite and dacite have higher MgO, TiO₂, Na₂O, and K₂O than orogenic calc-alkaline lavas with comparable SiO₂ contents (Gill 1981; Ewart 1982). The dacitic compositions and most of the trachyandesites fall within the high-K acid region of Gill's (1981) SiO₂ vs K₂O diagram.

TABLE 6.2. Major and trace element abundances in Eocene lavas of the Kilauea suite

SAMPLE	47	55	59	60	62	67	69	130	151	152	21	33	49	75
SiO ₂	68.21	56.68	57.58	72.91	54.87	73.30	68.83	68.15	68.28	68.56	72.42	55.64	55.13	55.38
TiO ₂	0.72	1.00	0.95	0.31	1.07	0.32	0.70	0.54	0.58	0.68	0.34	1.00	0.91	0.81
Al ₂ O ₃	15.23	15.94	15.52	14.38	14.85	14.92	14.93	16.41	15.00	14.63	14.68	14.99	18.04	15.21
FeO	4.31	7.86	6.98	2.51	10.02	1.65	4.18	3.34	3.70	4.36	2.14	8.26	6.71	5.44
MgO	1.18	3.68	4.89	0.56	5.50	0.38	0.96	0.91	0.98	0.68	0.57	5.42	2.64	2.53
MnO	0.06	0.12	0.10	0.05	0.15	0.02	0.06	0.07	0.07	0.11	0.03	0.11	0.11	0.10
CaO	3.13	6.34	4.34	1.92	7.53	1.48	1.05	2.29	3.17	2.77	0.96	7.03	6.84	5.18
Na ₂ O	2.87	2.23	3.20	2.67	2.06	3.28	2.78	2.89	3.09	2.43	2.85	1.92	2.24	3.04
K ₂ O	3.35	3.51	3.57	3.66	2.83	3.63	4.43	4.51	1.92	3.84	4.74	3.96	3.67	1.47
P ₂ O ₅	0.26	0.56	0.55	0.10	0.48	0.10	0.24	0.16	0.21	0.25	0.13	0.59	0.47	0.42
LOI	0.82	2.29	2.61	0.64	0.94	0.61	1.84	0.51	2.77	1.81	0.85	1.68	3.31	3.51
TOTAL	100.14	100.11	100.29	99.71	100.30	99.69	100.00	99.78	99.77	100.12	99.71	100.63	100.07	100.09
NORMS														
q	27.66	7.82	5.07	36.73	3.74	35.52	29.61	25.67	32.49	30.18	33.42	3.72	6.20	21.24
or	19.51	20.77	21.15	21.67	16.76	21.50	26.22	26.71	11.38	22.72	28.08	23.45	21.72	8.72
ab	24.28	18.87	27.07	22.59	17.43	27.75	23.52	24.45	26.14	20.56	24.11	16.24	18.95	25.72
an	14.24	23.13	17.47	8.97	22.93	7.12	3.73	10.58	14.50	12.23	4.26	20.60	28.38	23.21
di	—	4.03	0.71	—	9.67	—	—	—	—	—	—	8.93	2.49	—
by	9.77	20.14	23.22	5.58	25.77	3.48	9.02	7.64	8.41	8.78	4.84	22.83	16.33	15.13
il	1.37	1.90	1.80	0.59	2.03	0.61	1.33	1.03	1.10	1.29	0.65	1.90	1.73	1.54
ap	0.62	1.33	1.31	0.24	1.14	0.24	0.57	0.38	0.50	0.59	0.32	1.40	1.12	1.00
c	1.68	—	—	2.74	—	—	4.19	2.90	2.53	1.99	3.31	—	—	0.12
zr	0.04	0.05	0.04	0.04	0.04	0.04	0.05	0.05	0.05	0.05	0.04	0.04	0.05	0.06
cr	—	0.01	0.07	0.01	0.03	0.01	0.01	—	—	—	—	0.06	0.02	0.02
TOTAL	99.51	98.05	97.93	99.15	99.54	99.27	98.25	99.42	97.10	98.41	99.03	99.18	96.99	96.74
Rb	93	65	121	91	79	123	92	132	80	72	157	117	75	71
K	27809	20138	29636	30383	23493	30134	36775	37439	15938	31877	39348	32873	30466	12203
Ba	1342	2098	1877	1165	1658	1235	1916	2307	1221	1182	1768	2021	2080	1559
Pb	9	5	<5	8	9	8	6	10	5	13	12	12	15	5
Ta	12	<5	7	5	<5	15	<5	18	<5	<5	23	11	10	<5
U	9	9	10	8	7	11	<5	7	8	7	9	8	13	9
Nb	32	18	19	10	24	12	17	24	15	17	17	33	34	18
Sr	1119	399	1084	275	855	1117	260	723	390	343	963	986	1294	688
Zr	221	256	206	192	191	199	249	284	252	244	205	220	237	276
Ti	4316	5995	5695	1858	6415	1918	4197	3237	3477	4077	2038	5995	5455	4856
Y	25	31	20	24	26	10	28	21	29	28	13	25	23	31
V	80	151	166	30	225	34	91	42	79	86	28	174	138	115
Cr	102	56	328	<10	149	<10	25	<10	<10	<10	<10	291	92	88
Ni	100	<10	78	<10	17	<10	34	<10	<10	<10	<10	49	<10	<10
Al ₂ O ₃ /TiO ₂	21.2	13.9	16.3	46.4	13.9	46.6	21.3	30.4	25.9	21.5	43.2	15.0	19.8	18.8
CaO/TiO ₂	4.4	6.3	4.6	6.2	7.0	4.6	1.5	4.2	5.5	4.1	2.8	7.0	7.5	6.4
TiO ₂ /P ₂ O ₅	2.8	1.8	1.7	3.1	2.2	3.2	2.9	3.4	2.7	2.7	2.6	1.7	1.9	1.9
Rb/Sr	0.1	0.2	0.1	0.3	0.1	0.1	0.4	0.2	0.2	0.2	0.2	0.1	0.1	0.1
Ti/Zr	19.5	23.4	27.7	9.7	33.6	9.6	16.9	12.3	13.8	16.7	9.9	27.3	23.0	17.6
Ti/Y	172.6	293.4	284.8	77.4	246.7	191.8	149.9	154.1	119.9	145.6	156.8	239.8	237.2	156.6
K/Rb	6.4	4.9	5.2	16.4	3.7	15.7	8.8	11.6	4.6	7.8	19.3	5.5	5.6	2.5
K/Rb	299.0	448.3	245.0	333.9	297.4	245.0	399.7	283.6	199.2	442.7	250.7	281.0	406.2	171.9
Zr/Y	8.8	8.2	10.3	8.0	7.4	19.9	8.8	10.2	8.7	8.7	15.8	8.8	10.3	8.9
Zr/Nb	6.9	14.2	10.8	19.2	8.0	16.6	14.7	11.0	16.8	14.4	12.1	6.7	7.0	15.5
(Zr/Nb) _N	0.48	0.99	0.76	1.34	0.6	1.2	1.0	0.8	1.2	1.0	0.8	0.5	0.5	1.1

TABLE 6.3. Major element abundances in the Eocene lavas of the Kelowna suite

Major element abundances in the Eocene lavas of the Kelowna suite																	Kamloops suite																
SAMPLE	1	2	3	4	5	8	9	10	13	14	15	16	17	18	19																		
SiO ₂	68.84	63.18	67.18	55.17	56.49	58.95	54.25	54.92	59.61	62.49	62.86	61.04	62.19	57.91	49.17																		
TiO ₂	0.66	0.94	0.60	0.97	1.03	0.90	1.03	1.03	0.92	0.70	0.80	0.77	0.73	1.00	1.83																		
Al ₂ O ₃	14.52	14.90	15.40	18.06	17.93	15.50	15.46	15.96	17.85	17.97	15.23	16.81	16.74	16.81	15.21																		
FeO	3.37	4.55	3.46	6.92	5.71	4.51	7.07	6.93	4.67	3.32	4.73	4.92	4.53	5.24	10.32																		
MgO	0.53	2.24	1.07	4.89	2.65	1.82	6.08	5.63	1.66	0.84	2.57	2.16	1.95	3.18	7.68																		
MnO	0.06	0.10	0.07	0.10	0.06	0.07	0.12	0.11	0.07	0.07	0.09	0.09	0.08	0.08	0.15																		
CaO	2.72	4.07	3.28	7.13	4.91	3.16	8.79	8.34	3.07	1.62	5.23	4.67	4.34	3.94	9.27																		
Na ₂ O	3.62	3.41	4.47	3.23	4.17	4.59	2.73	2.85	4.10	4.72	4.30	3.65	3.72	3.65	3.33																		
K ₂ O	3.81	2.51	1.93	4.68	4.60	6.18	2.74	3.12	6.67	6.96	1.54	4.22	4.39	6.05	0.95																		
Cr ₂ O ₃	0.00	0.00	0.00	0.03	0.01	0.00	0.02	0.02	0.00	0.00	0.01	0.00	0.00	0.01	0.04																		
P ₂ O ₅	0.24	0.55	0.21	0.59	0.50	0.43	0.49	0.48	0.37	0.23	0.42	0.35	0.33	0.52	0.34																		
LOI	1.78	3.57	2.51	1.36	2.11	1.05	0.00	0.00	0.82	0.74	2.79	1.37	1.17	1.41	1.39																		
TOTAL	100.15	100.02	99.57	99.93	100.30	99.59	98.81	98.89	99.81	99.66	100.57	100.05	100.18	99.80	99.68																		

NORMS

q	24.89	20.05	23.26	---	---	---	0.14	---	---	---	15.86	9.89	---	---	27.66	
or	22.52	14.84	11.41	---	27.19	36.53	16.19	18.44	39.42	41.14	9.10	25.95	35.76	5.61	19.84	
ab	30.63	28.85	37.82	---	35.28	38.84	23.10	24.11	34.69	39.94	36.38	31.47	30.88	28.17	24.28	
an	11.93	1.60	14.90	22.73	16.62	3.44	21.84	21.55	10.61	6.53	17.71	16.02	11.62	23.75	14.24	
di	---	---	---	---	3.73	7.91	15.22	13.71	1.86	---	4.57	2.82	3.75	16.47	---	
by	6.53	12.57	8.16	---	7.74	4.38	19.17	18.44	8.11	7.16	11.66	10.70	10.70	0.02	9.77	
il	1.25	1.79	1.14	1.84	1.96	1.71	1.96	1.96	1.75	1.33	1.52	1.39	1.90	3.48	1.37	
ap	0.57	1.30	0.50	1.40	1.18	1.02	1.16	1.14	0.88	0.54	0.99	0.78	1.23	0.81	0.62	
c	0.07	0.49	0.50	---	---	---	---	---	---	0.28	---	---	---	---	1.68	
ne	---	---	---	---	---	---	---	---	1.69	---	---	---	2.56	19.95	---	
cl	---	---	---	---	---	---	---	---	---	---	---	---	---	0.04	---	
TOTAL	98.38	96.48	97.68	---	98.08	96.13	98.79	99.39	99.01	98.83	97.79	99.02	98.41	98.27	99.51	
Al ₂ O ₃ /TiO ₂	22.0	15.8	20.5	18.6	17.4	17.2	15.0	15.5	19.4	25.7	19.0	21.8	22.9	16.8	8.3	
CaO/TiO ₂	4.1	4.3	5.4	7.4	4.0	3.5	8.5	8.1	3.3	2.3	6.5	6.1	5.9	3.9	5.1	
TiO ₂ /P ₂ O ₅	2.8	1.7	2.9	1.6	2.1	2.1	2.1	2.1	2.5	3.0	1.9	2.2	2.2	1.9	5.4	

Though no tholeiitic lavas were found in the Kelowna area, some of the olivine basalts of the Kamloops suite may represent potential parental compositions (Ewing 1981).

K-metasomatism is known to affect the alkali distribution in volcanic suites stranded atop extensional faults of the Basin and Range Province in Nevada (Kerrick 1988; Roddy et al. 1988). This is notably illustrated by anomalous K/Na ratios ranging from 27 to 117 where part of the K is inherited from the alkaline rocks underlying the Basin and Range extensional faults. Despite their proximity to the alkaline rocks of the Okanagan complex, the Kelowna lavas yield reasonably stable K/Na ratios ranging between .5 and 2 considered as primary values (Nockolds et al. 1978), suggesting that the Kelowna lavas were not severely metasomatised.

Despite a probable mobility of alkalis, the Eocene lavas of the Kelowna and Kamloops suites are grouped as a continuous calc-alkaline series on the AFM diagram (Fig. 6.6a). Eocene olivine basalts of the Salmon River suite (Church and Evans 1982) plot at the mafic end of this trend and may represent parental magma compositions.

The major element data of the Eocene suites may be tentatively discriminated on a multicationic element scheme $R1-R2$ (de la Roche et al. 1980; Fig. 6.6b) which shows that the Kelowna lavas (open symbols) follow a sequential mixing trend (A) confined to Field 2 (subduction regime). The arrow on trend (A) indicates the compositional evolution of the Kelowna sequence with time. This evolution may result from either, the crystallization of clinopyroxene, olivine, and calcic plagioclase from a tholeiitic basalt magma producing a magma of intermediate composition. Alternatively trend (A) may be produced by the periodic mixing of an amphibolitic restite with a felsic anatectic melt. The data from the Kamloops suite (diagonally half-filled symbols; compiled from this study and Ewing (1981)) and from the Penticton suite (black symbols; compiled from Church 1973) indicate slightly different compositional trends (Fig. 6.6b). Many lavas of the Kamloops suite follow the mixing trend of the Kelowna suite but some magmas define trend (B) lining up in Field 3 (Post-collision exhumation). These rocks probably represent another mixing line involving metasedimentary restites and anatectic melts (de la Roche et al. 1980). A third trend (C) is defined by lavas of the Penticton suite (Church 1973) and certain lavas of the Kamloops suite. These rocks follow and cross the silica saturation plane. Trend (C) indicates that fractionation of augite and/or alkali feldspar and/or biotite has occurred

Figure 6.6:

(a) AFM diagram comparing Eocene lavas of three different areas straddling the Okanagan valley fault (Penticton data from Church 1973; Kamloops data from this study and Ewing 1981; Salmon River Basalts data from Church and Evans 1982). Symbols are as defined in diagram (b) $R1-R2$ multicationic element scheme showing a compilation of most Eocene lavas along the Okanagan valley. Tectonic domains (de la Roche et al. 1980) are: 1; Mantle Fractionates, 2; Pre-Plate Collision, 3; Post-Collision exhumation, 4; Late-orogenic, 5; Anorogenic, 6; Syn-Collision granitic melts, 7; Post-Orogenic, Am; amphibolitic restite. Trend A includes all Kelowna lavas which define a sequential mixing line between ?olivine basalt parental magma (mantle) and/or an amphibolitic restite towards silicic and alkaline compositions with time. Trend B includes several Kamloops lavas which define a mixing line between anatectic granitoids and metapelitic restites and/or fractionated basaltic magmas. Trend C includes most Penticton lavas and some of the Kamloops lavas and indicates the fractional crystallisation of augite + K-feldspar $\pm$ biotite.

MAGMATIC SUITES CORYELL LADYBIRD KELOWNA PENTICTON KAMLOOPS SALMON RIVER

Eocene Olivine Basalts _____ Δ _____ Δ
 Intermediate Volcanics _____ $\square$ _____ $\square$
 Felsic Volcanics _____ $\circ$ _____ $\bullet$
 Feldspar Porphyry _____ $\diamond$
 Miocene Plateau Basalts _____ $\diamond$
 Granite _____ $+$
 Syenite _____ $\times$

from a tholeiitic basaltic parental magma (de la Roche et al. 1980; Batchelor and Bowden 1985). These minerals are all present as phenocrysts in mafic lavas of the Penticton suite (Church 1973). Based on the various trends in Figure 6.6 it may be proposed that the Penticton lavas were formed almost exclusively by crystal fractionation from a basaltic source whereas the Kelowna and Kamloops suites were mainly produced by magma mixing between a felsic melt and a crustal contaminant and/or a basaltic source. Feldspar porphyries collected in the Kelowna and Penticton areas plot at the alkaline pole of the Penticton trend (C) consistent with the crystal fractionation of hornblende and oligoclase, the principal constituents of the porphyries.

The compositions of Eocene intrusives, Coryell syenites (Little 1961) and Ladybird granites (*see* Chapter 4) from the northern Okanagan region plot in the field of partial melting (Field 6, Fig. 6.6b) at the extremity of mixing trends (A) and (B). This indicates that these intrusive rocks were probably involved in the magmatic evolution of Eocene lavas as suggested by their coeval ages of crystallisation (*see* Chapter 4).

Intermediate lavas of south-central British Columbia were probably produced by batch mixing of a felsic melt with a mafic crustal contaminant and/or an alkali olivine basalt that may approximate the composition of the Salmon River basalts (Church and Evans 1982; Fig. 6.6b). The proportions of mafic to acid magma required to generate the Kelowna lavas can be estimated from the mixing trend (A) on Figure 6.6b as 40% felsite, and 60% mafic.

6.2.2 Trace element abundance patterns

Chondrite normalized abundances of Ba, Th, K, Rb, Sr, Nb, Zr, Ti and Y are plotted in Figure 6.7 in order of increasing compatibility with mantle (peridotite) mafic phases from Nb to Y, for the elements with high ionic potential (believed to be relatively immobile in a fluid phase; Pearce 1982), and from K to Ba, including Sr, for elements with low ionic potential (mobile in a fluid phase; Pearce 1982). The Nb, Zr, Ti, and Y have similar chemical behavior in the mantle as the rare-earth elements La, Sm, Eu and Tb respectively so that they should display equivalent chondrite normalized REE and non-REE abundances (Sun et al. 1979; Bougault and Treuil 1980).

The Kelowna lavas have generally high Sr (275-1294 ppm) and Ba (1165-2307 ppm) contents that are well above the abundance of most arc basalts and resemble the compositions

Figure 6.7:

Incompatible element concentrations from the Kelowna lavas normalized against chondrite in approximate order of their increasing bulk solid/melt partition coefficients. Normalizing values for K(120), Rb(0.35), Sr(11), Ti(620), Y(2.2), Zr(5.6), and Nb(0.39) are from Sun et al. (1979). Note the high LILE abundances of the Kelowna lavas and Nb and Ti negative anomalies relative to the neighbouring elements. Spidergrams of the Kelowna lavas (shaded area) overlap the compositions of several rock types lying in the footwall of the Okanagan valley fault (?anatectic granitoids, mylonitic gneisses, amphibolite gneisses). Average MORB and Columbia River plateau basalt (Flanagan 1973) compositions are drawn for reference. Miocene lavas of the Kelowna area follow the trend of Columbia River plateau basalts (CPRB).

of typical mid- to lower crustal rocks (Hildreth and Moorbath 1988).

The Kelowna lavas on the multielement diagram (Fig. 6.7) delineate a narrow range of compositional variations showing a marked enrichment in the mobile elements Rb, K, and Sr (LILE) and depleted HFS contents (Nb, Zr, Ti, Y) compared to MORBs. Individual patterns cross one another indicating that the source magmas were probably mixed with variable proportions of crustal contaminants. Felsic lavas exhibit greater LILE and depleted HFS contents than intermediate lavas but both rock types exhibit negative Nb and Ti relative to the neighbouring HFS elements. These anomalies may be caused by refractory Ti-bearing minerals, the stabilities of which are enhanced by elevated fO_2 (Wood et al. 1979; Briquet et al. 1984). Alternatively these anomalies may reflect the presence of deep crustal melts that are typified by low Nb, Ti and HREE contents (Weaver and Tarney 1980; McMillan and Dungan 1988).

The trace element pattern of the Kelowna lavas indicates that these rocks are LREE-enriched to as much as 300 X chondrite values whereas HREE would be slightly depleted below MORB compositions (Winchester and Floyd 1977). The characteristics of the trace element abundances resemble those of recent Andean calc-alkaline andesites (Pearce 1982; Hildreth and Moorbath 1988) but they seem more compatible with those of calc-alkaline lavas formed in intracontinental rift settings such as the Rio Grande (McMillan and Dungan 1988). The latter are thought to be produced by the contamination and/or melting of a tholeiitic magma of mantle origin by another magma resulting from the partial melting of the continental crust (McMillan and Dungan 1988). This phenomenon generates very high LILE concentrations of approximately 500 X chondrite (Fig. 6.7) that is 5 times more than typical calc-alkaline lavas formed as volcanic arcs (Green 1980; Pearce 1982) and generally reflects the continental crustal composition.

The calc-alkaline nature of the Kelowna lavas is confirmed with the ratios $Zr/Y > 7$ and $Ti/Zr < 33$ (Gélinas et al. 1984). On a $[Zr/Y]/Zr$ logarithmic diagram all of the Kelowna lavas plot in the upper part of the "Within Plate" field (Pearce and Norry 1979) but they plot mainly in the ARC field of the Ti versus Zr discriminant diagram (Fig. 6.8; Pearce 1982) similarly to the lavas of the Kamloops suite (Thorkelson 1989). The Kelowna lavas plot also in the ARC field of Shervais' (1982) V vs Ti diagram (Fig. 6.9) and define a linear trend indicating magnetite fractionation in a stratigraphic order from trachyandesites to rhyodacites. This suggests

Figure 6.8: Ti versus Zr discriminant diagram (Pearce 1982). Symbols identical to those in Figure 6.3.

Figure 6.9: *V versus Ti discriminant diagram (Shervais 1982) showing a magnetite fractionation trend from intermediate to felsic compositions in the Kelowna suite. Hexagons are Miocene plateau basalts. Other symbols for Eocene lavas are identical to those in Figure 6.3.*

that the Kelowna lavas evolved sequentially from a common magma source. The Ti/V ratios in these lavas range around 50 indicating a moderate source-rock fO_2 (Shervais 1982) whereas magnetite fractionation indicates high Fe^{3+}/Fe^{2+} ratios that may be related to high fH_2O .

The compatible elements Ni, Cr, and V all decrease rapidly with increasing SiO_2 . Only three samples of trachyandesite (33, 59, 62 in Table 6.3) have Cr contents exceeding 170 ppm and represent the least contaminated magmas of the Kelowna suite. These rocks have Ni, Cr, and V contents comparable to those of the most mafic lavas of the Kamloops suite (Ewing 1981).

6.2.3 Fractional crystallisation

Fractional crystallisation of observed phenocrysts (olivine $\pm$ pyroxene $\pm$ plagioclase $\pm$ magnetite $\pm$ biotite $\pm$ hornblende) may account for certain compositional trends in the Kelowna lavas as was proposed for the Penticton and Kamloops volcanic suites (Church 1973; Ewing 1981). The nearly constant presence of normative plagioclase and its increasing abundance in more felsic lavas of the Kamloops suite has been used to suggest that plagioclase was not efficiently fractionated whereas the progressive declining pyroxene contents in the same order clearly supports its fractionation (Tables 6.2, 6.3). The appearance of modal hornblende confirms that fractionation occurred in a hydrous environment (Ewing 1981b).

MgO in the Kelowna lavas probably varies as a result of low pressure fractionation of orthopyroxene. CaO and Al_2O_3 contents tend to increase with increasing MgO suggesting that plagioclase also controls their fractionation. Plagioclase and pyroxene fractionation is also consistent with the systematic variation of FeO and SiO_2 with MgO (Table 6.2). It is noticeable that the Kelowna lavas have Al_2O_3/TiO_2 ratios similar to MORB and/or chondrite whereas their CaO/ TiO_2 ratios are nearly similar to or less than chondrites (Sun et al. 1979).

All major- and trace-element concentrations have been plotted on logarithmic variation diagrams using Th as a reference incompatible element with a very small bulk partition coefficient (Fig. 6.10) assuming Th is completely incompatible and the bulk Kd for the trace-elements are constant. Kd values are probably constant but they would be affected by changes in proportions of separating phases during fractionation. If a lava suite is co-magmatic and generated solely by a fractional crystallisation process, Th plots will approximate a straight line with each

Figure 6.10: (a) Logarithmic major element variations against Th content. Visually estimated best-fit lines through the data points. (b) Logarithmic trace-element variation diagrams against Th content. (c) Logarithmic compatible element variations against Th content. Symbols are the same as those in Figure 6.3.

crystallisation stage having a constant bulk K_d . Th will generally show a very low K_d between plagioclase, pyroxene, olivine, and liquid (Villemant et al. 1981) but it may show high magnetite/liquid distribution coefficients (Mann 1983) and may be less incompatible in the more evolved magmas involving significant magnetite crystallisation.

Most of the major and trace element variation trends of the Kelowna lavas show regular correlations approximated by visually estimated "best fit" lines with variable Th contents (Fig. 6.10). Samples that vary systematically from the "best fit" evolving trend are those containing up to 10% xenocrysts of clinopyroxene and/or plagioclase. Straight lines cannot be traced on the second halves of figure 6.10b because these samples contain magnetite. Nevertheless, it appears that the Kelowna lavas follow a liquid line of descent from andesite through to rhyolite.

Major element versus Th variations are reasonably regular from andesite to rhyolite compositions (Fig. 6.10a). SiO_2 , Na_2O , and K_2O show positive correlations with Th and are progressively enriched from andesite to rhyolite whereas Al_2O_3 , MgO , and CaO are depleted in the same order. FeO and TiO_2 are slightly depleted to stable from andesite to trachyandesite and undergo significant depletion in the more siliceous compositions. It is noticeable that a marked inflection in the FeO , TiO_2 , MgO , CaO coincides with similar trachyandesitic compositions (Th ~ 11-12 ppm) which correlates with observed changes in phenocryst assemblages.

The incompatible elements (Rb, Zr, Nb) show weak positive correlations with Th, being progressively enriched from andesite to rhyolite (Fig. 6.10b) with enrichment factors of Rb x 10, Zr x 20, Nb x 2.5. Compatible elements (Y, Cr, V, Sr) are progressively depleted throughout the volcanic suite with inflection points at similar andesitic compositions to those shown by major element trends (Fig. 6.10a). Cr varies little in mafic lavas but together with TiO_2 , FeO , and V it drops rapidly in the felsites consistent with the increasing abundance of magnetite and biotite xenocrysts and the fractionation trend shown on the Ti versus V diagram (Fig. 6.9).

Increasing incompatible element concentrations together with decreasing compatible element trends (Fig. 6.10) is consistent with a model involving their fractionation of the observed phenocrystic phases (calcic plagioclase+augite+olivine+hypersthene±magnetite±apatite) from a parental mafic magma. The constancy of the ratios of the highly incompatible trace elements (K, Rb, Zr) from andesite to rhyolite is a feature diagnostic of co-magmatic volcanic series

Figure 6.11: Straight-line variation trends of highly incompatible trace element concentrations vs. Th content in the Kelowna lavas. Symbols are the same as those in Figure 6.3.

generated by crystal fractionation (Fig. 6.11). These variation trends do not pass through the origin of either components indicating that these ratios are not constant. This may reflect several possibilities: (1) that one or other of these elements was not truly incompatible; (2) fractionation was not in a closed system or; (3) an initially mixed magma source (Treuil and Varet 1973).

The Kelowna lavas are dividable into two distinct crystallisation stages which correlate with observed changes in phenocryst assemblages. The earliest extrusion of andesite to trachyandesite is marked by the separation of calcic plagioclase+augite+olivine consistent with an increase in SiO_2 , FeO, TiO_2 , Na_2O , and K_2O and incompatible element contents and the decrease in Al_2O_3 , MgO, CaO, Cr, and Ni contents with fractionation. The final extrusion of dacite to rhyolite components is marked by the inflection in the SiO_2 , FeO, TiO_2 , MgO, CaO, Na_2O , V, Cr, and Ni variation trends at similar andesitic compositions ($\text{SiO}_2 \sim 58\%$) reflecting the onset of crystallization of hornblende+biotite+magnetite.

Fractionation trends are reported on Y/Zr, TiO_2/Zr and Nb/Zr diagrams (Fig. 6.12) which show a consistent evolution from one ratio to the other. These trends match adequately with theoretical fractionation trends developed by Pearce and Norry (1979). In the Y/Zr diagram (Fig. 6.12a), the crystallization of clinopyroxene+olivine+plagioclase in the andesitic magma increases both the Zr and Y contents (vector 1) whereas amphibole, biotite, apatite and zircon crystallisation (vector 2) tend to deplete Y in the residual felsic magma. The mineral vectors defined on the TiO_2/Zr and the Nb/Zr diagrams (Figs. 6.12b, c) confirms that biotite, amphibole and particularly magnetite and zircon fractionated in the felsic magmas. The abundance of Ti-bearing phases such as ilmenite and magnetite in felsic lavas is consistent with their greater magnetic susceptibility than the mafic magmas (*see* Chapter 7). As expected the Kelowna and Kamloops suites follow similar fractionation trends which noticeably are comparable to those of Chilean lavas expelled in a context of recent crustal overthickening (Pearce and Norry 1979).

Two samples of quartz-oligoclase porphyry collected in the regions of Trepanier and Shingle Creek (Little 1961) plot as a continuous series with decreasing Ti, Nb and Zr contents from the felsic lavas of the Kelowna suite (Fig. 6.12). This suggests that they were residual liquids of the fractionation process, consistent with the isotopic age data (Table 6.1) that indicates that porphyries were extruded near the end of the magmatic event.

Figure 6.12: *Observed and modelled fractionation trends of the Kelowna lavas on (a) Y-Zr; (b) Ti-Zr; (c) Nb-Zr. Data from other volcanic suites showing comparable magmatic evolutions: Chile (Pearce and Norry 1979); Kamloops (Ewing 1981). Modelled fractionation vectors from Pearce and Norry (1979) 1: $\text{Plag}_{0.5} + \text{Cpx}_{0.5} + \text{Ol}_{0.1} + \text{Mag}_{0.05}$; 2: $\text{Plag}_{0.6} + \text{Bio}_{0.15} + \text{Amph}_{0.2} + \text{Mag}_{0.05} \pm \text{Zr}$. Symbols are the same as those in Figure 6.3 except the darkened triangles which represent quartz-oligoclase porphyries.*

6.2.4 *Crustal contamination*

Elemental and isotopic variations in the Eocene lavas of south-central British Columbia were probably caused by a combination of processes. Ewing (1981b) had proposed that isotopic variations (Rb-Sr) amongst lavas of the Kamloops suite (Ewing 1981b) indicate mixing of mid-crustal rocks with ?mantle derived basalt. Evidences for mantle parental magmas in the Kelowna lavas are difficult to establish. On the other hand the origin from crustal melts is supported by petrographic evidence such as: mafic units contain embayed quartz xenocrysts rimmed by devitrified glass; felsic units contain multiple populations of unzoned but strained plagioclase sometimes containing biotite inclusions; compositions of plagioclase range between An_{60} and An_{90} .

Very high Sr and Ba concentrations that are particularly enriched in the dacitic magmas of the Kelowna suite (Table 6.2) suggest these may have been produced by as much as 50% melting (McMillan and Dungan 1988) of mid-crustal rocks. Evidence of crustal assimilation also includes the marked enrichment in the mobile elements Rb, K, and Sr (LILE) and the overlap of chondrite normalized spidergrams between the Kelowna lavas and typical mid-crustal rocks of the Okanagan complex (anatectic granite, amphibolite gneiss and mylonitic gneiss; Fig. 6.7).

Simple mechanical mixing from a basaltic magma assimilating crustal material is unlikely to produce magmas more felsic than silicic andesite because large proportions of crustal material lead to a thermal quenching of the basaltic magma as it reaches equilibrium temperatures of the mixture (Frost and Mahood 1987). In order to reach the composition of the felsic magmas in the Kelowna suite a model implying simultaneous crustal assimilation and fractional crystallization of a hybrid magma is therefore more realistic and tentatively modelled.

The average K_2O content of the Kelowna lavas is 3% but it reaches 6% in Kamloops lavas (Table 6.2) and indicates the assimilation of large volumes of granite and/or gneissic material enriched in K and related elements. Granites and gneisses of the Okanagan complex contain on average less than 5% K_2O (Table 6.4). If we assume that an original mafic magma contained less than 0.5% K_2O then to generate a contaminated magma with 3% K_2O requires 56% melting of the Okanagan complex. Assuming that granite and gneiss which contain on average 65% SiO_2 (Table 6.4) were totally assimilated by an uncontaminated basaltic magma with 45% SiO_2 , then the SiO_2 content of the contaminated magma would reach at the most 56%. This magma mixing

process does not the compositions of the felsic members of the Kelowna suite. It is therefore proposed that felsic lavas were produced by the assimilation of extremely large volumes of crustal material and/or that parental magmas were more felsic than basalt.

The assimilation of anatectic melts from various protolithic compositions is expected because of the temporal overlap that exists between syn-extensional granitoids in the Okanagan complex and syn-extensional extrusive magmas. The assimilation of several magma types would have been facilitated by ongoing extensional processes which dilated the crust.

Table 6.4 Average composition (XRF) of anatectic granite¹ and amphibolitic gneiss² forming the Okanagan complex³ underlying the Kelowna outlier.

SAMPLE	Granite	Gneiss
SiO ₂	71.02	58.20
TiO ₂	0.23	0.93
Al ₂ O ₃	15.66	16.20
FeO	1.45	6.63
MgO	0.52	3.19
MnO	0.02	0.10
CaO	1.53	5.98
Na ₂ O	4.32	4.06
K ₂ O	4.96	2.25
P ₂ O ₅	0.06	0.29
Trace elements in ppm		
Ba	2047	1116
Pb	33	13
Zr	195	195
Sr	1353	772
Rb	141	52
Ni	---	10
Cr	67	90
Y	10	20
Nb	8	4
Zn	32	121
V	14	173
Rb/Sr	0.104	0.067

¹: Biotite ± hornblende ± titanite granite represents ~75% of the Okanagan complex near Kelowna

²: At least 80% hornblende ± feldspar ± titanite represents ~15% of the Okanagan complex near Kelowna

³: 10% of the Okanagan complex near Kelowna is composed of calcsilicates-pelites-semipelites

7. *Miocene and Pleistocene basalts*

Tholeiitic flood basalts of the Kelowna area were expelled during three age intervals around 14 Ma, 6 Ma, and 760,000 a. (Church 1982a; Mathews 1987). Based on their age and chemistry these rocks are correlatable with several formations of the Columbia River Plateau basalts (Mangan et al. 1986) and the basaltic lavas of the Western Cascades (McBirney et al. 1974). It has been proposed that these lavas were emplaced along crustal-scale breaks and deep seated lithospheric fractures which may have led to the creation or were created by "windows" in the subducting Farallon plate (Jachens and Griscorn 1983; Thorkelson and Taylor 1989). The occurrence of Miocene basalts near the Okanagan valley at Kelowna and Kamloops (Ewing 1981a) seems to suggest that they were emplaced through attenuated continental crust.

Miocene-Pliocene lavas on both sides of the Okanagan valley (Fig. 6.1) near Kelowna cover uranium-rich paleoplacers (Boyle 1982). Altogether these rocks are currently lying flat at an elevation exceeding 1.5 km. More recent basaltic flows of Pleistocene age (Church 1982a) are lying in the canyons of Lambly Creek and east of Westbank near the shoreline of Okanagan Lake. The present elevation of the Miocene lavas and their alluvial substrata in the Okanagan Highlands may suggest that this region has been uplifted by nearly 1 km (Boyle 1982) assuming the lavas extruded 14 Ma ago (Church and Suesser 1983; Mathews 1987) at lower levels than the present-day topography. The Mio-Pliocene uplift event may have affected the entire Okanagan region and reached a value of approximately 0.2 mma^{-1} , that is at least an order of magnitude less than the Eocene exhumation rate that affected the Okanagan complex (see Chapter 4).

