**Structure and tectonic setting of the
Okanagan Valley fault system
in the Shuswap Lake area,
southern British Columbia**

by
Bradford J. Johnson

A thesis submitted to
the Faculty of Graduate Studies and Research
in partial fulfilment of
the requirements for the degree of
Doctor of Philosophy

Department of Earth Sciences
Carleton University
Ottawa, Ontario
August 1994

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Frontispiece: Photograph looking southward across the Eagle River valley toward Mara Lake and the Okanagan Valley. The section in Fig. 2.4 crosses the southernmost of two prominent peninsulas on the west (right) side of Mara Lake. Both peninsulas and the Hunters Range (east of the lake) are underlain by footwall rocks of the Okanagan - Eagle River fault. The Enderby Cliffs (arrow) and Larch Hills (west of Mara Lake) are in the hanging wall. From Mara Lake, the fault trace (dotted) passes through the town of Sicamous and traverses the lower slopes of the Shuswap Range along the Eagle River valley, well below the trees in foreground. This perspective shows that the dip of the fault is about 20° or less.

ABSTRACT

One of the major low-angle normal fault systems in the southern Omineca belt that accommodated Early to Middle Eocene crustal extension, the Okanagan Valley fault system is characterized by west-dipping, en échelon, ductile-brittle detachment faults that delimit the western margin of the Shuswap metamorphic core complex. Migmatitic paragneisses in the uppermost few hundred metres to 2 km of the footwall of the system display upper amphibolite-facies mylonitic fabrics that are shown to consistently indicate relative westward movement of the hanging wall. These mylonitic rocks are interpreted as exhumed former parts of a mid-crustal shear zone, inferred from Lithoprobe reflection profiles, into which the detachment faults root.

Between the en échelon Okanagan - Eagle River and Adams - North Thompson fault segments is the newly discovered Shuswap Lake transfer zone, a 45-km-wide left step in the system within which there is no obvious brittle fault separating footwall from hanging wall. Within the transfer zone, two-mica leucogranites of the Pukeashun suite display mylonitic fabrics that consistently indicate west-northwesterly directed shear sense and are interpreted collectively as a component of the Okanagan Valley fault system. A U-Pb zircon date of 56 ± 1 Ma from a synextensional leucogranite dike places a new minimum age on the initiation of extensional shearing on the west flank of the Shuswap complex. The mylonitic foliation in the granite has been warped into a broad antiform that is interpreted as having resulted from isostatic rebound during tectonic denudation. The geometric relationships between folded mylonites and en échelon brittle detachments can be explained by kinematic models that invoke progressive warping and abandonment of the shear zone and excisement of the hanging wall during extension. These processes are inferred to have been driven by rapid exhumation of the granite relative to footwall rocks outside of the transfer zone, perhaps reflecting melt-enhanced extreme attenuation of the upper crust above the granite.

In the Eagle River area, footwall rocks of the Okanagan Valley fault system consist of migmatitic paragneiss and schist, amphibolite, calcsilicate gneiss and thin impure marbles, which are tentatively correlated with Neoproterozoic and older rocks of the ancestral North American miogeocline. These rocks were deformed and metamorphosed in upper amphibolite facies during Late Cretaceous to early Tertiary time as they were carried northeastward on the Monashee décollement. Mylonitic fabrics recording northeast-directed shear sense related to thrusting are developed within these rocks adjacent to the décollement, in contrast to the west-directed mylonitic fabrics related to extensional shearing on the Okanagan - Eagle River fault at higher structural levels.

In the hanging wall of the Okanagan - Eagle River fault are micaceous quartzite, pelitic and semipelitic schists, calcareous schist, marble, and phyllite. Most of these strata are tentatively correlated with lower Paleozoic units of the miogeocline, although some may belong to Permo-Triassic assemblages of accreted terranes. Three phases of deformation postdate emplacement of Devonian orthogneiss in the hangingwall rocks. Structures of the two latest phases are similar in style, orientation, and metamorphic relations to structures within miogeoclinal rocks that formed during Middle Jurassic to mid-Cretaceous deformation.

Dates obtained from twelve amphiboles by $^{40}\text{Ar}/^{39}\text{Ar}$ methods indicate that footwall rocks cooled below closure temperatures for ^{40}Ar in hornblende ($\sim 500^\circ\text{C}$) between Late Paleocene and Middle Eocene time. These dates and early Tertiary cooling rates of at least $15^\circ\text{C}/\text{m.y.}$ determined for the footwall are consistent with a cause and effect relationship between extensional shearing and rapid exhumation. Hangingwall amphiboles yielded relatively complex age spectra, with minima indicating Late Cretaceous and older cooling dates.

A crustal cross section through the Omineca belt at the latitude of the Trans-Canada Highway is presented, and the extensional faults are restored with the assumption that all displacement on them occurred after regional thrusting ceased. Drawing upon structural, metamorphic and thermochronological data from this study and from other studies in the region, it is shown that the Okanagan - Eagle River fault may have accommodated ~ 30 km of displacement, that total extension across the Omineca belt in the Eocene was likely to have been 20-25%, and that the crust was over 50 km thick prior to Eocene extension. Comparison of the crustal geometry in the present and restored sections suggests that extensional strain was likely to have been partitioned such that the upper crust was most highly attenuated above the central Shuswap complex, whereas the lower crust was most greatly stretched beneath the Intermontane and western Omineca belts.

ACKNOWLEDGEMENTS

Fieldwork was supported by British Columbia Geoscience Research Grants RG87-04 and RG88-02, Natural Sciences and Engineering Research Council of Canada operating grant A2693 and Lithoprobe Supporting Geoscience Grant 85 to R. L. Brown, and by British Columbia Geoscience Research Grant RG89-26 and Geological Society of America Research Grant 4474-90 to the author. I was guided toward life-supporting contracts in 1991 and 1993 by Sharon Carr and Rob Timlin.

Lise Bender assisted me faithfully in the field in 1988 and 1989; I thank her for always using the best of common sense in unpredicted situations, for maintaining a smiling sense of humour even through stretches of weather that were most fit for slugs, and for mapping some of the least appetizing roadcuts of Eagle Bay rocks. I also thank Vicki McNicoll, Kellie Emon and Lindsay Hall for assistance at different times in 1987 and 1990, during which many side orders of toast, assorted John Prine and Bob Marley tunes, wicked folds and cosmic wilderness experiences were shared.

On the technical side, thanks go to several people for help in acquiring and processing laboratory data: Randy Parrish and the staff of the Geochronology Section of the Geological Survey of Canada (notably Pat Hunt and Klaus Santowski) for teaching me U-Pb tactics and for pumping out numbers; Doug Archibald at Queen's University for making the $^{40}\text{Ar}/^{39}\text{Ar}$ lab available and for guiding me through the use of all those valves; Peter Jones for preliminary microprobe analyses of amphiboles; Anett Briggs for cranking out quality thin sections; Lisel Currie for providing a program for Th corrections; Lois Hardy for helping with assorted logistical things; Lindsay Hall for helping with data tabulation; and Louis Buschman for helping to relieve the Tory basement blues by loaning me a computer to use at home.

I appreciate discussions with many people concerning local and regional geology, and aspects of geology in general: Dick Brown, Sharon Carr, Rob Scammell, Jim Crowley, Murray Journeay, Andrew Okulitch, Randy Parrish, Paul Schiarizza, Philip Simony, Roberto Bosdachin, Rob Harrap, Marc Bardoux, Dave Parkinson, Ken Daughtry, Maurice Colpron, Simon Hanmer, Vicki McNicoll, Dick Armstrong, Derek Thorkelson, etc. Also, I express gratitude to Vicki, Randy, Sharon and Rob Scammell for reading and commenting on chapters; and to R.L.B. for enduring numerous drafts of each chapter and sharing many a 2-hour skull session with me. I especially thank Robsville, Sharon, Jim and Dick for many ideas shared in and out of the field.

My response to bombardment with data has been notably less than instantaneous, and in this regard I am very grateful to Dick Brown for being patient while my ideas were slowly diffused out onto paper and incrementally released. Finally, I humbly thank the Faculty of Graduate Studies and Research for having mercy.

ORIGINAL CONTRIBUTION

Following a brief reconnaissance period in the summer of 1986, I undertook a 1:50 000 scale mapping program that involved 10 months of mapping over four summers (1987-1990). My data from parts of 9 map sheets (82 L/10, 11, 14, 15, and 16; 82 M/2, 3, 4, and 6) have been compiled on the map and cross sections in this thesis; additional reconnaissance mapping to the north with Rob Scammell (82 M/11 and 14; 83 D/3) and Dick Brown (82 M/12) has not been included. The cross sections were constructed by projecting structural data by hand from my field maps. I collected and cut the oriented samples used in the kinematic study. I collected the samples used for geochronology, and did all of the required steps leading to, and including, mineral separation. For the U-Pb samples this concluded with handpicking and abrasion at the Geological Survey of Canada in Ottawa. For the $^{40}\text{Ar}/^{39}\text{Ar}$ study, I also weighed, loaded and ran the samples through step heating and mass spectrometer analysis in the geochronology laboratory at Queen's University.

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CHAPTER 1

Introduction

PURPOSE OF THIS INVESTIGATION

The boundary of the Shuswap metamorphic complex (Fig. 1.1 *a*) has long been recognized as a sharp transition between upper amphibolite-facies rocks within the complex and rocks of greenschist facies that overlie it (see Okulitch 1984 for a review). Mapping and geochronology during the past decade have demonstrated that the Shuswap and other metamorphic complexes in the southern Omineca belt are bounded by low to moderate-angle Eocene normal faults, which are responsible for the juxtaposition of rocks of different metamorphic grades and with different cooling histories (Parrish 1984; Parrish et al. 1985, 1988; Parkinson 1985; Bardoux 1985, 1993; Tempelman-Kluit and Parkinson 1986; Carr 1986; Carr et al. 1987).