7.1 Petrochemistry

Major, minor and trace elements of eight samples of Miocene and Pleistocene basalts were analysed by XRF (Table 6.5). No stratigraphic order was established in these rocks which exhibit contrasting compositions with the Eocene lavas. They are subalkaline vitrophyric columnar basalts containing phenocrysts of olivine and/or orthopyroxene coexisting with plagioclase in a partly devitrified matrix. Thin section observations suggest that the Miocene fractionating trends were probably controlled by the crystallization of plagioclase + clinopyroxene + olivine + magnetite and minor crustal assimilation as indicated by the presence of composite xenocrysts.

Table 6.5. Major and trace element abundances in Miocene basalts of the Kelowna region.

SAMPLE	6	7	11	12	45	61	64
SiO ₂	60.55	57.20	54.91	53.66	49.77	48.14	48.52
TiO ₂	0.83	0.89	1.21	1.12	2.65	1.85	1.94
Al ₂ O ₃	17.50	18.86	15.96	15.12	16.71	14.64	14.89
FeO	4.45	5.04	5.97	7.19	13.00	13.46	12.66
MgO	1.86	1.95	5.45	7.00	3.98	8.75	8.98
MnO	0.09	0.15	0.07	0.07	0.22	0.17	0.16
CaO	3.43	4.39	8.33	7.80	6.27	9.15	9.21
Na ₂ O	3.87	4.12	3.21	2.96	3.77	2.09	2.35
K ₂ O	5.46	4.72	3.21	2.94	1.93	0.68	0.86
P ₂ O ₅	0.31	0.41	0.54	0.53	1.07	0.30	0.40
LOI	2.11	2.07	0.00	0.00	0.88	1.56	1.01
TOTAL	100.46	100.40	98.92	98.47	100.25	100.79	100.98
NORM							
q	4.88	----	----	----	----	----	----
or	32.27	27.90	18.97	17.38	11.42	4.02	5.09
ab	32.74	34.86	27.16	25.04	31.90	17.68	19.88
an	14.26	19.03	19.66	19.29	23.00	28.58	27.56
di	0.62	2.56	14.81	13.02	1.15	12.45	12.95
hy	11.29	10.49	12.75	14.25	10.19	17.66	12.27
il	1.58	1.69	2.30	2.13	5.03	3.51	3.68
ap	0.73	0.97	1.28	1.26	2.54	0.71	0.95
ol	----	0.85	1.95	6.06	14.25	14.65	17.63
zr	----	----	----	----	0.07	0.03	0.03
cr	----	----	----	----	----	0.09	0.09
TOTAL	98.37	98.35	98.89	98.42	99.56	99.38	100.13
Rb	----	----	----	----	25	6	11
K	45325	39182	26647	24406	16022	5645	7139
Ba	----	----	----	----	685	198	297
Pb	----	----	----	----	<5	<5	<5
Th	----	----	----	----	<5	<5	<5
U	----	----	----	----	10	11	11
Nb	----	----	----	----	46	20	29
Sr	----	----	----	----	689	482	473
Zr	----	----	----	----	333	130	143
Ti	4976	5336	7254	6714	15887	11091	11630
Y	----	----	----	----	43	22	21
V	----	----	----	----	130	242	226
Cr	----	----	----	----	<10	429	417
Ni	----	----	----	----	<10	127	158
Al ₂ O ₃ /TiO ₂	21.1	21.2	13.2	13.5	6.3	7.9	7.7
CaO/TiO ₂	4.1	5.6	6.9	6.9	2.4	4.9	4.8
TiO ₂ /P ₂ O ₅	2.7	2.2	2.2	2.1	2.5	6.2	4.8
Rb/Sr	----	----	----	----	0.04	0.01	0.02
K/Ti	9.1	7.3	3.7	3.6	1.0	0.5	0.6
Ti/Y	----	----	----	----	369.5	504.1	553.8
Ti/Zr	----	----	----	----	47.7	85.3	81.3
Zr/Y	----	----	----	----	7.7	5.9	6.9
Zr/Nb	----	----	----	----	7.2	6.5	4.9
(Zr/Nb) _n	----	----	----	----	0.5	0.4	0.3

Two samples (61, 64 on Table 6.5) plot in the MORB field whereas a third one (45) plots in the WPB field of the Ti versus Zr diagram (Fig. 6.8). However, on the Ti versus V diagram samples 61 and 64 plot outside the MORB field and are only slightly depleted in V with respect to typical Columbia River Plateau basalts (Mangan et al. 1986; Fig. 6.8). A geochemical similarity with the Columbia River basalts is more clearly expressed on the chondrite normalized multielement diagram (Fig. 6.7). It is clear from this diagram that Columbia River basalts show similar but more attenuated spidergrams than Eocene lavas. For example, LILE abundances are approximately 100 X chondrite values (typical of volcanic arc trends) and HFS (hygromagmatophile) concentrations match those of MORB.

8. *Summary and discussion*

Syn-extensional andesite-dacite volcanoes were extruded along a volcanic front straddling the Okanagan valley on the west flank of an overthickened crustal welt in the southern Omineca Belt (Monger et al. 1982; Brown et al. 1986). Near Kelowna, the first eruptive product, the Trépanier dacite probably extruded ca. 54 Ma at the time anatectic melts were cooling below 750°C in the Okanagan complex. This was soon followed by the extrusion of Marron andesites and trachyandesites which heralded the onset of Eocene volcanism, ca. 53 Ma. (Church 1973, 1979b; Pearson and Obradovich 1977; Ewing 1981; McMechan 1983). Large volumes of Marron lavas erupted in less than 2 Ma and were followed by the intrusion of quartz-oligoclase porphyries related to the Coryell suite. During this time metamorphic minerals were cooling below 550°C in the middle crust (cooling temperature for Ar in hornblende; see Chapter 4).

Heterogeneous Eocene lava sequences formed an estimated volume of 100,000 km³ distributed from south-central British Columbia to north-central Washington (Snook 1965; Muessig 1967; Church et al. 1983; Ewing 1981a; Monger 1982; Little 1983; McMechan 1983; Carr and Parkinson 1989). The Marron-Sanpoil suite forms at least 60% of the bulk volume of the Eocene lavas which is approximately one third the volume of Miocene basalts expelled in the Columbia River Plateau which also extruded in nearly 2 Ma (Mangan et al. 1986).

The volume of Eocene lavas produced in south-central British Columbia requires relatively large magma chambers. The inferred source, the Coryell-type syenites, form several large

intrusive bodies exposed more than 90 km east of Kelowna (Church 1973; Little 1983; Carr and Parkinson 1989) but only small plugs of Coryell syenites have been mapped in proximity of the lavas straddling the Okanagan valley (Little 1961). However, based on the elemental, modal and temporal evolution of the Kelowna lavas it is proposed that anatectic melts of the Ladybird suite (59-55 Ma; Parrish et al. 1988; Carr 1990; *see* Chapter 4) were as good a source as Coryell intrusions. Plutonic sheets of the Ladybird suite are probably abundant in the Okanagan complex in the vicinity of the Eocene outliers bounding the Okanagan valley structure (*see* Chapter 4). Consequently the magnitude of displacement across the Okanagan valley fault based on magmatic affinities is probably less than 90 km (Tempelman-Kluit and Parkinson 1986; *see* Chapter 5).

Evidence that extensional faulting was contemporaneous with eruption of the Marron lavas includes; the Marron overlaps a greater structural relief and a larger surface than any older and younger volcanic sequence; feeder dykes of Marron composition nowhere cut footwall mylonites; numerous normal faults cut through all Marron flows; angular unconformities are locally developed within the Marron sequence. Perhaps active normal faulting in the roof of the Marron magma chamber decompressed the magma allowing a larger proportion of lava to escape to the surface. Large flows were probably deposited along paleotopographic fault scarps.

Middle Eocene sedimentary sequences in the Kelowna area are widely separated but remarkably similar in sequential order of their lithotypes. Clast types and structures record the progressive unroofing of mid-crustal rocks. Bedding in the basins is variable and locally rotated in the plane of the Okanagan valley fault suggesting that motion on the Okanagan valley fault occurred mainly before alluvial sedimentation.

Dacitic to rhyodacitic lavas and tuffs form the last eruptions in this volcano-sedimentary episode. These lavas evolved from similar sources as the Marron lavas following a fractionated evolution from a parental magma already contaminated with a significant amount of crustal material. These rocks erupted 48 Ma ago forming a preserved volume of about 100 km³ but it is expected that large volumes of co-magmatic ash flows and tuffs have been eroded away. Evidence of post-extensional extrusion is mainly constrained by the presence of many feeder dykes of similar composition cutting through footwall mylonites underlying the Joe Rich klippe.

The Kelowna outlier serves as an example of the linkage between magmatism and regional

extension across the southern Omineca Belt. The demonstration of close magmatic-tectonic interaction suggest that magmatism had an important role in extensional processes or vice versa similarly to the situation prevailing in the highly extended parts of the Basin and Range province (Gans et al. 1989). Well-dated *cogenetic* volcanic, plutonic and hydrothermal events in both upper and middle crustal rocks bounding the Okanagan valley fault suggest that the bulk of extensional shearing affected mid-crustal melts slightly after or during the eruption of voluminous lavas of intermediate composition. It is also proposed that extensional strain rates peaked at the time of major anatectic melting and intermediate volcanism ca. 52 Ma. Ductile shearing on the Okanagan valley fault stopped approximately 48 Ma ago while the last felsic flows were extruded, from evolved magma chambers, through and onto the Okanagan Complex.

The presence of large magma chambers at the top of the middle crust is required to create enough heat to melt and assimilate hosting amphibolite lithologies. Melting must have been efficient in order to assimilate sufficient sialic material and create density variations facilitating the rapid diapirism of large volumes of mixing and fractionating magmas.

Our data suggests that there is little, if any, input of basalt from a mantle source to generate the lavas of the Okanagan region (*see* Thorkelson and Taylor 1989). It is therefore proposed that voluminous chambers of *in situ* melts coalesced at the bottom of the upper crust and provided a zone of ductility by thermal softening in the middle crust leading toward the propagation of extensional structures.

Compositional trends amongst the Eocene lavas suggest that the felsic members of the Kelowna suite are genetically related to the intermediate lavas by fractional crystallisation. Mafic magmas which occur in the Penticton region further south (Church 1973) and in the Kamloops area further north (Ewing 1981b) may represent parental compositions to the Kelowna suite.

Crustal contributions to the magma genesis in the Eocene can be further demonstrated by an apparent but regular distribution of $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the Kamloops lavas (Ewing 1981b). These range from 0.704 to 0.708 from west to east across a NW-SE trending transition line traced amongst several outliers of the Kamloops suite without noticeable mineralogical changes (Ewing 1981b). The transition line between sialic type ratios and uncontaminated ratios in the Kamloops suite has been used to delineate the western margin of the Precambrian continental crust of

western North America (Ewing 1981b). Perhaps the isotopic ratios reflect the composition of middle to lower crustal contaminants rather than deep crustal boundaries. Furthermore this practice seems obsolete now that it is shown that the North American crust extends west of the Okanagan valley (Cook et al. 1992). It is noticeable that gneisses of the Vaseaux Formation (Okanagan complex) underlying the Kelowna and Penticton volcanic suites exhibit $^{87}\text{Sr}/^{86}\text{Sr}$ ratios that are ranging widely between .703 to .725 (Armstrong et al. 1991). These ratios would reflect in the crustal melts leading eventually to the extrusion of Eocene lavas. Therefore it seems unwise to use the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios to decipher the position of the continental margin in the Eocene.

The pattern of transition from andesitic to dacitic volcanism in the Kelowna volcanic sequence is typical of several major Tertiary volcanic fields of the southwestern United States and Western Mexico (Wark et al. 1990). Furthermore it appears that mineral fractionation vectors recorded by the Kelowna lavas resemble those of the northern Andean volcanic fields which formed over thickened continental crust (Hildreth and Moorbath 1988). Based on these similarities it seems consistent that Eocene lavas in south-central British Columbia were derived from anatectic melts themselves formed in an overthickened crust. Thus the inference that south-central British Columbia was underlain by abnormally thick crust (Brown et al. 1986) prior to the middle Eocene is corroborated by the presence and the composition of the Eocene lavas.

Eocene volcanism also occurred at a time of slowed convergence between the Kula-Farallon and North American plates between 59 and 50 Ma (Verplank and Duncan 1987; Lonsdale 1988; *see* Chapter 8). Slowed plate convergence may have resulted in decreased flux of mafic magmas from the mantle leading to the probable stagnation of residual reservoirs in the crust that were forced to differentiate and mix with mid-crustal anatectic melts. However, from the present data we are unable to detect directly the presence of these mantle sources.

In contrast with Eocene lavas the Miocene basalts are more typical of a volcanic arc setting though they also show evidence of crustal contamination (Carlson et al. 1981) which is consistent with the idea that the crust became thinner in the Miocene than in the Eocene.

9. *Conclusions*

Shield volcanoes erupted high-K calc-alkaline andesitic to dacitic lavas in south-central British Columbia in the middle Eocene soon after the tectonic transition from compression to crustal extension. Radiometrically constrained evidence suggests that anatectic melts affected by extensional shear features at mid-crustal levels crystallized between 56 and 51 Ma ago while the largest volumes of intermediate magmas, the Marron lavas, erupted. This temporal correlation indicates that upper crustal volcanism and mid-crustal melting were simultaneous processes which reflects in the compositional modes of the syn-extensional volcanic suite.

There is evidence that the Kelowna lavas, and most Eocene lavas of south-central British Columbia, were produced by the interaction of differentiation processes dominated by the assimilation of mid-crustal material and fractional crystallization. Evidence of crustal assimilation is demonstrated by: (1) trace element spidergrams that are typical of lavas contaminated by amphibolite-facies gneisses (Thompson et al. 1982) and overlapping with trace-element contents of the anatectic granitoids and mylonitic gneisses presently underlying the lavas; (2) similar spidergrams amongst all lithologies of the Kelowna suite suggest they interacted with similar contaminants and that none of the volcanic units escaped the effects of contamination, the last and most felsic units seem to have assimilated more crustal material increasing their LILE contents and reducing their HFS elements; (3) although complicated by plagioclase fractionation, the high Sr content (260-1295 ppm) of most volcanic units most likely reflects partial melting of plagioclase-rich crustal rocks; (4) the wide compositional ranges of values such as K/Rb and Ti/Y from neighbouring lava flows are indicative of crustal contributions; (5) variable $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in lavas of the Kamloops suite (Ewing 1981b) are similar to compositional variations in mid-crustal rocks (Armstrong et al. 1991) indicating also a crustal contribution to the Eocene magmas. Variable but probably high degrees of partial melting of mid-crustal rocks produced felsic liquids inherited from variable protolithic compositions. These melts were mobilized in the crust and mixed to generate magmas of intermediate compositions in which some fractional crystallisation of pyroxene, biotite and magnetite are readily identifiable by major and trace element variation within continuous volcanic series.

Indirect evidence from other Eocene outliers in south-central British Columbia suggests that parental magmas such as subduction-related olivine tholeiites may have mixed with at least 50% melts collected amongst the anatectic restites formed at mid-crustal levels. Diapirism of the Eocene magmas was probably enhanced by increasing assimilation of sialic crust creating buoyant felsic compositions capable of penetrating the low-density upper half of the continental crust. Based on chemical and regional arguments it is proposed that sources for the Eocene volcanic rocks in south-central British Columbia are both Coryell-type syenites and leucogranites of the Ladybird suite exposed presently in the immediate neighbourhood of the Eocene edifices.

The Eocene volcanic suite of south-central British Columbia may be interpreted as an intracontinental volcanic sequence and thus it is not a typical volcanic arc sequence as proposed by Ewing (1981) for the Kamloops suite. The trace-element ratios of the Kelowna lavas closely match those of calc-alkaline Andean lavas formed over thickened continental crust (Pearce and Norry 1979). The heterogeneity of the volcanic sequences throughout south-central British Columbia provides a good indication of the heterogeneity of the sources of primitive magmas which is consistent with the context of an overthickened crust containing anatectic melts.

It may be suggested that Tertiary volcanism in south-central British Columbia evolved from calc-alkaline to basaltic by decreasing the amount of hybridization, removing lower and middle crustal partial melts in the Eocene magmas and thinning the crust to its present thickness (32 km) in the Pliocene. This evolution is consistent with the postulate that the crust was overthickened as the present Chilean setting prior to the Eocene and evolved progressively, after extension, to a typical arc - back-arc setting in the Pliocene.

Thick alluvial sedimentary deposits interdigitated with the Eocene lavas show an inverse stratigraphy starting at the bottom with hanging wall clasts being replaced by footwall clasts at the top of the section. This indicates that sedimentation occurred during extension and that the footwall of the Okanagan valley was progressively unroofed to finally breach to the surface at the end of the sedimentary cycle. Felsic lavas at the top of the Kelowna outlier were expelled above and into these unconsolidated strata around 48 Ma at the end of the denudation event.

1. Introduction

Success in paleomagnetic work depends critically on one being able to determine the attitude of the magnetization horizontal plane. The magnetization horizontal plane (or more briefly the paleomagnetic horizontal) is the horizontal plane at the time the rocks were magnetized. In this paper we describe studies of coeval middle Eocene rocks lying in the hanging wall and footwall of the Okanagan valley fault (*see* Chapter 2) to investigate this problem.

In bedding sequences with a primary magnetization, the geologically determined bedding plane is a good approximation to the paleomagnetic horizontal. In bedded sequences whose magnetization is entirely secondary, the present horizontal usually is assumed to approximate the paleomagnetic horizontal because the dispersion of paleodirections is markedly better before than after correction for bedding attitudes. This simple -either/or- rationale, while still applicable to rock units with uncomplicated histories, is now known to be inapplicable in many instances because magnetization processes can occur over an interval of time that extends from before to after deformation. As the list of rock units with fairly simple thermal and chemical histories is exhausted, and those with more complex histories come to be studied, paleomagnetists will encounter more and more frequently such extended magnetization processes, and will have to face the concomitant problem of determining the paleomagnetic horizontal.

One of the hallmarks of rock units with such extended magnetization histories is that the dispersion of paleodirections is high before and after correction for bedding - higher, that is, than can be reasonably anticipated from paleosecular variation. Certain studies have shown that partial correction for bedding attitudes produces acceptable dispersions (Schwartz and van der Voo 1984) as if magnetizations were acquired at a certain stage during the deformation process, perhaps even controlled by it. This is unlikely to be generally true because many different combinations of primary, syn-deformational, and post-deformational magnetization may be expected to occur within single rock units. For example, a recent study has revealed magnetizations acquired before and during deformation, by two quite different processes, because the two magnetizations had different unblocking temperatures (Miller and Kent 1986).

Rocks may be expected to acquire their magnetization at deposition or during diagenesis by a variety of thermal and chemical processes, and it is the variety of these processes and their temporal relationship to deformation that creates the difficulty in determining the paleomagnetic horizontal. The first part of this chapter comprises a description of the magnetization of a volcanic sequence in the vicinity of Kelowna. The rocks have been variably tilted, and have both pre- and post-tilting magnetizations. The scalar properties of these magnetizations are indistinguishable, and they must, therefore, be separated by their directional characteristics. A rationale for dealing with this situation is presented and its tectonic consequences discussed.

The second topic in this chapter concerns the paleomagnetism of Eocene anatectic melts formed at middle crustal levels (Carr 1990). At first sight, observations from such deep-seated rocks may appear difficult to interpret, but, somewhat paradoxically, the problem of determining the paleomagnetic horizontal may be less severe here than in their upper crustal, bedded extrusive rocks. This is because middle-crustal rocks acquire a magnetization thermally through exhumation and cooling; their magnetizations are acquired by a relatively simple regional process rather than by variable local processes of deposition and diagenesis. In this situation, the paleomagnetic horizontal is the horizontal plane within the crust as it exhumes and cools. Hence, if the paleofield is known for the period of time in question, then tilts that have occurred after original uplift will be evidenced by departures of the paleomagnetic directions from those expected. From the departures, tilts can be calculated. The data in the second part of this chapter was collected by other workers in the vicinity of the Valhalla complex near Castlegar located 150 km east of Kelowna. These data are introduced in order to suggest a means of estimating crustal tilts in plutonic and high-grade metamorphic rocks associated with large-scale extensional faults.

2. *Geology and sampling*

The Eocene lavas of southeast British Columbia are an erosional remnant of a once large volcanic belt extending from Kamloops in British Columbia (Ewing 1981a) south to Republic in Washington (Heller et al. 1987). Total accumulation is estimated to be about $1 \times 10^6 \text{ km}^3$, assuming a length of 450 km, a width of 100 km, and an average thickness of 2 km.

The rocks studied are mainly from the middle Eocene (54-52 Ma) Marron Formation of

south-central British Columbia (Little 1961; Monger 1968; Church 1973, 1982b). Erosional remnants of these lavas occur in the hanging wall of the Okanagan valley fault near Kelowna. The Okanagan valley fault (Figs. 7.1-3) marks the boundary between the Omineca (high-grade) and Intermontane (low-grade) belts (Ross 1981; Monger et al. 1982). It is 450 km in length and is one of several very large low-angle extension faults that occur in the region of Eocene extension in south-eastern British Columbia (Parrish and Carr 1986; Haugerud 1986; Tempelman-Kluit and Parkinson 1986; Brown and Journeay 1987; Parrish et al. 1988; Carr 1990).

Mineral-stretching lineations in footwall gneisses underlying the Okanagan valley fault are parallel, and trending consistently N285E. Mylonites associated with the Okanagan valley fault contain kinematic indicators that suggest a westerly sense of motion of the hanging wall relative to the footwall (*see* Chapter 2). U-Pb dates on zircons and monazites from both westerly and easterly dipping normal faults across the Omineca Belt indicate that ductile deformation associated with motions on them that occurred between 59 and 50 Ma (Parkinson 1985; Parrish et al. 1988; Carr 1990; *see* Chapter 4). Some later motion may have occurred on brittle faults that cut the mylonite zones. One of the major faults, the Slocan Lake fault (Carr 1986), has been traced into the lower crust as an easterly dipping seismic reflector (Cook et al. 1988, 1992). Displacements on the main faults may have been several tens of kilometers (Carr 1985; Bardoux 1986; Tempelman-Kluit and Parkinson 1986; Parrish et al. 1988).

Footwall rocks are typically amphibolite-grade gneiss, into which have been intruded bodies of Eocene age - the Ladybird leucogranites and the Coryell syenites - that are essentially coeval with the lavas of the Marron Formation (Table 7.1) and may be feeders to them (Church 1973; Tempelman-Kluit and Parkinson 1986). Above the Okanagan valley fault, the hanging wall comprises low-grade late Paleozoic eugeosynclinal rocks cut by Jurassic plutons. K-Ar mica dates of Jurassic age (Medford 1975) are preserved in the plutons, indicating that they, and the hanging wall rocks, have not been heated above 300°C since emplacement. In contrast, K-Ar dates from micas and hornblende and fission-track dates on apatite and titanite from footwall gneisses indicate that they were heated to a temperature above 500°C until 52 Ma, and then cooled rapidly to 100°C within about 7 Ma (Medford 1975; Mathews 1981; Parkinson 1985). Such metamorphic rocks might be expected to preserve a record of the middle Eocene geomagnetic field.

Figure 7.1: *Relation of sampling region of Figure 7.2 to major tectonic belts of the Canadian Cordillera. K: Kelowna.*

Figure 7.2: *Zone of Eocene extension, southeastern British Columbia, compiled from Parrish et al. (1988). 1, middle Eocene volcanics and volcanoclastic rocks; 2, middle Eocene plutons, syenite, and granite; 3, rocks that predate extension-metamorphic core-complexes, Paleozoic to early Eocene granitoids and metavolcanics; 4, major Eocene low- to moderate-angle extension faults: Okanagan Valley fault (OVF), Slocan Lake fault (SLF), Valkyr shear zone (VSZ), Columbia River fault (CRF), Granby fault (GF), Greenwood fault system (GWF), Kettle River fault (KF), Newport fault (NF), Purcell Trench fault (PTF). Major core-complexes are Kettle - Grand Forks complex (KC), Okanagan complex (OC), Valhalla complex (VC). FB are major sampling regions of Fox and Beck (1985). C: Castlegar, G: Greenwood, K: Kelowna, N: Nelson, P: Penticton, R: Republic and the Republic graben. The hypothetical section below is drawn at a somewhat smaller horizontal scale, but with vertical and horizontal scales the same. Note that SLF is the only fault that has been seismically imaged to the lower crust (Cook et al. 1988). Other faults are extrapolated from surface observations though now modified on the LITHOPROBE profile (Brown et al. 1990). VSZ is omitted from the section because it predated the Eocene intrusions sampled at sites TSP5 - TSP7 and TSP51, data from which are discussed herein.*

The sampling region near Kelowna is shown in Figure 7.3. The lavas are exposed in long semicontinuous sections, capped by late middle Eocene lacustrine sedimentary rocks (see Chapter 6), which provide excellent determinations of bedding attitude. Only few faults cut the sequence allowing attitudes to be extrapolated from adjacent localities when bedded rocks are absent.

Around Kelowna, the Eocene section is approximately 4 km thick and is divisible into three main formations (see Chapter 6; Fig. 7.4). The lowest, the Marron Formation, is composed principally of potassic and calc-alkaline lavas, and consists of four members, three of which may be correlated stratigraphically with similar units to the south in the White Lake basin (Church 1973, 1982b). Fresh biotites provided a K-Ar age of 53.6 ± 2.2 Ma (Table 7.1). Paleomagnetic samples were obtained from 26 sites in the Marron Formation and provided most of the data (Table 7.2).

A lenticular body of an unnamed quartz-orthoclase rhyolitic porphyry underlies the Marron Formation itself underlain by the Trepanier dacite. This unit (site 21 in Table 7.2 and Fig. 7.3) has a minimum thickness estimated at 500 m, and has yielded a K-Ar date of 50.5 ± 1.0 Ma from unaltered biotite (see Chapter 4). Locally, andesitic lavas overlie the Trepanier dacite. The andesites are petrographically similar to the Shatford Creek Member of Church (1982a, 1982b) which lies at the bottom of the Marron Formation (Fig. 7.3). These andesites were sampled at sites 44 and 45 north of Kelowna.

The Shatford Creek member is followed by a thick sequence of trachytes and trachyandesites that correspond petrographically and geochemically to the Kitley Lake Member of Church (1973). Within the Kitley Lake Member, ash-flow layers occur adjacent to many sites, providing excellent bedding control. Gradationally overlying the Kitley is the Nimpit Lake Member, which consists of trachytes and scoriaceous trachyandesites. The Kitley Lake and Nimpit Lake members are often hematitic.

On top of the Marron Formation is a 1.2 km thick sequence of interbedded pyroclastic debris flows and lacustrine, fluvial, and alluvial-fan deposits, which are stratigraphically correlated with the White Lake Formation of Church (1973) and which contain palynomorphs of late middle Eocene age (McIntyre, unpublished, 1985). These rocks are tilted about 20° east into

Figure 7 3:

Geological sketch map with sampling sites. Geology from Little (1961), Church (1973), and Bardoux (1985). Rock units: 1, Pleistocene basalt and Miocene andesite; 2, Eocene dacites and rhyodacites (Marana Formation); 3, middle Eocene debris flow and lacustrine sediments (White Lake Formation equivalents); 4, Eocene trachyandesites (Nimpit Lake Member); 5, Eocene trachytes (Kitley Lake Member); 6, Eocene andesites (Shatford Creek Member); 7, Quartz-oligoclase porphyry and Trepanier dacite; 8, Mesozoic granodiorites; 9, late Paleozoic metavolcanics; 10, Mesozoic and Cenozoic plutonic rocks; 11, highly strained orthogneiss and paragneiss of the Okanagan gneiss complex. The Okanagan valley fault is marked by half circles.

Figure 7.4: *Magnetostratigraphic column of Eocene volcanic rocks of the Kelowna outlier. Sampling sites numbered as in Figure 7.3 and Tables 7.2 and 7.3. Stratigraphy after Church (1973) and Bardoux (1987). Radiometric dates from the marron are given in Table 7.1. Other dates (all K-Ar) are as follows: 54 ± 2 Ma Marron (Chapter 4); 50 ± 1 Ma Trépanier porphyry; 46 Ma Marama, biotite (Medford 1975); 48 Ma Marama (Church 1980a)/ 48 ± 1 Ma Marama (Chapter 4). N, R, and T denote normal and reversed polarity and transitional directions. The 26 sites in the black sequence (Marron Formation) all have normal polarity. Trépanier porphyry and Marama dacite are reversed.*

Table 7.1: K-Ar dates from the Marron Formation

Rock unit	Material	Age (Ma)
Kitley Lake Member, Carrot Mountain ^a	Whole rock	44 ± 2
Kitley Lake Member ^a	Biotite	52 ± 6
Kitley Lake Member ^a	Biotite	53
Mafic phonolite, Riddle Creek ^a	Biotite	53 ± 2
Shingle Creek porphyry ^a	Biotite	52 ± 2
Sanpoil volcanics ^b	13 dates	48 – 53

^aChurch (1978, 1981, 1982a).

^bMarron equivalent in Washington (Pearson and Obradovich 1977).

the Okanagan valley fault and provide excellent bedding attitudes. At the bottom of the White Lake Formation, thick pyroclastic debris flows carry large cobbles in a fine-grained tuffaceous and mudstone matrix. The cobbles are derived exclusively from the hanging wall. Seven Eocene volcanic cobbles were sampled for a conglomerate test at site 36 (Fig. 7.3).

The youngest Eocene lavas are fine-grained dacites. They occur predominantly east of Kelowna as circular plugs and sills (sites 7, 13, Fig. 7.3) cutting footwall gneisses, and are probably feeders to outliers of lava of dacitic composition. A K-Ar age of 48.6 ± 0.7 Ma has been obtained from biotite in these rocks (Fig. 7.4). Taking into consideration this age and their petrography, the dacites are tentatively correlated with the Marama Formation (Church 1973).

Hornblende gneisses underlying the Okanagan valley fault were also sampled at sites 2, 3, 6, and 14 (Fig. 7.3). Most gave random magnetization, and demagnetization did not improve groupings. However, some coherent reversed magnetizations were observed at sites 3 and 6, directed nearly parallel to the average mineral-stretching lineation (5° at 110° E), indicating that the remnant magnetization in these rocks has been entirely controlled by the mineral fabric and does not record the paleofield.

Syenitic bodies of mid-crustal origin were sampled 100 km to the east (Table 7.2) of Kelowna. Sample TSP51 was collected near Ymir in the hanging wall of the easterly-dipping Slocan Lake normal fault (Carr 1986), whereas TSP5, 6 and 7 are diorites and a syenite dike collected in the footwall along Lower Arrow Lake (Fig. 7.2). The bodies are late kinematic to post-kinematic with respect to Eocene ductile shearing along the Slocan Lake fault. TSP51 provided a U-Pb zircon age of 51.5 ± 0.5 Ma and diorites of TSP5-6 provided 52 Ma U-Pb zircon ages (Parrish et al. 1988). These rocks are considered coeval with the Marron Formation.

3. *Procedures and magnetizations*

Altogether, a total of about 200 cores were obtained from 41 sites in Eocene rocks. Cores were oriented by magnetic and solar compasses. Generally, two specimens were cut from each core. Tilt corrections were applied by simple rotation about strike. Geological mapping shows that the rocks have been tilted as large panels. Complex folding is absent in the Kelowna area and plunge corrections are not needed.

Table 7.2: Sampling sites, rock types and bedding estimates

Site	Rock type	BD/BDDA(RB)	UTM grid reference
M1	Trachyandesite (K)	29/143(Z)	319100E, 5534400N
2	Gneiss	16/016	337230E, 5525900N
3	Gneiss	04/220(?)	338600E, 5528100N
M4	Trachyandesite (K)	29/143(Z)	314450E, 5533850N
M5	Trachyandesite (K)	29/143(Z)	314200E, 5533900N
6	Gneiss	19/036	40900E, 5526400N
m7	Dacite dyke	—	340875E, 5526400N
M8	Trachyandesite (K)	45/075(X)	321400E, 5530900N
M9	Gneiss tuff (K)	45/075(X)	321550E, 5531050N
M10	Trachyandesite (N)	30/143(Y)	305150E, 5524750N
M11	Trachyandesite (N)	30/143(Y)	305400E, 5526250N
M12	Trachyandesite (K)	10/315(Z)	305500E, 5532300N
m13	Dacite dyke	—	347600E, 5508300N
14	Gneiss	07/028	338800E, 5534650N
m15	Dacite sill complex	13/304(X)	331350E, 5524000N
M16	Trachyandesite (N)	55/046(X)	308850E, 5521000N
M17	Trachyandesite (N)	55/046(X)	308950E, 5521100N
M18	Trachyandesite (K)	10/260(Z)	313250E, 5531250N
M19	Trachyandesite (K)	44/315(Z)	309550E, 5534050N
M20	Trachyandesite (K)	55/130(Z)	309900E, 5537450N
T21	Porphyry	10/090(Z)	302700E, 5520750N
M31	Trachyandesite (K)	29/143(Z)	318500E, 5537800N
M32	Trachyandesite (K)	29/143(Z)	319300E, 5536250N
M33	Trachyandesite (K)	55/130(X)	310700E, 5538600N
M34	Trachyandesite (K)	55/130(Y)	310650E, 5538850N
M35	Trachyandesite (K)	10/316(Z)	305800E, 5531700N
W36	Pyroclastic breccia	28/075(X)	309300E, 5521700N
M37	Trachyandesite (N)	55/047(X)	309150E, 5521400N
m38	Dacite	—	348750E, 5528300N
m39	Dacite	—	348800E, 5528500N
M40	Trachyandesite (K)	10/261(Y)	314100E, 5537350N
M41	Trachyandesite (K)	10/261(Y)	313550E, 5538800N
M42	Trachyandesite (N)	45/076(Y)	321800E, 5531750N
M43	Trachyandesite (N)	45/076(Y)	322300E, 5531650N
M44	Andesite (S)	37/161(X)	326300E, 5538800N
M45	Andesite (S)	37/161(Y)	325750E, 5537850N
M46	Ash flow (N)	55/046(X)	309150E, 5521350N
M47	Lahar (N)	55/046(X)	309050E, 5521300N
M48	Lahar (N)	55/046(X)	310000E, 5521250N
M49	Trachyandesite (K)	63/104(X)	323600E, 5534850N
M50	Ash flow (S)	63/104(X)	323500E, 5534400N
TSP51	Unaltered Coryell syenite	—	484900E, 5456800N
TSP5	Pump metadiorite	—	442050E, 5465050N
TSP6	Pump metadiorite	—	442300E, 5465000N
TSP7	Altered syenite dyke	—	442050E, 5465050N

NOTES: (1) M, Marron Formation; W, White Lake Formation; m, Marama Formation equivalent; TSP, Eocene "intrusions"; T, Trepanier porphyry. (2) Rock types are followed by K, Kitley, N, Nimpit, S, Shatford Creek denoting the member to which they belong (Fig. 4). (3) BD and BDDA are the dip of bedding and the bedding down-dip azimuth, both in degrees; RB is the reliability factor for bedding estimates: X, very good, Y, good, Z, obtained by extrapolation from neighbouring sites. (4) Sites TSP5–TSP7 are amphibolite-grade metamorphic rocks.

Measurements were made on a Schonstedt spinner magnetometer (SSM2) controlled by an IBM/PC. Pilot specimens (at least two per site) were demagnetized thermally and by artificial fields (AF). Stepwise thermal demagnetization was carried out using a Schonstedt TSD-1 furnace, and AF demagnetization using a Schonstedt GSD-5 tumbling demagnetizer.