The present study was spurred by the recognition of mylonitic rocks along the western margin of the Shuswap complex near the town of Sicamous (Fig. 1.1 *a*) by Journeay and Brown (1986) and near Adams Lake by R. L. Brown (pers. comm. 1986). Mylonitic fabrics in both of these areas were observed to have asymmetry indicative of relative westward movement of the upper plate, consistent with the sense of shear documented to the south in the mylonitic footwall of a major low-angle Eocene normal fault in the Okanagan Valley (Bardoux 1985, Parkinson 1985, Tempelman-Kluit and Parkinson 1986). This highlighted the possibility that the

Okanagan Valley fault continued northward to the Shuswap Lake area and from there northwestward to, or beyond, Adams Lake. It also pointed to a need for detailed mapping of the high-grade rocks within the Shuswap complex east of Shuswap Lake. These rocks are part of a composite allochthon that was thrust northeastward on the Monashee décollement, a major Mesozoic to early Tertiary ductile shear zone (Brown et al. 1992a and references therein); fabrics in mylonitic rocks adjacent to the décollement zone indicate the opposite sense of shear (upper plate to the northeast) from that at Sicamous, and within the panel of high-grade rock between the two mylonitic zones there should either be no mylonitic fabric or fabric of one sense should be superimposed on the other.

The major questions addressed in this thesis are:

- 1) Are the mylonitic fabrics at Sicamous related to a low-angle normal fault along the margin of the Shuswap complex?
- 2) Is evidence for such a fault traceable from Sicamous to Adams Lake?
- 3) What are the changes in structural style beneath the mylonitic rocks at Sicamous; in particular, are there overprinting relationships between the west-directed mylonitic fabrics and the northeast-directed fabrics associated with the Monashee décollement?
- 4) Can the high-grade metamorphic rocks within the Shuswap complex near Sicamous be correlated with the lower-grade rocks that overly the complex, and can any of these rocks be correlated with strata of either the ancestral North American miogeocline or accreted terranes?

- 5) What are the ages of mylonitic strain, and is there a difference between the cooling histories of rocks within and above the Shuswap complex that can be explained in the context of a model of extensional tectonic denudation?
- 6) Can these data be used to construct a crustal cross section adjacent to the Trans-Canada Highway, and can such a section be palinspastically restored? If so, how do the results compare with other estimates of the magnitude of extension and crustal thinning in the Omineca belt, and what bearing does this have on existing models for low-angle extensional faulting?

GEOLOGICAL SETTING

Overview

Five distinct morphogeological belts parallel the length of the Canadian Cordillera (Gabrielse et al. 1991; Fig. 1.1 *b*). The Foreland belt is characterized by the thin-skinned thrust and fold belt of the Rocky Mountains, in which Proterozoic to Upper Jurassic platformal and miogeoclinal strata that were deposited on the rifted western margin of the ancestral North American continent, together with synorogenic continental foreland basin strata, were translated northeastward and shortened by at least 200 km during the latest Jurassic through early Tertiary (Price and Mountjoy 1970, Price 1981). The Omineca belt is the exhumed metamorphic-plutonic hinterland to the Foreland belt. It contains variably metamorphosed equivalents of Proterozoic

and Paleozoic miogeoclinal strata, exhumed basement culminations, slivers of accreted terranes, and Paleozoic to Tertiary plutons. A major component of the Omineca belt in southern British Columbia is the pericratonic Kootenay terrane (Monger and Berg 1984, Wheeler et al. 1991), which consists of metamorphic rocks that are probably equivalents of North American miogeoclinal strata (Gabrielse et al. 1991), but within which there is considerable stratigraphic uncertainty and, locally, evidence for pre-late Paleozoic deformation that is not recognized in known miogeoclinal and platformal rocks (Read and Wheeler 1976, Klepacki and Wheeler 1985). West of the Omineca belt are the Intermontane, Coast and Insular belts, which for the most part are comprised at the present erosion surface of Paleozoic to Early Jurassic island arc and oceanic terranes. These terranes were amalgamated to form the composite Intermontane and Insular superterrane (Fig. 1.1 c), which were accreted to the western margin of ancestral North America either successively during Early Jurassic through Early Cretaceous time (Monger et al. 1982, Price et al. 1985, Gabrielse and Yorath 1991) or en masse in the Jurassic (van der Heyden 1992).

Accretion of the Intermontane superterrane in the Early and Middle Jurassic involved large-scale northeast-directed thrusting along the suture, followed or accompanied by southwestward thrusting and backfolding along and inboard of this boundary in the Omineca belt (Struik 1981; Rees and Ferri 1983; Ross et al. 1985; Brown et al. 1986, 1993a; Price 1986; Rees 1987; Murphy 1987). Further convergence from Late Jurassic to early Tertiary was accommodated by northeast-directed thrusting and folding with attendant metamorphism of upper to middle crustal rocks in the Omineca belt (Simony et al. 1980; Price 1981; Brown et al. 1986, 1992a; Journeay 1986), and by thin-skinned thrusting and folding in the Foreland belt (Price and Mountjoy 1970, Price 1981). Proterozoic basement, exposed in windows through

thick allochthonous cover in the Omineca belt, was involved in this deformation in the hinterland (Simony et al. 1980; Morrison 1982; Brown et al. 1986; Journeay 1986; McDonough and Simony 1988; McDonough 1989; Coleman 1989, 1990; Murphy et al. 1991; McNicoll and Brown 1994). A major ductile thrust zone that juxtaposes basement complex and allochthonous cover in the southern Omineca belt is the Monashee décollement (Read and Brown 1981, Journeay and Brown 1986, Brown et al. 1992a). This is an important structure in the context of the present study because it was still active as a thrust until the onset of extensional collapse of the belt, and the processes of thrusting and extensional faulting may have occurred synchronously during a transition from predominantly contractional to predominantly extensional tectonics. A third important process in the tectonic evolution of the Cordillera was large-scale dextral strike-slip faulting during mid-Cretaceous through early Tertiary time (Gabrielse 1985).

Extension in the Omineca belt

Crustal extension of the southern Omineca belt occurred toward the end of a long period of Mesozoic to early Tertiary compression and orogenic thickening, and resulted in tectonic exhumation of high-grade metamorphic rocks from beneath low to moderate-angle Eocene normal fault systems. The exhumed footwalls of the normal fault systems constitute a structural and metamorphic culmination in south-central British Columbia and adjacent Washington known as the Shuswap metamorphic complex (Fig. 1.1). A composite of several smaller, northerly elongate, domal culminations that are each either partly or entirely bounded by outward-dipping Eocene normal faults, the Shuswap complex is the largest metamorphic core complex in the North American Cordillera (Coney 1980, Armstrong 1982). Of particular

importance among the Eocene normal fault systems are the west-side-down Okanagan Valley system (Tempelman-Kluit 1984, Bardoux 1985, Parkinson 1985, Parrish et al. 1985), which bounds the Shuswap complex on the west, and the east-side-down Columbia River - Slocan Lake - Champion Lake system (Read and Brown 1981, Parrish 1984, Corbett and Simony 1984), which defines the eastern boundary of the complex. Mylonitic shear zones exposed in the immediate footwalls of brittle detachments in these systems are interpreted as former distal mid-crustal components of the systems that were abandoned as they were exhumed, in accord with models for evolving low-angle extensional shear zones (e.g. G. H. Davis 1983, Wernicke 1985, G. A. Davis et al. 1986). Rapid cooling during tectonic exhumation of the Shuswap complex is evident from the consistent early Tertiary K-Ar mineral dates, which contrast with the generally Mesozoic cooling ages from the low-grade, brittlely deformed hangingwall rocks (see Parrish et al. 1988).

Shuswap complex

The Shuswap metamorphic complex is the most deeply exhumed part of the southern Omineca belt. Traditionally, the Shuswap complex has been defined as those rocks within the sillimanite zone of regional metamorphism (see Okulitch 1984). In the southern part of the complex, the boundary is defined by Eocene normal faults that juxtapose high-grade metamorphic rocks against hangingwall rocks of distinctly lower grade (Fig. 1.1) (Parrish et al. 1988). In the north, however, the sillimanite isograd does not everywhere coincide with a fault and is not everywhere of the same age, and the boundary of the complex as traditionally defined has no structural significance. In order to keep the nomenclature consistent with that used for other Cordilleran

metamorphic core complexes, Brown and Carr (1990) proposed that the term *Shuswap complex* be used only in reference to rocks that primarily lie in the footwall of Eocene normal faults, and that were cooled from high temperatures in the early Tertiary during exhumation along these faults.

At the deepest structural levels of the Shuswap complex are Paleoproterozoic gneisses considered to be parautochthonous North American basement (Parrish and Armstrong 1983, Parrish and Ross 1990, Armstrong et al. 1991, McDonough and Parrish 1991, Parkinson 1991), exposed in culminations of the Monashee and Malton complexes (Fig. 1.1). These basement complexes are footwall windows through a composite thrust sheet, termed Selkirk allochthon by Read and Brown (1981), which forms the upper plate to major northeast-directed thrusts (the Monashee and Malton décollements) that feed into structures of the Rocky Mountain foreland thrust-fold belt (Brown et al. 1986, 1992a; McDonough and Simony 1988). In its lower level, within the northern and central Shuswap complex, the Selkirk allochthon consists of paragneiss, sillimanite-bearing metapelite, amphibolite and marble, extensively intruded by sheets and dikes of pegmatite (Read and Brown 1981; Journeay 1986; Journeay and Brown 1986; Carr 1989, 1990, 1991, 1992; Scammell 1985, 1986, 1990, 1993; Bosdachin and Harrap 1988; Johnson 1989, 1990, this study; Bosdachin 1989; Harrap 1990; Coleman 1990). These rocks have in part been assigned to the pericratonic Kootenay terrane (Wheeler et al. 1991). The southern part of the Shuswap complex is dominated by orthogneiss (Okulitch 1979, 1984; Tempelman-Kluit 1989; Bardoux 1993), but includes paragneisses that probably are in part equivalent to rocks of the Intermontane superterrane (Okulitch 1984; Carr 1989, 1991).

In the hanging wall of the Okanagan Valley fault system, the Kootenay terrane is represented primarily by the Lower Cambrian (or older) to Mississippian Eagle Bay

assemblage (Okulitch 1979, Schiarizza and Preto 1987). Metamorphic grade in these rocks is mostly greenschist facies, although local staurolite occurrences near the east side of Shuswap Lake (Fyson 1970, Johnson 1990) indicate lower amphibolite facies. The Intermontane superterrane west of the Okanagan Valley consists of Paleozoic to Middle Jurassic volcanic and sedimentary rocks, in part metamorphosed to greenschist facies and overlain by mid-Cretaceous volcanic cover (Monger 1985, 1989; Monger and McMillan 1989; Thorkelson and Rouse 1989). All hangingwall assemblages are cut by Mesozoic plutons (Wheeler and McFeely 1991, Woodsworth et al. 1991 and references therein), and are overlain unconformably by Tertiary volcanic and sedimentary rocks (Monger 1968, Church 1973, Ewing 1981, Tempelman-Kluit and Parkinson 1986).

PREVIOUS MAPPING

The study area encompasses portions of areas mapped at one inch to 4 miles by Jones (1959), Campbell (1963), and Wheeler (1965), and at 1:250 000 by Okulitch (1979). Details of the structure in the Shuswap Lake and Mara Lake areas have been investigated by Fyson (1970) and by Nielsen (1978, 1982). The northwest corner of the accompanying map (Plate 1) adjoins the 1:100 000 scale map by Schiarizza and Preto (1987). The southern segment of the Okanagan Valley fault system has been mapped and studied to the south between Enderby and Kelowna by Bardoux (1985, 1993). To the east, geology of the Monashee Mountains has been mapped at 1:100 000 scale by Read (1979a) and by Höy and Brown (1981). Detailed maps by Bosdachin (1989) and Harrap (1990) delineate the boundary between the Monashee

complex and the overlying Selkirk allochthon near the Trans-Canada Highway.

APPROACH OF THIS INVESTIGATION

Following a brief reconnaissance period in the summer of 1986, I undertook a 1:50 000 scale mapping program along the margin of the Shuswap complex between Three Valley Gap and Adams Lake. This involved 10 months of mapping over four summers (1987-1990), focussing initially on the structure of the Shuswap complex between the Okanagan - Eagle River fault and the Monashee décollement (Johnson 1988), then following the margin of the complex to Adams Lake (Johnson 1989, 1990, 1991). Field data collected during this project are presented in the form of a map at 1:100 000 scale (Plate 1, in back pocket) and seven accompanying cross sections (Plates 2a-g, in back pocket); the sections are presented at 1:50 000 scale so that more structural data can be shown. These serve as the basis for description of the structure of the Okanagan Valley fault system (Chapter 2) and of lithology and structure of the various tectonostratigraphic assemblages and plutonic suites in the area (Chapter 3).

Mylonitic rocks from the margin of the complex were studied in the field, in hand specimen and in thin section (Appendix A). cursory petrographic examination of non-mylonitic rocks was done to determine general mineralogical and structural characteristics and to track specific metamorphic or structural problems.

The question of timing of extensional shearing and unroofing of the Shuswap complex was approached through (i) U-Pb geochronology of selected granitic intrusions that displayed useful field relations, and (ii) $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology of amphiboles from the footwall and hanging wall of the Okanagan Valley fault system

(Chapter 4). U-Pb methods were intended to be used to bracket timing of deformation events, whereas the $^{40}\text{Ar}/^{39}\text{Ar}$ work was aimed mainly at determining and comparing cooling ages.

Finally, these field and laboratory data were used in conjunction with other regional data and Lithoprobe seismic reflection profiles to aid in the construction of present and partially restored crustal cross sections through the Omineca belt adjacent to the Trans-Canada Highway (Chapter 5; Plate 3, in back pocket).

Figure 1.1: Tectonic assemblage map of the southeastern Canadian Cordillera (*a*), simplified from Wheeler and McFeely (1991) and modified based on Carr (1991) and this study. Shuswap core complex as defined by Brown and Carr (1990) is mostly bounded by Eocene normal faults (thick black lines with barbs on hanging wall side), including the Okanagan Valley (OV), Columbia River (CR), and Slocan Lake - Champion Lake (SL) fault systems. Okanagan - Eagle River and Adams - North Thompson segments of OV step en échelon across a transfer zone west of Shuswap Lake. The OV system transects the sillimanite isograd (Sil; from Read et al. 1991) north of Adams Lake. Black lines of medium weight represent mainly northeast-verging thrusts, except along the boundary between the Coast and Intermontane belts where the Fraser - Straight Creek strike-slip fault system is designated by opposing arrows. Southern Rocky Mountain Trench (SRMT) is a morphogeological lineament that defines the boundary between Omineca and Foreland belts. Unpatterned units in these belts are Mesoproterozoic to early Tertiary miogeoclinal and platformal strata. Unpatterned units in the Coast belt are Permian to Cretaceous marine sedimentary and volcanic rocks and metamorphic rocks of uncertain age. Jurassic and Paleozoic granitoid plutons in Omineca and Intermontane belts, and Cretaceous and Neogene volcanic rocks in Intermontane belt, are not shown (see Fig. 5.1, Chapter 5). Bv = Beaven fault, M = Malton complex, MD = Monashee décollement, Mo = Monashee complex, V = Valhalla complex. Inset maps on following page show (*b*) morphogeological belts (IS: Insular; C: Coast; IM: Intermontane; O: Omineca; F: Foreland) and (*c*) composite superterrane of the Canadian Cordillera. AB: Alberta; AK: Alaska; BC: British Columbia; WA: Washington; YT: Yukon Territory. Legend on following page.

a

IGNEOUS UNITS:

- Eocene volcanic and sedimentary rocks
- Middle Eocene syenite, quartz monzonite
- Paleocene-Early Eocene granite
- Cretaceous granite, granodiorite

STRATA ABOVE MAJOR TERTIARY NORMAL FAULTS:

- Intermontane Superterrane (Devonian-Jurassic)
- Kootenay Terrane (Neoproterozoic-Carboniferous)

SHUSWAP COMPLEX:

- undivided metamorphic rocks, in part Kootenay terrane
- Neoproterozoic-Cambrian cover of basement complexes
- Paleoproterozoic basement rocks

- Eocene normal fault
- other faults

CHAPTER 2

Structure of the Okanagan Valley fault system near Shuswap Lake, British Columbia

INTRODUCTION

Low-angle normal faults (Armstrong 1972, Wernicke 1981) have for some time been recognized as important structural features of the broadly domal culminations in the North American Cordillera known as metamorphic core complexes (for example Davis and Coney 1979; numerous papers in Crittenden et. al. 1980, and in Frost and Martin 1982). The Shuswap metamorphic complex of southeastern British Columbia and northern Washington, the largest of these culminations (Armstrong 1982, Coney 1980, Brown and Carr 1990), is partially enclosed by normal faults and associated low-angle ductile shear zones that dip outward along its flanks, and along which hangingwall rocks of relatively low metamorphic grade have been displaced to the east and west away from the culmination (Fig. 2.1) (Parrish et al. 1985, 1988; Brown and Journeay 1987). Of particular importance among these structures are the west-side-down Okanagan Valley fault system (Tempelman-Kluit 1984, Bardoux 1985, Parkinson 1985, Parrish et al. 1985), which bounds the Shuswap complex on the west, and the east-side-down Columbia River (Read and Brown 1981) and Slocan Lake - Champion Lake (Parrish 1984, Corbett and Simony 1984) faults, which define the eastern boundary of the complex.

Several lines of evidence indicate Early to Middle Eocene movement on these faults, and associated rapid uplift and exhumation. Among these are truncation and disruption of Eocene volcanic and sedimentary strata (Bardoux 1985, 1993; Parkinson 1985; Parrish et al. 1985; Tempelman-Kluit and Parkinson 1986), U-Pb ages of deformed and nondeformed intrusions (Parkinson 1985; Carr 1986, 1991; Carr et al. 1987; Parrish et al. 1988), K-Ar dates from the Shuswap complex (Ewing 1980, Mathews 1981, Armstrong 1982, Parrish et al. 1988 and references therein), and the presence of thick Eocene sedimentary basins (Ewing 1981).

Despite cognizance of a major low-angle normal fault in the Okanagan Valley, until recently the nature of the western boundary of the Shuswap complex north of the Okanagan Valley has been problematic. Faults were mapped locally along the boundary near Shuswap Lake, but were interpreted as high-angle brittle faults of small displacement (Jones 1959; Nielsen 1978, 1982; Okulitch 1979). This left open the possibility that the abrupt contrast in metamorphic grade across the boundary of the complex was an artifact of a localized steep paleogeotherm, a view held by earlier workers (Price and Mountjoy 1970, Fyson 1970). However, Journeay and Brown (1986) recognized that mylonitic high-grade rocks along the boundary near the town of Sicamous (Fig. 2.1) contained gently dipping shear foliation and asymmetric fabrics indicative of west-side-down shearing. On the basis of this observation, they proposed that the boundary at Sicamous is a low-angle ductile-brittle normal fault ("Eagle River detachment"), and that it is a northerly continuation of the Okanagan Valley fault.

This paper presents evidence that the Okanagan - Eagle River fault (defined here) is part of a west-side-down normal fault system (Okanagan Valley fault system) that delimits the western boundary of the Shuswap metamorphic complex at least as far north as Adams Lake (Fig. 2.1), and that most likely continues northward, joining

the North Thompson River fault, for a total length of over 450 km. The structure of the Okanagan Valley fault system between the Shuswap River valley and Adams Lake is documented, and it is shown that

- 1) asymmetrical shear fabrics in mylonites of the uppermost footwall (few hundred metres to about 2 km) throughout this region consistently indicate relative westward movement of the upper plate;
- 2) left-stepping en échelon segments of the brittle detachment define a WNW-trending transfer zone between Shuswap Lake and Adams Lake, within which steep brittle faults transect footwall mylonites; and
- 3) mylonitic footwall rocks that have been exhumed within this transfer zone consist predominantly of peraluminous granite and pegmatite, which were at least partly syntectonic and which served to localize strain and facilitate extensional shearing.

OKANAGAN VALLEY FAULT SYSTEM

The Okanagan Valley fault system consists of two major west-dipping, left-stepping, en échelon ductile-brittle fault segments that strike about N10°E (Fig. 2.1). The 290-km-long Okanagan - Eagle River segment is recognized from northernmost Washington to Shuswap Lake by juxtaposition of low to medium-grade hangingwall rocks against mylonitic, high-grade, gneissic footwall rocks of the Shuswap complex, within which asymmetric shear fabrics indicate relative westward displacement of the upper plate. Similar relationships occur at Adams Lake, from where the other major segment of the fault system continues northward for 170 km along the North

Thompson River. Between the Okanagan - Eagle River and Adams Lake - North Thompson segments is the 45-km-wide Shuswap Lake transfer zone. Named and documented herein, the Shuswap Lake transfer zone contains exhumed mylonites but no well-defined brittle detachment. Each of these components of the Okanagan Valley fault system is described in the sections that follow.