The magnetizations are of two types, designated A and B. A-type magnetizations have excellent end points, after removal of small soft magnetizations that are aligned approximately along the present Earth's field and are considered to be recent viscous remanent magnetizations (VRM). Orthogonal plots show regular linear decay to the origin above 300°C and after AF demagnetization in 20 mT and more (Figs. 7.5, 7.6). The direction of these high unblocking temperatures high coercive force magnetizations were obtained by choosing a demagnetization step placed centrally along the line; that is, A-type specimens were cleaned at a single field or temperature. This magnetization is referred to as the "principal" magnetization because it is the largest in magnitude and the most important geophysically.

B-type magnetization, like A-type, is composed of a small VRM and a large principal magnetization, but, in addition, there is commonly a very small magnetization, usually with a very high blocking temperature and very high coercivity. Decay lines on orthogonal plots are linear, but less well-defined than in A-type magnetizations (Fig. 7.7), and they do not always go precisely through the origin. The principal components of B-type magnetization have been determined by least-squares fitting to linear decay lines using the analysis of Kirschvink (1980) and the computer programme LINEFIND (Kent et al. 1983). The small, very high stability magnetizations are attributable to hematite and probably were produced by incipient recent weathering. Analytical details are given in the footnote to Table 7.3.

To estimate the time at which magnetizations were acquired a study has been made of a penecontemporaneous volcanic debris flow, and bedding tilt tests have been carried out. Seven petrographically different breccia cobbles were sampled in the Westbank outlier from a debris flow (site 36) derived from nearby Eocene lavas. With one exception, the natural remanent magnetizations are directed to the east of northeast and downward (Fig. 7.8). Clearly, the magnetizations are predominantly overprints, postdating formation of the debris flow, otherwise their directions would have been widely scattered. The exception (specimen 4A) has

Figure 7.5: *A₁ magnetization. Hematitic lahar, Nimpit Lake Member, site 47. Direction (top) and intensities (bottom) during thermal demagnetization. Orthogonal plot on north-south vertical plane (middle). Inclinations downward.*

Figure 7.6: *AF magnetization. Trachyandesite flow, Kitley Lake Member, site 33. Direction (top) and intensities (bottom) during AF demagnetization. Orthogonal plot on north - south vertical plane (middle). Inclinations downward.*

Figure 7.7: B2 magnetization. Green ash flow in Shatford Creek Member, site 44. Direction (middle) and intensity (top) changes during AF demagnetization. Orthogonal plots are shown in bottom of figure. Inclinations upward.

a principal magnetization with a steep negative inclination that is retained up to 600°C, and that, therefore, predates formation of the debris flow. During thermal demagnetization, the magnetization of the remaining specimens either remained unchanged in direction or migrated in different directions. Some of the "migrating" specimens reach end points (2C of Fig. 7.8), indicating that the overprint has been removed. Others (8B) do not, indicating incomplete removal. All "migrating" specimens have distributed unblocking temperatures, and all magnetizations removed during demagnetization are directed to the northeast and downwards. Evidently the debris flow was probably of a volcanogenic origin and formed at temperatures of about 200°C, which was sufficient to partially remagnetize some specimens and completely remagnetize others. The directions of the high-temperature magnetizations are scattered, indicating that they were acquired before the cobbles were incorporated in the debris flow.

4. *Magnetostratigraphy*

Church (1980b) has previously found that the natural remanent magnetization (undemagnetized) of 11 samples spread throughout much of the thickness of the Marron Formation had normal polarity, indicating that only one polarity chron was represented. Although most of our samples are normal, some have reversed polarities. Samples of the Shatford Creek Member at the bottom of the overlying Marron Formation (sites 44, 45) have directions that are neither normal nor reversed, and differ from one another by much more than would be expected from paleosecular variation. Passing upward into the Kitley Lake Member, the directions stabilize around the normal direction and retain the same polarity through the Marron. Hence, the widely varying directions at the bottom of the Marron Formation probably represent a reversal transition in the geomagnetic field that occurred at the time the lower Marron lavas were extruded. Because it takes only a few thousand years for the field to reverse (Hospers 1954), this transition could provide a very accurate stratigraphic marker horizon. The Trepanier porphyry, has reversed polarity (site 21, Table 7.3, Fig. 7.4). The dacites we have assigned to the Marama Formation of Church (1973) are the youngest rocks sampled. Two sites are reversely magnetized (38, 39) and three (7, 13, and 15) are normal. Church (1980b) observed normal polarity at one site in the Marama dacite.

Table 7.3: Results by sites

Site	Lat., long (°N, °W)	C (S)	M (A/m)	D_H, I_H (°)	D_B, I_B (°)	D_C, I_C (°)	k	α_{95} (°)	Type
M1	49.9, 119.5	5 (10)	0.51	346, +62	68, +79	37, +79	36	8	B1
M4	49.9, 119.6	6 (13)	0.40	16, +59	72, +64	20, +61	75	5	A3
M5	49.9, 119.6	6 (12)	1.18	10, +59	69, +67	14, +61	98	4	A3
m7	49.9, 119.2	6 (11)	0.18	357, +65	357, +65	—	63	6	B
M8	49.9, 119.5	7 (13)	0.31	282, +46	358, +71	331, +71	17	10	B1
M9	49.9, 119.5	5 (6)	0.43	28, +64	54, +25	—	17	20	B2
M10	49.8, 119.7	6 (12)	2.07	322, +38	321, +68	321, +62	14	12	A1
M11	49.9, 119.7	6 (12)	1.10	322, +35	321, +65	322, +59	131	4	A1
M12	49.7, 119.9	5 (9)	0.67	48, +70	22, +68	45, +70	121	5	A3
m13	49.7, 119.1	5 (10)	0.17	94, +49	94, +49	—	5	3	B
m15	49.8, 119.3	4 (8)	0.09	321, +68	315, +55	—	28	9	A2
M16	49.8, 119.7	3 (4)	0.60	261, +58	12, +57	358, +66	84	10	B1
M17	49.8, 119.7	4 (4)	0.54	334, +69	24, +26	347, +66	193	7	B2
M18	49.9, 119.6	4 (7)	1.34	43, +56	31, +63	34, +62	90	6	B1
M19	49.9, 119.7	8 (15)	1.18	19, +60	6, +55	17, +60	115	4	A3
M20	50.0, 119.7	4 (8)	0.56	27, +62	95, +36	38, +63	67	7	B2
T21	49.8, 119.7	4 (4)	1.66	125, -51	134, -59	—	41	15	A1
M31	50.0, 119.4	6 (12)	0.51	312, +84	256, +84	311, +61	93	5	A2
M32	50.0, 119.6	4 (5)	2.07	290, +34	270, +56	276, +52	30	14	A1
M33	50.0, 119.6	7 (12)	3.88	24, +72	108, +38	43, +73	83	4	A2
M34	50.0, 119.7	5 (10)	0.74	356, +74	114, +45	16, +77	43	7	A2
M35	49.9, 119.7	6 (10)	1.02	53, +84	349, +79	43, +84	87	5	A3
W36(1)	49.8, 119.7	7 (12)	0.49	41, +70	60, +44	—	147	6	B2
W36(2)		7 (12)	0.49	Random	—	—	2	—	—
M37	49.8, 119.7	6 (10)	0.79	257, +27	308, +62	306, +66	34	8	A1
m38	49.9, 119.1	5 (9)	0.01	153, -53	153, -53	—	31	9	B2
m39	49.9, 119.1			Reversed					
M40	50.0, 119.6	6 (12)	0.69	36, +68	12, +74	34, +69	100	4	A3
M41	50.0, 119.6	5 (6)	0.61	15, +66	352, +68	13, +66	172	5	A3
M42	49.9, 119.5	7 (14)	2.91	298, +33	340, +56	328, +54	120	4	A1
M43	49.9, 119.5	9 (14)	1.93	275, +26	301, +66	290, +59	59	5	A1
M44	50.0, 119.4	5 (10)	0.05	202, -51	263, -65	—	73	8	B2
M45	50.0, 119.4	3 (6)	0.16	313, +81	170, +61	—	13	19	BT
M46	49.8, 119.7	6 (7)	0.26	222, -14	220, +41	—	8	23	BT
M47	49.8, 119.7	6 (12)	1.17	272, +54	1, +53	348, +60	178	3	A1
M48	49.8, 119.7	5 (6)	0.27	216, +28	173, +79	—	55	9	AT
M49	50.0, 119.5	7 (12)	0.34	242, +25	187, +52	—	74	5	AT
M50	50.0, 119.5	7 (12)	0.09	312, +36	42, +65	13, +68	76	5	A1
TSP51	49.3, 117.2	5 (10)	0.62	57, +63	—	—	154	4	A?
TSP5	49.3, 117.8	6 (12)	—	68, +51	—	—	14	12	A?
TSP6	49.3, 117.8	5 (9)	—	65, +37	—	—	77	6	A?
TSP7	49.3, 117.8	5 (10)	—	52, +54	—	—	20	11	A?

NOTES: (1) C (S) is the number of cores (specimens). M is the mean intensity of natural remanent magnetization. D_H, I_H is the mean site direction calculated giving specimen unit weight and relative to the present horizontal; D_B, I_B is the same relative to bedding given in Table 1; and D_C, I_C is the same after tilting required to optimize precision. k and α_{95} are precision and radius of the circle of confidence (Fisher 1953). Magnetization type and category are given in the last column. (2) Experimental procedures are as follows: site number [demagnetization applied, as either a single value for A-type magnetization or a demagnetization range for B-type, used for line fitting (number of specimens)]. M1 [15–60 mT (10)]; M4 [80 mT (12), 600°C (1)]; M5 [20 mT (11), 500°C (2)]; m7 [20–80 mT (9), 100–600°C (2)]; M8 [20–60 mT (13)]; M9 [10–60 mT (3), 100–600°C (3)]; M10 [20 mT (11), 600°C (1)]; M11 [20 mT (11), 550°C (1)]; M12 [20 mT (8), 600°C (1)]; m13 [20–80 mT (9), 300–600°C (1)]; m15 [20 mT (9), 550°C (1)]; M16 [20–60 mT (3), 300–700°C (1)]; M17 [40–100 mT (3), 100–650°C (1)]; M18 [20–60 mT (6), 200–700°C (1)]; M19 [15 mT (15), 500°C (1)]; M20 [20–50 mT (8), 100–450°C (1)]; T21 [50 mT (3), 600°C (1)]; M31 [40 mT (11), 500°C (1)]; M32 [40 mT (4), 550°C (1)]; M33 [40 mT (15), 400°C (11)]; M34 [20 mT (9), 400°C (1)]; M35 [20 mT (9), 400°C (1)]; W36(1) [low-temperature component, 100–300°C (7), W36(2) [high-temperature component, 550°C]; M37 [40 mT (5), 400°C (5)]; m38 [10–40 mT (1), 500–600°C (8)]; m39 [general low stability does allow accurate removal of present field VRM, but the most stable part of the remanence is reversed]; M40 [10–60 mT (11), 500–550°C (1)]; M41 [10–80 mT (5), 150–500°C (1), 4 specimens (2 cores) have very high intensities of NRM, their directions diverged from main group, and have sun and magnetic compass orientation that differ by $<10^\circ$ —they are regarded as effect of lightning strikes and are rejected]; M42 [40 mT (13), 550°C (1)]; M43 [50 mT (13), 600°C (1)]; M44 [20–60 mT (2), 500–600°C (8)]; M45 [20 mT (2), 10–100 mT (1), 100–400°C (2), 300–650°C (1)]; M46 [20–80 mT (2), 500–600°C (5)]; M47 [40 mT (2), 550°C (10)]; M48 [80 mT (1), 600°C (5)]; M49 [20 mT (8), 550°C (4)]; M50 [10–60 mT (14), 300–500°C (1), 3 specimens (2 cores) were rejected because of high NRM's and 60° divergence from the main group—considered to be effect of lightning]; TSP51 [20 mT (9), 550°C (1)]; TSP5 [40–60 mT (9), 500–600°C (3)]; TSP6 [40–60 mT (4), 550–600°C (5)]; TSP7 [20–40 mT (5), 500–600°C (5)].

Figure 7.8: *Debris flow, site 36. The "conglomerate" test is negative for low-temperature magnetizations, positive for high-temperature magnetizations. Solid and shaded symbols denote downward (upward) inclination. Strike and dip of bedding are shown top left. Typical thermal demagnetization curves (bottom) and corresponding direction (middle top) are shown. The shaded area (4) in the lower diagram is the envelope enclosing all thermal demagnetization curves, except that from specimen 4a, and shows the distributed nature of unblocking temperatures. (1) and (2) are high-temperature end points in four cobbles, negative and positive inclination respectively. (3) is the envelope of directions of magnetizations (downward inclination) removed in the range 100-350°C.*

Therefore, at least two reversals occurred, one above and one within the base of the Marron Formation, confining it, in all probability, to one normal polarity chron. The age of the Marron Formation (54 Ma) is sufficiently well known to assign the normally magnetized Marron to magnetic anomaly 22 transitional between chrons C22 and C21. It appears that the 2 km of Marron lavas were extruded in not more than approximately 0.5 Ma, the time span of anomaly 22 (Harland et al. 1982). As already noted, the total accumulation of the Marron lavas may have been about $1 \times 10^6 \text{ km}^3$, which, if correct, yields an extrusion rate of about $1 \times 10^6 \text{ m}^3/\text{a}$, about equivalent to minimum rates estimated for the Columbia River basalts (Baksi and Watkins 1973).

5. *Analysis of data from lavas*

In Table 7.3, site mean directions are given with respect to present horizontal plane and to bedding. The exceptions are the dykes at sites 7 and 13, whose directions are given with respect to horizontal only; because no estimate of paleohorizontal can be made, they are excluded from further analysis. The within-site errors (α) range from 23 to 3° , but generally are between 5 and 9° . Sites 9 and 46 have errors of 20° or more, which is considered too large; they too are not considered further.

The data are summarized in Table 7.4 and Figure 7.9. In all calculations, each site has been given unit weight. In entries (a) and (b) of Table 7.4 and Figure 7.9, *all* data from the Marron Formation are given with respect to present horizontal and to bedding; tilt corrections do not improve the precision, which remains poor ($k = 8$ before and 9 after correction). Dispersion is much greater than that expected from paleosecular variation. According to the model of McFadden and McElhinny (1984), the precision k expected from a random sample of the geomagnetic field at these latitudes would be much higher - in the range $20 - 30$. The precisions observed from coeval cratonic units are consistent with this model, although the value from the Bearpaw Mountains is greater (Table 7.5). Certainly, all are much higher than the values ($k = 8$ or 9) obtained from the above analysis of the Marron lavas. It is possible that paleosecular variation was particularly large during the deposition of the Kelowna lavas, and that this interval was not represented on the craton. However, if this were so, the bedding tilts observed are such that the precision *in situ* would be less than that after tilt correction, which is

Table 7.4: Summary of the paleomagnetic data

Entry	Ref.	N	D, I	k	θ_{63}	α_{63}	α_{95}	Lat., long. (°N, °E)	dm, dp (°)
(a) Marron all sites	H	24	323, 64	8	29	7	12	65, 154	18, 15
(b) Marron all sites	B	24	020, 72	9	27	6	11	76, 290	19, 16
(c) Marron pretilt (category 1)	B	12	343, 69	18	19	6	11	79, 172	18, 15
(d) Entry (c) 80% tilt correction	B	12	331, 68	22	17	5	10	71, 163	16, 13
(e) Marron posttilt (category 2)	H	5	352, 70	31	15	7	14	84, 193	24, 21
(f) Marron posttilt (categories 2, 3)	H	12	011, 69	38	13	4	7	82, 306	12, 10
(g) Entry (f) 10% tilt correction	H	12	016, 70	42	13	4	7	80, 304	12, 10
(h) Marron shallow dips (category 3)	H	7	023, 67	58	11	4	8	75, 322	13, 11
(i) Marron shallow dips (category 3)	B	7	025, 71	32	14	6	11	74, 301	19, 16
(j) Marron (c) + (e) + (h) (model A)	—	24	358, 70	23	17	4	6	86, 218	11, 09
(k) Marron (c) + (e) + (h) (model B)	—	24	—	9	26	6	10	86, 228	—
(l) Marron (d) + (g) ^a (model A)	—	24	352, 70	23	17	4	6	83, 195	11, 09
(m) Marron (d) + (g) ^a (model B)	—	24	—	9	27	6	10	96, 230	21, 20
(n) Marron (j) + extras (model A)	—	28	357, 69	23	17	3	6	87, 204	10, 09
(o) Marron (j) + extras (model B)	—	28	—	9	27	5	10	87, 217	—
(p) Marron (l) + extras ^b (model A)	—	28	352, 69	21	18	3	6	84, 184	10, 09
(q) Marron (n) + extras ^b (model B)	—	28	—	9	27	5	10	85, 197	—
(r) Sanpoil volcanics	H	83	027, 66	10	25	3	5	72, 322	08, 07
(s) Sanpoil volcanics	B	83	013, 69	12	23	3	5	80, 296	08, 07
(t) Eocene "intrusives"	H	4	060, 52	66	10	5	11	43, 329	16, 11

Notes: (1) N is number of sites; ref. is reference; H, with respect to horizontal, B, with respect to bedding; D, I are declination and inclination of the mean; k is Fisher's precision; θ_{63} is circular standard deviation, α_{63} and α_{95} are errors in mean direction or paleopole at $P = 0.63$ and 0.95 ; lat., long. are the coordinates of the paleopole; dm, dp are the semiaxes of the paleopole error oval. (2) (n), (o), (p) and (q) include site 44 (Shaford Creek Member), site 21 (Trepanier porphyry), site 36(1) (White Lake Fm.), and site 38 (Marama dacite). (3) (r) and (s) have been calculated from Table 2 of Fox and Beck (1985), excluding sites for which $\alpha_{95} > 10^\circ$; (r) is to be compared with entry (a) of our analysis, and (s) with (b). (4) (t) Average of data from sites TSP5–TSP7 and TSP51 from Eocene "intrusives."

^aConsidered to be the best summary of Marron Formation.

^bConsidered to be the best summary of Eocene rocks of the Kelowna outlier.

Figure 7.9: Mean site directions. Solid (open) symbols denote downward (upward) inclination. Top left show all sites, including polarity-transition sites. Other stereograms are labelled as in Table 7.4. Stars in (p) are south-seeking directions in reversed sites.

not. This, and the evidence from the debris flow, indicate that some Marron rocks acquired their magnetization before, and others after, tilting.

We begin by adopting the procedure of Schwartz and van der Voo (1984), taking first *all* site directions *in situ*, and then successively applying to them an increasing percentage of the attitudinal correction (Fig. 7.10). The assumption made is that the rocks acquired their principal magnetization when tilted by the same proportional amount, so that the precision should become maximum when corrections by that amount are made. There is a maximum ($k = 12$) after 60% correction, but the improvement is not great, and the dispersion is still over twice that expected from paleosecular variation at these latitudes. As tilt correction is increased, the mean direction migrates clockwise, and is closest to the expected Eocene paleofield after 60% correction. The fact that the optimum fit is after 60% correction indicates that about one-half the sites have pretilt, the remainder posttilt, magnetization. This confirms the conclusion reached in the preceeding paragraph.

Having established that magnetizations were acquired over a substantial time, we proceed by supposing that there are two main intervals of acquisition, one before tilting, and a second, after. Magnetizations were assigned to one interval or the other by comparing their direction. In this way two categories of magnetization can be recognized. For example, it can be argued that the principal magnetization at site 47 is almost certainly pretilting because the direction, with respect to bedding, is close to the expected Eocene field, whereas the direction with respect to present horizontal implies improbable 90° anticlockwise rotation (Figs. 7.5, 7.11). With respect to bedding, the deviation from the expected Eocene field is small, being that expected either from paleosecular variation or from uncertainties in determining bedding attitude. In contrast, the principal magnetization at site 33 (Figs. 7.6, 7.11) is almost certainly posttilting, because the direction with respect to present horizontal are close to the expected Eocene direction, whereas with respect to bedding the directions indicate that site 33 has rotated clockwise by 100° relative to site 47 and was situated about 3000 km further south than site 47; the latter can be discounted because both sites belong to the same mapped volcanic field. Most sites can be readily assigned to one of these two categories. We recognize also category 3 magnetizations, which cannot, because of the low dips ($< 10^\circ$) at the sites from which they were obtained, be allocated to either

Figure 7.10: Variation of precision (k), standard deviation (θ_{63}), and mean direction with degree of unfolding in all sites. Marron Formation. EE is the expected Eocene paleofield direction (Table 7.5).

Figure 7.11: *Pretilting (category 1) (top) and posttilting (category 2) (bottom) magnetization. The strikes and tilts are given. WRH and WRB, with respect to horizontal and bedding respectively. EE is the expected Eocene field.*

Figure 7.12: Variation of precision (k), standard deviation (θ_{63}), and mean direction with degree of unfolding of category 1 sites, Marron Formation. EE is the expected Eocene field.

Figure 7.13: Variation of precision (k), standard deviation (θ_{63}), and mean direction with degree of unfolding for categories 2 and 3 sites, Marron Formation. EE is the expected Eocene field.

category. Magnetization type (A or B) and category (1, 2 or 3) are given in Table 7.3.

Incremental attitudinal corrections were then applied to the categories separately (Figs. 7.12, 7.13; entries (c)-(i), Table 7.4). For categories 2 and 3, the assumption that the magnetization is entirely posttilting is probable because the precision ($k = 38$) is comparable to that expected for paleosecular variations at these latitudes (McFadden and McElhinny 1984) and to those obtained from the craton (Table 7.5). Even so, partial unfolding yields a small increase after 10% correction (Fig. 7.13). The mean direction, at first quite close to the expected Eocene paleofield (EE on Fig. 7.13), migrates away from it as corrections are increased, generating a large and, we argue, erroneous apparent tectonic rotation. For category 1, the precision improves from 9 to 18 after total correction, with a maximum of 22 after 80% unfolding (Fig. 7.12). As corrections increase, and in contrast to categories 2 and 3, the mean direction migrates towards the expected Eocene paleofield direction, becoming essentially coincident with it. Hence we conclude that category 1 magnetizations are almost, but not entirely, pre-tilting.

Site directions after optimal tilting (entries (d), (g), Table 7.4) have been averaged to obtain entries (l) and (m). A notable feature of Figs. 7.10, 7.12, and 7.13 is the way in which erroneous declination anomalies (apparent tectonic rotations) are generated if incorrect tilt corrections are applied, that is, if incorrect assumptions about the paleomagnetic horizontal are made. To provide an overall summary for Eocene lavas of the Kelowna outlier (entry (p) of Table 7.4), data from four other sites also have been included. These are from a tuffaceous bed in the Kitley Member (site 9), the Shatford Creek Member (44), the Trepanier porphyry (21), and from the somewhat younger Marama dacite (38).

6. Data from Eocene "intrusives"

Data from Eocene intrusions of mid-crustal origin in the vicinity of the Slocan Lake fault show excellent internal coherence. An example from the unmetamorphosed syenite near Ymir is shown in Figure 7.14. The directions are consistent throughout the demagnetization range, the range of unblocking temperatures and coercive force being typical of those for fine-grained magnetite. The directions are to the northeast and downward, significantly different from those observed in the Kelowna outlier (Table 7.4).

7. *Tectonics*****

The paleomagnetic data may be analysed in a context of local tectonics with respect to the local evolution of extensional faults but also with respect to more global tectonics of the southern Canadian Cordillera.

7.1 Time of tilting of the Kelowna lavas

We have just shown that the magnetization process occurred over a period extending from the time of initial cooling following extrusion until after tilting. If this period had exceeded the length of a polarity chron (less than 1 Ma for the middle Eocene), then the polarity sequence of Figure 7.4 would not have been observed. Hence the time lapse between deposition and tilting probably did not exceed 1 Ma; tilting followed deposition very closely.

7.2 Absence of rotations

Paleopoles are available from the craton for the interval 53-47 Ma spanning the time within which the Kelowna lavas lie (Table 7.5). Other data of about this age are available, some a little older (Paleocene intrusions in Montana, 67-62 Ma, Jacobsen et al. 1980), others younger (Rattlesnake Hills intrusions in Montana, 44 Ma, Sheriff and Shive 1980). They have not been included to ensure that comparisons are made within the tightest possible time limits. Data from the Monterey intrusives of Virginia (47 Ma, Lovlie and Opdyke 1974) are not included because they are based on only six sites (see Diehl et al. 1983). To have included these data would not have significantly affected the analysis.

The mean cratonic paleopole and that for Eocene lavas of the Kelowna outlier are in excellent agreement (Fig. 7.15). This shows that there has been no *large general* rotation of the Kelowna outlier. Motion along the Okanagan valley fault appears to have been simple and extensional, without large systematic rotations about vertical axes, a result which agrees with the local structural evidence reviewed earlier, namely the evidence for simple east-southeast to west-northwest extension of the Okanagan valley fault without rotation about local vertical axes. However, between-site dispersions (Fig. 7.9(p)) are such that small rotations of several degrees among sampling localities are not excluded.

Figure 7.14: Thermal and AF demagnetization of Coryell syenite, site 51.

Table 7.5: Paleopoles from cratonic North America

Rock unit	Age (Ma)	N	k	Lat., long. (°N, °E)	A_{95} (°)
Bearpaw Mountains volcanics, Montana ^a	50–54	18	72	81, 198	6
Absaroka Supergroup volcanics, Wyoming ^b	48	19	21	84, 177	10
Highwood Mountains volcanics, Montana ^a	49–53	29	28	81, 167	7
Mean				83, 180	6

NOTES: (1) N is the number of sampling sites; k is the precision of directions (2) Paleofield direction expected at Kelowna is 352, 68, $\alpha_{95} = 4^\circ$.

^aDiel *et al.* (1983).

^bShive and Pruss (1977).

Fox and Beck (1985) have reached the opposite conclusion, based on an extensive study of the Sanpoil lavas from several grabens spread over 120 km in north-central Washington (FB on Fig. 7.2). Their rocks are coeval and petrologically similar to those we have sampled. Summaries of their data, relative to bedding and to present horizontal, have been compiled in Table 7.4 (entries (r), (s)). The precisions of the Sanpoil data before and after correction ($k = 10$ and 12 respectively, Table 7.4, entries (r) and (s)) are much less than those expected from paleosecular variation, and are comparable to those at Kelowna ($k = 8$ and 9 respectively, Table 7.4, entries (a), (b)).

As in our study, correction for bedding does not improve the precision significantly. Inclinations in the Sanpoil lavas, like those we have observed, are consistent with cratonic data, and indicate the absence of any significant latitudinal motion. After full correction for bedding tilt at all Sanpoil sites, the declinations are to the north-northeast (Table 7.4, entry (s)). Fox and Beck (1985) interpreted this as indicating a 25° clockwise rotation of their sampling area.

Likewise, our data, after correction of all sites for full geological tilt (Table 4, entry (b)), would indicate a clockwise rotation of 28° , not significantly different from that invoked by Fox and Beck (1985). We argue that, for the Kelowna lavas, this would not be a real tectonic effect, but would be an artifact of erroneous bedding corrections made on rocks whose magnetizations were acquired over a time spanning deposition and tilting. When precision is maximized in a manner consistent with the physical evidence, then data from the Kelowna lavas agrees with those from the craton, both in respect to their mean direction and to their precision; when precision is maximized, the data fit very well the cratonic field. The precision obtained is not significantly different, either from that predicted by models of the geomagnetic field over the past few million years, or from that observed in middle Eocene cratonic rocks. Our data are entirely consistent with the null hypothesis that, relative to the craton, there have been no large rotations about a vertical axis in the Kelowna area; there is no basis in the paleomagnetic observations from Kelowna to invoke significant systematic clockwise rotation. We wish to stress that the difference between our interpretation of the Kelowna data and Fox and Beck's does *not* arise from differences in observations (the data sets seem similar), but from the application of a different mode of analysis that makes different assumptions about the paleomagnetic horizontal.

Figure 7.15: Paleopoles for middle Eocene rocks of cratonic North America and Kelowna (K) compared. Standard errors ($P = 0.63$) are given. Cratonic data compiled in Table 7.5.

Figure 7.16: Mean direction of Eocene "intrusions" from core-complex compared with those expected from craton (EE) and observed at Kelowna. Arrow indicates the tilting proposed and corresponding direction change.

7.3 Tilting in the zone of Eocene extension in south-central British Columbia

Turning now to the discrepancy between data from the Eocene "intrusives" and those from the Kelowna outlier, this discrepancy could have been produced in one of two ways - by tilting of the sampling region to the west (275°) at $37 \pm 10^\circ$ (Fig. 7.16) or by clockwise rotation of $68 \pm 16^\circ$ combined with translation from the south of $19 \pm 12^\circ$. The latter can be discounted, because excellent paleolatitudinal conformity has been demonstrated to the immediate west at Kelowna (Fig. 7.15), to the far west in the Eocene Flores Formation of western Vancouver Island (Irving and Brandon 1990), and to the south in northern Washington (Fox and Beck 1985). Hence, a westerly tilt of 37° around a mean strike of 350° is the more probable explanation of the aberrant data from the Eocene "intrusives".

In a qualitative sense, the amount of tilt calculated from the paleomagnetic data are reasonable considering that these sites are located between large faults between which "trap door" rotations about horizontal axes may have occurred (Wernicke 1985). It appears, therefore, that in situations such as this, in which large rotations about vertical axes and translations can be ruled out, paleomagnetic directions can be used to provide quantitative estimates of regional tilt in rocks of middle and lower crustal origin which otherwise contain no signature of paleohorizontal.

7.4 Possible significance for interpretation of other paleomagnetic aberrancies in the Cordillera

The discordance observed in Eocene "intrusives" is similar to those observed in mid-Cretaceous intrusive rocks of the northern Cascades (Beck and Noson 1972; Irving et al. 1985) of the coast Plutonic Belt (Symons 1977), and of the Intermontane Belt (Armstrong et al. 1985). The latter have been interpreted as caused either by systematic tilting over this very wide region of the western Cordillera to the southwest at 30° , or by clockwise rotation of 60° and translation from the south by over 2000 km (Beck et al. 1982; Irving et al. 1985). This similarity could be a coincidence, and without common cause. On the other hand, the similarity could be taken to indicate a common cause, for example, systematic trapdoor tilting over a much wider area of the western Cordillera in Canada than previously has been contemplated. The fact that the extent of regional extension in southeastern British Columbia has only recently been recognized (Parrish

et al. 1988; Carr 1990) makes it possible that this idea might bear discussion. Such a hypothesis would imply the existence of very large crustal extension faults in the Intermontane and Coast Plutonic belts which are currently being recognized (see Butler et al. 1989 and discussion by Umhoefer and Magloughlin 1990; Coleman and Parrish 1990; Carr 1990).

8. Conclusions

Despite careful and widespread sampling throughout the Kelowna outlier syn-extensional tilting within Eocene lavas stranded atop the Okanagan valley fault have not been recorded. Instead it was discovered that the apparent spreading of paleomagnetic directions is an artefact caused by the protracted acquisition of magnetization before and/or after bedding rotation, that is before and/or after deformation of the volcanic sequence. Pre-tilting magnetization is a "normal" phenomenon but it may appear that in the context of extensional tectonics some rocks lying above the main denudation surface would be reheated to reasonably high temperatures (300°C) as indicated by the thermal resetting of biotites in granites underlying the Eocene lavas analysed herein. Syn-tectonic reheating may not be a pervasive and homogeneous phenomenon and seems to have the potential of generating post-tilting magnetizations in those minerals with relatively low to medium blocking temperatures. Thus sampling for paleomagnetic directions in a context of intense deformation accompanied by thermal anomalies seems to produce apparent rotations if reheating processes are omitted especially when dealing with upper crustal rocks. The situation is quite different when dealing with Eocene mid-crustal anatectic rocks because they likely underwent simple cooling during uplift. Therefore these rocks record unimodal paleodirections which in this case indicates that footwall rocks of the Slocan valley fault have been tilted 37° westerly which appears to offer an alternative explanation to anomalous paleodirections obtained in Mesozoic rocks of several structural belts of the Canadian Cordillera.

CHAPTER 8 Geodynamics of extension in south-central British Columbia

1. *Introduction*

Systematic studies on the kinematics and dynamics of the continental collision zone underlying the hinterland of the southern Canadian Cordillera, the Omineca Belt, reveal that extensional strain followed immediately at the end of major crustal thickening events (Ranalli et al. 1989; Brown and Carr 1990; Carr 1991, 1992). This paper considers certain mechanical and thermal conditions that may have affected the lithosphere at the transition from compressional to extensional tectonic regimes in the southern Omineca Belt during the Early Tertiary. The effects of subduction of the Kula-Farallon plates, critical crustal elevations, lithospheric delamination and the presence of anatectic melts at mid-crustal depths (Carr 1990) are examined. It is proposed that these three concomitant elements led to instantaneous, short-lived, stretching of the crust rather than long-term mechanisms such as: (1) the action of a mantle plume (Gough 1986; Wilson 1991; but see Chapple and Tullis 1977); or (2) back-arc spreading (Ewing 1982; Davis and Lewis 1984; Lewis et al. 1988) combined with the anomalous heating effects created by tearing the oceanic slab along subducted transform faults forming lithospheric "windows" and asthenospheric upwellings (Thorkelson and Taylor 1989).

It has been estimated that the crustal thickness of the southern Omineca Belt reached 50 to 60 km at the end of compression (Brown et al. 1986; Journeay 1986; Ranalli et al. 1989). Its present average thickness of 32 km (Cook et al. 1988, 1992; Brown et al. 1991) indicates a thinning factor of 1.4 to 1.7 X translating into 40% to 70% extension assuming 10% of erosion of the pre-extensional crust. This is nearly the horizontal stretching factor (80%) estimated by fault-to-fault reconstructions along the 50°N latitude assuming planar fault geometries (Parrish et al. 1988). Thinning factors may be larger if important volumes of magmatic rocks intruding at sub-crustal levels since the Eocene are omitted.

The southern Omineca Belt is characterized by high heat flow values of 80 mWm⁻² (Davis and Lewis 1984) and anomalous upper mantle densities with P_n velocities of 7.6 km/s (Olsen 1983). The high heat flow is herein considered as a relict of Eocene extension given the time to reach thermal equilibrium (Mareschal and Bergantz 1990).

2. *Tectonic setting*

Early Tertiary extension has developed mainly in the intracontinental setting of the southern Omineca Belt where four major metamorphic culminations, the Shuswap-Valhalla-Kettle River and Okanagan complexes exhibiting mid- to lower crustal rocks underlie major denudation faults (Fig. 8.1; Snook 1965; Lane 1984; Parkinson 1985; Tempelman-Kluit and Parkinson 1986; Carr et al. 1987; Parrish et al. 1988; Carr and Parkinson 1989; Johnson 1989b; Carr 1990). Rocks in the hanging walls of the denudation faults are at greenschist-grade and related to the allochthonous Intermontane Superterrane (Fig. 8.1). In the hanging wall of the Okanagan valley fault these rocks are capped by intermediate lavas and alluvial strata (*see* Chapter 6). The geochronometric and geochemical data obtained from anatectic magmas and intermediate lavas in the footwall and hanging wall of the Okanagan valley fault suggest that intrusive and extrusive magmatism occurred nearly simultaneously between 56 Ma and 51 Ma and likely are cogenetic.