Omission of section across the Okanagan Valley fault system is such that no stratigraphic cutoffs or structural piercing points have been correlated across it, and estimates of displacement have therefore been based on other methods of reasoning. A comparison of the cooling histories of footwall and hangingwall rocks led Parkinson (1985) to conclude that between 10 and 60 km of net slip has occurred in the southern Okanagan Valley. Near Shuswap Lake, the mismatch in metamorphic grade and cooling ages across the fault lead to an estimated net slip of about 30 km (Chapter 5). Such estimates based on thermal arguments rely on certain assumptions about the initial dip of the fault and on the paleogeothermal gradient. A less conservative figure of 60-100 km for horizontal extension (heave) across the fault in the Okanagan Valley, based on possible correlations between igneous rock bodies in hanging wall and footwall, was suggested by Tempelman-Kluit and Parkinson (1986). Such correlations are highly speculative, and the estimate of extension so obtained has been strongly contested (Okulitch 1987).

Okanagan - Eagle River segment

For the most part, this segment is delineated by a single major detachment fault, which will be referred to as the Okanagan - Eagle River fault. From the southern end of the Okanagan Valley fault system at Omak Lake, Washington (Goodge and Hansen 1983, Parrish et al. 1988), the detachment trace follows the

Okanagan Valley (Bardoux 1985, Parkinson 1985, Tempelman-Kluit and Parkinson 1986) northward to Sicamous, British Columbia (Parrish et al. 1985) (Figs. 2.1, 2.2). From there it follows the Eagle River valley northeast to Craigellachie Creek, where it turns northwestward, steepens, and continues toward the Anstey Arm of Shuswap Lake (Journeay and Brown 1986; Johnson 1989, 1990). Relatively minor northwest-striking faults mapped within the hanging wall north and south of Sicamous are interpreted as splays of the detachment (Fig. 2.2).

Mylonitic rocks occur throughout the upper 1 km or more of the footwall (Bardoux 1985, Tempelman-Kluit and Parkinson 1986; this study). The contact between these and the hangingwall rocks is a brittle detachment, which is exposed in a few localities in the Okanagan Valley (Parkinson 1985, Parrish et al. 1985), and in two localities east of Shuswap Lake (Johnson 1989, 1990). Its location elsewhere is generally constrained within tens of metres to a few hundred metres, and for the most part coincides with topographic depressions.

A prominent stretching lineation in footwall mylonites trends west-northwest in the southern Okanagan Valley (Parkinson 1985) and near Kelowna (Bardoux 1993), and westward along Mara Lake and the Eagle River (Johnson 1989, 1990; Fig. 2.3). The lineation is defined by aligned inequant minerals and mineral aggregates, elongate and pulled-apart feldspar grains, and quartz ribbons. Along Craigellachie and Four Mile creeks (Fig. 2.2), a mineral lineation defined mainly by micas and sillimanite plunges west-southwest. Mineral lineations of this orientation are ubiquitous well beneath the detachment in the Hunters Range, south of the Eagle River, where rock fabrics record strain that predates Eocene extension (Fig. 3.4 of Chapter 3); similarly, the lineations at Craigellachie and Four Mile creeks are thought to be unrelated to Eocene normal faulting.

Hanging wall / footwall relations

In the southern Okanagan Valley near Oliver, Parkinson (1985) noted that the intensity of the mylonitic fabric in the footwall increases upward toward the detachment and, correspondingly, that the intensity of brittle fracturing in the hanging wall increases downward. Near the detachment, both hangingwall and footwall rocks are brecciated, chloritized and silicified (Parkinson 1985, Parrish et al. 1985). Bardoux (1985) observed that the upper 50 m of the footwall in the Kelowna area is fractured, brecciated, chloritized and epidotized, whereas the hanging wall is less extensively fractured and altered. Along the Eagle River valley and south of Sicamous along Mara Lake, the uppermost several hundred metres of the footwall is predominantly nonannealed mylonite. Gradationally beneath the mylonites are gneissic protoliths in which strain is high but fabrics are annealed.

On a logging road near Four Mile Creek (Fig. 2.2), the contact between fractured and weathered sillimanite-muscovite-biotite schist and migmatite of the footwall and Devonian orthogneiss of the hanging wall is a 50-m-wide zone of cataclastic breccia and clay gouge (Johnson 1989) (Plate 4 *a*; station 397, Plate 1 and Appendix A). In the footwall of this zone, discrete slickensided shear planes in semipelitic schists and anastomosing shear surfaces in pelitic biotite schists are defined by fine-grained retrograde chlorite. Preserved within the brittle zone are relics of footwall mylonite that in thin section exhibit C/S fabric, muscovite fish, and oblique foliations in dynamically recrystallized quartz aggregates (structures described in the next section). Silicified mylonite in breccia zones is cut by anastomosing graphite seams up to a few millimetres wide, which are in turn cut by hematite veinlets and fractures (Plate 4 *b, c*). In the lower 10 m of the hanging wall, the orthogneiss is brecciated and intensely altered by albite-epidote-chlorite (Plate 4 *d*); above this

brecciation diminishes and discrete veins of quartz, epidote-chlorite, and carbonate-epidote cut the rocks. Pegmatite that intrudes both the footwall and hanging wall is nonfoliated to weakly foliated, and is intensely fractured and weathered within the fault zone.

A low-angle detachment is exposed near the village of Cambie in the Eagle River valley (Plate 1, station 845). Mylonitic impure marble in the footwall (designated as unit E?c in Plate 1) lies sharply beneath orthogneiss and amphibolite of the hanging wall. The latter are silicified, altered and veined by chlorite-carbonate-epidote, and intensely fractured in the lower 2-3 m of the hanging wall; they grade upward into gradually less altered granitoid gneiss of the Devonian Mount Fowler suite (unit Dg). The footwall marble has an east-dipping foliation that is oblique to the gently northeast-dipping detachment. Beneath the marble are semipelite migmatites within which mylonitic fabric is developed throughout a structural thickness of about 300 m, and gradationally beneath these are feldspathic gneisses with annealed fabric. The covered contact between the marble and the migmatites strikes obliquely to the foliation in both units, and is interpreted as being another fault. Petrologically, the marble more closely resembles units in the Eagle Bay assemblage than the diopsidic marbles typical of the Shuswap complex, and hence the more significant fault may be that between the marble and migmatites.

Mylonites

Migmatitic pelite, semipelite, and pegmatite along the east side of Mara Lake and along the Eagle River valley exhibit nonannealed mylonitic fabrics indicative of grain-size reduction by dynamic recrystallization during noncoaxial flow (Plates 4 f, 5). Asymmetric structures such as C/S fabrics, rotated feldspar porphyroclasts with

recrystallized tails, and extensional shear bands have been used in the field and in thin section to deduce the sense of shear (Plate 5; Appendix A). In each of 17 localities where two or more independent shear-sense criteria were observed, and in 6 others where well-developed C/S fabrics or porphyroclast systems alone were considered reliable, relative westward movement of the upper plate (hereinafter called west-directed shear) is consistently indicated (Fig. 2.2). Multiple indicators of west-directed shear have also been documented at 4 localities in the central part of the Hunters Range.

The mylonitic foliation and lineation are defined by the peak metamorphic mineral assemblage garnet + sillimanite + K-feldspar + biotite + plagioclase + quartz + leucosome. Sillimanite is oriented parallel to the stretching lineation, occurs locally in pressure shadows adjacent to rotated garnets and K-feldspar porphyroclasts, and occurs along C surfaces that cut across the S foliation, implying that the peak metamorphic assemblage grew before and during the early stages of west-directed shearing (Plate 4 *g-h*). Chlorite is common as a retrograde product of biotite, especially in discrete shears and on C surfaces, along extensional shear bands together with retrograde white mica, and as pressure fringes on garnet and feldspar porphyroclasts. Local brittle shears display slickensides with west-side-down chlorite fibre lineations subparallel to the stretching lineation. Millimetre-scale fractures filled with chlorite and sericite cut across the foliation and across sillimanite grains, perpendicular to the lineation (Plate 4 *h*). Collectively these relationships indicate that retrogression was synkinematic, that the rocks were further sheared as they were exhumed, and that extension and chlorite growth continued as the rocks reached brittle upper-crustal levels.

C/S fabrics (Berthé et al. 1979a, 1979b; Lister and Snoke 1984) consist of

discrete shear (C) surfaces that represent displacement discontinuities and oblique foliation (S) surfaces that parallel the flattening (XY) plane of finite strain (Plate 5 *a*, *b*). Asymmetrical extensional shear bands (Hanmer and Passchier 1991; called C' surfaces by Berthé et al. 1979b, shear band foliation by White et al. 1980, extensional crenulation cleavage by Platt and Vissers 1980) are discrete shears that cut across the mylonitic foliation and that, unlike C surfaces, are oblique with respect to the bulk flow plane (Plates 4 *f*, 5 *c*, 5 *h*). In these rocks the bulk flow plane is approximately parallel to compositional layering, as strain gradients occur normal to the layering. The S foliation is defined by flattened feldspar porphyroclasts, ribbons and lenticular domains of recrystallized quartz, and lenticular leucosome pods, separated from one another by anastomosing biotite and sillimanite. C planes are marked by trains of platy minerals (biotite, chlorite, sillimanite) and by fine-grained dynamically recrystallized quartz and feldspar. Shear bands also are defined by platy minerals (chlorite, biotite, white mica) and recrystallized quartz.

Ribbons and lenses of dynamically recrystallized quartz commonly exhibit internal oblique foliations defined by elongate grain shapes and alignment of grain boundaries (Plate 5 *d*). Obliquity of this type of foliation with respect to C planes is a function of the instantaneous, rather than the finite, strain field, because the orientation is progressively reset by recrystallization throughout the strain path (Law et al. 1984, Lister and Snoke 1984, Hanmer and Passchier 1991).

Feldspar porphyroclasts with dynamically recrystallized tails, including both σ -type (Plate 4 *f*, Plate 5 *a*, *b*, *g*) and δ -type (Plate 5 *e*) porphyroclast systems of Passchier and Simpson (1986), were observed at several localities. At one locality near Sicamous a complex σ - δ system has been noted (Plate 5 *f*), and others have been identified in thin section. δ -type systems tend to form at high shear strains if the ratio

of recrystallization rate to shear strain rate is low; complex systems can result when this ratio is increased during deformation (Passchier and Simpson 1986).

Some flattened feldspar porphyroclasts have been partially replaced by myrmekite, which in some cases is asymmetrically disposed along the long sides of the clasts. Simpson (1985) has argued that, as a result of strain, myrmekite forms preferentially along the edges of K-feldspars that face the instantaneous shortening (minimum elongation) direction in amphibolite-facies quartzofeldspathic mylonites. Hanmer and Passchier (1991) introduced the term "quarter structure" for such asymmetrical structures that form in opposite material quarters of porphyroclasts and that are dependent on the attitudes of the instantaneous strain axes, but they caution that myrmekite quarter structures may not always form synkinematically.

Other structures that were used as kinematic indicators include muscovite fish (Lister and Snoke 1984; Plate 5 *b*), lenticular backrotated foliation segments (foliation fish of Hanmer 1986), and asymmetric microfolds (Berthé et al. 1979a; Plate 5 *g*).

Results of the kinematic analysis described above imply unequivocally that the mylonites formed in response to west-directed noncoaxial shear. Heterogeneous strain is indicated by local strain gradients on the outcrop scale. Asymmetrical extensional shear bands, which require a component of extension within the plane of flow for their development (Platt and Vissers 1980), indicate that the strain regime deviated from ideal simple shear for at least part of the strain history. Because the shear bands are defined by retrograde chlorite and white mica, this deviation need only have been in late stages of the unroofing history of the rocks, and may have been insignificant on the scale of a km-thick shear zone.

Late retrogression and brittle overprint are most pronounced close to the brittle detachment. For example, about 150 metres beneath the topographic depression that

represents the inferred position of the detachment on the west side of Mara Lake (Fig. 2.4, Plate 2c), migmatitic semipelite mylonites exhibit slickensided C planes, spaced about 2 millimetres apart, defined by chlorite, hematite, and relict sillimanite. Thin sections show that strong sericitization of flattened feldspar porphyroclasts and almost complete replacement of biotite by chlorite have occurred. Microfractures filled with quartz and hematite cut feldspars, garnets, and matrix (Plate 4 e).

Brittle detachment

Except along Four Mile and Craigellachie creeks where it dips about 55° to the southwest, the brittle detachment dips less than 30° . Near Sicamous, field relationships imply that the detachment has a low dip. Firstly, the closely spaced, slickensided shear planes in mylonites beneath the detachment along Mara Lake dip 25° west-northwest (Fig. 2.4; Plate 2c). Secondly, retrograded ultramylonites are exposed high on the northwestern slope of the Hunters Range facing the Eagle River (Plate 1 and Appendix A, stations 240, 241, 291); this suggests that prior to erosion the detachment was hundreds of metres or less above the present topographic surface and, therefore, that the 15° topographic slope mimics the morphology of the detachment surface (*frontispiece*). Further south, klippen of Eocene volcanic and sedimentary rocks overlie footwall mylonites between Enderby and Mabel Lake (Mathews 1981, Bardoux 1993) (Fig. 2.2), and the map pattern constrains the dip of the unexposed detachment as being gentle to subhorizontal. East of Vernon, subhorizontal segments of the detachment are dropped down to the east by steep north-striking normal faults, resulting in exposure of medium-grade hangingwall klippen across the Okanagan Highland (Carr 1991, Brown et al. 1992a) (Fig. 2.1). Three-point solutions constrain the dip of the detachment to between subhorizontal and

about 10-20° westward in the southern Okanagan Valley (Parkinson 1985, Bardoux 1993). In the subsurface, Lithoprobe seismic reflections correlated with the fault and with footwall mylonites west of Vernon dip less than 10° westward and flatten to subhorizontal in the middle crust (Carr 1991, Cook et al. 1992, Brown et al. 1992a).

Collectively, these data suggest that the detachment dips 10-25° where it intersects the present erosion surface along the Okanagan and Eagle River valleys. Downdip to the west, it roots in a subhorizontal mid-crustal shear zone. Updip to the east, it rolls over to subhorizontal for up to 70 km and may steepen to a southwest-dipping breakaway zone or ramp that parallels the southwest flank of the Monashee complex. The Beaven fault (70 km east of Vernon, Fig. 2.1), which has been correlated with the Okanagan Valley fault (Carr 1991), and the Four Mile - Craigellachie Creek section of the fault northeast of Sicamous may be remnants of a relatively steep zone that have escaped erosion.

Cross sections drawn through the Sicamous area illustrate these regional geometric features of the detachment, as well as some local aberrations, which in combination are responsible for the deep reentrant in the fault trace northeast of Sicamous. They also illustrate the geometric relationships of the detachment to footwall mylonites.

Plate 2d, drawn at a low angle to the strike of the fault system, shows a broad, asymmetrical, synformal warp in the fault beneath the northeastern Shuswap Range, across which the fault steepens toward Craigellachie Creek. Warps like this that parallel the extension direction in low-angle normal faults have been interpreted in various ways. They may mimic preexisting structure in either the lower (John 1987, Davis and Lister 1988) or upper plate (Spencer and Reynolds 1991), or they may be caused by folding in response to either late tectonic compression (Davis et al. 1980,

Yin and Dunn 1992) or syntectonic decompressional expansion of the lower plate (Spencer 1982). Late compressional folding is not a likely cause for the Shuswap Range reentrant because the only other warp in the fault surface of comparable scale is 120 km south near Kelowna (Bardoux 1985; Fig. 2.1), hence the wavelength is nonuniform, and because there are no other warps comparable in both scale and geometry. Older structure in the immediate footwall is also unlikely as a major control because these rocks were at temperatures of the order of 500°C in the Early Eocene ($^{40}\text{Ar}/^{39}\text{Ar}$ data presented in Chapter 4) and would have been quite ductile. Spencer and Reynolds (1991) have argued that the lower plate of a detachment is more likely to flow around and conform to the structure of the colder upper plate as it is pulled out from underneath. Structures in the Eagle Bay assemblage, and in particular the relatively stiff Mount Fowler orthogneiss, are at low angles to the trend of the reentrant and to the extension vector. These upper-plate rocks in the Shuswap Range were at about 500°C in the Late Cretaceous (Chapter 4) and did not cool below about 300°C until late Middle Eocene (Johnson 1989 and references therein; Chapter 5); this part of the upper plate therefore may not have been seriously fractured and weakened when extension began, and it would have been strong relative to footwall rocks.

Preexisting basement structure such as a major southwest-dipping ramp in the Monashee décollement (Brown et al. 1992a) may have controlled the orientation of the section of the fault along Craigellachie Creek. Both this section and the Beaven fault strike subparallel to the southwestern margin of the Monashee complex, oblique to the overall northerly strike of the Okanagan - Eagle River detachment, and both are steeper than adjoining parts of the detachment. The reentrant may have therefore initiated as a primary groove or inflection in the fault surface. Other factors, such as

expansion of the footwall due to decompression (Spencer 1982) and inhomogeneous buoyant forces caused by emplacement of granite melts in the footwall (Yin 1991), may have modified its geometry syntectonically. This interpretation is consistent with the observation that foliation in the footwall parallels the detachment across this structure in the plane of section of Plate 2d.

Plate 2a is an east-west section across the Eagle River valley. Along the Eagle River the detachment generally dips northwest, but locally it and the underlying mylonites dip northeast. Similar dip changes could be produced by backrotation of the detachment against younger listric faults that cut it, but none with significant displacement have been recognized in the Eagle River area. Instead, the detachment and mylonitic foliation are evidently corrugated by two pairs of gently north-plunging warps with amplitudes of up to 250 m and wavelengths of 3.5-5.5 km (Plate 2a, just west of where section crosses the Eagle River). Synextensional isostatic uplift could have caused these warps to form. East of where the section crosses the Eagle River, the now-eroded detachment is depicted as paralleling the northwest slope of the Hunters Range as discussed earlier, eventually steepening updip as it does at Craigellachie Creek.

Migmatitic pelite and semipelite units in the central Hunters Range locally display west-directed mylonitic fabrics (locations shown in Fig. 2.2 and Plate 1). These fabrics occur in discrete shears up to a few metres wide, in which biotite is partially replaced by chlorite. Most of the shears that have been observed dip eastward (stations 173, 44, 891); these are associated with, and locally crosscut, overturned WSW-verging folds that have moderately to steeply ENE-dipping axial surfaces (F_3 folds, see Chapter 3). Although these shears could be extensional in origin (Johnson 1989, 1990), they are more likely minor backthrusts related to

movement of the Hunters Range rocks over a ramp in the Monashee décollement (see discussion of footwall structures in section on *regional structural and metamorphic relations* in Chapter 3). If these shears are extensional in origin, then presumably they have been backrotated from original westerly or subhorizontal dips. This is difficult to rectify in light of a palinspastic restoration of the extensional faults (Chapter 5), wherein the restored attitudes of structures in the Hunters Range are not much different than their present attitudes.

Adams - North Thompson segment

The northern part of Adams Lake and the lower Adams River valley are coincident with the boundary between sillimanite-bearing rocks, pegmatites, and two-mica granites of the Shuswap complex on the east and upper greenschist-facies rocks of the Eagle Bay assemblage, Devonian orthogneiss, and the mid-Cretaceous Baldy granite batholith on the west (Fig. 2.2). Pre-Late Cretaceous structures that dominate the map pattern in the Eagle Bay assemblage strike east-southeast (Schiarizza and Preto 1987), at a high angle to the contact with high-grade rocks. Migmatitic pelite mylonites with numerous indicators of WNW-directed shear sense (C/S fabrics, σ -type feldspar porphyroclast systems, muscovite fish, asymmetric microfolds, extensional shear bands, and oblique foliations in recrystallized quartz; Plate 5 *h*) have been documented at two localities on the east side of Adams Lake. A muscovite-biotite granite dike from one of these localities also shows WNW-directed C/S and oblique quartz fabrics (Plate 6 *a*). Quartzofeldspathic gneisses along Adams Lake are foliated, but are not lineated and lack asymmetric shear fabrics; as with the Eagle River segment, strain related to west-directed shear is thus apparently partitioned into the weaker metapelites and granite intrusions. Chlorite and quartz slickenlines in rocks

along Adams Lake consistently trend 290-300°, subparallel to the stretching lineation, and consistently have fibre asymmetry indicative of northwest-side-down normal slip.