Prior to extension, rocks underlying the southern Omineca Belt had been subjected to compression caused by the major collision between the oceanic Intermontane Superterrane and the paleocontinental margin of western North America in the Early Jurassic (Monger et al. 1982; Brown et al. 1986, 1990). Compression renewed when collision of a second Superterrane (Insular) occurred west of the Intermontane Superterrane creating an overthickened crustal welt at the site of the southern Omineca Belt (Fig. 8.1; Brown et al. 1986; Journeay 1986) in Late Cretaceous-Paleocene time (Monger et al. 1982; Carr 1990). Major thrust sheets issued from middle and lower crustal levels underlie the Omineca Belt (Journeay 1986). The imbrication of these sheets transferred to thick-skinned thrusting of supracrustal rocks in the Rocky Mountains (Simony et al. 1980; Price 1981; Brown et al. 1986, 1990). Structural styles and metamorphic assemblages observed in basement rocks of the southern Omineca Belt indicate that crustal blocks may have been overthickened by 25 km increasing the crustal thickness to nearly 60 km at the end of the compressional period (Journeay 1986; Ranalli et al. 1989; Brown and Carr 1990). This shortened and thickened crustal welt was presumably compensated by a crustal root and a thickened sub-crustal lithosphere (Ranalli et al. 1989).

Figure 8.1: *Tectonic reconstruction map of the paleocontinental configuration of western North America ca. 55 Ma. Intracontinental stripped and dotted pattern outlines high-grade gneiss domes defining the SC: Shuswap complex; OC: Okanagan complex; KC: Kettle complex; VC: Valhalla complex. M: Monashee dome in the Shuswap complex exhibits North American basement rocks. Early Tertiary transcurrent faults are NRMT: northern Rocky Mountain Trench (Gabrielse 1985); Y-P: Yalakom-Pasayten fault (Monger 1986); H: Harrison Lake fault (Monger 1986); FR: Fraser River fault (Monger 1986). The extended domain of the southern Omineca Belt was located only 300 km inland from the paleocontinental margin. Perhaps the accretionary prism underlying the Olympic Mountains was the locus of impingement between the subducting Kula-Farallon ridge and the paleocontinental margin (Engelbreton et al. 1985; Heller et al. 1987). Extensional structures are compiled from Spencer and Reynolds (1986), Heller et al. (1987) and Brown and Carr (1990). The Yellowstone hotspot site 55 Ma ago was impinging onto the coast of northern California (Engelbreton et al. 1985; Heller et al. 1987). Notice also that the Eocene coastline was lying more than 400 km inland, much closer to the extended terrane, and has been shifted westward due to Late Tertiary extension in southwestern U.S. (Spencer and Reynolds 1986). Inset: Plate configuration of Baja California 20 Ma ago (Tennyson 1989). Notice the position of the East Pacific rise with respect to extended terranes in southern California and western Arizona. The outline of Figure 8.2 crosses the Okanagan complex.*

Compression in the Omineca Belt ended 58 Ma ago (Carr 1992) and transferred nearly instantaneously to a system of ESE-WNW extension (Fig. 8.1; Carr 1990; *see* Chapter 2) that lasted approximately 10 Ma (*see* Chapter 4 p. 121). Syn-extensional erosion of the Omineca crustal welt seems to have provided large volumes of clastic sediments that were deposited eastward in the Cordilleran molasse between 70 and 45 Ma (Mack and Jerzykiewicz 1989; Leckie 1989) and westward within continental and coastal basins between 55 and 40 Ma (Johnson 1984; Tabor et al. 1984; Heller et al. 1987; Taylor et al. 1988). Large volumes of sediment derived from metamorphic terranes were also carried further west to the Pacific in the form of alluvial sequences and sedimentary melange engulfed in the accretionary prism underlying the Olympic Mountains in northwestern Washington. This prism probably formed at the front of the subducted Kula-Farallon ridge (Heller et al. 1987), a setting that may have resembled that of the present East Pacific Rise impinging on the coastline of Baja California (Tennyson 1989).

The absence of major Early Tertiary sedimentary sequences within the Omineca and the Intermontane Belts suggests that both regions were probably at high elevation. The elevation may have been high enough to be affected by gravitational collapse as modeled by Arthuyshkov (1973); this hypothesis will be tested in the following sections after structural cross-sections and palinspastic restorations removing the effects of the Eocene structures have been presented.

3. *Overall extension across the southern Omineca Belt*

Published data are used (Lane 1984; Carr et al. 1987; Parrish et al. 1988; Carr and Parkinson 1989; Cook et al. 1992; *see* Chapter 5) to infer the age, the geometry and the magnitude of displacement of each extensional fault across the Omineca Belt along the 49°30'N latitude. The amount of horizontal extension is evaluated by comparing the crustal geometries before and after the extensional event. The towns of Nelson and Princeton in the hanging walls of the Slocan Lake fault and the Okanagan valley fault respectively are the reference points between which extension has been determined (Figs. 8.2, 8.3). The estimate assumes that faults are listric and flatten at present mid-crustal depth except for the Slocan Lake fault which clearly transects the entire crust (Cook et al. 1988, 1992).

Figure 8.2: *Tectonic map of south-central British Columbia and boundary regions as modified from Potter et al. (1986), Parrish et al. (1988) and Cook et al. (1992). Cross sections of Figure 8.3 are drawn along the line of section shown. Metamorphic domes and faults are designated as follows: [west]> BDF: Boundary fault (McMechan 1983); OVF-ERF: Okanagan valley-Eagle River fault system (Parkinson 1985; Bardoux 1985; Tempelman-Kluit and Parkinson 1986; Brown and Journeay 1987; Johnson 1988, 1989a); OC: Okana(o)gan-complex (Snook 1965; Christie 1973; Parkinson 1985; Bardoux 1985); [north]> SC: Shuswap complex (Okulitch 1984); MD: Monashee décollement (Brown et al. 1992); MC: Monashee complex (Brown et al. 1986); CRF: Columbia River fault (Lane 1984; Lane et al. 1989); [center]> CF: Cherryville fault (Carr 1990); BF: Beaver fault (Carr 1990, 1991; Cook et al. 1992; Brown et al. 1992); [south]> GWF: Greenwood fault (Monger 1968; Little 1983; Parrish et al. 1988); GF: Granby fault (Little 1983; Carr and Parkinson 1989); KC: Kettle River complex (Cheney 1980; Armstrong et al. 1991); KF: Kettle River fault (Cheney 1980; Parrish et al. 1988); [east]> VSZ: Valkyr shear zone (Carr et al. 1987; Parrish et al. 1988); VC: Valhalla complex (Carr et al. 1987; Parrish et al. 1988); GCSZ: Gwillim creek shear zone (Carr et al. 1987; Parrish et al. 1988); SLF: Slocan lake fault (Carr et al. 1987; Parrish et al. 1988); PTF: Purcell trench fault (Parrish et al. 1988). TAB is the Terrane accretion boundary. Extensional faulting sequence is (ductile-brittle) VSZ-SLF = 59-56 Ma (Parrish et al. 1988); (ductile-brittle) OVF-ERF-BF: 56-51 Ma (see Chapter 4); (brittle) GWF-GF-KF: 51-50 Ma (Cheney 1980; Carr and Parkinson 1989).*

UPPER CRUST (hanging wall of extensional faults)

Stratified ? Paleozoic (meta) sediments and volcanics (accreted)

Mesozoic granitic rocks

Eocene volcanic and sedimentary rocks

MIDDLE TO LOWER CRUST

Coryell intrusives

Footwall rocks of North American affinity (include ? basement gneisses / ? Proterozoic paragneisses)

Late Paleocene-Eocene leucogranites

FAULTS Compressional shear structures
 Ductile-brittle Eocene extension
 Brittle Eocene extension
 Thrust
 TECTONIC ACCRETION BOUNDARY
SEISMIC LINES Lithoprobe 1988
 Cocorp 1984

Figure 8.3:

Present and restored crustal cross sections drawn along line shown in Figure 8.2. Certain Mesozoic and Cenozoic intrusions are omitted. Hadrynian and Paleozoic rocks of North American sources are illustrated without shading. Top, present crustal structure that consists of six listric normal faults dipping outward from the Omineca crustal welt that was created until approximately 60 Ma ago (Carr 1990) and restored in the middle section. The upper crust is dissected by major extensional listric faults that formed at different structural levels and during three distinctive periods. The east-dipping Valkyr shear zone (VSZ) underlying Valhalla complex (Carr 1985; Carr et al. 1987) was activated as a ductile structure between 59 and 56 Ma (Parrish et al. 1988). East and west dipping ductile normal faults (SLF, KF and OVF) formed between 56 and 51 Ma. Late brittle structures are west-dipping (GWF, GF) and were activated between 52 and 50 Ma (Carr and Parkinson 1989). High grade metamorphic rocks are exposed in the footwall of most of these faults but their relative age is only tentatively based on their palinspastic reconstruction. By the end of extension the distance between Nelson and Princeton on both sides of the section had increased by 52 % a slightly lower estimate than that of Parrish et al. (1988). Middle, cross section restored to approximately 60 Ma showing the relative position of the different generations of Eocene normal faults outlined by different patterns of dashed lines. Middle Eocene strata might have deposited first in a basin bounded by the breakaway zones of the SLF, KF, and OVF. Several other basins may have formed independently at slightly different times in the hanging wall of the GWF and the GF. Crustal shortening prior to extension involved basement thrusting in pyggyback style (Brown et al. 1986). Shortening accommodated is approximately 200 km along shallow dipping thrusts and décollements. MD is the Monashee décollement (Brown et al. 1986, 1992) and GCSZ is the Gwillim Creek shear zone underlying, respectively, the Monashee Terrane (Ranalli et al. 1989) and the Valhalla complex (Parrish et al. 1988). Numbers along approximate locations of basement thrusts correspond to a possible temporal sequence. Bottom, section palinspastically restored prior to major shortening of North American supracrustals, at the time of obduction of Quesnellia terrane and Mid- to Late-Jurassic anatectic plutonism. Back-thrusting occurred at the collision front involving both North American rocks and Quesnellia rocks. It is proposed that back-thrusting was compensated by basement underthrusting (Brown et al. 1986) creating an anomalous crustal thickness.

- Middle Eocene sedimentary and volcanic rocks
- Middle Eocene syenite
- Early Eocene Ladybird granite
- Mid-Cretaceous granite
- Mid-Jurassic granitic rocks
- Intermontane Superterrane
- Hadrynian-Paleozoic strata
- Proterozoic Purcell-Belt Supergroup
- Crystalline basement
- 59-56 Ma ductile-brittle normal fault
- 56-51 Ma ductile-brittle normal fault
- 52-50 Ma brittle normal fault
- Thrust fault
- Tectonic accretion boundary

Mid-Jurassic palinspastically restored

The present-day crustal cross-section (Fig. 8.3a) was drawn partly from data interpreted on segments of the COCORP and LITHOPROBE seismic profiles shown on both sides of the section line in Figure 8.2 (Potter et al. 1986; Cook et al. 1992). Though the section line runs east-west it is assumed to follow nearly parallel with the transport directions recorded on both eastward- and westward-dipping extensional faults. Surface dips of approximately 30° and displacements of 10-20 km are attributed to the Slocan Lake fault (Carr 1987; Carr et al. 1987), the Columbia River fault (Lane 1984; Lane et al. 1989), the Granby fault (Carr and Parkinson 1989), the Greenwood fault (Parrish et al. 1988) and the Purcell Trench fault (Parrish et al. 1988). A shallower dip of 15° and a displacement of approximately 70 km is assigned to the Okanagan valley fault (Bardoux 1986; Cook et al. 1992; *see* Chapter 5 p. 127).

Extension faults seem to breakaway from the same area above the crest of the compressional crustal welt in the restored section (Fig. 8.3b). Field observations and geochronometric data indicate that extensional structures formed in two sets between 59-55 Ma and between 52-50 Ma which seem to refer to a specific setting with respect to the overthickened crustal configuration in pre-Eocene time (Fig. 8.3b). The oldest ductile-brittle extension faults formed at mid-crustal levels and restore to the most outward position from the compressional crustal welt. These are the Valkyr shear zone-Slocan Lake fault system (Carr et al. 1987; Parrish et al. 1988), the Columbia River fault (Lane 1984; Lane et al. 1989), the Kettle River fault (Cheney 1980), and the Okanagan-Eagle River fault system (Parkinson 1985; Bardoux 1985; Tempelman-Kluit and Parkinson 1986; Johnson 1988). These faults were activated nearly simultaneously (Parrish et al. 1988; *see* Chapter 4) and were still active in the brittle regime 50 Ma ago when the second set of strictly brittle extensional faults were activated. The brittle structures are the Greenwood fault (Monger 1968; Little 1983), the Granby fault (Carr and Parkinson 1989), and the Cherryville fault (Carr 1990). Brittle faults appear to have formed immediately after a major period of erosion (Heller et al. 1987) above the apex of the already exhumed compressional crustal welt between 52 and 50 Ma ago (Fig. 8.3).

At the end of the extensional period the distance between the Kootenay Arc east of Nelson and the central part of the Intermontane Belt west of Princeton has been increased by 52% relative to the fixed pins. This magnitude of extension is slightly less than that estimated by

Parrish et al. (1988) because displacements on individual faults are conservative and restricted to the upper crust. In order to maintain a balanced crustal section, it is expected that upper crustal stretching was compensated by stretching of the same magnitude in the middle and/or lower crust. Perhaps some lower crustal seismic reflectors on the COCORP and LITHOPROBE profiles (Potter et al. 1986; Cook et al. 1992) represent extensional shears.

The details of subsurface features in present and restored sections (Figs. 8.3a, b) are speculative but many structures match important reflectors identified in the seismic profiles. These include a flat Moho reflection at 32-35 km, shallow west-dipping basement thrusts and non-reflective lithologies of Proterozoic-Paleozoic sequences and strata of the Intermontane Superterrane. Stratigraphic sequences in the subsurface include Aphebian basement overthrust by Proterozoic-Paleozoic strata via the Monashee décollement (Brown et al. 1992). These rocks are in turn overthrust by allochthonous rocks of the Intermontane Superterrane. Most of these lithologies have been identified sporadically at the surface (Little 1961; Monger 1968; Carr et al. 1987; Tempelman-Kluit 1988; Armstrong et al. 1991) along the cross-section line. Their relative thickness is nevertheless unpredictable because they were intruded by large Mesozoic and Cenozoic intrusions and submitted to several phases of compressional deformation (Brown et al. 1986) besides they were locally attenuated by extensional shears. The thicknesses of basement (10 km) and Proterozoic-Paleozoic lithologies (10-12 km) are inferred from a restored section drawn along the 51°N latitude by Brown et al. (1986). The pre-extensional section (Fig. 8.3b) reflects the crustal geometry at the close of the compressional period, about 60 Ma ago, and assumes that basement blocks were imbricated in a duplex style and accommodated as much as 200 km of shortening (Brown et al. 1986). One of the consequences of the constructions made from restoring the geometries and magnitudes of the extensional faults is that the restored crust must have been as thick as 55 km which is consistent with the interpretations made by Brown et al. (1986), Journeay (1986), Parrish et al. (1988) and Ranalli et al. (1989).

4. *Pre-extensional elevation of the southern Omineca Belt*

The pre-extensional crust in the Omineca Belt and the Intermontane Belt is expected to have been thicker than average with elevations comparable to the Andes and Asia (Molnar and

Lyon-Caen 1988). Predicting these elevations prior to extension is critical to evaluate stress fields over topographic features (Artuyshkov 1973; Molnar and Lyon-Caen 1988; England and Houseman 1989). Estimates of elevations are poorly constrained but it can be argued that the region was probably higher than the surrounding ranges simply from isostatic considerations. It is expected that doubling the crustal thickness with a lithospheric root would necessarily raise the elevation to 2 km. Knowing from LITHOPROBE (Cook et al. 1992) that extensional faults across south-central British Columbia flatten at mid-crustal depth, except the Slocan Lake fault, the present-day crust being thinner than normal may indicate that the lithosphere has been attenuated. If the lithosphere was attenuated during extension then it can be proposed that the syn-extensional elevation may have reached up to 4 km. This data is consistent with the fact that most Early Tertiary sedimentary sequences occur on the outskirts of the Omineca and the Intermontane belts (Johnson 1984; Heller et al. 1987; Mack and Jerzykiewicz 1989).

The pre-extensional physiography may be tentatively estimated from the mid-Eocene flora preserved in the alluvial sequences closest to the southern Omineca Belt. Though these flora were deposited in depressions they probably represent an assemblage of species living in the Eocene environment and climate. These consist of *Alnus* (alder), *Picea* (spruce), *Pinus* (pine), *Betula* (birch), *Tsuga* (douglas fir), *Sequoiapollenites* (coast redwood), *Liquidambar* (sweetgum), *Quercus* (oak), *Ulmus* (elm), *Thuja* (cedar), and *Tilia* (McIntyre, pers. comm. 1986). They characterise a rain forest made of deciduous hardwoods and diversified conifers (Axelrod 1957) typical of mountainous altitudes of at least 1 km which is probably a minimum value for the Eocene mountain range. Together the isostatic processes, the absence of Early Tertiary sedimentary strata, and the preserved fauna suggest that the Early Tertiary physiography of south-central British Columbia may have been about half the present average altitudes of the Greater Himalayas and the Central Andes (Molnar and Lyon-Caen 1988).

Most paleodrainage directions in Eocene sedimentary rocks on the west side of the southern Omineca Belt are from eastern sources (Heller et al. 1984). The lack of Early Tertiary strata in the foreland basin (Leckie 1989) may suggest that the Omineca Belt formed a major geographic barrier and a major drainage divide. Perhaps the pre-Eocene paleoslope was dipping gently westward from a ridge centered on the Monashee and Kettle complexes (Fig. 8.2). This

suggests that the Intermontane Belt was slightly lower than the Omineca Belt creating a westward dipping tapering crustal wedge.

5. *Gravitational collapse following major compressional uplift*

If only local stress differences caused by loads of wavelength on the order of the thickness of the lithosphere are implied a uniform two-dimensional load will tend to bring the system to an equilibrated crustal thickness (Bott 1965). It has long been recognized that isostatically-compensated uplift regions contain excess potential energy due to the accumulation of the relief itself and its compensating root leading to a state of extensional deviatoric stress (Evison 1960; Artuyshkov 1973; Molnar and Lyon-Caen 1988). Under the condition that horizontal compressional forces are small, the crust will tend to spread horizontally in order to diminish its potential energy. Large additional extensional stresses are added if the crust is underlain by light (hydrated) mantle (Arthuyskov 1973; Molnar and Lyon-Caen 1988).

Calculations were made to evaluate the deviatoric stress induced by topographic features with and without a crustal root (Appendix 4) at the end of the compressional event in south-central British Columbia in the Paleocene (Ranalli et al. 1989). Differential stresses are calculated assuming a layered elastic lithosphere with surface (topography) and internal (change in crustal thickness) loading using the method of Mareschal and Kuang (1986). Figure 8.4 illustrates the pre-extensional lithospheric conditions of the model with the distribution of the principal stresses and the contoured principal stress differences in MPa. Case 1 (Fig. 8.4a) shows the stress distribution for a topographic load of 3 km (Max. stress of ~100 MPa) compensated by an overthickened crust underlain by a thick lithospheric root (Ranalli et al. 1989). The noteworthy features of this interpretation are that the largest principal stress is concentrated in the most elevated region and that principal stresses are tensional in the upper crust and compressional in the underlying mantle. This indicates that extension would be impeded or be restricted to the upper crust under a stress difference reaching a maximum value of 60 MPa that is the same order of magnitude as the normal plate boundary stress estimated to be less than 50 MPa (Solomon et al. 1975; Forsyth and Uyeda 1975). Case 2 (Fig. 8.4b) involves the same topographic feature 3 km high compensated by a thickened crust but underlain by thinned to completely delaminated

Figure 8.4: *Case 1. Distribution of the stresses at the end of compression 58 Ma in south-central British Columbia. Modelled relief is 3 km (100 MPa), isostatic lithospheric root is 200 km. Case 2. Stress difference contours showing tensional conditions centered on the zone of highest relief.*

Stress difference contours
interval: 20 MPa

lithosphere (Bird 1979; Houseman et al. 1981). Here, it is noticeable that the stress field becomes tensional throughout the lithosphere with a maximum principal stress difference of 100 MPa at the highest topographic site, that is twice the maximum regional deviatoric stress.

The magnitude of the vertical principal stress depends mainly on the height of the anomalous relief, the thickness of the lithosphere, and the critical taper conditions (Dahlen and Suppe 1988). Owing to these conditions, tensile stress will occur during and/or immediately after a threshold relief has been formed by compressional forces. This mechanism is enhanced if the continental lithosphere is delaminated (Houseman et al. 1981; Ranalli et al. 1989) but it is not a necessary condition to initiate crustal spreading.

6. *Extension in the southern Omineca Belt and the Eocene plate tectonic setting*

Perhaps Eocene extension in the southern Omineca Belt reflects changes in the motions of the oceanic Kula and Farallon plates relative to North America in a hotspot reference frame. Significant changes in plate configuration and plate velocities between these plates have been determined using Euler poles (Engelbreton et al. 1985). The genealogy of the convergence vectors for the Kula-North America and the Farallon-North America pairs indicates that between 65 and 50 Ma convergence between the Kula-Farallon plates and North America was oblique and at the most rapid rate since the Mesozoic (Engelbreton et al. 1985; Fig. 8.1). Northward velocities of the Kula and Farallon plates relative to North America reached values of at least 15 cm-yr⁻¹ between 65 and 50 Ma. Both plates decelerated very rapidly to less than 10 cm-yr⁻¹ (Fig. 8.1) between 50 and 45 Ma. Altogether the velocity data for the plate pairs considered are consistent with the tectonic evolution of the southern Omineca Belt in the Late Cretaceous-Early Tertiary. The Late Cretaceous compressional event recorded in this region (Scammell 1991; Carr 1992) occurred while young and presumably buoyant lithosphere of the Farallon plate was accelerating and rotating to the northeast maintaining a compressional tectonic regime. On the other hand the Eocene extensional episode in the southern Omineca Belt culminated while the subducting Farallon plate was drastically decelerating causing a relaxation of the compressive stress field leading to tensional stresses. It is expected that plate boundary conditions in the Late Cretaceous-Early Tertiary produced in the southern Omineca Belt produced a northward-trending

horizontal stress component. This component may have added simultaneously to the intracrustal tensional stress field created by anomalous topography at the end of the compressional event. Altogether the orientations of these stress fields are in agreement with the direction of the west-northwest stretching directions recorded in the footwalls of the extensional faults (Fig. 8.1).

The location of the Kula-Farallon-North America triple junction in the Late Cretaceous-Early Tertiary cannot be precisely constrained by ocean floor data. Amongst several options it has been proposed that the Kula-Farallon ridge intersected the continental margin somewhere between northern California and northern Washington (see Engebretson et al. 1985, p. 27). Several geological features such Late Cretaceous northwest-trending dextral strike-slip faults, thick accretionary material underlying the Olympic mountains, abundant Eocene deltaic deposits, and north-trending extensional features in south-central British Columbia-northern Washington are consistent with the postulated position of the Kula-Farallon ridge in northern Washington (Fig. 8.1; Heller et al. 1987). Perhaps the tearing forces caused by this ridge may have forced extensional stretching trends (Fig. 8.1) to develop obliquely from most compressional structures in the southern Omineca Belt and would suggest that gravitational collapse along ancient compressive features is not a unique solution to Eocene extension.

7. *Review of some geophysical data across south-central British Columbia*

Detailed accounts of diverse geophysical data in south-central British Columbia are found in Caner (1970); Wickens (1971); Stacey (1973); Olsen (1983); Davis and Lewis (1984); Gough (1986); Cook et al. (1988); Brown et al. (1992). Many of these geophysical characteristics are compiled on Figure 8.5 modified from Monger (1989). The crustal profile runs generally parallel to the 50°N latitude and nearly parallel to the crustal section in Figure 8.3a.

The Omineca Belt is magnetically "quiet" indicating that the crust has been thinned (Haines et al. 1971). It is also affected by anomalously high electrical conductivity related to either the lower crust/upper mantle boundary or hydration and/or partial melting in the lower crust (Caner 1970). These data imply high temperature and decreasing density conditions in the lower crust and upper mantle underneath the Omineca Belt in agreement with the Bouguer anomaly profile (Stacey 1973). Low density upper mantle is outlined by a low-velocity layer

Figure 8.5: *Two-dimensional section across the physiographic belts of the Canadian Cordillera (PO: Pacific ocean; IB: Insular Belt; CPB: Coast Plutonic Belt; OB: Omineca Belt; RM: Rocky Mountains; AP: Alberta Plains). Distribution of the seismic refraction data. Distribution of the electrical resistivity and Bouguer profile. Heat flow data in mWm^{-2} (Lewis and Davis 1984; Lewis et al. 1988). Hatched line indicates the expected position of the base of the lithosphere. Note that the western part of the lithosphere under North America is missing corresponding to a probable mantle wedge gap. Distribution of the principal crustal seismic reflection data from the LITHOPROBE project. Data compiled and modified from Stacey (1973), Clowes et al. (1987), Lewis et al. (1988), Monger (1989), Brown et al. (1990), and Cook et al. (1992).*

Figure 8.6: *Regional tectonic map of the central Okanagan valley area. Dots with values locate heat flow stations with their respective values across the Okanagan valley fault on the west flank of the Omineca Belt (modified from Davis and Lewis 1984). Other extensional structures are: CF; Cherryville fault, GWF; Greenwood fault, GF; Granby fault.*

characterized by $V_p < 7.6$ to 8.0 kms⁻¹ and $V_s < 4.61$ kms⁻¹ through the depth range of 40 to 240 km (Fig. 8.5; Wickens 1971; Stacey 1973; Monger 1989). According to the LITHOPROBE data (Cook et al. 1992) the base of the Omineca Belt is underlain by a flat reflection Moho at 32 km dipping constantly to 52 km eastward underneath the main ranges of the Rocky Mountains but remaining at a relatively constant depth of 30 km underneath the Intermontane Belt.

Uniformly high heat flow data have been recorded on the west flank of the Omineca Belt across the Okanagan valley fault (Fig. 8.6) and reach 85 mWm⁻² when corrected for glaciation effects (Davis and Lewis 1984). Davis and Lewis (1984) suggested that the heat flow data in the Okanagan region is produced by continuing heat advection emanating from present-day subduction-driven fluid circulation because it links with a typical subduction-related heat flow profile further west (Fig. 8.5; Lewis et al. 1988). However, it seems improbable that subduction related heat advection would be carried 500 km inland from the nearest subduction zone located seaward of Vancouver Island with a slab dip of approximately 25° (Fig. 8.5; Clowes et al. 1987). The lack of earthquake foci under south-central British Columbia probably indicates that the slab is either completely melted or that it has been absorbed in the mantle west of the Omineca Belt. Gough (1986) and Wilson (1991) proposed that heat flow in south-central British Columbia relates to a mantle plume modeled by a highly conductive layer ($k = 10 \pm 5$ Wm.m)(Dragert and Clarke 1977; Gough et al. 1982; Majorowicz and Gough 1991) presumably located in the upper mantle. However conductive layers may as well relate to metasomatic fluids in the crust (Stacey 1973; Gough 1986). Clearly there is no consensus reached on the origin of high heat flow in the Omineca Belt and we will consider that it is a relict of intracontinental extension.

8. *High heat flow related to extension on the west flank of the Omineca Belt*

Conductive heating through a lithosphere heated from below requires at least 75 Ma to be perceptible in the surface heat flow (Mareschal and Bergantz 1990). This time is reduced to 50 to 30 Ma if the lithosphere is mechanically stretched to reach steady-state (Pollack and Chapman 1977; Mareschal and Bergantz 1990). Eocene extension in the southern Omineca Belt will therefore affect the heat flow and induce high geothermal conditions, perhaps 40-50°C/km.

Syn- and post-compressional erosion and extension has exhumed deep crustal rocks to the

4 of/de 4

surface (Fig. 8.3a). These rocks are less radiogenic than upper crustal rocks ($0.4 \mu\text{Wm}^{-1}$) and account for less than 20 mWm^{-2} of the steady-state heat flux. The reduced heat flow (q^*) for the western Omineca Belt then reduces to 60 mWm^{-2} (Gough 1986). It is assumed that the abundant Eocene plutons in the southern Omineca Belt and hydrothermal circulation from the subducting Juan de Fuca plate underneath the Canadian Cordillera account for 10% of the average regional heat flow, that is 10 mWm^{-2} (Blackwell 1978). Thus the present heat flow (q) from the mantle underlying the Okanagan region is taken to be approximately 50 mWm^{-2} .

The amount of extension, assuming mantle delamination, can be linked to the measured heat flow by an equation established by Hamdani et al. (1991) which could also be derived from the temperature calculations of Royden and Keen (1980):

$$(eq. 1) \quad q = q_0 \left[1 + 2 \sum_{n=1}^{\infty} \frac{\beta}{n} \sin \left(\frac{n\pi}{L} \cdot \frac{M}{L} \right) \exp \left(\frac{-n^2 \pi^2 K t}{L^2} \right) \right]$$

where M is the crustal thickness, β is the stretching factor, L is the lithospheric thickness, K is the thermal diffusivity ($10^{-6} \text{ m}^2 \text{ s}^{-1}$).

Figure 8.7 shows the heat flow decay curve based on equation 1 after 100% of extension that is thinning the crust by a factor of 2. Calculations were made using a program written by J. C. Mareschal (Appendix 5) for a lithospheric thickness of 120 km, a crustal thickness of 50 km, and equilibrium heat flow (without crustal heat production) of 40 mWm^{-2} . Implicit in the calculations is that extension is instantaneous and uniform. In the case that extension may be partitioned in two layers the same patterns would develop with an overall increase of the decaying heat flow curve. In either situation, extension instantaneously doubles the heat flow. If uniform extension occurs without lithospheric delamination, heat flow returns to non-equilibrium values of the order of 60 mWm^{-2} after 50 Ma (Fig. 8.7a). When including delamination, the heat flow increases abruptly and reaches a minimum of 120 mWm^{-2} after 15 Ma; the transient heat then decays and the heat flow reaches 65 mWm^{-2} after 50 Ma (Fig. 8.7b).

Both models in Figure 8.7 are two end members of a spectrum of interpretations and it is noticeable that the difference in heat flow 50 Ma after the extension is small and the two models become indistinguishable. Therefore one cannot verify with this method whether the lithosphere was or was not delaminated. Regardless of the mechanism of extension (homogeneous

Figure 8.7:

- a. The change in surface heat flow at time 50 Ma as a function of the heat flow perturbation if the heat flow were raised uniformly at the base of the crust (60 km thick) after instantaneous extension.*
- b. The surface heat flow as a function of time for 100% instantaneous extension for the combined effects of lithospheric stretching and dike injection. In either case of uniform crustal extension or crustal extension and delamination the residual heat flow anomaly after 50 Ma would be in excess of 40 mWm⁻² after 100% extension.*

HEAT FLOW NON UNIFORM STRETCHING

HEAT FLOW AFTER EXTENSION

lithospheric extension or extension with delamination), it appears that the present thermal characteristics found in the Okanagan region are compatible with a factor of extension smaller than 100% probably closer to 50% occurring 50 Ma ago. Alternatively, this heat flow could have been produced by less extension of a thinner crust accompanied by abundant magmatism.

9. Discussion

Extension in the southern Omineca Belt may have resulted from "passive" denudation responding to regional and local stress fields caused by critical lithospheric conditions (Wernicke et al. 1987; Sonder et al. 1987; Brown et al. 1991) but magmatism has probably played an "active" role (Gans et al. 1989) facilitating strain partitioning (Carr 1989).

9.1 The causality between anatectic melts and extension

Crustal thickening can lead to widespread crustal melting under favorable hydration conditions (England and Thompson 1986). The abundance of S-type melts in the southern Omineca Belt (Carr 1990) suggest that rocks buried during compression were dominantly pelitic lithologies. Less abundant refractory I-type melts require that the thermal regime in the southern Omineca Belt was at least 75 mWm^{-2} which is consistent with the expected heat flow extrapolated from the present surface heat flow in the Okanagan complex. The thickening process in the Late Cretaceous was probably capable of producing dehydration melting of amphibolite with consequent production of calc-alkaline lavas (England and Thompson 1986; see Chapter 6).

The continental lithosphere of the southern Omineca Belt was probably strong enough to support a plateau-like physiography until the Late Cretaceous. However large volumes of melts derived from the lower crust migrated upwards to the middle crust in the Early Tertiary and probably produced ductility contrasts across the brittle-ductile decoupling zone (Sibson 1977). Carr (1990, 1992) demonstrated that the oldest melts (62-58 Ma) are affected by late compressional fabrics in the Thor-Odin region whereas younger melts (58-51 Ma) nucleate extensional features. Some of these melts may have concentrated along and reactivated segments of ancient compressional structures (Ivins et al. 1990) such as the accretionary boundary between the Intermontane Superterrane and North America along which the Okanagan valley normal fault

seems to root in restored sections. This may explain why the Okanagan valley fault has a shallower dip than most extensional structures across the southern Omineca Belt (Fig. 8.3).

Tectonomagmatic processes across the Okanagan valley fault have occurred nearly simultaneously during extension. Syn-extensional melts underlying the fault and undeformed calc-alkaline lavas overlying the fault provided U-Pb zircon dates of 56 ± 1 Ma and K-Ar cooling ages of 54 ± 2 Ma respectively (see Chapter 4) suggesting that both magmatic rock types are cogenetic. This magmatic evolution is consistent with the crustal thickening/thinning model of England and Thompson (1986) in which calc-alkaline lavas can be produced by isobaric cooling of hot lower crustal rocks during the exhumation period.

9.2 Comparing extensional settings in other orogenic belts

There are many examples of extension taking place in an orogenic belt (Molnar and Tapponier 1978; Allègre 1984; Channell 1986; Molnar and Lyon-Caen 1988; Channell and Mareschal 1989; England and Houseman 1989; Wang et al. 1989). In each case the potential energy conditions probably created tensional deviatoric stresses over elevated regions isostatically compensated by crustal and lithospheric roots. These stresses may have been of the same magnitude (100 MPa) as stresses resulting from lithospheric decoupling (Ranalli et al. 1989) and exceeded regional tectonic stresses estimated at 30 MPa (Forsyth and Uyeda 1975; Chapple and Tullis 1977; Bott 1982).

Certain plate boundary conditions, mainly oblique and decelerating subduction, may have added to the process of crustal extension in south-central British Columbia. Evidence is growing that when a ridge and/or a transform fault intersects a passive or an active continental margin it creates deep-seated lithospheric dislocations within that margin leading to tectonic denudation or complete continental rifting. Tectonic denudation appears to form principally along active continental margins which exhibit earlier compressive features. Perhaps the tectonic setting of southern British Columbia and northern Washington in the Early Tertiary resembled that of New Zealand in the Late Cretaceous (Weissel et al. 1977; Gibson et al. 1988) and southern Chili at Present (Dalziel and Brown 1989). Extensional domains in those regions developed inland from the site of subducting oceanic plates creating major compressed zones evolving to sialic

denudation and strike-slip faulting under decelerating and rotational plate convergence conditions.

The southern Omineca Belt and the Basin and Range Province share many geological and geophysical similarities. The seismic velocities and electrical conductivities recorded in both regions are comparable (Smith 1978) and could indicate that the lithosphere has been completely delaminated (Bird 1979) or converted into asthenospheric material by phase changes (Mareschal and Gliko 1991; Hamdani et al. 1991). High heat flow values also characterize both regions (Lachenbruch and Sass 1977, 1978; Blackwell 1978; Pollack and Chapman 1977) and probably relate to the extensional event (Mareschal and Bergantz 1990). The Basin and Range has undergone as much as 60-100% extension in the central portion (Wernicke et al. 1982). As the present crustal thickness ranges around 30 km (Smith 1978) the region must originally have resembled in elevation and crustal thickness (Coney and Harms 1984) the Andean and Asian plateaux which are presently extending (Molnar and Lyon-Caen 1988).

10. Conclusions

Extension of the continental crust in south-central British Columbia between 59-46 Ma may have been significantly controlled by the plate dynamics between North America, the Kula and the Farallon plates. Northeast trending crustal thickening in the Late Cretaceous transferred to east-west extension in the Early Tertiary during a major plate reorganization caused by the obliquity and decelerating subduction of the Farallon plate which progressively reduced compressional stress conditions. This change in the balance of boundary forces was accompanied by a combination of "passive" and "active" mechanisms leading to the state of higher gravitational energy and lithospheric stretching.

It is proposed that the Omineca and the Intermontane Belts were forming elevated ranges in the Early Tertiary. The combination of critical elevations, the presence of abundant anatectic melts at mid-crustal levels, and mantle lithospheric delamination led crustal extension to focus mainly, but not uniquely, on the region of maximum crustal shortening and maximum elevation.

Present-day heat flow data on the west flank of the Omineca Belt can relate to Early Tertiary extension and is consistent with the magnitude of extension estimated with restored cross-sections in the order of 50%.

Appendix 1 Summary of published isotopic data in the Okanagan valley

Table A. Published K-Ar Data (Okanagan gneiss complex)

Sample No.	Ref.	Rock Type	Mineral	Date Ma*	Lat. (N)	Long. (W)
=====						
Inkaneep 2	(1)	Gneiss	Bi	56.4	49°09.8'	119°29.8'
Anar. Mtn.	(2)	Phyllite	WR	53.6	49°01.7'	119°21.3'
JR-1	(3)	Rhomb.Por.	WR	42.4	49°16.7'	119°29.0'
JR-1	(3)	Rhomb.Por.	Bi	45.3	49°16.7'	119°29.0'
JR-1	(3)	Rhomb.Por.	Hb	48.4	49°16.7'	119°29.0'
M-15	(3)	Pegmatite	Ms	48.9	49°15.9'	119°32.8'
Osoyoos 2	(4)	Gneiss	Bi	50.3	49°01.4'	119°22.5'
1-50	(5)	Dike	WR	48.2	49°36.7'	119°34.6'
1-148	(5)	Gneiss	Hb	49.3	49°43.9'	119°31.3'
1-150	(5)	Gneiss	Hb	50.7	49°44.5'	119°31.0'
1-160	(5)	Dike	WR	49.8	49°39.2'	119°33.8'
1-178	(5)	Gneiss	Hb	51.8	49°42.6'	119°36.0'
3-5	(5)	Gneiss	Hb	53.0	49°51.6'	119°17.2'
3-8	(5)	Paragneiss	Hb	54.2	49°47.5'	119°08.0'
3-10	(5)	Paragneiss	Hb	61.0	49°16.5'	119°31.2'
3-13	(5)	Gneiss	Hb	53.3	49°26.6'	119°34.5'
3-21	(5)	Gneiss	Bi	52.7	49°42.8'	119°16.5'
3-22	(5)	Grd.	Bi	52.7	49°15.0'	119°10.5'
76-1	(6)	Gneiss	Hb	48.4	49°39.8'	119°36.1'
78-82	(6)	Gneiss	Bi	47.8	49°17.5'	119°30.0'
78-83	(6)	Gneiss	Ms	45.5	49°17.5'	119°30.0'
78-84	(6)	Gneiss	Bi	45.6	49°00.6'	119°24.3'
78-85	(6)	Gneiss	Hb	63.5	49°00.6'	119°24.3'
80-54	(6)	Gneiss	Hb	48.8	49°39.6'	119°36.1'
80-55	(6)	Gneiss	Ms	59.4	49°09.7'	119°30.3'

(1) Mathews and Armstrong, unpublished; (2) Soregaroli and Christopher, unpublished; (3) Ross (1974); (4) Fox et al. (1975, 1977); (5) Medford (1975); (6) Wanless et al. (1978) and Stevens et al. (1982).

* All dates corrected to decay constants recommended by Steiger and Jäger (1977)

Table B. Published fission-track analytical data* (Okanagan gneiss complex)

No.	Rock type	Mineral	Apparent age	Lat.(N)	Long.(W)
1-145	Grd. Gneiss	Apatite	41.2 $\pm$ 7.7	49°41.2'	119°34.4'
1-148	Grd. Gneiss	Apatite	47.8 $\pm$ 3.1	49°43.9'	119°31.3'
1-150	Grd. Gneiss	Apatite	42.4 $\pm$ 3.1	49°44.5'	119°31.0'
1-154	Grd. Gneiss	Apatite	61.9 $\pm$ 7.2	49°39.7'	119°34.3'
1-173	Grd. Gneiss	Apatite	53.6 $\pm$ 6.4	49°39.9'	119°39.0'
1-178	Grd. Gneiss	Sphene	56.8 $\pm$ 7.8	49°42.6'	119°36.0'
2-83	Grd. Gneiss	Apatite	43.8 $\pm$ 2.7	49°37.6'	119°34.0'
3-5	Augen Gneiss	Apatite	46.4 $\pm$ 4.6	49°51.6'	119°17.2'
3-5	Augen Gneiss	Sphene	46.1 $\pm$ 11.5	49°51.6'	119°17.2'
3-8	Paragneiss	Apatite	53.0 $\pm$ 4.4	49°47.5'	119°08.0'
3-10	Paragneiss	Apatite	47.9 $\pm$ 5.8	49°16.5'	119°31.2'
3-12	Orthogneiss	Apatite	63.1 $\pm$ 10.4	49°16.1'	119°22.3'
3-13	Grd. Gneiss	Apatite	45.2 $\pm$ 3.9	49°26.0'	119°34.5'
3-13	Grd. Gneiss	Sphene	51.9 $\pm$ 11.4	49°26.0'	119°34.5'
3-17	Augen Gneiss	Apatite	35.6 $\pm$ 3.5	49°31.0'	119°32.7'
3-21	Grd. Gneiss	Apatite	47.2 $\pm$ 3.9	49°42.8'	119°16.5'
3-22	Valhalla Gd.	Apatite	48.0 $\pm$ 4.6	49°15.0'	119°10.5'
3-22	Valhalla Gd.	Sphene	59.5 $\pm$ 12.2	49°15.0'	119°10.5'

* Medford 1978

Appendix 2.

Analytical data of various mineral
phases in the Okanagan gneiss complex

GARNET

SPLE OF1	A1	A2	A3	A4	A5	A6	A7	A8	A9	A10	A11	A12	A13	A14	A15	A16
FeO	25.29	26.98	27.49	27.84	27.68	28.41	28.61	29.39	29.41	28.43	28.74	28.30	27.06	29.08	30.79	30.01
MnO	12.36	10.51	9.63	9.54	9.33	8.57	8.41	8.34	8.24	8.21	8.33	8.89	9.44	8.15	6.85	7.35
Na2O	0.04	0.01	0.01	0.01	0.05	0.01	0.04	0.02	0.08	0.05	0.01	0.01	0.05	0.05	0.01	0.04
Al2O3	20.92	21.19	20.87	20.91	21.11	21.31	20.91	21.02	21.06	21.41	21.07	21.24	20.97	21.16	21.11	21.41
SiO2	37.52	37.15	36.69	36.47	36.48	36.82	36.53	36.31	36.36	36.56	36.55	36.25	36.39	35.92	36.79	36.84
MgO	2.09	2.44	2.61	2.71	2.77	2.84	2.91	2.89	2.84	2.96	2.81	2.69	2.54	2.93	3.14	2.94
CaO	2.22	2.16	1.95	1.81	1.69	1.69	1.72	1.59	1.82	1.86	1.84	2.05	2.33	1.71	1.55	1.73
TiO2	0.01	0.01	0.03	0.01	0.03	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.03	0.01
TTL	100.44	100.44	99.29	99.28	99.13	99.65	99.12	99.76	99.83	99.47	99.33	99.43	98.78	98.99	100.28	100.31
Cations per 24 oxygens																
Fe	3.39	3.63	3.74	3.79	3.77	3.84	3.89	3.99	3.99	3.85	3.89	3.85	3.69	3.98	4.15	4.03
Mn	1.68	1.43	1.33	1.32	1.28	1.17	1.16	1.17	1.13	1.12	1.14	1.22	1.31	1.13	0.93	1.01
Na	0.01	0.00	0.00	0.00	0.01	0.00	0.01	0.01	0.03	0.01	0.00	0.00	0.00	0.00	0.00	0.00
Al	3.96	4.01	4.01	4.01	4.05	4.06	4.01	4.02	4.03	4.09	4.03	4.07	4.04	4.08	4.01	4.05
Si	6.03	5.97	5.97	5.94	5.94	5.96	5.95	5.91	5.91	5.92	5.94	5.91	5.94	5.87	5.93	5.93
Mg	0.51	0.58	0.63	0.65	0.67	0.68	0.70	0.69	0.68	0.71	0.67	0.65	0.61	0.71	0.75	0.71
Ca	0.38	0.37	0.34	0.32	0.29	0.29	0.31	0.27	0.31	0.32	0.32	0.35	0.41	0.29	0.27	0.29
TTL	15.97	16.01	16.02	16.04	16.03	16.01	16.04	16.08	16.09	16.04	16.03	16.06	16.03	16.09	16.06	16.04
Xalm	56.78	60.28	61.94	62.36	62.59	64.12	64.26	65.02	65.12	64.04	64.55	63.27	61.31	64.99	67.99	66.81
Xpyr	8.43	9.72	10.49	10.82	11.17	11.42	11.64	11.38	11.22	11.87	11.21	10.71	10.26	11.69	12.34	11.66
Xsps	28.36	23.81	21.98	21.65	21.38	19.59	19.14	19.13	18.49	18.72	18.94	20.14	21.67	18.44	15.32	16.59
Xgro	5.71	6.19	5.54	5.17	4.81	4.86	4.95	4.43	5.17	5.37	5.28	5.88	6.76	4.86	4.27	4.95

SPLE OF1	A17	A18core	A19	A20	A21	A22	A23	A24	A25	A26	A27	A28	A29	A30	A31	A32
FeO	31.03	31.98	31.62	31.45	31.67	30.54	27.54	26.42	28.05	26.93	26.29	26.07	27.71	27.57	27.32	26.19
MnO	6.29	5.37	4.79	4.88	5.31	6.56	10.39	11.46	9.85	10.76	11.55	11.25	9.96	10.42	10.35	11.91
Na2O	0.04	0.04	0.08	0.02	0.06	0.04	0.04	0.01	0.03	0.01	0.01	0.04	0.02	0.02	0.04	0.04
Al2O3	21.15	21.31	21.29	21.09	21.19	21.08	21.37	21.09	21.19	21.51	21.01	20.89	21.35	21.19	21.19	21.19
SiO2	36.19	35.98	36.45	36.67	36.37	36.66	36.18	36.76	36.41	35.91	36.18	36.82	36.23	36.43	36.28	36.66
MgO	3.34	3.37	3.37	3.46	3.32	3.26	2.18	1.94	2.65	2.23	2.04	2.02	2.46	2.51	2.52	1.96
CaO	1.53	1.58	1.44	1.51	1.52	1.54	2.11	2.01	1.69	1.95	2.02	2.29	2.23	2.11	2.04	2.13
TTL	99.58	99.66	99.11	99.12	99.48	99.69	99.82	99.71	99.88	99.32	99.12	99.39	99.97	100.25	99.75	100.12
Cations per 24 oxygens																
Fe	4.21	4.34	4.29	4.26	4.29	4.13	3.74	3.59	3.81	3.68	3.61	3.54	3.76	3.73	3.71	3.54
Mn	0.87	0.74	0.66	0.67	0.73	0.89	1.43	1.57	1.35	1.49	1.61	1.55	1.37	1.42	1.42	1.63
Na	0.01	0.01	0.02	0.01	0.02	0.01	0.01	0.00	0.01	0.00	0.00	0.01	0.01	0.01	0.01	0.01
Al	4.04	4.07	4.08	4.03	4.05	4.02	4.09	4.00	4.05	4.14	4.05	4.01	4.08	4.04	4.05	4.06
Si	5.88	5.84	5.92	5.94	5.91	5.93	5.88	5.97	5.90	5.86	5.92	5.99	5.88	5.91	5.89	5.94
Mg	0.81	0.81	0.81	0.83	0.81	0.78	0.52	0.47	0.64	0.54	0.49	0.49	0.59	0.61	0.61	0.47
Ca	0.27	0.28	0.25	0.26	0.26	0.26	0.36	0.35	0.29	0.34	0.35	0.39	0.39	0.37	0.36	0.37
TTL	16.09	16.11	16.04	16.03	16.07	16.06	16.07	16.01	16.06	16.06	16.04	16.01	16.07	16.08	16.07	16.03
Xalm	68.47	70.38	71.30	70.71	70.51	67.93	61.66	59.96	62.46	60.79	59.49	59.26	61.55	60.85	60.85	58.91
Xpyr	13.13	13.22	13.56	13.88	13.18	12.91	8.72	7.86	10.52	8.99	8.22	8.18	9.73	9.88	9.99	7.87
Xsps	14.07	11.98	10.96	11.11	11.96	14.76	23.59	26.34	22.22	24.61	26.48	25.91	22.41	23.29	23.36	27.13
Xgro	4.33	4.37	4.02	4.23	4.22	4.36	6.02	5.81	4.78	5.56	5.75	6.62	6.28	5.97	5.75	6.01

SPLE	A30	A31	A32	A33	A34	A35	A36	A37	A38	A39	A40	A41	A42	A43	A44r	B45r
FeO	29.69	30.79	30.35	29.62	30.38	30.12	29.97	30.36	29.46	29.62	29.98	30.20	29.77	29.74	30.52	29.51
MnO	4.59	5.15	5.06	5.23	5.38	5.51	5.80	5.98	5.90	6.13	6.79	5.91	6.35	6.83	6.09	6.99
Na2O	0.06	0.04	0.00	0.04	0.01	0.04	0.03	0.05	0.02	0.04	0.02	0.03	0.04	0.00	0.02	0.04
Al2O3	21.42	21.32	21.37	21.38	21.39	21.09	20.87	20.90	20.87	20.93	21.02	20.94	21.11	20.94	21.08	20.47
SiO2	36.52	36.99	37.19	36.43	36.94	37.01	37.17	36.77	36.78	36.79	36.72	36.96	36.14	36.54	36.90	36.73
MgO	4.22	4.32	4.13	3.97	4.04	3.82	3.87	3.75	3.63	3.37	3.00	3.74	3.41	3.35	3.77	3.03
CaO	1.85	1.88	2.06	2.11	2.06	2.00	1.99	1.89	1.90	1.88	1.84	1.63	1.48	1.35	1.52	1.73
TTL	98.37	100.49	100.17	98.78	100.21	99.60	99.70	99.72	98.52	98.78	99.36	99.46	98.30	98.69	99.92	99.00
Cations per 24 oxygens																
Fe	4.03	4.10	4.05	4.01	4.06	4.04	3.97	4.05	4.00	4.02	4.07	4.06	4.07	4.06	4.09	4.02
Mn	0.63	0.69	0.68	0.71	0.73	0.75	0.79	0.82	0.81	0.84	0.93	0.81	0.88	0.94	0.83	0.95
Na	0.02	0.01	0.00	0.01	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.00	0.00	0.01
Al	4.09	4.01	4.02	4.08	4.03	4.00	3.95	3.98	4.00	4.00	4.01	3.98	4.07	4.00	3.99	3.93
Si	5.92	5.90	5.94	5.90	5.91	5.95	5.97	5.93	5.98	5.98	5.96	5.96	5.92	5.96	5.93	5.99
Mg	1.02	1.03	0.98	0.96	0.96	0.91	0.92	0.90	0.88	0.81	0.73	0.90	0.83	0.81	0.90	0.74
Ca	0.32	0.32	0.35	0.37	0.35	0.34	0.34	0.33	0.33	0.33	0.32	0.28	0.26	0.24	0.26	0.30
TTL	16.03	16.08	16.04	16.05	16.06	16.04	16.02	16.06	16.01	16.01	16.03	16.03	16.05	16.02	16.05	16.01
Xalm	67.12	66.71	66.74	66.27	66.51	66.80	65.84	66.47	66.43	66.94	67.28	67.12	67.41	67.08	67.23	66.93
Xpyr	17.00	16.72	16.20	15.85	15.78	15.12	15.39	14.78	14.60	13.59	11.99	14.91	13.75	13.47	14.85	12.23
Xspe	10.52	11.32	11.27	11.85	11.94	12.14	13.08	13.42	13.50	14.03	15.43	13.36	14.56	15.60	13.63	15.83
Xgro	5.34	5.02	5.80	6.01	5.76	5.46	4.40	4.32	5.38	5.41	5.29	4.09	4.28	3.78	3.95	3.25
SPLE*5A*	Al3ar	Al6ar	Al7ar	Blr	B2r	B3	B4r	B5	B6	B7	B8	B9	B10	B11	B12	B13
FeO	29.94	29.51	29.18	29.71	30.27	30.42	29.94	30.06	29.79	29.59	29.52	29.15	29.69	29.45	29.25	29.33
MnO	6.45	6.64	6.01	7.18	5.64	6.69	6.86	5.23	4.66	4.69	4.89	5.37	5.33	5.49	5.65	5.47
Na2O	0.00	0.03	0.01	0.00	0.04	0.04	0.00	0.03	0.03	0.01	0.04	0.22	0.04	0.03	0.01	0.03
Al2O3	21.35	21.29	21.27	20.57	20.56	20.77	20.38	20.83	20.93	20.74	21.03	20.79	20.59	20.90	20.84	21.00
SiO2	38.41	38.16	38.06	36.66	37.40	36.99	37.21	37.70	37.61	37.61	37.49	37.21	37.72	37.65	37.41	37.62
MgO	3.47	3.53	3.71	2.83	3.84	3.29	3.42	4.33	4.74	4.60	4.83	4.69	4.66	4.53	4.06	4.17
CaO	2.05	2.00	2.06	1.64	1.82	1.81	1.60	1.90	1.99	2.08	2.06	1.96	2.04	2.11	2.00	2.04
TTL	101.71	101.19	100.29	98.64	99.58	100.04	99.45	100.08	99.76	99.34	99.86	99.59	100.07	100.30	99.22	99.67
Cations per 24 oxygens																
Fe	3.92	3.88	3.88	4.06	4.05	4.09	4.02	3.99	3.96	3.94	3.92	3.91	3.93	3.91	3.92	3.91
Mn	0.86	0.89	0.81	0.99	0.77	0.92	0.94	0.71	0.63	0.63	0.66	0.73	0.72	0.74	0.77	0.74
Na	0.00	0.01	0.00	0.00	0.01	0.01	0.00	0.01	0.01	0.00	0.01	0.06	0.01	0.01	0.00	0.01
Al	3.96	3.96	3.98	3.96	3.90	3.95	3.88	3.91	3.93	3.91	3.95	3.92	3.86	3.91	3.94	3.95
Si	6.04	6.03	6.05	5.99	6.01	5.96	6.01	6.01	5.99	6.02	5.97	5.96	6.00	5.99	6.01	6.01
Mg	0.81	0.83	0.88	0.69	0.92	0.79	0.82	1.02	1.12	1.09	1.14	1.11	1.10	1.07	0.97	0.99
Ca	0.34	0.34	0.35	0.28	0.31	0.31	0.28	0.32	0.34	0.36	0.35	0.34	0.36	0.36	0.34	0.35
TTL	15.96	15.97	15.95	15.99	15.99	16.02	15.98	15.99	16.00	15.98	16.02	16.06	15.99	16.01	15.98	15.98
Xalm	65.95	65.32	65.54	67.16	66.45	66.67	65.81	65.51	65.02	64.92	64.21	63.70	63.64	63.88	65.08	65.07
Xpyr	13.74	14.03	14.85	11.51	15.42	13.08	13.82	17.21	18.81	18.43	19.03	18.60	18.52	17.88	16.29	16.65
Xspe	14.50	14.97	13.68	16.58	12.88	15.10	15.75	11.83	10.51	10.67	10.95	12.11	12.02	12.32	12.89	12.43
Xgro	5.02	5.05	5.83	3.91	2.76	3.63	1.83	3.32	3.97	3.91	4.45	3.36	2.43	3.82	4.59	4.89
Xadu	0.79	0.64	0.10	0.84	2.49	1.52	2.79	2.12	1.70	2.08	1.37	2.22	3.39	2.10	1.15	0.96

SPLE*5A*	B14	B15	B16core	B17	B18	B19	B20*	B21*	B22*	B23	B4*	B5*	C2	C3	C4	D1*
FeO	29.20	29.68	28.86	29.50	29.92	29.99	30.84	30.18	30.20	29.46	29.96	29.55	29.15	29.76	29.64	30.60
MnO	5.59	5.56	5.34	5.19	4.82	5.18	5.37	6.21	5.89	5.27	6.71	6.48	6.54	4.72	6.37	6.21
Na2O	0.06	0.02	0.04	0.05	0.02	0.02	0.03	0.00	0.03	0.06	0.06	0.00	0.08	0.08	0.01	0.04
Al2O3	20.84	20.87	20.83	20.95	20.85	20.96	20.78	20.97	20.58	20.88	21.33	21.42	20.93	21.51	21.35	21.55
SiO2	37.34	37.95	37.66	37.78	37.82	37.45	37.49	37.45	37.52	37.74	37.98	37.90	36.35	37.10	36.91	36.74
MgO	4.21	4.56	4.69	4.61	4.85	4.27	4.38	3.74	3.79	4.75	3.35	3.64	3.25	4.31	3.30	3.62
CaO	2.11	2.08	2.07	2.13	2.04	1.84	1.63	1.62	1.74	2.15	1.86	1.88	1.60	1.99	1.84	1.76
TTL	99.41	100.73	99.48	100.21	100.33	99.71	100.52	100.28	99.77	100.33	101.32	100.91	97.90	99.50	99.47	100.52
Cations per 24 oxygens																
Fe	3.91	3.90	3.84	3.90	3.95	4.00	4.09	4.04	4.03	3.89	3.96	3.92	4.00	3.99	4.00	4.10
Mn	0.76	0.75	0.72	0.69	0.65	0.70	0.72	0.84	0.80	0.71	0.90	0.87	0.91	0.64	0.87	0.84
Na	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.00	0.00	0.01	0.02	0.00	0.03	0.02	0.00	0.01
Al	3.94	3.89	3.92	3.92	3.90	3.95	3.90	3.95	3.90	3.90	3.97	6.00	4.04	4.06	4.06	4.07
Si	5.99	6.00	6.01	6.00	6.00	5.98	5.99	6.02	6.00	6.00	6.01	4.00	5.96	5.94	5.96	5.88
Mg	1.00	1.07	1.11	1.09	1.14	1.01	1.04	0.89	0.90	1.12	0.79	0.86	0.79	1.02	0.79	0.86
Ca	0.36	0.35	0.35	0.36	0.35	0.32	0.28	0.28	0.30	0.37	0.31	0.32	0.28	0.34	0.32	0.30
TTL	16.00	15.99	15.99	16.00	16.00	16.00	16.02	16.00	15.99	16.01	15.99	15.98	16.02	16.03	16.01	16.07
Xalm	64.49	63.66	63.22	64.03	64.31	66.04	66.15	66.55	66.30	63.38	66.36	65.68	66.84	66.52	66.89	67.14
Xpyr	16.83	17.99	18.73	18.25	19.11	16.98	17.24	14.82	15.23	18.73	13.28	14.43	13.29	17.16	13.28	14.15
Sspe	12.69	12.45	12.12	11.67	10.79	11.71	12.00	14.00	13.46	11.80	15.11	14.59	15.18	10.68	14.55	13.79
Xgro	4.60	3.24	4.01	4.10	3.24	4.10	2.03	3.46	2.48	3.67	4.83	5.24	4.69	5.57	5.21	4.89
SPLE*5A*	D2	D3core	D41	D5	D6	D7*	*5B*	Al*	A2	A3	A4core	A5	A7	A8*	C1*	C2*
FeO	29.99	30.47	28.10	30.32	31.13	29.76	*	30.69	30.06	30.59	29.96	30.11	30.30	30.58	29.56	30.42
MnO	4.73	4.66	4.53	4.87	4.72	5.69	*	5.90	5.05	5.44	5.57	5.47	4.90	5.37	6.64	6.11
Na2O	0.02	0.04	0.11	0.01	0.04	0.00	*	0.00	0.00	0.00	0.02	0.00	0.02	0.00	0.03	0.00
Al2O3	21.52	21.83	22.29	21.58	21.69	21.46	*	21.24	21.44	21.26	21.54	21.53	21.48	21.48	20.73	20.70
SiO2	37.01	37.31	38.59	36.91	36.82	37.27	*	36.88	37.16	37.16	37.28	37.24	36.54	36.54	37.38	37.29
MgO	4.64	4.84	5.07	4.73	4.52	3.89	*	3.67	4.45	3.87	4.25	4.49	4.60	4.09	3.52	3.36
CaO	1.97	2.05	2.01	1.95	1.96	1.87	*	1.66	1.99	1.98	2.09	1.98	2.10	1.60	1.78	1.90
TTL	99.89	101.20	100.73	100.37	100.90	99.97	*	100.03	100.16	100.35	100.73	100.85	99.90	99.67	99.65	99.78
Cations per 24 oxygens																
Fe	4.00	4.02	3.67	4.04	4.14	3.98	*	4.12	4.01	4.09	3.98	3.99	4.07	4.12	3.97	4.09
Mn	0.64	0.63	0.60	0.66	0.63	0.77	*	0.80	0.68	0.73	0.75	0.74	0.67	0.73	0.90	0.83
Na	0.01	0.01	0.03	0.00	0.01	0.00	*	0.00	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00
Al	4.05	4.05	4.10	4.05	4.06	4.04	*	4.02	4.03	4.01	4.03	4.02	4.06	4.08	3.93	3.93
Si	5.91	5.88	6.02	5.88	5.85	5.96	*	5.93	5.93	5.94	5.92	5.91	5.86	5.89	6.01	6.01
Mg	1.10	1.14	1.17	1.12	1.07	0.92	*	0.88	1.06	0.92	1.01	1.06	1.10	0.98	0.84	0.80
Ca	0.34	0.35	0.33	0.33	0.33	0.32	*	0.29	0.34	0.33	0.36	0.34	0.36	0.28	0.31	0.33
TTL	16.05	16.08	15.94	16.09	16.11	16.01	*	16.05	16.05	16.04	16.05	16.07	16.11	16.07	15.98	15.99
Xalm	65.83	65.60	63.43	65.65	66.96	66.39	*	67.70	65.83	67.24	65.34	65.21	65.65	67.43	65.54	67.14
Xpyr	18.14	18.58	20.41	18.26	17.33	15.46	*	14.44	17.38	15.14	16.54	17.34	17.78	16.06	14.15	13.48
Sspe	10.51	10.17	10.37	10.67	10.29	12.85	*	13.17	11.21	12.11	12.31	11.99	10.76	12.00	15.18	13.92
Xgro	5.47	5.65	5.75	5.42	5.41	5.24	*	4.68	5.56	5.41	5.78	5.40	5.79	4.48	3.53	3.72

SPLE*SA*	D2	D3core	D41	D5	D6	D7r	*SB*	A1r	A2	A3	Alcore	A5	A7	A8r	C1r	C2r
FeO	29.99	30.47	28.10	30.32	31.13	29.76	*	30.69	30.06	30.59	29.96	30.11	30.30	30.58	29.56	30.42
MnO	4.73	4.66	4.53	4.87	4.72	5.69	*	5.90	5.05	5.44	5.57	5.47	4.90	5.37	6.64	6.11
Na2O	0.02	0.04	0.11	0.01	0.04	0.00	*	0.00	0.00	0.00	0.02	0.00	0.02	0.00	0.03	0.00
Al2O3	21.52	21.83	22.29	21.58	21.69	21.46	*	21.24	21.44	21.26	21.54	21.53	21.48	21.48	20.73	20.70
SiO2	37.01	37.31	38.59	36.91	36.82	37.27	*	36.88	37.16	37.16	37.28	37.24	36.54	36.54	37.38	37.29
MgO	4.64	4.84	5.07	4.73	4.52	3.89	*	3.67	4.45	3.87	4.25	4.49	4.60	4.49	3.52	3.36
CaO	1.97	2.05	2.01	1.95	1.96	1.87	*	1.66	1.99	1.98	2.09	1.98	2.10	1.60	1.78	1.90
TTL	99.89	101.20	100.73	100.37	100.90	97.97	*	100.03	100.16	100.35	100.73	100.85	99.90	99.67	99.65	99.78
Cations per 24 oxygens																
Fe	4.00	4.02	3.67	4.04	4.14	3.98	*	4.12	4.01	4.09	3.98	3.99	4.07	4.12	3.97	4.09
Mn	0.64	0.63	0.60	0.66	0.63	0.77	*	0.80	0.68	0.73	0.75	0.74	0.67	0.73	0.90	0.83
Na	0.01	0.01	0.03	0.00	0.01	0.00	*	0.00	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00
Al	4.05	4.05	4.10	4.05	4.06	4.04	*	4.02	4.03	4.01	4.03	4.02	4.06	4.08	3.93	3.93
Si	5.91	5.88	6.02	5.88	5.85	5.96	*	5.93	5.93	5.94	5.92	5.91	5.86	5.89	6.01	6.01
Mg	1.10	1.14	1.17	1.12	1.07	0.92	*	0.88	1.06	0.92	1.01	1.06	1.10	0.98	0.84	0.80
Ca	0.34	0.35	0.33	0.33	0.33	0.32	*	0.29	0.34	0.33	0.36	0.34	0.36	0.28	0.31	0.33
TTL	16.05	16.08	15.94	16.09	16.11	16.01	*	16.05	16.05	16.04	16.05	16.07	16.11	16.07	15.98	15.99
Xalm	65.83	65.60	63.43	65.65	66.96	66.39	*	67.70	65.83	67.24	65.34	65.21	65.65	67.43	65.54	67.14
Xpyr	18.14	18.58	20.41	18.26	17.33	15.46	*	14.44	17.38	15.14	16.54	17.34	17.78	16.06	14.15	13.48
Xape	10.51	10.17	10.37	10.67	10.29	12.85	*	13.17	11.21	12.11	12.31	11.99	10.76	12.00	15.18	13.92
Xgro	5.47	5.65	5.75	5.42	5.41	5.24	*	4.68	5.56	5.41	5.78	5.40	5.79	4.48	3.53	3.72
SPLE*SB*																
FeO	30.58	29.79	29.97	29.27	30.36	30.94	30.44	29.69	30.40	31.30	30.30	29.86	29.86	31.44	30.60	30.05
MnO	4.69	5.02	5.64	5.53	5.48	5.55	5.45	5.34	4.88	5.33	7.49	6.36	5.62	4.89	4.88	4.57
Na2O	0.00	0.03	0.01	0.01	0.00	0.00	0.01	0.03	0.02	0.00	0.00	0.01	0.04	0.03	0.01	0.02
Al2O3	20.77	20.59	20.63	20.77	20.81	20.62	20.72	20.78	20.97	20.56	20.62	21.56	21.66	21.37	21.39	21.19
SiO2	37.78	37.73	37.76	37.78	37.20	37.73	37.40	37.60	37.76	37.32	37.21	38.32	38.20	37.53	36.78	37.36
MgO	4.59	4.52	4.34	4.31	3.74	3.63	3.89	4.34	4.55	4.22	2.66	3.43	3.73	4.13	3.99	4.57
CaO	1.98	1.97	2.02	2.00	2.06	1.97	2.10	2.04	1.77	1.75	1.47	1.83	1.95	1.88	1.82	1.95
TTL	100.41	99.71	100.38	99.55	97.57	100.49	100.04	99.85	100.36	100.50	99.78	101.39	101.07	101.27	99.47	99.73
Cations per 24 oxygens																
Fe	4.05	3.95	3.97	3.90	4.07	4.11	4.06	3.95	4.03	4.16	4.09	3.93	3.94	4.15	4.12	4.03
Mn	0.63	0.68	0.76	0.75	0.75	0.75	0.74	0.72	0.66	0.72	1.03	0.85	0.75	0.66	0.67	0.62
Na	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.00	0.00	0.01	0.01	0.00	0.01
Al	3.89	3.87	3.87	3.91	3.95	3.88	3.91	3.91	3.93	3.87	3.93	4.00	4.02	3.99	4.06	3.99
Si	6.00	6.03	6.00	6.04	5.98	6.02	5.99	6.01	6.00	5.96	6.02	6.04	6.03	5.94	5.92	5.97
Mg	1.08	1.07	1.03	1.03	0.90	0.86	0.93	1.03	1.08	1.00	0.64	0.81	0.88	0.97	0.96	1.09
Ca	0.34	0.34	0.34	0.34	0.35	0.34	0.36	0.35	0.30	0.30	0.25	0.31	0.33	0.32	0.31	0.33
TTL	15.99	15.97	15.99	15.97	16.01	15.98	16.00	16.00	16.00	16.03	15.98	15.95	15.96	16.05	16.04	16.03
Xalm	65.72	64.75	64.33	64.36	66.87	67.30	66.28	64.80	66.04	66.58	67.75	66.72	66.79	67.98	68.02	66.19
Xpyr	18.12	18.13	17.20	17.28	14.89	14.52	15.46	17.28	17.97	16.59	10.78	13.68	14.88	16.02	15.81	18.03
Xape	10.53	11.44	12.72	12.59	12.40	12.59	12.29	12.09	10.97	11.91	17.23	14.39	12.74	10.77	10.98	10.25
Xgro	2.86	2.65	2.51	3.71	4.47	2.76	3.77	3.71	3.18	1.56	2.73	5.19	5.58	4.64	5.20	5.16
Xadu	2.76	3.03	3.23	2.05	1.39	2.83	2.19	2.13	1.84	3.36	1.51	0.00	0.00	0.59	0.00	0.37