These structural and metamorphic relationships suggest that the main detachment of the Okanagan Valley fault system must follow Adams Lake and the lower reaches of the Adams River northward. It is approximately on strike with the North Thompson fault (Campbell 1968, Pell and Simony 1981, Pell 1984) and is inferred to somehow be linked to it. The North Thompson fault juxtaposes lower to upper amphibolite-facies rocks of the Cariboo Mountains on the west (Pell and Simony 1981, Pell 1984, Dechesne et al. 1984, Currie and Simony 1987) against upper amphibolite-facies rocks in the Monashee Mountains on the east (Sevigny and Simony 1989, Digel et al. 1989). To the north it forms the western boundary of the Malton gneiss complex (Morrison 1979; Fig. 2.1). Net slip is at least 4 km in the North Thompson River area (Pell and Simony 1981), and is between 0.5 and 5 km at the Southern Rocky Mountain Trench where the fault apparently merges with a northwest-striking, southwest-side-down normal fault (Murphy 1990a, Murphy et al. 1991).

Shuswap Lake transfer zone

The Shuswap Lake transfer zone consists of several components that either collectively or individually, and in either case diachronously, served to transfer displacement between the Okanagan - Eagle River and Adams - North Thompson fault segments. Between Shuswap Lake and Adams Lake, strain related to westerly directed shearing is most evident in granites of the Pukeashun suite, which are well exposed at Pukeashun Mountain (Fig. 2.2; Plate 1). These granites are heterogeneously sheared and are concentrated in a zone up to 2 km or more in structural thickness; this zone, here called the Pukeashun shear zone, crops out in a

wide belt north of (and structurally below) the sillimanite isograd. Granite and pegmatite form about 80% of the total volume of rock within the shear zone (50-100% at any given exposure); paragneiss, amphibolite, and migmatitic schist are preserved as screens. Above and below the zone, medium-grained granite forms less than 50% of the total rock volume and is much less affected by ductile strain. The lower (northern) boundary of the shear zone has yet to be studied in detail but is apparently gradational. The top of the zone is gradational at Pukeashun Mountain. Near the east shore of Adams Lake, strain increases southward in the granite, and at the southern margin of the zone it is mylonite that breaks readily into plates along C planes (Plate 6 *b*; station 725, Plate 1). Sillimanite-muscovite-biotite schists and paragneisses south of the granite mylonite are intruded by sparse, relatively weakly foliated granites, and are cut by steep brittle faults. These faults strike west-northwestward, parallel to lithological contacts, and may be strike-slip transfer faults. Steep northerly striking faults also occur in the step between the major en échelon segments of the Okanagan Valley fault system and may have aided in displacement transfer.

Pukeashun shear zone

The Pukeashun granite suite consists of multiple intrusions of muscovite granite pegmatite and medium-grained, leucocratic, muscovite and garnet-biotite-muscovite monzogranites. Contacts between different dikes and sheetlike intrusions of granite and pegmatite are commonly sharp, but some pegmatitic segregations have diffuse contacts with enveloping granites. Sheets of granite are concordant to variably discordant with respect to the foliation and to compositional layering in the country rocks. U-Pb analyses of zircon and monazite from granite protomylonite of one slightly discordant dike are consistent with an Early Eocene age of 56 ± 1 Ma (sample

J89-762, Chapter 4).

Strain in the Pukeashun granites is inhomogeneous. Although some of the pegmatites and the most discordant of granite dikes have no penetrative fabric, most of the granite is strongly to mildly foliated and well lineated. The foliation dips westward to northwestward near Adams Lake, and dips 10-45° east-southeast between Pukeashun Mountain and the Seymour Arm of Shuswap Lake (Plate 1). The lineation, defined by elongate feldspar grains and quartz ribbons, plunges 5-30° east-southeast in the Pukeashun Mountain area (mean 121°-13°; Fig. 2.3).

Where the foliation is strongest the texture is mylonitic. Well developed C/S fabrics were noted at 4 localities in the field and in thin sections from 4 additional localities (Plate 6 *b-g*). Muscovite fish, oblique foliations in dynamically recrystallized quartz ribbons, and myrmekite quarter structures were also seen in thin section. In each case the sense of shear indicated was to the west-northwest (i.e. toward a mean azimuth of 301°). Apophyses and thin tails of discordant and otherwise weakly foliated leucogranite dikes are flattened parallel to the foliation and locally display excellent C/S fabrics (Plate 6 *e, f*). This relationship suggests that the Pukeashun suite is at least in part syntectonic with respect to west-northwesterly shearing.

East of Pukeashun Mountain, the granite generally is moderately to weakly foliated and lineated, and commonly is highly fractured. Ductile strain is concentrated in discrete mylonite zones a few metres wide. Mylonitic muscovite granite and highly chloritized semipelitic schists were collected from one of these zones just west of the Seymour Arm of Shuswap Lake (station 1097, Plate 1). The granite bears a strong stretching lineation, and in thin section exhibits three independent structures that indicate west-directed shear (Appendix A; Plate 5 *d*). This shear zone and the weakly foliated muscovite-biotite granite outside of it are overprinted by extensive brittle

fractures in a zone at least 200 m wide.

Structures above the Pukeashun granite

South of and structurally above the Pukeashun granite between Adams Lake and Kwikoit Creek (Fig. 2.2; Plate 1), psammitic paragneisses and coarse-grained biotite-muscovite schists containing fibrous sillimanite (Lichen assemblage; see Plate 1) are exposed within 200 metres of chloritic phyllites of the Eagle Bay assemblage. Pegmatites are abundant in the sillimanite-bearing rocks, but are absent in the phyllites and in other Eagle Bay rocks to the south. On the basis of the contrasting metamorphism and pegmatite content, truncation and local omission of the Devonian Mount Fowler orthogneiss, and truncation of the garnet isograd in the Eagle Bay assemblage (Johnson 1990; Plate 1), the sillimanite-bearing rocks and the phyllites are interpreted as being in faulted contact. This contact was previously interpreted as a low-angle brittle splay of the Adams Lake detachment (Johnson 1990). In this interpretation, the Mount Fowler orthogneiss lies entirely within the hanging wall, on both sides of a narrow footwall culmination of high-grade schist and paragneiss. In favour of this interpretation, the lower part of the orthogneiss and the subjacent schists in small outcrops near Kwikoit Creek show intense chlorite and sericite alteration, are locally silicified and brecciated, and are cut by veinlets of quartz, chlorite-sericite, and carbonate; slickensides and epidote veinlets persist in the orthogneiss above the contact.

An alternative and simpler interpretation is that the high-grade and low-grade rocks west of Kwikoit Creek are juxtaposed by a steeply to moderately dipping, northeasterly striking normal fault, as shown in Figure 2.2 and Plate 1. The fault is inferred to have southeast-side-down slip and to separate contacts between the high-

grade rocks, the orthogneiss, and the low-grade rocks in an apparent left-lateral sense. The northeast strike of the fault is consistent with the orientation of lineaments and high-angle normal faults discussed later; the style is reflected at the mesoscale by brittle normal shears that parallel steeply southeast-dipping foliation in the Mount Fowler orthogneiss northwest of Kwikoit Creek, which display oblique foliation defined by micas (Plate 6 *h*). To account for the possibility that minor shearing occurred on a low-angle fault or faults in or marginal to the orthogneiss, the fault west of Kwikoit Creek is depicted in Plate 1 as flattening to a gentler dip in the north and following the lower contact of the orthogneiss.

Between Lichen Mountain and Shuswap Lake, no sharp metamorphic or structural discontinuity is evident between high- and low-grade metamorphic rocks. Instead, the sillimanite isograd passes through an apparently intact succession of coarse-grained mica schists and thinly layered feldspathic quartzites on Lichen Mountain. A sheet of Mount Fowler orthogneiss more than 1.5 km thick separates these rocks from structurally higher upper to lower greenschist-facies rocks of the Eagle Bay assemblage. Pegmatites are abundant in the schists between the Mount Fowler gneiss and the Pukeashun granite (both above and below the sillimanite isograd), but are rare within the orthogneiss and are absent in the low-grade rocks above it. Neither the lower nor the upper contacts of the orthogneiss are demonstrably faulted between Lichen Mountain and Shuswap Lake; the orthogneiss clearly intrudes the underlying rocks of the Lichen assemblage north of Mount Fowler, apparently intrudes the overlying Eagle Bay assemblage near Mount Mobley, and apparently intrudes both underlying and overlying rocks on the southwest flank of Lichen Mountain.

Northwest-directed extensional shear bands and slickenfibres were observed in

one of several discrete, southeast-dipping, cm-scale shears that cut across the foliation in the Devonian orthogneiss near Mount Fowler. Apart from the zones of high strain in the Pukeashun granite, these constitute the only evidence of west-directed shearing between Lichen Mountain and Shuswap Lake, and are likely to have accommodated minor Tertiary extension. The main tectonite fabric in the orthogneiss, however, is considered to be a result of older strain; it is everywhere annealed, which would not be expected if it were the same age as the nonannealed fabric in the Pukeashun granite because the orthogneiss was relatively cold when the granite was emplaced and sheared ($^{40}\text{Ar}/^{39}\text{Ar}$ data, Chapter 4; see below).