SPLE*5B*	D4	D5	D6	D7core	D8	D9	D10r	D11r	*7*	A1r	A2	A3	A4	A6core	A7	A8
FeO	30.33	29.97	29.84	29.83	30.13	29.74	30.41	30.90	*	32.14	31.85	31.23	31.38	30.83	30.59	31.79
MnO	5.04	4.97	5.12	5.24	5.29	4.94	4.33	4.59	*	4.94	4.24	4.34	4.66	4.58	4.69	4.32
NaO	0.03	0.00	0.02	0.03	0.04	0.03	0.05	0.00	*	0.04	0.03	0.02	0.00	0.05	0.14	0.02
Al2O3	21.08	21.07	21.29	21.15	21.24	21.33	21.35	21.42	*	21.30	21.39	21.31	21.50	21.34	21.26	21.24
SiO2	37.15	37.54	37.39	37.38	37.49	37.22	37.27	37.56	*	37.31	37.50	37.36	37.29	37.68	37.54	37.54
MgO	4.40	4.32	4.43	4.16	4.23	4.36	4.64	4.22	*	4.16	4.66	4.61	4.55	4.44	4.35	4.55
CaO	1.97	2.06	2.00	2.03	1.82	1.99	1.96	1.76	*	1.16	1.36	1.58	1.17	1.65	1.59	1.37
TTL	100.01	99.91	100.09	99.82	100.24	99.61	100.06	100.45	*	100.97	100.98	100.45	100.57	100.40	100.30	100.86
Cations per 24 oxygens																
Fe	4.05	3.99	3.97	3.97	4.00	3.98	4.05	4.11	*	4.26	4.20	4.14	4.17	4.10	4.06	4.20
Mn	0.68	0.67	0.69	0.02	0.71	0.67	0.58	0.62	*	0.65	0.57	0.58	0.62	0.62	0.63	0.58
Na	0.01	0.00	0.01	0.71	0.01	0.01	0.02	0.00	*	0.01	0.01	0.01	0.00	0.02	0.04	0.01
Al	3.98	3.96	4.00	3.99	4.03	4.01	4.01	4.01	*	3.99	3.99	3.99	4.03	4.00	3.98	3.97
Si	5.94	5.99	5.96	5.98	5.97	5.96	5.94	5.97	*	5.94	5.94	5.94	5.93	5.97	5.99	5.95
Mg	1.05	1.03	1.05	0.99	1.00	1.04	1.10	1.00	*	0.99	1.10	1.09	1.08	1.05	1.03	1.08
Ca	0.34	0.35	0.34	0.35	0.31	0.34	0.33	0.30	*	0.20	0.23	0.27	0.20	0.28	0.27	0.23
TTL	16.05	16.00	16.03	16.02	16.02	16.02	16.05	16.01	*	16.05	16.05	16.04	16.04	16.03	16.02	16.04
Xalm	66.01	65.88	65.57	65.97	66.33	65.99	66.76	68.16	*	69.89	68.84	68.02	68.66	67.73	67.69	68.90
Xpyr	17.24	17.09	17.37	16.46	16.69	17.26	18.14	16.61	*	16.18	18.08	17.95	17.75	17.41	17.24	17.75
Xspe	11.21	11.17	11.43	11.79	11.84	11.10	9.63	10.25	*	10.70	9.33	9.61	10.32	10.21	10.56	9.56
Xgro	4.62	4.94	5.45	5.36	4.75	5.63	5.37	4.97	*	2.78	3.32	4.05	3.23	4.45	4.01	2.87
SPLE*7*	A9r	B10r	B11	B12	B13	B14core	B15	B16	B17	B18r	*47A*	A12r	A13	A14	A15	A16
FeO	31.00	31.51	31.02	30.20	30.39	30.59	30.10	31.26	31.79	31.23	*	33.40	33.08	32.78	31.51	31.01
MnO	6.68	6.44	4.98	5.64	5.69	6.01	5.53	5.04	5.89	6.32	*	2.83	2.56	2.34	2.38	2.75
Na2O	0.00	0.01	0.04	0.03	0.02	0.00	0.00	0.01	0.02	0.01	*	0.03	0.00	0.04	0.04	0.04
Al2O3	21.60	21.32	21.27	21.10	21.07	21.06	21.06	21.04	21.04	21.05	*	21.19	21.11	21.18	21.06	21.16
SiO2	37.15	37.08	37.14	37.25	37.14	37.17	37.26	37.30	37.30	37.21	*	37.08	36.80	35.55	36.39	36.85
MgO	3.43	3.45	4.37	4.18	4.11	4.20	4.23	4.16	3.58	3.44	*	4.46	4.96	4.89	5.72	5.65
CaO	1.06	1.15	1.44	1.60	1.53	1.60	1.67	1.59	1.30	1.25	*	0.86	0.95	1.19	1.55	1.73
TTL	100.95	100.98	100.27	100.00	100.04	100.64	99.83	100.39	100.95	100.51	*	99.92	99.45	98.99	98.71	99.19
Cations per 24 oxygens																
Fe	4.14	4.21	4.14	4.03	4.06	4.07	4.03	4.17	4.24	4.18	*	4.48	4.44	4.43	4.24	4.15
Na	0.90	0.87	0.67	0.76	0.77	0.81	0.75	0.68	0.80	0.86	*	0.38	0.35	0.32	0.33	0.38
Al	0.00	0.00	0.01	0.01	0.01	0.00	0.00	0.00	0.01	0.00	*	0.01	0.00	0.01	0.01	0.01
Si	4.06	4.02	4.01	3.98	3.98	3.96	3.98	3.96	3.96	3.98	*	4.00	4.00	4.03	4.00	4.00
Mg	5.93	5.93	5.93	5.96	5.95	5.93	5.97	5.96	5.96	5.97	*	5.95	5.91	5.90	5.87	5.90
Ca	0.82	0.82	1.04	1.00	0.98	1.00	1.01	0.99	0.85	0.82	*	1.06	1.19	1.17	1.38	1.35
TTL	16.03	16.06	16.06	16.03	16.04	16.06	16.02	16.04	16.04	16.02	*	16.04	16.07	16.08	16.11	15.08
Xalm	69.54	69.01	67.82	66.33	66.52	65.91	66.16	67.96	69.18	68.71	*	73.77	72.25	72.23	68.25	57.15
Xpyr	13.52	13.48	17.08	16.50	16.15	16.34	16.69	16.32	14.07	13.57	*	17.54	19.39	19.20	22.21	21.94
Xspe	14.96	14.29	11.06	12.64	12.73	13.30	12.42	11.23	13.13	14.20	*	6.33	5.69	5.22	5.26	6.08
Xgro	2.95	3.20	3.83	3.88	3.66	3.16	4.02	3.29	2.44	2.81	*	2.24	2.27	3.32	3.75	4.21
Xadu	0.04	0.02	0.21	0.56	0.74	1.29	0.70	1.19	1.18	0.71	*	0.11	0.40	0.03	0.53	0.61

SPL	A33	A34	A7a	A8a	A14a	A15a	*5A*	A1r	A2r	A3r	A4	A5	A6	A7	A8	A9
FeO	26.95	26.73	26.68	26.61	27.86	27.98	*	28.88	29.38	29.22	29.81	29.88	30.09	29.70	29.76	29.93
MnO	11.11	11.63	10.94	10.73	9.77	9.09	*	7.05	6.66	6.37	6.17	5.57	5.50	5.01	5.00	4.76
Na2O	0.03	0.02	0.05	0.01	0.03	0.01	*	0.02	0.01	0.03	0.01	0.02	0.02	0.03	0.00	0.00
Al2O3	20.84	20.89	21.36	21.42	21.42	21.44	*	21.17	21.18	21.10	21.16	21.28	21.31	21.35	21.61	21.25
SiO2	36.19	36.73	38.42	38.11	38.08	37.71	*	37.68	37.61	37.54	37.44	37.83	37.84	37.30	37.36	37.82
MgO	2.05	2.05	2.42	2.51	2.71	2.77	*	3.26	3.34	3.54	3.64	3.75	4.12	4.14	4.27	4.25
CaO	2.27	2.19	1.88	2.02	1.83	1.69	*	1.98	2.07	2.09	2.00	2.07	2.04	2.04	2.01	2.01
TiO2	0.06	0.01	0.05	0.03	0.01	0.01	*	0.00	0.00	0.00	0.00	0.05	0.02	0.01	0.00	0.00
TTL	99.49	99.66	101.84	101.42	101.71	100.71	*	100.05	100.31	100.03	100.39	100.47	101.22	99.57	100.02	100.30
Cations per 24 oxygens																
Fe	3.69	3.64	3.52	3.53	3.69	3.73	*	3.87	3.93	3.91	3.98	3.97	3.97	3.96	3.97	3.97
Mn	1.54	1.52	1.46	1.44	1.31	1.23	*	0.96	0.90	0.87	0.83	0.75	0.74	0.68	0.67	0.64
Na	0.00	0.00	0.01	0.00	0.01	0.00	*	0.01	0.00	0.01	0.00	0.01	0.01	0.01	0.00	0.00
Al	4.01	4.01	3.97	4.01	3.99	4.03	*	3.99	3.99	3.98	3.98	3.98	3.97	4.03	4.05	3.98
Si	5.92	5.97	6.06	6.04	6.03	6.02	*	6.03	6.01	6.01	5.98	6.01	5.98	5.97	5.95	6.01
Mg	0.51	0.49	0.57	0.59	0.63	0.66	*	0.78	0.79	0.84	0.87	0.88	0.97	0.99	1.01	1.02
Ca	0.39	0.38	0.32	0.34	0.31	0.29	*	0.33	0.35	0.36	0.34	0.35	0.34	0.35	0.34	0.35
TTL	16.07	16.02	15.95	15.95	15.98	15.97	*	15.97	15.98	15.99	16.01	15.98	16.01	16.01	16.01	15.99
Xalm	60.24	60.24	60.01	59.81	62.05	63.18	*	65.09	65.69	65.44	66.04	66.66	65.98	66.34	66.12	66.42
Xpyr	8.19	8.25	9.71	10.01	10.71	11.15	*	13.11	13.31	14.13	14.40	14.89	16.10	16.50	16.92	16.98
Xspe	25.13	25.18	24.92	24.43	22.03	20.79	*	16.09	15.07	14.45	13.85	12.59	12.22	11.33	11.24	10.73
Xgro	6.34	6.29	5.17	5.72	5.21	4.88	*	5.69	5.72	5.55	5.17	5.70	4.84	5.82	5.72	5.44
SPL	A10	A11	A12	A13	A14	A15	A16	A17	A18	A19	A23core	A25	A26	A27	A28	A29
FeO	30.46	30.11	30.62	29.94	30.30	30.27	30.01	29.58	31.06	30.78	30.27	30.32	30.02	30.02	30.31	30.39
MnO	4.86	4.73	4.57	4.74	4.37	4.37	4.27	4.46	4.61	4.35	4.34	4.58	4.49	4.25	4.63	4.69
Na2O	0.05	0.08	0.01	0.00	0.02	0.03	0.00	0.01	0.00	0.02	0.00	0.06	0.02	0.01	0.03	0.05
Al2O3	21.59	21.40	21.59	21.51	21.32	21.40	21.64	21.15	21.13	21.53	21.41	21.64	21.70	21.53	21.20	21.36
SiO2	37.59	37.39	37.43	37.52	37.40	37.59	37.73	37.43	37.26	37.18	37.32	36.85	37.13	37.31	37.22	36.57
MgO	4.38	4.32	4.40	4.50	4.50	4.60	4.48	4.49	3.63	3.82	4.27	4.45	4.55	4.42	4.43	4.22
CaO	2.09	2.15	2.15	2.11	2.14	2.03	2.05	2.07	2.06	2.05	2.04	2.07	2.10	2.06	2.00	1.96
TTL	101.03	100.20	100.77	100.32	100.09	100.32	100.20	99.24	99.85	99.89	99.69	99.77	100.01	99.60	99.95	99.25
Cations per 24 oxygens																
Fe	4.03	4.01	4.06	3.97	4.03	4.01	3.98	3.96	4.17	4.12	4.04	4.06	4.00	4.01	4.05	4.10
Mn	0.65	0.64	0.61	0.64	0.59	0.59	0.57	0.60	0.63	0.59	0.59	0.59	0.61	0.57	0.63	0.64
Na	0.01	0.03	0.00	0.00	0.01	0.01	0.00	0.00	0.00	0.01	0.00	0.02	0.01	0.03	0.01	0.02
Al	4.02	4.01	4.03	4.03	4.00	4.00	4.04	3.99	4.00	4.06	4.03	4.08	4.07	4.05	3.99	4.06
Si	5.94	5.96	5.93	5.95	5.96	5.97	5.98	5.99	5.98	5.95	5.97	5.90	5.91	5.96	5.95	5.90
Mg	1.03	1.02	1.04	1.06	1.07	1.09	1.06	1.07	0.88	0.91	1.01	1.06	1.05	1.05	1.06	1.02
Ca	0.35	0.37	0.37	0.36	0.36	0.35	0.35	0.36	0.34	0.35	0.35	0.35	0.36	0.35	0.34	0.34
TTL	16.04	16.04	16.04	16.02	16.03	16.02	15.99	16.00	16.01	16.00	16.01	16.05	16.04	16.01	16.04	16.07
Xalm	66.39	66.42	66.80	65.86	66.62	66.51	66.78	66.08	69.34	69.08	67.54	66.89	66.19	66.96	66.69	67.25
Xpyr	17.03	16.98	17.10	17.63	17.64	18.04	17.78	17.89	14.59	15.29	16.95	17.48	17.87	17.57	17.37	16.67
Xspe	10.74	10.56	10.09	10.56	9.73	9.73	9.63	10.09	10.43	9.89	9.80	9.79	10.02	9.59	10.31	10.52
Xgro	5.84	6.00	6.01	5.95	5.89	5.67	5.77	5.82	5.52	5.48	5.72	5.82	5.92	5.85	5.23	5.56

SPLE*47A* A17core A18 A19 A20r A7ar A9ar A9ar A9ar B1r B2 B3 B4 B5core B6 B7 B8 B9																	
FeO	30.56	29.69	31.16	31.86	33.19	32.49	32.56	33.42	31.58	30.51	30.37	30.39	30.38	30.39	30.55	30.91	
MnO	3.62	4.10	3.06	2.14	2.63	2.63	2.42	2.92	2.81	3.12	3.29	3.62	3.42	2.96	2.85	2.65	
Na2O	0.02	0.11	0.03	0.02	0.00	0.02	0.03	0.03	0.00	0.02	0.01	0.06	0.00	0.03	0.00	0.03	
Al2O3	21.06	20.91	20.97	20.99	21.65	21.49	21.36	21.19	21.57	21.41	21.26	21.35	21.63	21.67	21.28	21.37	
SiO2	36.94	36.67	36.56	36.85	37.88	39.46	38.58	37.34	37.68	37.15	37.25	37.31	37.30	37.42	37.22	37.27	
MgO	5.36	5.06	5.59	6.06	4.83	5.19	5.54	4.52	5.66	5.39	5.32	5.27	5.24	5.42	5.42	5.40	
CaO	2.02	2.00	1.84	1.10	0.96	1.04	1.03	0.76	1.60	2.05	1.99	2.04	1.95	1.81	1.81	1.95	
TTL	99.63	98.55	99.22	98.95	101.16	101.33	101.53	100.20	100.92	99.96	99.52	100.07	99.99	99.70	99.13	99.58	
Cations per 24 oxygens																	
Fe	4.07	4.00	4.17	4.26	4.37	4.24	4.18	4.46	4.15	4.07	4.05	4.03	4.04	4.04	4.08	4.12	
Mn	0.49	0.56	0.42	0.29	0.35	0.35	0.32	0.40	0.38	0.42	0.44	0.49	0.47	0.40	0.39	0.36	
Na	0.01	0.04	0.01	0.01	0.00	0.01	0.01	0.01	0.00	0.01	0.00	0.02	0.00	0.01	0.00	0.01	
Al	3.97	3.98	3.97	3.95	4.02	3.96	3.93	3.99	4.00	4.02	3.99	4.00	4.05	4.06	4.01	4.02	
Si	5.91	5.93	5.88	5.91	5.97	6.02	6.02	5.97	5.93	5.92	5.94	5.92	5.92	5.94	5.95	5.94	
Mg	1.27	1.21	1.34	1.45	1.14	1.21	1.29	1.08	1.33	1.28	1.27	1.25	1.24	1.28	1.29	1.28	
Ca	0.35	0.35	0.32	0.19	0.16	0.17	0.17	0.13	0.27	0.35	0.34	0.35	0.33	0.31	0.31	0.33	
TTL	16.08	16.08	16.11	16.07	16.02	15.98	15.98	16.03	16.06	16.06	16.05	16.06	16.05	16.03	16.03	16.05	
Xalm	65.63	65.12	66.58	68.57	72.66	70.87	70.12	73.53	67.72	66.46	66.35	65.89	66.39	66.97	67.26	67.61	
Xp/r	20.80	19.99	21.60	23.63	18.85	20.34	21.63	17.79	21.72	20.94	20.79	20.45	20.41	21.29	21.27	21.05	
Xspe	7.97	9.21	6.73	4.74	5.82	5.85	5.38	6.54	6.13	6.88	7.31	7.98	7.73	6.62	6.36	5.88	
Xgro	4.37	4.76	3.72	1.46	2.63	2.19	1.21	1.71	4.09	5.73	5.24	5.24	5.47	5.12	5.10	5.45	
Xadu	1.22	0.91	1.37	1.61	0.04	0.75	1.67	0.43	0.33	0.00	0.32	0.43	0.00	0.00	0.00	0.01	
SPLE*47A* B10c B15ar B16ar *47B* B11r B12 B13 B14 B15 B16 B19 B20core B21 B22 B23 B24																	
FeO	31.69	32.21	32.45	*	33.18	31.74	31.62	31.03	31.72	32.03	32.15	30.49	30.87	30.16	30.86	30.25	
MnO	2.66	2.58	2.60	*	3.30	2.45	2.48	2.75	3.01	3.06	3.24	4.13	4.30	4.45	3.94	3.17	
Na2O	0.01	0.02	0.00	*	0.02	0.00	0.00	0.00	0.01	0.04	0.04	0.01	0.08	0.03	0.01	0.03	
Al2O3	21.44	21.07	21.05	*	21.42	21.37	21.67	21.46	21.51	21.20	21.32	21.38	21.58	21.58	21.36	21.31	
SiO2	36.81	38.06	38.16	*	31.34	37.59	37.68	38.01	37.66	37.73	37.62	37.52	37.73	37.74	37.72	38.12	
MgO	5.25	5.41	5.42	*	4.24	5.65	5.56	5.67	5.48	4.59	4.76	4.96	4.93	4.80	4.94	5.24	
CaO	1.54	0.98	1.40	*	1.11	1.61	1.60	1.52	1.32	1.27	0.92	1.74	1.86	1.81	1.87	2.03	
TTL	99.40	100.39	101.15	*	100.61	100.52	100.63	100.43	100.74	99.92	100.13	100.22	101.38	100.57	100.71	100.18	
Cations per 24 oxygens																	
Fe	4.25	4.25	4.24	*	4.42	4.20	4.17	4.08	4.16	4.26	4.28	4.04	4.05	3.98	4.07	3.99	
Mn	0.36	0.35	0.34	*	0.45	0.33	0.33	0.37	0.40	0.41	0.44	0.56	0.57	0.60	0.53	0.42	
Na	0.00	0.01	0.00	*	0.01	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.02	0.01	0.00	0.01	
Al	4.05	3.92	3.90	*	4.02	3.98	4.03	3.98	4.00	3.98	4.00	4.00	4.00	4.02	3.98	3.96	
Si	5.90	6.01	6.00	*	5.95	5.94	5.94	5.99	5.94	6.01	5.99	5.96	5.93	5.97	5.96	6.22	
Mg	1.25	1.27	1.27	*	1.01	1.33	1.31	1.33	1.29	1.09	1.13	1.17	1.16	1.13	1.17	1.23	
Ca	0.26	0.16	0.24	*	0.19	0.27	0.27	0.26	0.22	0.22	0.16	0.30	0.31	0.31	0.32	0.34	
TTL	16.07	15.98	16.00	*	16.04	16.05	16.04	16.01	16.01	15.99	16.01	16.03	16.06	16.02	16.03	15.98	
Xalm	69.32	70.08	69.21	*	72.92	68.35	68.61	67.55	68.57	71.19	71.37	66.59	66.47	66.25	66.83	66.46	
Xp/r	20.48	21.36	21.12	*	16.61	21.82	21.52	22.10	21.16	18.27	18.83	19.37	18.99	18.78	19.23	20.69	
Xspe	5.89	5.79	5.75	*	7.34	5.36	5.46	6.09	6.61	6.92	7.27	9.15	9.42	9.90	8.72	7.11	
Xgro	4.31	0.95	1.53	*	3.12	3.65	4.39	3.80	3.39	3.22	2.39	4.70	4.78	5.09	4.52	4.98	
Xadu	0.00	1.82	2.38	*	0.01	0.83	0.02	0.47	0.26	0.40	0.14	0.15	0.34	0.00	0.70	0.76	

SPL#478*	B25r	B28	B29	B30	B31	B32r	B33r	D38r	D39	D40	D41core	D42	D43	D44	D45	D46r
FeO	30.39	30.40	30.70	30.68	30.69	32.68	32.99	32.82	31.27	30.71	30.10	32.42	32.60	31.06	31.57	33.26
MnO	2.74	4.53	4.26	3.88	2.86	2.59	2.53	2.78	2.31	2.54	3.12	2.72	2.58	3.04	2.60	2.66
Na2O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.05	0.01	0.03	0.03	0.00	0.02	0.02	0.04	0.04
Al2O3	21.74	21.34	21.65	21.65	21.34	21.49	21.43	21.40	21.48	21.47	21.54	22.05	21.29	21.58	21.37	21.48
SiO2	37.73	37.92	38.15	37.99	37.35	37.55	37.72	37.45	36.39	37.51	38.22	37.24	37.24	37.26	36.75	36.88
MgO	5.56	4.55	4.62	4.97	5.51	4.82	4.73	4.71	5.72	5.55	5.31	5.36	5.03	5.39	5.65	4.62
CaO	1.86	1.94	2.08	2.04	1.39	1.42	1.17	1.46	1.53	1.45	1.86	1.20	0.96	1.82	1.00	1.29
TTL	100.03	100.68	101.50	101.02	99.61	100.34	100.44	100.96	99.77	98.15	99.48	102.00	99.76	100.19	99.00	100.27
Cations per 24 oxygens																
Fe	4.01	4.01	4.02	4.03	4.07	4.34	4.38	4.33	4.15	4.15	4.00	4.22	4.36	4.12	4.24	4.44
Mn	0.37	0.61	0.57	0.49	0.39	0.35	0.34	0.37	0.31	0.35	0.42	0.36	0.35	0.41	0.35	0.36
Na	0.00	0.00	0.41	0.00	0.00	0.00	0.00	0.02	0.00	0.01	0.01	0.00	0.01	0.01	0.01	0.01
Al	4.05	3.98	4.00	4.01	3.99	4.03	4.01	3.99	4.02	4.09	4.04	4.05	4.00	4.03	4.05	4.05
Si	5.96	6.00	5.98	5.97	6.00	5.94	5.96	5.96	5.95	5.88	5.97	5.95	5.95	5.91	5.91	5.89
Mg	1.31	1.07	1.08	1.17	1.30	1.14	1.12	1.11	1.35	1.34	1.26	1.24	1.19	1.27	1.34	1.10
Ca	0.31	0.33	0.35	0.34	0.24	0.24	0.20	0.25	0.26	0.25	0.32	0.20	0.16	0.31	0.17	0.22
TTL	16.01	16.00	16.01	16.01	15.99	16.04	16.03	16.03	16.04	16.07	16.01	16.02	16.04	16.07	16.07	16.08
Xalm	66.85	66.54	66.85	66.81	67.91	71.50	72.55	71.41	68.33	68.19	66.71	70.09	71.08	67.42	69.40	72.58
Xpyr	21.80	17.87	17.95	19.32	21.76	18.80	18.54	18.38	22.28	21.98	20.98	20.67	19.75	20.85	21.90	17.98
Xspe	6.11	10.12	9.40	8.12	6.42	5.73	5.64	6.16	5.12	5.71	7.01	5.96	5.76	6.69	5.71	5.87
Xgro	5.22	4.90	5.79	5.70	3.80	3.97	3.22	3.53	4.25	4.12	5.29	3.22	2.59	5.00	2.80	3.50
Xadu	0.01	0.57	0.01	0.00	0.11	0.00	0.05	0.53	0.01	0.00	0.00	0.06	0.01	0.04	0.00	0.07
SPL#478*	D47r	E1r	E2	E3	E4	E5	EScore	E9	E10r	*45*	Alr	A2	A3	A4core	A5	A6
FeO	32.54	32.88	32.68	31.70	32.30	33.10	32.92	32.25	32.60	*	35.09	35.90	35.69	34.98	35.65	35.97
MnO	2.52	3.69	2.19	2.06	2.24	3.60	2.87	2.35	2.46	*	1.15	1.54	1.89	2.00	1.55	1.05
Na2O	0.03	0.03	0.00	0.03	0.04	0.00	0.03	0.06	0.01	*	0.04	0.01	0.03	0.03	0.03	0.03
Al2O3	21.24	20.94	21.38	21.36	21.56	20.76	21.42	21.37	21.16	*	20.79	20.87	20.99	20.51	20.78	20.97
SiO2	37.36	36.91	37.84	37.38	37.59	37.64	37.14	37.55	37.38	*	37.22	37.08	36.75	36.94	37.07	36.74
MgO	4.95	3.56	5.46	5.76	5.29	4.02	4.52	5.33	4.66	*	3.23	3.09	2.78	2.54	2.75	3.07
CaO	1.07	1.06	1.32	1.25	1.32	1.01	1.14	1.09	0.87	*	2.38	2.27	2.53	2.49	2.52	2.40
TTL	99.71	99.13	100.88	99.57	100.08	100.19	100.04	100.00	99.28	*	99.98	100.75	100.73	99.54	100.37	100.23
Cations per 24 oxygens																
Fe	4.34	4.46	4.30	4.22	4.25	4.43	4.40	4.28	4.37	*	4.71	4.80	4.80	4.73	4.83	4.84
Mn	0.34	0.51	0.29	0.28	0.31	0.49	0.39	0.32	0.33	*	0.16	0.21	0.26	0.28	0.21	0.14
Na	0.01	0.01	0.00	0.01	0.01	0.00	0.00	0.02	0.00	*	0.01	0.00	0.01	0.01	0.01	0.01
Al	4.00	4.00	3.97	4.00	4.03	3.91	4.04	4.00	4.00	*	3.94	3.95	3.98	3.93	3.94	3.98
Si	5.97	5.98	5.96	5.95	5.96	6.02	5.94	5.97	6.00	*	5.99	5.95	5.92	6.00	5.97	5.93
Mg	1.18	0.86	1.28	1.37	1.25	0.96	1.08	1.26	1.14	*	0.78	0.74	0.67	0.61	0.66	0.74
Ca	0.18	0.18	0.22	0.21	0.22	0.17	0.19	0.19	0.15	*	0.41	0.39	0.44	0.43	0.43	0.42
TTL	16.03	16.01	16.03	16.05	16.03	15.98	16.04	16.03	16.00	*	16.01	16.04	16.07	16.00	16.03	16.05
Xalm	71.81	74.25	70.37	69.41	70.53	72.88	72.60	70.80	72.95	*	77.70	78.00	77.84	78.00	78.41	78.80
Xpyr	19.53	14.31	21.15	22.50	20.77	16.06	17.79	20.88	18.99	*	12.92	12.13	10.90	10.24	10.91	12.09
Xspe	5.63	8.44	4.82	4.58	5.00	8.17	6.40	5.24	5.58	*	2.61	3.44	4.21	4.60	3.50	2.34
Xgro	2.83	2.89	2.72	3.40	3.68	0.87	3.21	2.91	2.43	*	5.35	4.83	6.21	5.35	5.65	6.04
Xadu	0.20	0.10	0.94	0.11	0.02	2.03	0.00	0.17	0.05	*	1.43	1.60	0.85	1.82	1.52	0.74

SPL*45*	A6	A7r	A8	A9	A11	A12r	A13	A14r	A15r	B16r	B17r	B18	B19	B20core	B23	B24
FeO	35.77	36.32	35.60	35.80	35.35	36.81	35.72	36.15	36.10	34.95	36.14	35.79	35.79	35.61	36.23	35.74
MnO	1.05	0.86	1.81	1.49	1.12	0.96	1.88	1.10	0.73	0.90	0.84	1.35	2.09	2.21	1.41	1.16
Na2O	0.03	0.01	0.05	0.03	0.02	0.01	0.02	0.03	0.01	0.01	0.01	0.01	0.04	0.04	0.01	0.06
Al2O3	20.97	20.85	20.97	20.85	20.92	21.08	20.86	20.80	21.22	20.96	20.88	20.94	20.86	20.94	20.94	21.17
SiO2	36.74	36.80	36.72	36.32	36.74	36.40	36.30	36.91	37.10	36.04	36.87	36.75	36.86	36.51	36.43	36.50
MgO	3.07	3.36	2.66	2.94	3.15	3.02	2.64	3.13	3.49	3.12	3.18	2.77	2.47	2.37	2.58	2.88
CaO	2.40	1.82	2.48	2.38	2.41	2.21	2.78	2.36	1.93	1.90	1.99	2.56	2.46	2.73	2.77	2.67
TTL	100.23	100.09	100.60	99.87	99.78	100.53	100.31	100.52	100.61	98.20	100.28	100.18	100.65	100.48	100.39	100.19
Cations per 24 oxygens																
Fe	4.84	4.90	4.80	4.86	4.78	4.97	4.84	4.85	4.83	4.79	4.91	4.83	4.82	4.81	4.90	4.82
Mn	0.14	0.12	0.25	0.21	0.15	0.13	0.26	0.15	0.10	0.12	0.12	0.19	0.29	0.30	0.19	0.16
Na	0.01	0.02	0.01	0.01	0.01	0.00	0.01	0.01	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.01
Al	3.99	3.97	3.98	3.99	3.99	4.01	3.98	3.94	4.00	4.05	3.97	3.99	3.96	3.99	3.99	4.03
Si	5.93	5.94	5.92	5.90	5.94	5.88	5.86	5.94	5.94	5.91	5.94	5.94	5.94	5.90	5.89	5.90
Mg	0.74	0.81	0.64	0.71	0.76	0.73	0.64	0.75	0.83	0.76	0.76	0.67	0.59	0.57	0.62	0.69
Ca	0.41	0.32	0.43	0.41	0.42	0.38	0.48	0.41	0.33	0.33	0.34	0.44	0.42	0.47	0.47	0.46
TTL	16.07	16.05	16.06	16.09	16.05	16.11	16.10	16.05	16.05	16.02	16.05	16.05	16.05	16.08	16.09	16.09
Xalm	78.80	79.68	78.77	78.42	78.24	80.03	77.82	78.62	79.31	80.01	79.96	78.79	78.66	78.11	79.05	78.62
Xpyr	12.09	13.27	10.50	11.57	12.46	11.75	10.31	12.31	13.67	12.72	12.53	10.93	9.76	9.33	10.08	11.30
Xspe	2.34	1.93	4.05	3.33	2.52	2.11	4.18	2.45	1.64	2.09	1.89	3.03	4.71	4.94	3.13	2.59
Xgro	6.04	4.01	6.15	5.90	6.24	5.77	6.81	4.92	5.26	4.57	4.51	6.61	5.75	7.00	7.09	7.46
Xadu	0.74	1.11	0.54	0.77	0.55	0.34	0.88	1.69	0.13	0.60	1.10	0.63	1.14	0.63	0.65	0.03
SPL*45*	B26	B27	B28r	*E4*	Alr	A2r	A3	A4c	A5	A6	A7	A8	A9	A10	Al1r	Bl1r
FeO	36.64	36.38	35.64	*	35.81	35.56	36.16	35.49	36.69	36.01	35.50	35.84	35.95	36.14	36.38	36.60
MnO	1.10	1.07	0.81	*	1.13	1.05	1.12	1.18	1.18	1.05	1.01	1.08	1.02	1.13	1.43	1.86
Na2O	0.01	0.00	0.05	*	0.06	0.01	0.00	0.03	0.05	0.01	0.00	0.01	0.01	0.03	0.03	0.05
Al2O3	21.28	21.10	21.43	*	21.12	21.09	21.08	21.01	21.05	20.92	21.15	21.29	20.86	21.04	20.93	21.01
SiO2	36.70	36.33	37.07	*	37.01	37.17	37.21	37.42	36.20	37.16	37.21	37.38	37.40	37.02	37.19	37.30
MgO	3.10	3.06	3.26	*	4.08	4.09	4.05	4.02	3.36	4.05	4.19	4.10	4.22	4.03	3.78	2.89
CaO	2.28	2.16	2.64	*	1.12	1.11	1.09	1.11	1.12	1.06	1.10	1.05	1.12	1.18	1.10	0.98
TTL	100.10	100.10	100.94	*	100.33	100.07	100.73	100.26	99.64	100.27	100.25	100.75	100.58	100.60	100.84	100.72
Cations per 24 oxygens																
Fe	4.84	4.93	4.76	*	4.79	4.75	4.82	4.74	4.99	4.82	4.75	4.77	4.79	4.83	4.85	4.90
Mn	0.15	0.15	0.11	*	0.15	0.14	0.15	0.16	0.16	0.14	0.14	0.15	0.14	0.15	0.19	0.25
Na	0.00	0.00	0.02	*	0.02	0.00	0.00	0.01	0.10	0.00	0.00	0.00	0.00	0.01	0.01	0.02
Al	4.02	4.03	4.03	*	3.99	3.99	3.97	3.96	4.04	3.95	3.99	4.00	3.93	3.97	3.95	3.98
Si	5.88	5.88	5.92	*	5.93	5.96	5.95	5.99	5.89	5.96	5.96	5.96	5.97	5.93	5.95	5.99
Mg	0.74	0.74	0.78	*	0.98	0.98	0.97	0.96	0.81	0.97	1.00	0.97	1.00	0.96	0.90	0.69
Ca	0.39	0.37	0.45	*	0.19	0.19	0.19	0.19	0.19	0.18	0.19	0.18	0.19	0.20	0.18	0.17
TTL	16.10	16.10	16.07	*	16.06	16.03	16.05	16.01	16.10	16.03	16.04	16.04	16.02	16.07	16.05	16.01
Xalm	79.30	79.64	78.09	*	78.33	78.36	78.62	78.23	80.99	78.69	78.17	78.59	77.92	78.45	78.89	81.44
Xpyr	11.97	11.96	12.74	*	16.01	16.15	15.85	15.94	13.21	15.95	16.45	16.04	16.61	15.76	15.84	11.54
Xspe	2.41	2.37	1.61	*	2.51	2.35	2.49	2.66	2.64	2.36	2.26	2.42	2.28	2.51	3.18	4.22
Xgro	6.25	6.02	7.30	*	2.54	2.65	2.00	2.16	3.16	1.65	2.66	2.70	1.18	2.18	1.49	2.17
Xadu	0.08	0.02	0.06	*	0.61	0.49	1.04	1.01	0.00	1.33	0.46	0.25	2.00	1.10	1.62	0.62

SPLE*04*	B13	B14	B15	B16	B18c	B19	B21	B22	B21r
FeO	35.80	35.68	35.87	35.85	35.66	36.24	35.52	35.90	36.31
MnO	1.24	1.08	1.08	1.12	1.14	1.18	1.19	1.25	1.76
Na2O	0.03	0.06	0.01	0.03	0.03	0.02	0.02	0.05	0.03
Al2O3	20.79	20.84	21.12	20.81	20.83	20.99	21.05	20.86	20.86
SiO2	37.28	37.37	37.23	37.28	37.47	37.21	36.64	36.53	36.42
MgO	3.83	3.98	4.07	4.01	4.10	4.02	3.89	3.57	3.42
CaO	0.96	0.99	0.99	1.01	1.01	1.12	1.12	1.06	1.10
TTL	99.95	100.03	100.42	100.11	100.24	100.82	99.43	99.22	99.93
Cations per 24 oxygens									
Fe	4.79	4.77	4.78	4.79	4.76	4.82	4.80	4.87	4.91
Mn	0.02	0.02	0.15	0.15	0.15	0.16	0.16	0.17	0.24
Na	0.01	0.02	0.00	0.01	0.01	0.01	0.01	0.01	0.01
Al	3.94	3.94	3.98	3.94	3.93	3.95	4.02	4.00	3.99
Si	5.99	6.00	5.96	5.98	6.00	5.94	5.93	5.94	5.91
Mg	0.92	0.95	0.97	0.96	0.98	0.96	0.94	0.86	0.83
Ca	0.17	0.17	0.17	0.17	0.17	0.19	0.19	0.18	0.19
TTL	16.01	16.01	16.03	16.02	16.00	16.05	16.06	16.06	16.08
Xalm	79.14	78.90	78.82	78.66	78.22	78.52	78.77	79.93	79.55
Xpyr	15.31	15.88	16.01	15.93	16.32	15.74	15.38	14.21	13.45
Xspe	2.81	2.44	2.42	2.53	2.57	2.62	2.66	2.82	3.93
Xgro	1.22	1.28	2.18	1.20	1.06	1.67	3.19	2.65	2.24
Xadu	1.52	1.50	0.57	1.67	1.82	1.45	0.00	0.39	0.83

BIOTITE

SPL#	OPl	Alr	A2al	A3al	A4al	A9al	Allal	A12al	A13am	A14ar	A8bl	A3cm	Aldr	A2dr	A3drc	A4drc	A10dr
FeO	19.78	18.51	18.92	18.77	20.17	20.77	20.89	21.18	21.75	18.81	21.62	19.47	20.01	19.68	19.38	20.08	
MnO	0.49	0.28	0.37	0.28	0.56	0.68	0.48	0.56	0.71	0.42	0.53	0.51	0.53	0.54	0.44	0.64	
Na2O	0.16	0.19	0.15	0.15	0.11	0.12	0.16	0.12	0.11	0.15	0.12	0.13	0.16	0.13	0.11	0.11	
Al2O3	18.63	19.89	19.27	21.03	18.75	18.79	18.61	17.99	17.49	19.13	18.57	19.27	19.43	19.07	19.08	19.45	
SiO2	36.01	35.38	36.19	35.36	35.83	36.18	35.88	35.84	35.47	36.49	35.96	35.43	35.97	35.61	35.72	36.16	
MgO	8.45	8.51	8.89	8.49	8.15	7.93	8.18	7.97	7.87	8.76	7.99	8.49	8.53	8.32	8.42	8.47	
K2O	10.18	9.99	10.07	9.81	10.24	10.27	10.22	10.24	10.11	10.36	10.26	10.25	10.21	10.31	10.31	10.05	
CaO	0.01	0.11	0.04	0.13	0.01	0.03	0.02	0.03	0.04	0.04	0.01	0.02	0.01	0.02	0.04	0.02	
TiO2	2.68	2.32	2.33	2.21	1.56	1.98	2.19	2.56	2.81	2.35	2.68	2.71	2.91	2.69	2.57	1.91	
OH	3.95	3.93	3.97	3.99	3.91	3.95	3.94	3.92	3.91	3.98	3.97	3.95	4.01	3.95	3.95	3.98	
TTL	100.33	99.11	100.21	100.21	99.26	100.71	100.56	100.41	100.23	100.45	101.79	100.23	101.79	100.21	100.01	100.88	

Cations per 22 oxygens

Fe	2.51	2.36	2.39	2.36	2.59	2.63	2.66	2.71	2.79	2.37	2.73	2.47	2.49	2.49	2.46	2.53	
Mn	0.06	0.04	0.05	0.04	0.07	0.09	0.06	0.07	0.09	0.05	0.07	0.06	0.07	0.07	0.06	0.08	
Na	0.04	0.06	0.05	0.04	0.03	0.04	0.03	0.03	0.03	0.04	0.03	0.04	0.04	0.04	0.03	0.03	
Al	3.33	3.57	3.42	3.73	3.39	3.36	3.33	3.23	3.17	3.39	3.29	3.44	3.42	3.41	3.41	3.45	
Aliv	0.78	0.96	0.88	1.04	0.89	0.85	0.79	0.71	0.62	0.88	0.72	0.82	0.79	0.82	0.84	0.89	
Si	5.45	5.39	5.45	5.31	5.51	5.49	5.45	5.47	5.45	5.49	5.42	5.37	5.37	5.41	5.42	5.44	
Mg	1.91	1.92	1.99	1.91	1.86	1.79	1.85	1.81	1.79	1.96	1.79	1.92	1.89	1.88	1.91	1.89	
K	1.97	1.94	1.94	1.88	2.01	1.98	1.98	1.99	1.98	1.99	1.99	1.98	1.95	1.97	1.99	1.93	
Ca	0.00	0.02	0.00	0.02	0.00	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.00	
Ti	0.31	0.26	0.26	0.25	0.18	0.23	0.25	0.29	0.32	0.27	0.29	0.31	0.33	0.31	0.29	0.22	
OH	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	
TTL	17.58	17.56	17.56	17.53	17.64	17.62	17.64	17.63	17.65	17.56	17.64	17.61	17.59	17.59	17.59	17.59	

SPL#*OPl* Alldrc A2er *5A*

FeO	20.25	20.64	* 17.88	17.98	17.98	18.01	16.06	15.66	17.85	17.79	18.11	18.81	18.37	15.87	15.81	15.36	
MnO	0.67	0.56	* 0.18	0.26	0.26	0.16	0.12	0.00	0.23	0.17	0.12	0.26	0.20	0.12	0.12	0.12	
Na2O	0.12	0.11	* 0.16	0.14	0.14	0.24	0.30	0.29	0.17	0.14	0.16	0.15	0.19	0.29	0.28	0.31	
Al2O3	19.21	18.42	* 19.65	19.77	18.94	19.19	18.98	19.12	19.46	19.46	19.43	19.41	19.73	18.74	18.82	18.98	
SiO2	35.76	35.65	* 35.83	35.62	35.29	36.29	35.56	35.41	34.88	35.93	34.37	34.37	35.70	36.08	35.56	36.01	
MgO	8.35	8.11	* 10.11	9.87	12.08	11.30	10.87	9.65	9.65	9.90	9.98	10.43	10.07	11.24	11.03	11.29	
K2O	10.35	10.23	* 10.12	10.09	6.70	9.49	9.44	9.85	10.10	10.04	8.66	9.78	9.78	9.58	9.39	9.76	
CaO	0.02	0.00	* 0.04	0.02	0.13	0.04	0.02	0.02	0.08	0.06	0.00	0.06	0.09	0.02	0.04	0.05	
TiO2	1.86	2.78	* 2.54	2.43	3.09	4.45	4.35	2.26	2.26	2.35	2.76	2.03	2.23	4.24	4.18	4.29	
OH	3.95	3.94	* 4.00	3.99	3.99	3.99	4.08	4.00	3.92	3.92	4.00	3.90	4.00	4.00	4.00	4.04	
TTL	100.51	100.43	* 100.52	100.18	98.64	101.31	99.19	98.54	98.77	98.77	100.55	98.10	100.36	100.20	99.23	100.22	

Cations per 22 oxygens

Fe	2.57	2.63	* 2.23	2.26	2.26	2.26	1.97	1.96	2.28	2.27	2.27	2.41	2.30	1.97	1.98	1.90	
Mn	0.09	0.07	* 0.02	0.03	0.03	0.02	0.01	0.00	0.03	0.02	0.02	0.03	0.02	0.01	0.01	0.01	
Na	0.03	0.03	* 0.04	0.04	0.04	0.07	0.08	0.08	0.05	0.04	0.04	0.04	0.05	0.08	0.08	0.09	
Al	3.43	3.31	* 3.46	3.50	3.35	3.35	3.31	3.35	3.44	3.50	3.42	3.51	3.48	3.27	3.32	3.31	
Aliv	0.86	0.73	* 0.86	0.73	0.73	0.73	0.73	0.73	0.73	0.73	0.73	0.73	0.73	0.73	0.73	0.73	
Si	5.42	5.42	* 5.36	5.35	5.29	5.32	5.32	5.32	5.40	5.32	5.32	5.27	5.35	5.35	5.33	5.33	
Mg	1.88	1.84	* 2.25	2.21	2.70	2.46	2.42	2.19	2.19	2.25	2.22	2.38	2.25	2.48	2.46	2.49	
K	2.01	1.99	* 1.93	1.93	1.28	1.77	1.80	1.91	1.91	1.97	1.91	1.69	1.87	1.81	1.79	1.84	
Ca	0.00	0.00	* 0.01	0.00	0.02	0.01	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	
Ti	0.21	0.32	* 0.29	0.27	0.34	0.49	0.48	0.25	0.25	0.26	0.31	0.23	0.25	0.47	0.47	0.47	
OH	2.00	2.00	* 2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	
TTL	17.66	17.61	* 17.60	17.60	17.35	17.45	17.44	17.59	17.59	17.56	17.58	17.60	17.61	17.47	17.47	17.49	

SPLE*5A*	A7c	A8c	A9m	A9aZ	A10aZ	A11aZ	B12m	B13F	B14Z	B15I	B17m	B18F	B1aZ	B2aZ	B3aZ	B7aZ
FeO	18.71	18.08	17.18	18.65	18.03	18.39	17.68	16.97	16.95	16.01	17.31	16.79	17.75	17.56	17.84	16.31
MnO	0.25	0.19	0.29	0.22	0.18	0.12	0.20	0.10	0.19	0.16	0.17	0.22	0.13	0.20	0.19	0.13
Na2O	0.12	0.20	0.21	0.20	0.17	0.17	0.18	0.24	0.16	0.25	0.20	0.20	0.21	0.19	0.21	0.17
Al2O3	19.20	19.89	20.09	19.77	19.33	19.39	19.43	20.35	20.28	19.24	19.62	20.40	19.97	19.90	19.83	20.13
SiO2	35.08	35.16	34.92	35.66	35.78	35.42	34.95	34.37	35.09	35.14	35.11	34.54	35.98	36.15	36.06	35.75
MgO	9.75	9.99	9.22	10.28	10.28	10.25	10.05	10.71	11.02	11.64	10.20	10.95	10.53	10.50	10.85	11.00
K2O	9.86	10.15	10.29	10.18	9.97	9.84	10.05	10.19	9.52	9.88	10.02	10.19	10.11	9.97	10.07	10.18
CaO	0.03	0.04	0.03	0.01	0.05	0.08	0.01	0.05	0.13	0.02	0.01	0.03	0.04	0.04	0.04	0.06
TiO2	2.46	2.42	3.28	2.44	2.30	2.36	2.66	1.52	1.15	3.32	2.65	1.42	2.02	1.95	1.94	1.68
OH	3.94	3.98	3.96	4.02	3.98	3.98	3.94	3.92	3.95	4.00	3.96	3.94	4.02	4.02	4.03	3.99
TTL	99.40	100.89	99.47	101.45	100.08	100.00	99.15	98.43	98.44	99.67	99.24	98.67	100.73	100.49	101.05	99.59

Cations per 22 oxygens

Fe	2.38	2.27	2.17	2.32	2.27	2.32	2.25	2.17	2.15	2.01	2.19	2.14	2.21	2.19	2.22	2.07
Mn	0.03	0.02	0.03	0.03	0.02	0.01	0.02	0.01	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02
Na	0.04	0.05	0.06	0.06	0.05	0.05	0.05	0.07	0.04	0.07	0.05	0.06	0.06	0.06	0.06	0.05
Al	3.44	3.53	3.58	3.47	3.42	3.44	3.48	3.66	3.62	3.40	3.50	3.65	3.50	3.49	3.47	3.56
Alvi	0.78	0.83	0.86	0.78	0.80	0.78	0.79	0.91	0.96	0.66	0.81	0.91	0.86	0.89	0.83	0.92
Si	5.33	5.29	5.28	5.31	5.38	5.34	5.31	5.24	5.32	5.26	5.31	5.25	5.36	5.39	5.36	5.36
Mg	2.21	2.24	2.07	2.28	2.30	2.30	2.27	2.43	2.49	2.60	2.30	2.48	2.33	2.33	2.40	2.45
K	1.91	1.95	1.98	1.93	1.91	1.89	1.94	1.98	1.84	1.89	1.90	1.97	1.92	1.90	1.91	1.94
Ca	0.00	0.01	0.00	0.00	0.01	0.01	0.00	0.01	0.02	0.00	0.00	0.00	0.01	0.01	0.01	0.01
Ti	0.28	0.27	0.37	0.27	0.25	0.26	0.30	0.17	0.13	0.37	0.30	0.16	0.23	0.21	0.21	0.19
OH	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
TTL	17.63	17.66	17.57	17.68	17.63	17.64	17.65	17.77	17.67	17.64	17.63	17.77	17.65	17.61	17.67	17.66

SPLE*5A*	B4F	D9F	D10F	D11F	E6m	*5B*	A11F	A12I	A14m	C19F	C20I	C21I	C22I	C24F	C25m	C1aZ
FeO	19.25	17.70	18.16	18.15	18.25	*	17.90	16.64	18.36	18.19	17.49	17.62	18.08	18.13	18.44	18.19
MnO	0.20	0.24	0.27	0.27	0.22	*	0.20	0.08	0.20	0.18	0.14	0.13	0.18	0.20	0.17	0.25
Na2O	0.12	0.20	0.22	0.19	0.16	*	0.15	0.56	0.11	0.17	0.31	0.31	0.20	0.16	0.12	0.19
Al2O3	19.78	19.47	19.70	20.14	19.46	*	19.61	18.60	19.44	20.03	19.22	18.92	18.97	19.60	19.35	19.58
SiO2	34.39	36.02	35.73	36.06	36.11	*	35.88	36.62	36.34	35.77	36.70	36.77	35.93	36.00	35.87	35.68
MgO	11.33	9.98	10.03	9.95	9.97	*	9.73	10.56	9.72	10.03	10.40	10.45	10.03	9.77	9.70	10.06
K2O	7.74	10.20	10.10	10.23	10.15	*	10.05	9.37	10.08	9.58	9.71	9.48	9.47	9.91	9.95	9.98
CaO	0.06	0.00	0.05	0.00	0.00	*	0.01	0.01	0.03	0.06	0.00	0.00	0.04	0.06	0.01	0.04
TiO2	1.09	2.35	2.12	1.76	2.91	*	2.73	4.01	2.88	2.02	3.57	3.66	3.45	2.68	2.92	2.34
OH	3.91	3.99	3.99	4.01	4.03	*	4.00	4.04	4.03	3.99	4.07	4.06	4.01	4.01	4.00	3.99
TTL	97.88	100.16	100.38	100.77	101.28	*	100.25	100.49	101.20	100.02	101.61	101.40	100.35	100.50	100.53	100.29

Cations per 22 oxygens

Fe	2.47	2.22	2.28	2.27	2.27	*	2.24	2.06	2.28	2.28	2.15	2.17	2.26	2.27	2.31	2.28
Mn	0.03	0.03	0.03	0.03	0.03	*	0.03	0.01	0.02	0.02	0.01	0.02	0.02	0.03	0.02	0.03
Na	0.03	0.06	0.07	0.05	0.05	*	0.04	0.16	0.03	0.04	0.09	0.09	0.05	0.05	0.03	0.05
Al	3.57	3.44	3.49	3.54	3.41	*	3.46	3.24	3.41	3.54	3.33	3.29	3.34	3.45	3.41	3.46
Alvi	0.78	0.83	0.86	0.78	0.80	*	0.78	0.91	0.96	0.66	0.81	0.91	0.86	0.89	0.83	0.92
Si	5.27	5.40	5.36	5.38	5.37	*	5.37	5.42	5.40	5.36	5.40	5.42	5.37	5.38	5.37	5.36
Mg	2.58	2.23	2.24	2.21	2.20	*	2.17	2.33	2.15	2.24	2.28	2.30	2.23	2.17	2.16	2.25
K	1.51	1.95	1.93	1.95	1.92	*	1.92	1.77	1.91	1.83	1.82	1.78	1.81	1.89	1.90	1.91
Ca	0.01	0.00	0.01	0.00	0.00	*	0.00	0.00	0.00	0.01	0.00	0.00	0.01	0.01	0.00	0.01
Ti	0.12	0.27	0.24	0.19	0.32	*	0.30	0.44	0.32	0.22	0.40	0.41	0.39	0.30	0.33	0.26
OH	2.00	2.00	2.00	2.00	2.00	*	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
TTL	17.59	17.61	17.65	17.64	17.58	*	17.56	17.46	17.54	17.57	17.49	17.47	17.50	17.55	17.56	17.63

SPLE*58*	C2acr	C3ar	Dir	D2r	D3r	D4m	D5r	*7*	A1r	A2r	A3m	Ar	A5r	A6i	A7i	B10r
FeO	18.13	18.73	18.59	18.43	18.43	18.08	18.17	*	17.38	17.24	17.74	18.02	17.83	17.92	18.44	18.80
MnO	0.31	0.15	0.13	0.21	0.13	0.17	0.20	*	0.11	0.19	0.16	0.23	0.17	0.24	0.23	0.30
Na2O	0.12	0.17	0.22	0.24	0.19	0.15	0.14	*	0.35	0.34	0.27	0.18	0.16	0.35	0.25	0.17
Al2O3	19.61	19.58	19.87	19.95	19.59	19.52	19.75	*	19.06	18.96	19.67	20.03	20.17	19.34	19.87	19.53
SiO2	35.63	35.16	35.69	36.03	35.81	35.94	35.92	*	35.06	34.80	36.03	36.16	36.03	35.44	35.89	35.15
MgO	9.82	9.84	9.94	10.08	9.86	9.93	9.93	*	9.33	9.04	9.44	9.13	9.48	9.54	9.45	8.86
K2O	10.00	9.90	9.98	9.92	9.96	9.93	10.06	*	9.97	9.52	10.18	10.23	10.15	10.45	10.20	9.92
CaO	0.03	0.02	0.02	0.01	0.00	0.02	0.01	*	0.02	0.07	0.01	0.05	0.00	0.01	0.00	0.00
TiO2	2.32	2.19	2.37	2.30	2.58	2.66	1.81	*	2.44	2.34	2.85	2.36	2.44	1.72	2.08	2.82
OH	3.98	3.95	4.01	4.03	4.00	4.00	3.98	*	3.89	3.84	4.00	4.00	4.01	3.92	3.99	3.94
TTL	99.96	99.69	100.80	101.22	100.56	100.40	99.97	*	97.62	96.35	100.36	100.38	100.44	98.94	100.39	99.49
Cations per 22 oxygens																
Fe	2.28	2.37	2.32	2.29	2.31	2.26	2.29	*	2.24	2.24	2.22	2.26	2.23	2.29	2.31	2.39
Mn	0.04	0.02	0.02	0.03	0.02	0.02	0.03	*	0.01	0.03	0.02	0.03	0.02	0.03	0.03	0.04
Na	0.03	0.05	0.06	0.07	0.06	0.04	0.04	*	0.11	0.10	0.08	0.05	0.04	0.10	0.07	0.05
Al	3.48	3.49	3.50	3.49	3.46	3.44	3.50	*	3.46	3.48	3.47	3.54	3.55	3.48	3.51	3.49
Alv1	0.85	0.82	0.82	0.82	0.82	0.82	0.82	*	0.87	0.90	0.87	0.95	0.94	0.89	0.91	0.91
Si	5.37	5.32	5.33	5.36	5.36	5.38	5.40	*	5.40	5.42	5.39	5.41	5.38	5.41	5.39	5.37
Mg	1.92	1.91	1.90	1.88	1.90	1.89	1.93	*	1.96	1.99	1.94	1.95	1.94	2.03	1.95	1.92
K	0.01	0.00	0.00	0.00	0.00	0.00	0.00	*	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00
Ca	0.01	0.00	0.00	0.00	0.00	0.00	0.00	*	0.00	0.01	0.00	0.01	0.00	0.00	0.00	0.00
Ti	0.26	0.25	0.27	0.26	0.29	0.30	0.20	*	0.28	0.27	0.32	0.27	0.27	0.20	0.23	0.32
OH	2.00	2.00	2.00	2.00	2.00	2.00	2.00	*	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
TTL	17.60	17.66	17.63	17.61	17.59	17.56	17.62	*	17.61	17.56	17.56	17.55	17.55	17.71	17.62	17.57
SPLE*7*																
FeO	18.48	18.69	* 17.49	15.77	15.44	17.42	18.70	18.11	18.51	18.23	17.34	15.83	15.50	17.68	20.75	
MnO	0.21	0.21	* 0.05	0.06	0.05	0.10	0.17	0.13	0.13	0.17	0.09	0.04	0.07	0.17	0.30	
Na2O	0.16	0.19	* 0.22	0.17	0.28	0.30	0.13	0.17	0.16	0.17	0.29	0.26	0.31	0.18	0.10	
Al2O3	19.57	19.26	* 19.20	18.95	19.16	18.78	18.81	19.10	18.70	18.91	18.60	18.89	18.99	18.50	20.42	
SiO2	35.49	35.37	* 36.17	35.42	35.64	35.81	35.92	35.57	35.88	36.34	36.31	36.73	36.76	36.20	33.93	
MgO	9.22	9.06	* 10.50	12.08	12.18	10.95	10.06	9.83	10.76	10.56	10.56	11.62	11.81	9.96	8.99	
K2O	10.17	10.04	* 10.11	9.82	9.85	9.94	10.34	10.22	10.24	10.17	9.68	9.96	9.88	9.95	10.08	
CaO	0.03	0.00	* 0.00	0.02	0.02	0.03	0.00	0.01	0.00	0.03	0.03	0.09	0.03	0.00	0.02	
TiO2	2.33	2.95	* 2.84	3.09	3.24	2.59	2.97	3.27	2.36	2.94	2.78	3.83	3.74	3.21	0.51	
OH	3.95	3.95	* 4.02	4.00	4.02	3.99	4.01	3.99	3.98	4.04	3.99	4.08	4.08	3.98	3.88	
TTL	99.61	99.73	* 100.60	99.57	99.90	99.92	101.11	100.41	100.51	101.57	99.75	101.26	101.15	99.83	99.04	
Cations per 22 oxygens																
Fe	2.34	2.37	* 2.18	1.98	1.92	2.19	2.34	2.28	2.33	2.26	2.18	1.94	1.90	2.22	2.68	
Mn	0.03	0.03	* 0.01	0.01	0.01	0.01	0.02	0.02	0.02	0.02	0.01	0.01	0.01	0.02	0.03	
Na	0.05	0.06	* 0.06	0.11	0.08	0.09	0.04	0.05	0.05	0.05	0.09	0.07	0.09	0.03	0.03	
Al	3.49	3.44	* 3.37	3.34	3.36	3.22	3.31	3.38	3.31	3.30	3.29	3.26	3.28	3.28	3.71	
Alv1	0.76	0.64	* 0.76	0.64	0.66	0.70	0.69	0.72	0.68	0.68	0.68	0.72	0.68	0.68	0.96	
Si	5.38	5.36	* 5.39	5.30	5.31	5.38	5.37	5.34	5.37	5.39	5.44	5.39	5.39	5.44	5.24	
Mg	2.08	2.04	* 2.33	2.69	2.70	2.45	2.24	2.20	2.41	2.33	2.36	2.54	2.58	2.23	2.07	
K	1.97	1.94	* 1.92	1.88	1.87	1.91	1.97	1.96	1.96	1.92	1.85	1.87	1.85	1.91	1.98	
Ca	0.00	0.00	* 0.00	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.01	0.01	0.01	0.00	0.01	
Ti	0.26	0.34	* 0.32	0.35	0.36	0.29	0.33	0.37	0.27	0.33	0.31	0.42	0.41	0.26	0.06	
OH	2.00	2.00	* 2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	
TTL	17.61	17.57	* 17.59	17.66	17.62	17.75	17.53	17.60	17.71	17.61	17.55	17.51	17.51	17.52	17.84	

SPL#47A*	B11ac	B12ar	*47B*	B5r	B6i	B7i	B8r	B9i	B10a	C13a	C1a	D6r	D7i	D8i	D9r	D10a
FeO	18.80	18.13	*	18.24	16.96	17.06	20.77	16.58	17.42	17.33	17.59	17.16	15.04	15.38	18.31	18.02
MnO	0.19	0.18	*	0.18	0.06	0.07	0.21	0.06	0.11	0.13	0.09	0.20	0.07	0.06	0.14	0.09
Na2O	0.10	0.22	*	0.22	0.30	0.35	0.22	0.43	0.21	0.28	0.31	0.15	0.40	0.32	0.20	0.30
Al2O3	18.57	18.77	*	18.49	18.59	18.23	18.32	17.99	18.37	19.26	14.55	19.43	18.51	18.44	19.47	18.27
SiO2	35.38	35.86	*	36.29	36.00	36.14	35.19	35.85	35.70	35.04	36.34	36.17	36.55	36.69	36.10	36.25
MgO	10.43	10.33	*	10.62	11.22	11.32	8.79	12.61	11.11	10.77	10.78	9.62	12.27	11.88	10.92	10.82
K2O	10.22	10.07	*	9.98	9.94	9.78	10.11	9.37	10.04	10.14	10.16	10.04	9.97	9.87	10.13	10.12
CaO	0.00	0.00	*	0.00	0.02	0.01	0.00	0.01	0.03	0.01	0.00	0.02	0.10	0.00	0.01	0.03
TiO2	2.54	3.27	*	2.77	4.32	3.40	2.98	3.14	3.04	2.76	2.83	3.13	4.45	4.32	1.52	3.12
OH	3.97	4.01	*	4.01	4.06	4.01	3.94	4.01	3.98	3.97	4.06	4.00	4.11	4.08	4.01	4.02
TTL	100.34	100.83	*	100.79	101.46	100.39	100.52	100.05	100.01	99.69	101.73	99.91	101.78	101.03	100.82	101.03
Cations per 22 oxygens																
Fe	2.38	2.27	*	2.28	2.09	2.13	2.64	2.07	2.19	2.19	2.17	2.15	1.83	1.89	2.29	2.25
Mn	0.02	0.02	*	0.02	0.01	0.01	0.03	0.01	0.01	0.02	0.01	0.02	0.01	0.01	0.02	0.01
Na	0.07	0.06	*	0.06	0.09	0.10	0.06	0.12	0.06	0.08	0.09	0.04	0.11	0.09	0.06	0.09
Al	3.31	3.31	*	3.26	3.24	3.21	3.29	3.16	3.25	3.43	3.40	3.43	3.18	3.19	3.43	3.51
Alvi	0.65	0.67	*													
Si	5.35	5.35	*	5.42	5.32	5.39	5.35	5.35	5.37	5.29	5.36	5.42	5.38	5.39	5.39	5.40
Mg	2.34	2.30	*	2.36	2.47	2.52	1.99	2.80	2.49	2.42	2.37	2.15	2.66	2.60	2.43	2.40
K	1.97	0.92	*	1.90	1.87	1.86	1.96	1.78	1.93	1.95	1.91	1.92	1.85	1.85	1.93	1.92
Ca	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	0.00
Ti	0.29	0.37	*	0.31	0.48	0.38	0.34	0.35	0.34	0.31	0.31	0.35	0.49	0.48	0.17	0.35
OH	2.00	2.00	*	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
TTL	17.73	17.61	*	17.62	17.56	17.60	17.67	17.66	17.65	17.70	17.63	17.49	17.52	17.50	17.72	17.64
SPL#47B*																
FeO	17.13	17.25	*	20.74	18.35	18.13	*	20.89	20.48	20.44	20.91	19.80	19.53	18.74	21.43	21.20
MnO	0.14	0.09	*	0.10	0.04	0.04	*	0.00	0.09	0.00	0.01	0.06	0.05	0.01	0.05	0.10
Na2O	0.16	0.23	*	0.26	0.34	0.13	*	0.31	0.30	0.24	0.31	0.30	0.37	0.38	0.32	0.33
Al2O3	19.32	19.49	*	19.86	20.30	19.62	*	18.85	18.74	18.90	20.16	19.00	19.15	19.00	18.65	19.51
SiO2	35.98	35.94	*	35.32	36.38	37.05	*	35.27	35.48	35.59	35.28	35.93	36.34	36.05	35.39	34.90
MgO	10.21	10.13	*	9.13	9.20	9.27	*	7.88	7.88	8.24	8.92	8.77	8.65	8.94	8.16	8.22
K2O	10.07	10.19	*	9.68	9.69	5.36	*	9.63	9.58	9.47	8.78	8.48	9.44	9.58	9.32	9.30
CaO	0.00	0.02	*	0.00	0.00	0.11	*	0.00	0.00	0.00	0.02	0.06	0.00	0.00	0.04	0.02
TiO2	2.69	3.07	*	1.16	1.89	1.29	*	3.81	3.90	3.48	2.03	3.76	3.68	3.68	3.42	3.09
OH	3.99	4.01	*	3.95	4.01	3.98	*	3.96	3.96	3.96	3.98	3.99	4.02	4.00	3.96	3.96
TTL	99.69	100.42	*	100.19	100.20	98.98	*	100.82	100.41	100.33	100.40	100.17	101.22	100.37	100.74	100.62
Cations per 22 oxygens																
Fe	2.15	2.15	*	2.63	2.29	2.28	*	2.64	2.59	2.58	2.64	2.49	2.46	2.35	2.71	2.68
Mn	0.02	0.01	*	0.01	0.02	0.03	*	0.00	0.01	0.00	0.00	0.00	0.01	0.00	0.01	0.01
Na	0.05	0.07	*	0.08	0.10	0.04	*	0.09	0.09	0.07	0.09	0.09	0.11	0.11	0.09	0.09
Al	3.42	3.43	*	3.55	3.58	3.49	*	3.36	3.34	3.37	3.58	3.36	3.36	3.36	3.33	3.48
Alvi			*	0.91	1.02	1.06	*	0.68	0.71	0.75	0.89	0.75	0.77	0.76	0.68	0.68
Si	5.41	5.37	*	5.36	5.44	5.58	*	5.33	5.37	5.38	5.32	5.39	5.41	5.41	5.36	5.28
Mg	2.29	2.26	*	2.07	2.05	2.08	*	1.82	1.78	1.85	2.00	1.96	1.92	2.00	1.84	1.85
K	1.93	1.94	*	1.87	1.85	1.80	*	1.85	1.85	1.83	1.69	1.62	1.79	1.83	1.80	1.80
Ca	0.00	0.00	*	0.00	0.00	0.02	*	0.00	0.00	0.00	0.00	0.01	0.00	0.00	0.01	0.00
Ti	0.30	0.34	*	0.13	0.21	0.14	*	0.43	0.44	0.39	0.23	0.42	0.41	0.41	0.39	0.35
OH	2.00	2.00	*	2.00	2.00	2.00	*	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00	2.00
TTL	17.56	17.57	*	17.70	17.52	17.44	*	17.53	17.48	17.48	17.55	17.36	17.44	17.47	17.53	17.57

SPL*4*	B2l	B3r	B4r	B5r	B7m	B8m
FeO	18.77	22.89	21.80	21.37	21.25	20.59
MnO	0.06	0.08	0.08	0.05	0.09	0.08
Na2O	0.38	0.29	0.28	0.32	0.26	0.36
Al2O3	19.46	18.63	19.05	19.11	19.56	19.76
SiO2	35.61	34.04	34.69	36.11	35.81	35.46
MgO	9.44	7.89	7.81	8.12	7.98	8.15
K2O	9.38	8.33	9.27	9.41	9.59	9.25
CaO	0.04	0.03	0.02	0.02	0.00	0.00
TiO2	4.52	3.24	2.88	2.79	3.36	3.39
OH	4.06	3.88	3.91	3.99	4.01	3.99
TTL	101.92	99.31	99.78	101.29	101.92	101.05
Cations per 22 oxygens						
Fe	2.32	2.96	2.79	2.68	2.65	2.58
Mn	0.01	0.01	0.01	0.01	0.01	0.01
Na	0.11	0.09	0.08	0.09	0.08	0.10
Al	3.38	3.39	3.44	3.38	3.44	3.49
Alvi						
Si	5.28	5.25	5.32	5.42	5.34	5.32
Mg	2.07	1.82	1.78	1.82	1.78	1.82
K	1.77	1.64	1.81	1.80	1.83	1.77
Ca	0.01	0.00	0.00	0.00	0.00	0.00
Ti	0.50	0.38	0.33	0.31	0.38	0.38
OH	2.00	2.00	2.00	2.00	2.00	2.00
TTL	17.45	17.54	17.58	17.52	17.51	17.49

PLAGIOCLASE

SPLE*OF1*	Alam	A2am	Al5am	Al6am	A5br	A6br	Al2br	Al3br	*5A*	Al4r	Al5cr	B6r	B8cr	B1m	B2m	B3m
FeO	0.37	0.11	0.01	0.03	0.05	0.08	0.26	0.18	*	0.29	0.14	0.45	0.22	0.04	0.01	0.03
NaO	0.22	7.85	7.59	7.88	8.15	8.11	8.27	8.21	*	6.61	6.92	6.56	6.79	6.17	6.42	0.07
Al2O3	24.02	23.85	22.39	23.59	24.28	24.44	24.64	24.21	*	26.20	25.76	26.68	26.08	26.69	26.58	26.12
SiO2	62.16	61.91	59.96	60.75	61.25	61.31	61.08	61.18	*	58.01	58.32	57.80	57.45	58.16	58.69	58.69
K2O	0.22	0.41	0.38	0.33	0.28	0.39	0.24	0.27	*	0.19	0.24	0.27	0.28	0.35	0.30	0.36
CaO	5.64	5.56	5.39	5.87	5.62	5.86	6.09	5.94	*	8.63	8.20	8.72	8.23	9.04	8.80	8.36
TTL	100.71	99.68	95.76	98.51	99.65	100.23	100.63	99.98	*	99.98	99.62	100.15	99.52	99.83	100.27	100.31
Cations per 8 oxygens																
Fe	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.01	*	0.01	0.01	0.02	0.01	0.00	0.00	0.00
Na	0.71	0.68	0.68	0.68	0.71	0.69	0.71	0.71	*	0.57	0.60	0.57	0.59	0.53	0.56	0.58
Al	1.24	1.25	1.22	1.25	1.27	1.28	1.28	1.27	*	1.38	1.36	1.41	1.38	1.41	1.39	1.37
Si	2.74	2.75	2.77	2.74	2.73	2.72	2.69	2.72	*	2.60	2.62	2.57	2.61	2.58	2.59	2.61
K	0.01	0.02	0.02	0.02	0.02	0.02	0.01	0.02	*	0.01	0.01	0.02	0.02	0.02	0.02	0.02
Ca	0.27	0.27	0.27	0.28	0.26	0.28	0.29	0.28	*	0.41	0.40	0.42	0.40	0.43	0.42	0.40
TTL	4.99	4.97	4.97	4.98	4.99	4.99	5.01	5.01	*	4.99	5.00	5.01	5.00	4.99	4.99	4.99
Xalb	71.57	70.15	70.15	69.46	71.24	69.84	70.14	70.37	*	57.44	59.06	56.79	58.95	54.13	55.92	57.91
Xor	1.28	2.35	2.29	1.94	1.59	2.21	1.32	1.51	*	1.11	1.35	1.54	1.60	2.02	1.74	2.07
Xan	27.15	27.51	27.53	28.61	27.17	27.93	28.55	28.12	*	41.46	39.04	41.68	39.46	43.84	42.34	40.02
SPLE*5A*	D8m	E7m	*5B*	Al5m	Al7r	Al8m	C4ar	C5ar	C26m	D6r	*7*	A9m	Bl*	B2m	B15r	*
FeO	0.05	0.06	*	0.04	0.04	0.00	0.42	0.37	0.04	0.05	*	0.05	0.07	0.00	0.01	*
Na2O	7.01	6.36	*	6.47	6.80	6.70	6.31	6.80	6.76	6.61	*	8.10	8.12	8.30	8.21	*
Al2O3	25.73	26.54	*	25.93	25.68	25.62	26.68	26.17	25.88	26.06	*	23.79	23.46	23.20	23.51	*
SiO2	59.26	57.52	*	58.48	59.62	58.57	57.14	58.25	58.66	58.54	*	61.53	60.98	61.78	60.74	*
K2O	0.28	0.41	*	0.34	0.27	0.39	0.24	0.26	0.31	0.30	*	0.32	0.34	0.48	0.32	*
CaO	7.79	9.09	*	8.08	7.75	7.78	8.89	8.28	8.05	8.30	*	5.95	5.77	5.16	5.75	*
TTL	100.13	99.98	*	99.34	100.18	99.13	99.77	100.12	99.72	99.90	*	99.76	98.75	98.93	98.55	*
Cations per 8 oxygens																
Fe	0.00	0.00	*	0.00	0.00	0.00	0.01	0.01	0.00	0.00	*	0.00	0.00	0.00	0.00	*
Na	0.61	0.55	*	0.56	0.59	0.59	0.55	0.59	0.59	0.57	*	0.70	0.71	0.72	0.72	*
Al	1.35	1.40	*	1.37	1.35	1.36	2.57	2.61	1.37	1.37	*	1.25	1.24	1.23	1.25	*
Si	2.64	2.58	*	2.63	2.65	2.64	1.42	1.38	3.63	2.62	*	2.74	2.74	2.77	2.74	*
K	0.02	0.02	*	0.02	0.02	0.02	0.01	0.01	0.02	0.02	*	0.02	0.02	0.03	0.02	*
Ca	0.37	0.43	*	0.39	0.37	0.38	0.43	0.40	0.39	0.40	*	0.28	0.28	0.25	0.28	*
TTL	4.99	5.00	*	4.98	4.97	4.98	5.00	5.00	4.99	4.99	*	4.99	5.00	4.99	5.00	*
Xalb	60.99	54.61	*	57.99	60.42	59.52	55.47	59.00	59.24	58.00	*	69.84	70.42	72.39	70.78	*
Xor	1.61	2.30	*	2.02	1.56	2.29	1.37	1.33	1.80	1.75	*	1.80	1.93	2.76	1.81	*
Xan	37.41	43.09	*	39.99	38.02	38.19	43.16	39.67	38.96	40.35	*	28.36	27.65	24.85	27.41	*