If, as it appears, no major displacement and structural omission have occurred across the boundaries of the Mount Fowler orthogneiss, then the contrast in metamorphic character of the rocks above and below it requires another explanation. As described earlier, the rocks beneath the orthogneiss are coarse-grained, generally contain fibrous sillimanite, and are intruded by pegmatites, whereas the rocks above it lack pegmatite and are of upper to lower greenschist facies. It is suggested that the orthogneiss acted as a mechanical barrier to melts emplaced from below, and that it insulated the overlying rocks from heat introduced during emplacement of these melts while allowing fibrolite to grow in the underlying rocks. This hypothesis requires testing, but is currently permissible. Above the orthogneiss, the metamorphic peak in the Eagle Bay assemblage predated emplacement of the Baldy batholith (Schiarizza and Preto 1987), which is mid-Cretaceous in age (Calderwood et al. 1990). Hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ cooling dates from the Mount Fowler orthogneiss and from rocks associated with fibrolite-bearing schists between the orthogneiss and the Pukeashun granite are mid- to Late Cretaceous (Chapter 4), implying that those rocks cooled below about 500°C (Harrison 1981) by then. The fibrolite, which can grow at still

lower temperatures (Kerrick 1987), could be mid-Cretaceous or younger. The Pukeashun granite suite is in part earliest Eocene, but ages of the older components of the suite and of the abundant pegmatites are uncertain; granites of Early Cretaceous to Paleocene age are widespread at deeper structural levels in the Shuswap complex to the north (Scammell 1992, 1993; Sevigny et al. 1989), where arguments can be made for a mid-Cretaceous thermal peak (Scammell 1993). Magmatism and fibrolite growth therefore could have been contemporaneous.

Evidence for a steep transfer fault zone

At Four Mile Creek (Fig. 2.2; Plate 1), the brittle detachment of the Okanagan - Eagle River fault strikes northwesterly and dips moderately southwest. The foliation in footwall rocks there dips moderately to subvertically. Rocks along the shore of Anstey Arm that are inferred to belong to the hanging wall of the main detachment (Mount Fowler orthogneiss and metasedimentary units designated Lc? and Lu?, Plate 1) display steep WNW-striking foliation and are intruded by abundant pegmatite, both unusual characteristics for hangingwall rocks elsewhere in the Shuswap Range.

Approximately along strike to the west-northwest, other steeply dipping structures occur within the narrow and very poorly exposed domain of metasedimentary rocks between the Mount Fowler gneiss and the Pukeashun granite (the Lichen assemblage; see Plate 1 and structural descriptions in Chapter 3). On the east shore of Seymour Arm, steeply foliated biotite-muscovite schists occur ~100 m south of highly strained sheets of muscovite leucogranite and pegmatite interleaved with migmatitic schist and paragneiss (station 1102, Plate 1). Foliation in the high-strain zone dips moderately to the south-southwest and a strong stretching lineation

plunges 11° toward 280° ; petrographic observation of the leucogranite revealed no kinematic indicators. Further along strike near Adams Lake, schists and paragneisses adjacent to the southern margin of the Pukeashun granite are cut by steep brittle faults that strike west-northwestward and that are characterized by zones up to 100 m wide of fracturing, intense weathering, and cataclasis (stations 424, 425, and 746, Plate 1 and Appendix A).

These observations constitute equivocal evidence for a steep WNW-striking zone of displacement linking the Okanagan - Eagle River and Adams - North Thompson normal fault segments, and offer an objective for future field mapping. The zone would appear to be mainly brittle near Adams Lake but ductile or ductile-brittle near Shuswap Lake. To be kinematically compatible with the Okanagan Valley fault system it would have to have a right-lateral sense of displacement.

High-angle brittle faults

Both the footwall and the hanging wall of the Okanagan Valley fault system are cut by brittle high-angle faults that strike north-northwest in the Hunters Range to north-northeast between Shuswap Lake and Adams Lake, subparallel to post-Early Eocene dike swarms and to pronounced topographic lineaments. These faults include the west-side-down Perry River, Three Valley, Crowfoot - Ross Creek, and Cross Creek faults and an east-side-down fault at Mount Mara. The largest faults have up to about 2 km of dip slip, as estimated from downplunge projections of hangingwall and footwall rocks of the Crowfoot and Cross Creek faults (normal faults respectively east and west of Lichen Mtn. on Plate 2e) and by offset of the muscovite-out isograd across the Three Valley fault (Johnson 1990; Plate 2a). The Anstey and Seymour arms of Shuswap Lake, both north-trending topographic lineaments across which map

units appear to be offset (Fig. 2.2; Plate 1), could be high-angle faults with up to 4 km of west-side-down dip slip partitioned between them, as suggested by projection of sparse data from the area (Plate 2e).

INTERPRETATION

Like low-angle normal faults associated with other core complexes in the North American Cordillera, the Okanagan - Eagle River segment of the Okanagan Valley fault system consists of a low-angle brittle detachment that truncates a wide mylonite zone in its footwall. Strain increases toward the detachment, and the immediate footwall is affected by synkinematic retrograde metamorphism and superimposed brittle structures. Similar field relationships are exhibited along northern Adams Lake. These relationships are consistent with models for extensional simple shear zones (for example G. H. Davis 1983, Wernicke 1985, G. A. Davis et al. 1986), which predict that footwall mylonites developed at high metamorphic grade in the middle crust are overprinted by lower-grade ductile and eventually brittle shears as they are tectonically exhumed.

The Shuswap Lake transfer zone is a step between en échelon detachment fault segments, across which mylonitic footwall rocks of the Pukeashun shear zone have been exhumed. Foliation across the Pukeashun shear zone defines a broad, SSW-plunging antiform. On both limbs of the antiform, mylonitic fabrics in the Pukeashun granite indicate WNW-directed shear. Sheared granite on the southeast-dipping limb is interpreted as the backrotated part of a westerly rooted normal-slip shear zone. However, the southeast dip of the foliation is consistent with thrust-sense shear, and so

this possibility must be considered. The rocks lie above and behind a footwall ramp of the Monashee décollement, a major northeast-directed ductile thrust that was still active in the latest Paleocene (Brown and Journeay 1987, Carr 1992, Brown et al. 1992a). Thrust displacement across transitions from ramp to flat induces a component of simple shear in the upper plate of opposite sense to that of net displacement (Sanderson 1982), which could cause minor backthrusts to form (Fig. 2.5 a).

Objections to this interpretation for shear fabrics in the Pukeashun granite are many. First, rocks above the zone of west-directed shear in the Pukeashun granite are of lower metamorphic grade than rocks beneath it, a relationship which can occur across a thrust only if it cuts parallel to, or down through, isogradic surfaces. This is geometrically possible only if the isogradic surfaces originally dipped eastward, and this is inconsistent with the field observation that, in both hangingwall and footwall rocks, there is no significant change in metamorphic grade across the width of the transfer zone. Second, thrusting in the Omineca belt during the late Mesozoic and early Tertiary was directed northeastward, and major westerly verging backthrusts of Paleocene or younger age have not been documented. Finally, the main objection to this interpretation is that west-side-down normal slip must have somehow been transferred between mylonites of the Okanagan - Eagle River and Adams - North Thompson segments of the Okanagan Valley fault system when they were active. The mismatch in metamorphic grade across these fault segments requires up to tens of kilometres of slip, and the Pukeashun granite contains the only documented ductile strain between the two segments that is kinematically and temporally compatible with extension. Strain in the Pukeashun granite is therefore considered to have been caused by Early Eocene and possibly older normal-sense shear.

Low-angle normal faults and their mylonitic footwalls in Cordilleran core

complexes typically are folded into broad, doubly plunging antiforms (Coney 1980). These generally are thought to result from uplift driven by isostasy in response to tectonic denudation (Rehrig and Reynolds 1980, Hyndman 1980, Spencer 1984, Wernicke and Axen 1988, Davis and Lister 1988). They have been mathematically simulated as such in both elastic (Spencer 1984, Buck 1988) and viscous (Wdowinski and Axen 1992) two-dimensional models, and in three-dimensional elastic models (Yin 1991). Isostatically induced warping is favoured by an initial listric geometry of the detachment fault, nonuniform extension of the upper plate, and initial nonuniform topography (Spencer 1984), and can be enhanced by uncompensated crustal roots beneath the extended terrane and by buoyant forces induced by emplacement of syntectonic granites (Yin 1991).

Geometric - kinematic models

Several possible interpretations for evolution of the Shuswap Lake transfer zone involve progressive abandonment and tilting or broad warping of the Pukeashun granite mylonites during crustal extension (Figs. 2.5 *b-e*). In most cases, warping is attributed to isostasy. In addition, if the granites were emplaced laterally rather than through underlying rocks, buoyant forces could enhance warping. The key differences between the various interpretations are in the kinematic, geometric, and temporal relationships between the Okanagan - Eagle River (O-ER) and Adams - North Thompson (A-NT) segments and the Pukeashun shear zone.

The simplest interpretation is that the O-ER, Pukeashun, and A-NT segments are part of a contiguous Okanagan Valley fault that has been folded into a broad antiform-synform pair (Fig. 2.5 *b*). It seems unlikely that mylonites both updip (O-ER) and downdip (A-NT) from the granite should have brittle faults superimposed on

them if the granite does not. However, mylonitic Pukeashun granite near Adams Lake is juxtaposed against metasediments and weakly foliated granites that are cut by brittle zones of cataclasite and gouge, and high structural levels of the granite on the backrotated limb near Seymour Arm are cut by discrete chloritic shears and fractures. These constitute the only documented evidence for a brittle detachment at the top of the Pukeashun shear zone. At Pukeashun Mountain where exposure of the transition is best, the top of the shear zone is gradational, and thus evidence against such a detachment seems more compelling. Furthermore, the brittle zones near Adams Lake are steeply dipping and probably are strike-slip transfer faults as discussed previously. The kinematic relationship shown in Figure 2.5 *b*, though not favoured, is still possible if the Pukeashun granite was emplaced into the upper crust and if all of the displacement occurred while the granite was still hot and ductile relative to brittle rocks beneath the O-ER and A-NT.

A second possible interpretation is that the Pukeashun mylonites are the distal (formerly downdip) part of the O-ER segment, cut by and uplifted in the footwall of the A-NT segment (Fig. 2.5 *c*). This is consistent with absence of a brittle detachment above the Pukeashun granite, as motion on the O-ER and in the Pukeashun shear zone could have ceased before the latter was dissected by the A-NT segment and uplifted to high crustal levels. The A-NT could have either incised the Pukeashun shear zone or captured and continued to utilize the west-dipping part of the zone. Identical models have been proposed to explain the origin of backrotated mylonites that diverge downward from overlying younger detachments in the Valhalla complex, 150 km to the southeast (Carr 1986, Carr et al. 1987; Fig. 2.1), and in detachment systems of the Colorado River region in California (Davis and Lister 1988), although in these systems the faults that originally fed displacement into the backrotated mylonites have not been

positively identified. Reynolds and Lister (1990) have suggested that such a model is applicable to detachment systems in many Cordilleran core complexes. They call upon warping of the initial detachment to cause abandonment of the backrotated limb, whereupon the former upper plate becomes progressively excised by new detachments that continue to feed displacement into the normal-dipping limb of the shear zone. Davis and Lister (1988) have argued that isostatically induced warping can promote either excisement of the upper plate or incisement of the lower plate during progressive extension. The presence of low-grade discrete shears in the upper part of and structurally above the Pukeashun granite, including chloritic shears near the base of the Devonian orthogneiss near Kwikoit Creek and in the orthogneiss at Mount Fowler, are consistent with excisement concomitant with exhumation.