SPL#47A*	A3m	A4r	A4c	A5ar	A6ar	B1r	B4r	B5m	B13acr	B14ar	*47B*	B14m	C2m	C4m	C5m	D14r	E4m
FeO	0.16	0.15	0.17	0.67	0.30	0.04	0.00	0.06	0.01	0.15	*	0.07	0.00	0.07	0.09	0.16	0.07
Na2O	8.10	7.74	7.65	7.75	7.86	7.63	7.62	7.57	7.62	7.78	*	7.47	7.56	7.60	7.60	7.91	7.35
Al2O3	24.61	24.87	24.39	24.07	24.13	24.87	24.87	24.48	23.28	23.88	*	24.55	24.87	24.74	24.81	24.63	24.67
SiO2	61.09	60.32	60.39	59.69	59.78	60.51	59.63	59.50	59.62	59.63	*	59.63	60.83	61.00	60.79	60.47	60.44
K2O	0.23	0.28	0.20	0.22	0.18	0.30	0.27	0.40	0.58	0.11	*	0.31	0.31	0.32	0.22	0.26	0.31
CaO	6.18	6.54	6.48	6.29	6.25	6.46	6.44	6.33	5.89	6.22	*	6.58	6.60	6.40	6.58	6.22	6.70
TTL	100.52	99.94	99.32	98.71	98.54	99.85	98.86	98.36	97.06	97.79	*	98.62	100.31	100.14	100.11	99.67	99.59
Cations per 8 oxygens																	
Fe	0.01	0.01	0.01	0.03	0.01	0.00	0.00	0.00	0.00	0.01	*	0.00	0.00	0.00	0.00	0.10	0.00
Na	0.70	0.67	0.66	0.68	0.69	0.66	0.66	0.66	0.68	0.69	*	0.65	0.65	0.65	0.65	0.68	0.64
Al	1.29	1.31	1.29	1.28	1.29	1.31	1.32	1.31	1.26	1.28	*	1.31	1.30	1.29	1.30	1.30	1.30
Si	2.70	2.69	2.70	2.70	2.70	2.69	2.68	2.69	2.73	2.71	*	2.69	2.70	2.71	2.70	2.70	2.70
K	0.01	0.02	0.01	0.01	0.01	0.02	0.02	0.02	0.03	0.01	*	0.02	0.02	0.02	0.00	0.01	0.02
Ca	0.29	0.31	0.31	0.30	0.30	0.31	0.31	0.31	0.29	0.30	*	0.32	0.31	0.30	0.31	0.30	0.32
TTL	5.00	5.00	4.99	5.00	5.00	4.99	5.00	5.00	5.00	4.99	*	4.99	4.98	4.98	4.98	5.00	4.98
Xalb	69.45	67.10	67.35	68.18	68.77	66.95	67.09	66.78	67.70	68.93	*	66.03	66.26	66.98	66.76	68.69	65.30
Xor	1.29	1.59	1.14	1.25	1.03	1.73	1.56	2.34	3.38	0.62	*	1.80	1.79	1.84	1.27	1.46	1.80
Xan	29.30	31.31	31.51	30.56	30.20	31.32	31.36	30.88	28.92	30.44	*	32.17	31.95	31.18	31.97	29.85	32.90

ILMENITE

SPLE*QF1*	A1	A2	A3	A4	*45B*	B11	B21	B31	B7F	*04*	A4	A6
FeO	33.58	33.12	31.98	26.49	*	46.01	46.22	46.66	46.33	*	43.96	45.01
TiO2	51.55	51.60	47.39	39.12	*	51.81	51.41	51.67	50.95	*	50.67	50.54
MnO	13.09	12.10	10.12	7.99	*	0.42	0.45	0.13	0.43	*	1.96	1.48
SiO2	0.07	0.06	3.83	10.65	*	0.01	0.00	0.03	0.07	*	0.45	0.39
Al2O3	0.05	0.07	2.56	4.88	*	0.00	0.00	0.00	0.00	*	0.39	0.24
WGO	0.01	0.06	1.14	2.26	*	0.00	0.00	0.00	0.00	*	0.08	0.04
TTL	97.35	97.01	97.03	95.35	*	98.26	98.08	98.50	97.77	*	97.52	98.69
Cations per 6 oxygens												
Fe	1.45	1.43	1.33	1.05	*	1.98	1.99	2.00	2.00	*	1.88	1.91
Ti	2.00	2.00	1.77	1.40	*	2.00	1.99	1.99	1.98	*	1.96	1.98
Mn	0.53	0.53	0.43	0.32	*	0.02	0.02	0.00	0.02	*	0.09	0.06
Si	0.00	0.00	0.19	0.51	*	0.00	0.00	0.00	0.00	*	0.02	0.02
Al	0.00	0.00	0.15	0.27	*	0.00	0.00	0.00	0.00	*	0.02	0.01
Mg	0.00	0.00	0.08	0.16	*	0.00	0.00	0.00	0.00	*	0.00	0.00
TTL	3.99	3.98	3.96	3.87	*	4.00	4.00	4.00	4.00	*	3.99	3.99

ROUTILE

SPLE*45B*	B41	B5r	B6r	B8r	A1r	A21	A3m	A4r	A6m	*04*	A1	A2	A3
FeO	0.44	0.09	0.59	0.13	0.22	0.30	0.23	0.21	0.27	*	0.59	1.64	1.39
TiO2	97.24	97.77	97.34	97.62	98.23	96.30	95.91	97.66	98.14	*	98.18	96.78	97.36
SiO2	0.09	0.05	0.09	0.13	0.15	0.25	0.22	0.09	0.04	*	0.00	0.00	0.00
TTL	97.77	97.92	98.09	97.88	98.78	97.82	97.24	98.22	98.63	*	98.91	98.61	98.93
Fe	0.01	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.01	*	0.01	0.02	0.02
Ti	0.99	1.00	0.99	1.00	1.00	0.99	0.99	1.00	0.99	*	0.99	0.98	0.99
Si	0.00	0.00	0.00	0.00	0.00	0.01	0.01	0.00	0.00	*	0.00	0.00	0.00
TTL	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	*	1.00	1.00	1.01

K-FELDSPAR

SPLE*OF1*	Alam	Al7am	Al8am	*5A*	El2m	*47A*	A7cr	A8r	B2r	B3m	*47B*	B12cr	B13r	D13r
Na2O	1.16	0.98	1.06	*	0.95	*	1.24	1.39	1.78	1.88	*	1.94	1.59	2.64
Al2O3	18.53	17.84	17.88	*	18.45	*	18.62	18.31	18.71	18.79	*	18.31	18.36	18.86
SiO2	63.86	63.42	63.25	*	64.45	*	62.52	62.81	64.21	63.34	*	62.82	62.54	65.03
K2O	15.49	15.63	15.72	*	15.79	*	15.55	15.11	14.75	14.56	*	14.23	14.70	13.50
CaO	0.04	0.05	0.06	*	0.00	*	0.11	0.11	0.10	0.09	*	0.10	0.07	0.06
TTL	99.14	97.95	97.99	*	99.75	*	98.11	97.93	99.62	98.63	*	97.47	97.28	100.25
Cations per 8 oxygens														
Na	0.10	0.09	0.10	*	0.08	*	0.11	0.13	0.19	0.17	*	0.18	0.15	0.23
Al	1.01	0.99	0.99	*	1.00	*	1.04	1.02	1.02	1.04	*	1.02	1.03	1.02
Si	2.98	2.99	2.99	*	2.98	*	2.95	2.97	2.97	2.96	*	2.97	2.97	2.98
K	0.92	0.94	0.95	*	0.93	*	0.94	0.91	0.87	0.87	*	0.86	0.89	0.79
Ca	0.00	0.00	0.00	*	0.00	*	0.01	0.01	0.00	0.00	*	0.01	0.00	0.00
TTL	5.02	5.02	5.03	*	5.02	*	5.05	5.04	5.03	5.04	*	5.04	5.04	5.03
Xalb	10.18	8.70	9.31	*	8.42	*	10.76	12.17	15.47	16.31	*	17.09	14.11	22.85
Xor	89.63	91.07	90.40	*	91.58	*	88.72	87.32	84.08	83.26	*	82.43	85.56	76.87
Xan	0.19	0.23	0.29	*	0.00	*	0.52	0.52	0.46	0.42	*	0.48	0.34	0.28

MUSCOVITE

SPLE*OF1*	A3am	A5al	A6al	A7al	Al10al	A9br	A4cm	*45B*	Alm	A21	A6m	B71	B9mr
FeO	1.47	1.66	1.58	1.62	1.65	1.47	1.57	*	1.23	1.37	1.15	1.37	1.91
Na2O	0.35	0.51	0.56	0.43	0.57	0.35	0.55	*	0.77	0.85	0.88	0.78	0.76
Al2O3	35.38	33.92	33.76	33.27	34.36	34.09	34.51	*	35.39	36.35	35.09	36.23	34.46
SiO2	46.81	46.34	46.18	46.60	46.06	46.21	46.23	*	46.93	46.43	46.29	46.52	47.22
MgO	0.81	0.91	0.84	0.96	0.85	0.79	0.77	*	0.72	0.46	0.74	0.48	0.80
K2O	10.91	11.24	11.08	11.25	11.10	10.91	11.10	*	10.34	10.22	10.05	10.30	10.20
TiO2	0.74	0.56	0.59	0.56	0.24	0.63	0.84	*	0.38	0.48	0.57	0.37	0.75
OH	4.45	4.46	4.44	4.45	4.45	4.45	4.49	*	4.53	4.55	4.49	4.54	4.54
TTL	101.01	99.65	99.02	99.26	99.32	98.97	100.07	*	100.32	100.80	99.27	100.60	100.67
Cations per 24 oxygens													
Fe	0.16	0.18	0.18	0.18	0.18	0.17	0.17	*	0.14	0.15	0.13	0.15	0.21
Na	0.09	0.01	0.14	0.11	0.15	0.09	0.14	*	0.20	0.22	0.23	0.19	0.19
Al	5.49	5.36	5.36	5.28	5.44	5.41	5.42	*	5.51	5.64	5.52	5.63	5.37
Alvi	3.68	3.58	3.59	3.55	3.64	3.63	3.60	*	3.71	3.75	3.69	3.76	3.76
Si	6.16	6.21	6.23	6.27	6.19	6.22	6.17	*	6.20	6.11	6.18	6.13	6.24
Mg	0.16	0.18	0.17	0.19	0.17	0.16	0.15	*	0.14	0.09	0.15	0.10	0.16
K	1.63	1.92	1.90	1.93	1.90	1.87	1.89	*	1.74	1.71	1.71	1.73	1.72
Ti	0.07	0.06	0.05	0.06	0.03	0.06	0.08	*	0.04	0.05	0.06	0.04	0.07
OH	2.00	2.00	2.00	2.00	2.00	2.00	2.00	*	2.00	2.00	2.00	2.00	2.00
TTL	15.9-	16.07	16.05	16.05	16.08	15.99	16.04	*	15.97	15.98	15.97	15.97	15.96

STAURITIDE

SPL#45B*	A1	A2	B1	B2	*04*	A2	A3
FeO	13.17	12.74	11.66	12.51	*	13.28	13.78
MnO	0.11	0.15	0.12	0.15	*	0.09	0.11
Al2O3	53.98	53.69	53.44	53.11	*	54.37	54.16
SiO2	27.99	28.20	28.16	28.33	*	26.95	27.11
MgO	1.44	1.53	1.45	1.52	*	1.76	1.75
TiO2	0.52	0.38	0.48	0.37	*	1.01	1.08
OH	2.16	2.15	2.13	2.14	*	2.15	2.16
TTL	99.39	98.83	97.47	98.15	*	99.72	100.18
Fe	1.53	1.49	1.37	1.47	*	1.54	1.60
Mn	0.01	0.02	0.01	0.02	*	0.01	0.01
Al	8.84	8.82	8.86	8.78	*	8.90	8.85
Si	3.89	3.93	3.96	3.97	*	3.74	3.76
Mg	0.30	0.32	0.30	0.32	*	0.36	0.36
Ti	0.05	0.04	0.05	0.04	*	0.10	0.11
OH	1.00	1.00	1.00	1.00	*	1.00	1.00
TTL	15.63	15.62	15.56	15.60	*	15.70	15.70

SILLIMANITE

SPL#5B*	Al6r	D9r	*04*	A1	A3	B1
FeO	0.34	0.20	*	0.24	0.23	0.24
Al2O3	59.82	61.86	*	62.13	61.49	61.92
SiO2	36.74	37.75	*	37.34	37.55	37.27
K2O	0.05	0.01	*	0.03	0.00	0.03
TTL	97.40	99.86	*	99.81	99.30	99.56
Cations per 20 oxygens						
Fe	0.01	0.01	*	0.02	0.02	0.01
Al	1.96	1.97	*	1.92	1.87	1.98
Si	1.02	1.02	*	1.04	1.08	1.01
K	0.01	0.00	*	0.00	0.00	0.00
TTL	3.00	2.99	*	3.00	2.98	3.00

Appendix 3.**Geochronology sample descriptions and locations**

Undeformed granites of the Upper Crustal layer

MBX32: This hornblende-biotite granodiorite was collected along a logging road cut climbing into an erosional remnant underlying Eocene trachyandesites near the headwaters of McCulloch Creek north of Westbank. The Jurassic age agrees with several other isotopic dates determined on several other granitic bodies lying westward of Okanagan Lake (Peto and Armstrong 1976). The intrusive age is also considered coeval with that of MBX39 and suggest similar conclusions with respect to the thermal resetting bounding the Okanagan valley fault.

(GSC no. 3898)

Lat./Long.: 49°56'35"/119°35'55"

K-Ar age: 157.2 ± 2.4 Ma (Hornblende)

Ar-Ar age: 159.9 ± 0.7 Ma (Biotite, Integrated)

160.4 ± 0.6 Ma (Biotite, Plateau)

MBX39: Undeformed biotite granodiorite lying east of mylonitic gneisses east of Arlington lakes north of Carmi. Biotite rarely exceeds 10 % of the modal composition and occurs as small aggregates of individual crystals with ragged rims. The textures of the granitic assemblage indicates that it was recrystallized. The age is interpreted as the crystallization age and suggests that contrarily to surrounding gneisses this granitoid body was not reheated above 300°C since its time of emplacement. The crystallization age nearly coincides with that of several other plutonic bodies lying west of the Okanagan valley fault (see Peto and Armstrong 1976).

(GSC no. 3894)

Lat./Long.: 49°37'52"/119°03'30"

K-Ar age: 158.9 ± 2.4 Ma (Biotite)

Strained gneisses of the Middle Crustal layer

MXOF01: Garnet-sillimanite-biotite schist in the high strain zone of the Okanagan valley fault preserving unequivocal kinematic indicators suggesting a westerly sense of shear.

Lat./Long.: 49°15'18"/119°30'10"

Ar-Ar age: 50.1 ± 1.0 Ma (Muscovite, total fusion)

47.0 ± 11.5 Ma (Muscovite, total fusion)

MBX5: (GSC no. 3902, 3903)	Amphibolite gneiss interleaved with well banded orthogneiss at the Siemalt Speedway midway between Hydraulic Lake and Highway 33. Overlain unconformably by Eocene dacitic lava flows and conglomerate. The rock is coarse grained containing euhedral hornblende with pyroxene cores in a feldspar matrix. Medford obtained a K-Ar hornblende age of 54.4 ± 2 Ma on an altered paragneiss located approximately 1 km north of this locality. The sampling site lies in the high strain Zone A (Bardoux chapter 2) of the Okanagan valley fault and demonstrates that the Eocene thermal event is laterally extensive across the Okanagan complex.
Lat./Long.:	49°46'08"/119°07'07"
K-Ar age:	51.3 $\pm$ 1.1 Ma (Hornblende) 49.3 $\pm$ 0.9 Ma (Biotite)
MX19E:	Muscovite-biotite-bearing fine grained mylonitic leucogranite showing diagnostic kinematic fabrics suggesting top-to-the-west sense of shear. Sample was classified as typical Ladybird-type (<i>see</i> Carr 1990). The interpreted age of the rock is 56 Ma.
Lat./Long.:	50°27'40"/118°46'30"
U-Pb age:	53.9 $\pm$ 0.2 Ma (Zircon) 54.3 $\pm$ 0.6 Ma (Zircon) 60.6 $\pm$ 0.2 Ma (Zircon) 53.3 $\pm$ 0.2 Ma (Monazite) 53.0 $\pm$ 0.4 Ma (Monazite)
Ar-Ar age:	48.9 $\pm$ 0.9 Ma (Muscovite, Integrated) 48.7 $\pm$ 0.6 Ma (Muscovite, Plateau) 49.2 $\pm$ 0.6 Ma (Muscovite, Integrated) 49.2 $\pm$ 0.4 Ma (Muscovite, Plateau)
MX42E:	Zone A mylonitized muscovite-garnet-biotite pegmatite underlying the Trinity Hills klippe.
Lat./Long.:	50°30'10"/118°45'15"
Ar-Ar age:	51.7 $\pm$ 0.5 Ma (Muscovite, total-fusion)
MX48: (GSC. no. 3897, 3900):	Coarse grained amphibolitic body interleaved with paragneissic sections correlated with the Vaseaux Lake Formation (Bostock 1941; Christie 1973) of early Proterozoic age (Armstrong et al. 1991). The sample lies in structural Zone B (Bardoux chapter 2) and contains 60% subhedral chloritized hornblende and 20% of both biotite and feldspar. The Ar-Ar data is considered more reliable than the K-Ar as it is much less contaminated by atmospheric Ar.
Lat./Long.:	49°48'15"/119°14'50"
K-Ar age:	47.7 $\pm$ 0.8 Ma (Biotite) 44.1 $\pm$ 1.2 Ma (Hornblende)
Ar-Ar age:	48.9 $\pm$ 0.5 Ma (Biotite, Integrated) 48.7 $\pm$ 0.4 Ma (Biotite, Plateau) 51.5 $\pm$ 0.9 Ma (Hornblende, Integrated) 50.4 $\pm$ 0.9 Ma (Hornblende, Plateau)

MX49.1:	Amphibolitic lenses interleaved with metasediments along the abandoned CP railway at the headwaters of Pooley Creek. Similar structural position as MX48.
Lat./Long.:	49°45'30"/119°19'30"
Ar-Ar age:	48.7 ± 4.4 Ma (Hornblende, Integrated) 48.5 ± 2.4 Ma (Hornblende, Plateau) 61.9 ± 3.2 Ma (Hornblende, total-fusion) 48.5 ± 0.6 Ma (Biotite, total-fusion)
MXE100:	Zone B foliated muscovite-bearing pegmatite cutting sillimanite-K-feldspar metapelites underlying Eocene volcanics of Camels Hump near Reiswig.
Lat./Long.:	50°15'03"/118°45'15"
Ar-Ar age:	52.7 ± 0.3 Ma (Muscovite, total-fusion)
MBX229:	Zone A mylonitized paragneiss containing medium grained hornblende-rich bands (GSC. no. 3899) interleaved with calcsilicate lenses along Lakeshore Road, 2 km north of the entrance gate to Okanagan Mountain Provincial Park. The sampled section is correlated with the Vaseaux Formation (Bostock 1941). The rock is intensely chloritized as it projects immediately beneath the cataclastic zone of the Okanagan valley fault. The isotopic hornblende date agrees with the general characteristic that the amphibolite grade rocks underlying the Okanagan valley fault have been affected by a heating event in the early middle Eocene. However, the large quantity of atmospheric Ar in this sample may constitute a considerable source of error. Note that the isotopic dates determined in the gneiss is nearly coeval with that of Eocene volcanic sample MBX2 which lies nearby this site across Okanagan Lake.
Lat./Long.:	49°46'35"/119°34'15"
K-Ar age:	51.4 ± 0.9 Ma (Hornblende)
Ar-Ar age:	56.4 ± 5.1 Ma (Hornblende, Integrated) 47.3 ± 3.7 Ma (Hornblende, Plateau) 55.2 ± 5.7 Ma (Hornblende, Integrated) 51.6 ± 4.6 Ma (Hornblende, Plateau)
MX261:	Zone B strained granoblastic biotite microgranite interleaved with paragneisses along the abandoned CP railway 5 km west of Hydraulic Lake.
Lat./Long.:	49°45'30"/119°22'00"
U-Pb age:	50.0 ± 1.8 Ma (Thorite) 49.8 ± 2.0 Ma (Thorite)

MX300: Hornblende-bearing diorite collected along the shoreline of Vaseaux Lake near Okanagan Falls. A concordant zircon U-Pb isotopic age of 95 Ma was obtained from a nearby outcrop of diorite (see Parkinson 1985).

Lat./Long.: 49°25'10"/119°34'30"

Ar-Ar age: 49.1 ± 2.3 Ma (Hornblende, Integrated)
48.7 ± 1.0 Ma (Hornblende, Plateau)
56.1 ± 2.1 Ma (Hornblende, total fusion)

Cataclastic gneisses

MX03: Brecciated mylonitic gneiss from the Hydraulic Lake area. Represents the highest structural level of the Okanagan valley fault.

Lat./Long.: 49°47'00"/119°13'10"

Ar-Ar age: 50.6 ± 0.8 Ma (Muscovite, Integrated)
50.5 ± 0.6 Ma (Muscovite, Plateau)
52.1 ± 4.1 Ma (Muscovite, total fusion)
52.2 ± 2.1 Ma (Muscovite, total fusion)

MX10:

Lat./Long.: 49°47'30"/119°05'30"

Ar-Ar age: 48.5 ± 3.1 Ma (Whole rock, Integrated)
48.5 ± 1.4 Ma (Whole rock, Plateau)

MX21: Zone A, retrograded mylonitic hornblende gneiss along the road to the Big White ski resort. The rock analysed underlies a dacitic Eocene outlier of which a feeding dike (MBX4) was dated 48.6 ± 0.7 Ma on biotite. Comparing the ages of either units it appears that the cataclastic zone of the Okanagan valley fault has not been reset to temperatures in excess of 350°C since the dacitic lavas were extruded.

Lat./Long.: 49°47'30"/119°04'10"

Ar-Ar age: 50.2 ± 0.4 Ma (Muscovite, total-fusion)
50.1 ± 0.2 Ma (Muscovite, total-fusion)

MX78: Brecciated mylonite collected south of Black Knight Mountain at approximately 100 m below the top surface of the Okanagan valley fault.

Lat./Long.: 49°52'30"/119°17'30"

Ar-Ar age: 59.1 ± 2.8 Ma (Whole rock, Integrated)
59.7 ± 1.0 Ma (Whole rock, Plateau)

MBX133: (GSC no. 3901)	Intensely chloritized mylonitic hornblende gneiss located at high structural levels in the Kelowna quarry located near the intersection of Kloo Creek and Mission Creek. The rock is medium to fine grained and contains unannealed extensional strain fabrics and thus lies at the bottom of the cataclastic Zone A ₁ . Hornblende porphyroclasts are partly retrograded to chlorite and pulled apart. The sampled unit is considered as a strained equivalent to the nearby granitoid resembling sample MBX39. The Eocene date agrees with the regional thermal cooling event.
Lat./Long.:	49°49'35"/119°21'52"
K-Ar age:	55.2 ± 1.3 Ma (Hornblende)

Volcanics

MBX1: (GSC no. 3895)	Quartz-orthoclase-biotite rhomb porphyry at the base of the Eocene Westbank volcanic outlier. It contains up to 50 % subhedral phenocrysts in an aphanitic felsic matrix. The unit has a reversed paleomagnetic polarity (Bardoux and Irving 1989). Similar hypabyssal rocks occur further south near Shingle Creek and Beavercreek (Little 1961). The age is interpreted as a crystallization age. It is slightly older than another analysis obtained by Church (1981) of 47.7 ± 1.4 Ma on altered biotite extracted from a layered rhyolitic dome lying above the porphyry, 500 meters eastward of the present site.
Lat./Long.:	49°48'15"/119°44'35"
K-Ar age:	50.5 ± 1.0 Ma (Biotite)
MBX2: (GSC. no. 3896)	Porphyritic trachyandesite cutting through Mesozoic hornblende granitoid sample MBX32. unit is correlated with the Kitley Lake Member of the Marron Formation (Church 1973). The age determined is slightly older than three other dates obtained by Church (1981) from biotite fractions in similar rocks (53 ± 2 Ma) in both the Westbank outlier and the White Lake Basin near Penticton. The data is interpreted as a crystallization age.
Lat./Long.:	49°54'30"/119°36'56"
K-Ar age:	53.6 ± 2.2 Ma (Biotite)
MBX4: (GSC no. 3893)	This dacitic dyke cuts through deformed coarse grained granodiorite such as MBX39 along the abandoned CP railway near the intersection of highway 33 and Devil Creek. The dyke contains biotite and hornblende phenocrysts in an aphanitic matrix and is petrographically similar to rhyodacites underlying the Joe Rich Klippe (Bardoux chapter 6). Biotites in the latter were dated 46 ± 2 Ma (Mathews, uncorrected unpublished data, 1964). Bardoux and Irving (1989) correlated the unit with the Maramba Formation of Church (1973) and found an inverse paleomagnetic polarity. The undeformed texture of the dyke suggests that regional ductile deformation associated with the Okanagan valley fault had ceased by 49 Ma ago.
Lat./Long.:	49°42'35"/119°06'50"
K-Ar age:	48.6 ± 0.7 Ma (Biotite)

Appendix 4. Stress induced by topography and density heterogeneities in the lithosphere (written by J.-C. Mareschal)

The stress induced by surface and internal loading in a layered elastic lithosphere is determined by the equilibrium condition:

$$\nabla \cdot \mathbf{T} = 0 \quad (\text{A.1})$$

where $\mathbf{T}$ is the stress tensor. It is related to the displacement vector $\mathbf{u}$ by the stress-strain relationship:

$$\mathbf{T} = \lambda \nabla \cdot \mathbf{u} \mathbf{I} + \mu (\nabla \mathbf{u} + \nabla \mathbf{u}^T) \quad (\text{A.2})$$

where $\mathbf{I}$ is the identity tensor and λ and μ are the Lamé parameters. Replacing the stress by the displacement in the equilibrium condition yields:

$$\mu \nabla^2 \mathbf{u} + (\lambda + \mu) \nabla \nabla \cdot \mathbf{u} = 0 \quad (\text{A.3})$$

This equation implies that each component of the displacement vector must satisfy the biharmonic equation:

$$\nabla^4 \mathbf{u} = 0 \quad (\text{A.4})$$

The boundary conditions specify the loading at the surface and on each interface. The stress and displacement in any layer k will be denoted by $\mathbf{T}^k$ and $\mathbf{u}^k$. At the surface, $z = 0$, the normal stress is equal to the surface load:

$$T_{zz}^1 = g \rho_1 (H^0(x) - u_z^1(x)) \quad (\text{A.5})$$

where ρ_1 is the density of the upper crust and $H^0(x)$ is the topography. The shear stress vanishes:

$$T_{xz}^1 = 0 \quad (\text{A.6})$$

Across any interface between two layers ($z = z_k$), the shear stress is continuous, and the load (topography of the interface) is equal to the jump in normal stress across the interface:

$$T_{xz}^k = T_{xz}^{k+1} \quad (\text{A.7})$$

$$T_{zz}^k - T_{zz}^{k+1} = g (\rho_{k+1} - \rho_k) (H^k(x) - u_z^k(x)) \quad (\text{A.8})$$

Also, the elastic displacements must be continuous across each interface:

$$u_{x,z}^k = u_{x,z}^{k+1} \quad (\text{A.9-10})$$

These equation and boundary conditions are more easily solved in Fourier transform domain. The general solution of the biharmonic equation (A.4) is:

$$u_{x,z} = (A_{x,z} + B_{x,z} kz) \exp(kz) + (C_{x,z} + D_{x,z} kz) \exp(-kz) \quad (\text{A.11})$$

where k is the wavenumber. The indices referring to layer are dropped whenever they are not necessary. The constants $A_{x,z}$, $B_{x,z}$, $C_{x,z}$, and $D_{x,z}$ are not independent because the components of the displacement are related by the lower order equation (A.3). Introducing the solution (A.11) into the equation (A.3) and eliminating yields:

$$u_x = (A + Bkz) \exp(kz) + (C + Dkz) \exp(-kz) \quad (A.12)$$

$$-iu_z = \left\{ A - \frac{\lambda + 3\mu}{\lambda + \mu} B + Bkz \right\} \exp(kz) - \left\{ C + \frac{\lambda + 3\mu}{\lambda + \mu} D + Dkz \right\} \exp(-kz) \quad (A.13)$$

The components of the stress tensor are given by:

$$\frac{T_{xx}}{2ik\mu} = - \left\{ A + \frac{\lambda}{\lambda + \mu} B + Bkz \right\} \exp(kz) - \left\{ C - \frac{\lambda}{\lambda + \mu} D + Dkz \right\} \exp(-kz) \quad (A.14)$$

$$\frac{T_{zz}}{2ik\mu} = \left\{ A - \frac{\lambda + 2\mu}{\lambda + \mu} B + Bkz \right\} \exp(kz) + \left\{ C + \frac{\lambda + 2\mu}{\lambda + \mu} D + Dkz \right\} \exp(-kz) \quad (A.152)$$

$$\frac{T_{xz}}{2k\mu} = \left\{ A - \frac{\mu}{\lambda + \mu} B + Bkz \right\} \exp(kz) - \left\{ C + \frac{\mu}{\lambda + \mu} D + Dkz \right\} \exp(-kz) \quad (A.16)$$

There are 4 constants A , B , C , and D for each layer. They are determined by the boundary conditions. In practice, the constants and the stress are determined in Fourier domain; the transform of stress is inverted by a fast Fourier transform algorithm.

Appendix 5. Heat flow following extension and delamination (written by J.-C. Mareschal)

The temperature perturbation following extension of the upper part of the lithosphere and delamination of the lower part is obtained by solving the one dimensional heat equation with boundary and initial conditions:

$$\frac{1}{\kappa} \frac{\partial T}{\partial t} = \frac{\partial^2 T}{\partial z^2} \quad (B.1)$$

where κ is the thermal diffusivity of the lithosphere and heat sources are neglected.

At equilibrium and if heat sources are neglected, the temperature gradient is constant, and the temperature at the base of the lithosphere is T_L . The equilibrium temperature T_e and heat flow Q_e are thus:

$$T_e = T_L \frac{z}{L} \quad Q_e = K \frac{T_L}{L} \quad (B.2-3)$$

where K is the thermal conductivity.

Between the surface and the base of the crust ($z = M$ after extension), the lithosphere has been extended by a factor β , the gradient has thus been multiplied by β . Below the crust, the asthenosphere at temperature T_L has replaced the lithosphere. The initial temperature perturbation $\theta(t=0, z)$ is thus:

$$(\beta - 1)T_L \frac{z}{L} \quad \text{for } z < M \quad (B.4)$$

$$T_L \frac{z}{L} \quad \text{for } M < z < L \quad (B.5)$$

The temperature perturbation vanishes at the surface ($z=0$) and at the base of the lithosphere ($z=L$). At any time ($t>0$), temperature and heat flow are continuous at the crust mantle boundary.

Solving the heat equation with the above initial and boundary conditions is easily performed with Laplace transforms. The Laplace transform of the heat equation with its initial condition is:

$$\frac{s}{\kappa} \theta - \frac{1}{\kappa} \theta(t=0, z) = \frac{d^2 \theta}{dz^2} \quad (B.6)$$

The solution for the crust, $0 < z < M$, is:

$$\theta_1 = A \sinh\left(\sqrt{\frac{\kappa}{s}} z\right) + B \cosh\left(\sqrt{\frac{\kappa}{s}} z\right) - \frac{1}{s} (\beta - 1) T_L \frac{z}{L} \quad (B.7)$$

and for the mantle, $M < z < L$:

$$\theta_2 = C \sinh\left(\sqrt{\frac{\kappa}{s}} z\right) + D \cosh\left(\sqrt{\frac{\kappa}{s}} z\right) - \frac{1}{s} T_i \left(1 - \frac{z}{L}\right) \quad (B.8)$$

where A, B, C, and D are constants determined by the boundary conditions. Solving for the constants yields the solution in Laplace transform domain, for the crust:

$$\begin{aligned} \theta_1 = & \left\{ \frac{T_L}{s} (1 - \beta M/L) \cosh\left(\sqrt{\frac{s}{\kappa}} (M - L)\right) \dots \right. \\ & \left. + \sqrt{\frac{\kappa}{s^3}} \frac{\beta T_L}{L} \sinh\left(\sqrt{\frac{s}{\kappa}} (M - L)\right) \right\} \frac{\sinh\left(\sqrt{\frac{z}{\kappa}} z\right)}{\sinh\left(\sqrt{\frac{z}{\kappa}} L\right)} + \frac{1}{s} (\beta - 1) T_i \frac{z}{L} \end{aligned} \quad (B.9)$$

and, for the mantle:

$$\begin{aligned} \theta_2 = & \left\{ \frac{T_L}{s} (1 - \beta M/L) \cosh\left(\sqrt{\frac{s}{\kappa}} M\right) \dots \right. \\ & \left. + \sqrt{\frac{\kappa}{s^3}} \frac{\beta T_L}{L} \sinh\left(\sqrt{\frac{s}{\kappa}} M\right) \right\} \frac{\sinh\left(\sqrt{\frac{z}{\kappa}} (z - L)\right)}{\sinh\left(\sqrt{\frac{z}{\kappa}} L\right)} + \frac{1}{s} T_i \left(1 - \frac{z}{L}\right) \end{aligned} \quad (B.10)$$

The surface heat flow perturbation is directly obtained as:

$$\begin{aligned} \frac{Q(s)}{Q_e} = & \left\{ \frac{L}{\sqrt{\kappa s}} (1 - \beta M/L) \cosh\left(\sqrt{\frac{s}{\kappa}} (M - L)\right) \dots \right. \\ & \left. + \frac{\beta}{s} \sinh\left(\sqrt{\frac{s}{\kappa}} (M - L)\right) \right\} \frac{1}{\sinh\left(\sqrt{\frac{z}{\kappa}} L\right)} + \frac{1}{s} (\beta - 1) \end{aligned} \quad (B.11)$$

The inversion of the Laplace transform following standard techniques yields the total surface heat flow (Q_e + perturbation) as:

$$\frac{Q(t)}{Q_e} = 1 + 2 \sum_{n=1}^{\infty} \left\{ \frac{\beta}{n\pi} \sin\left(n\pi \frac{M}{L}\right) + \left(1 - \beta \frac{M}{L}\right) \cos\left(n\pi \frac{M}{L}\right) \right\} \exp\left(-n^2 \pi^2 \kappa \frac{t}{L^2}\right) \quad (B.12)$$

The standard result for extension without delamination (McKenzie, 1978) is obtained as a special case by observing that, in this situation $M = L / \beta$.

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