Two final possibilities involve incisement and backrotation of the Pukeashun shear zone in the hanging wall of a younger, listric O-ER segment. In one case the Pukeashun shear zone and the A-NT segment are contiguous, and have been folded into a broad rollover antiform (Fig. 2.5 *d*). This is unlikely, for reasons outlined in regard to the interpretation shown in Figure 2.5 *b*. In the other case, the Pukeashun shear zone is cut by both the A-NT and O-ER segments, which presumably merge into a common shear zone in the middle crust (Fig. 2.5 *e*). One tectonic scenario that would lead to this geometry is a variation of the model illustrated in Figure 2.5 *c*, wherein the processes of warping, abandonment and excisement lead to reactivation of the normal-dipping proximal part of the abandoned breakaway fault and to incisement of its footwall; in other words, the O-ER in Figure 2.5 *c* becomes the O-ER in Figure 2.5 *e*. This scenario is consistent with available data.

A review of the timing constraints on shearing along the different segments shows that each of the interpretations shown in Figure 2.5 is admissible. The only

direct constraints come from dates on three sheared intrusions from the footwall and on detached Middle Eocene volcanic rocks. Two sheared intrusions in the Okanagan Valley are 50 ± 2 Ma (Rb-Sr whole-rock isochron; Solberg 1976) and 51 ± 1 Ma (U-Pb zircon; Parkinson 1985). The other dated intrusion is a 56 ± 1 Ma dike from the Pukeashun granite suite (U-Pb zircon; Chapter 4). Volcanic rocks carried in the hanging wall have given Middle Eocene K-Ar dates (see summary in Tempelman-Kluit and Parkinson 1986). These data imply that the last motion in the Okanagan Valley was Middle Eocene or later and that some motion in the Pukeashun shear zone was earliest Eocene. Rapid cooling of footwall rocks between Late Paleocene and Middle Eocene, indicated by K-Ar dates on hornblende and micas, is consistent with tectonic denudation in Early to Middle Eocene (Parrish et al. 1988). In the North Thompson River area, cooling ages and geobarometric constraints favour mainly Middle Eocene (51-45 Ma) movement (Sevigny et al. 1990).

The following conceptual model is based mainly on the interpretation offered in Figure 2.5 *e*: The Pukeashun granite, emplaced in the Paleocene to Early Eocene, ponded in a tabular zone beneath the Mount Fowler orthogneiss. Heat from the granite caused this zone to soften and allowed the granite to localize and facilitate extensional shearing. Tectonic denudation of this shear zone caused the zone to warp into a broad antiform. Locking and abandonment of parts of the shear zone, as the granite cooled and was warped, led to excisement of its hanging wall by splays that cut progressively up structural section, the youngest splay being the Adams - North Thompson fault segment. Abandonment of the Pukeashun shear zone may also have led to the development of a west-northwesterly striking transfer fault system linking the en échelon detachment segments. By Middle Eocene time, the Pukeashun shear zone was incised by the brittle Okanagan - Eagle River detachment and numerous

smaller brittle faults. The present map pattern (Figs. 2.1, 2.2; Plate 1) shows the superposition of brittle faults on mylonites that were progressively abandoned and exhumed as a consequence of tectonic denudation.

DISCUSSION

Bartley and Glazner (1991) proposed that regional extension oblique to an active magmatic belt controlled the en échelon arrangement of the Miocene Colorado River and central Mojave Desert extended terranes in southeastern California and Arizona. Their model invokes a mid-crustal shear zone into which the left-stepping en échelon segments of the system merge, and involves clockwise rotation of the crustal block in the transfer zone between the segments. A different model for the same region presented by Martin et al. (1993) calls upon an inferred strike-slip transfer fault between the en échelon segments to accommodate displacement transfer.

The Shuswap Lake transfer zone may have elements in common with both of these models and is important because, unlike the Colorado River - Mojave Desert system, the sheared magmatic roots of the zone are exposed as a result of deep exhumation. The WNW trend of the stretching lineation in the Pukeashun granite and the NNE strike of high-angle normal faults and porphyry dikes in the transfer zone are consistent with clockwise rotation relative to the footwall of the Eagle River segment of the fault system, where lineations related to extension trend westward and high-angle faults and dikes strike north to NNW. However, the different dike orientations could simply reflect preferential development of different regionally conjugate sets, and the dikes may even postdate motion of the Okanagan Valley fault system.

Paleomagnetic studies of the dikes could test whether or not the difference in orientations has anything to do with rotation. With regard to the second model cited above, further detailed mapping within the transfer zone may uncover more substantial evidence for, or against, the existence of a transfer fault. In any case, synextensional magmatism is considered to have played a key role in development of the transfer zone.

Carr (1992) has argued that a southwest-dipping ramp in the Monashee décollement (Brown et al. 1992a, Cook et al. 1992) influenced the production and distribution of Paleocene-Eocene leucogranites near the south end of the Monashee complex (Fig. 2.1). Tertiary leucogranites and pegmatites are rare in the footwall of the Monashee décollement; in contrast, pegmatite forms up to half the volume or more of the upper plate of the décollement beneath major normal faults (Read and Brown 1981; Journeay and Brown 1986; Journeay 1986; Scammell 1985, 1986, 1990, 1993; Sevigny and Ghent 1986; Carr 1989, 1991, 1992; Johnson 1989, 1990; Coleman 1990) and Paleocene to Eocene leucogranites are abundant in many parts of this structural panel (Carr et al. 1987; Carr 1986, 1989, 1990, 1992; Parrish et al. 1988; Sevigny et al. 1989; Murphy 1990b; Scammell 1992, 1993; Bardoux 1993). Granitic melts are likely to have contributed to the nucleation of extensional shear zones in the middle crust during the Paleocene and Eocene along a magmatic front (Carr 1992) that would have trended approximately northwestward, parallel to the ramp in the Monashee décollement. Obliquity of this magmatic front to the approximately west-northwest direction of regional crustal extension, and hence to the northerly strike of normal faults that formed in the brittle upper crust (Fig. 2.1), may have had an influence on the left-stepping geometry of the Shuswap Lake transfer zone. In addition, the relative strength of the melt-free rocks in the footwall of the ramp may have inhibited the

propagation of extensional shears into the Monashee complex.

Development of the transfer zone was probably dynamically linked to differential uplift of the Pukeashun shear zone relative to adjacent parts of the footwall of the system. Spencer (1984) has shown that isostatic rebound of the footwall of a low-angle normal fault, in response to tectonic unloading, can lead to pronounced warping of the fault if tectonic attenuation within the hanging wall is heterogeneously distributed (see also Fig. 5.8, Chapter 5). The crest of the antiformal warp should occur beneath the locus of most extreme hangingwall extension, and the degree of arching will be greatest for faults that have a listric breakaway. Whether the Pukeashun shear zone had a listric breakaway is uncertain but is permissible, and extreme attenuation of the hanging wall is likely to have occurred above synextensional magmas of the Pukeashun granite. Arching would be expected to be further enhanced if the zone into which the Pukeashun granite was emplaced was initially very gently dipping or subhorizontal. The main point here is that extreme thinning of the hanging wall of the Pukeashun shear zone could have led to more rapid exhumation and uplift of the Pukeashun granite than of adjacent parts of the footwall, and that this would have in turn driven the processes of excisement and transfer faulting.

CONCLUSIONS

1. The Okanagan Valley fault system delimits the western margin of the Shuswap metamorphic complex from northern Washington to Adams Lake. From there it crosses the metamorphic culmination and continues northward for a total

strike length of over 450 km.

2. Along this entire length, footwall rocks of relatively high metamorphic grade are juxtaposed against lower-grade hangingwall rocks.
3. From the Okanagan Valley to at least as far north as Adams Lake, mylonites in the footwall exhibit asymmetric shear fabrics that consistently indicate relative westward motion of the upper plate.
4. The Okanagan - Eagle River and Adams - North Thompson segments of the system are delineated by west-dipping ductile-brittle normal faults. Between them is the Shuswap Lake transfer zone, within which there is no obvious brittle detachment, but across which mylonitic footwall rocks have been exhumed. Ductile shearing in the transfer zone was localized and accommodated by early Tertiary two-mica granites.
5. Mylonitic foliation in the granite dips southeastward across much of the 45-km width of the transfer zone. Several kinematic models that invoke tilting or warping are compatible with this geometry. Interpretations other than that shown in Figure 2.5 *b* predict that one or both of the en échelon fault segments would have terminations that have yet to be documented. The models can be tested by further mapping, thermobarometric studies, and more comprehensive geochronometry.

Figure 2.1: Tectonic assemblage map of the southeastern Canadian Cordillera, simplified from Wheeler and McFeely (1991) and modified based on Carr (1991) and this study. Shuswap core complex as defined by Brown and Carr (1990) is mostly bounded by Eocene normal faults (thick black lines with barbs on hanging wall side), including the Okanagan Valley (OV), Columbia River (CR), and Slocan Lake - Champion Lake (SL) fault systems. Okanagan - Eagle River and Adams - North Thompson segments of OV step en échelon across a transfer zone west of Shuswap Lake. The OV system transects the sillimanite isograd (Sil; from Read et al. 1991) north of Adams Lake. Southern Rocky Mountain Trench (SRMT) is a morphogeological lineament that defines the boundary between Omineca and Foreland belts. Black lines of medium weight represent mainly northeast-verging thrusts, except along the boundary between the Coast and Intermontane belts where the Fraser - Straight Creek strike-slip fault system is designated by opposing arrows. Southern Rocky Mountain Trench (SRMT) is a morphogeological lineament that defines the boundary between Omineca and Foreland belts. Unpatterned units in these belts are Mesoproterozoic to early Tertiary miogeoclinal and platformal strata. Unpatterned units in the Coast belt are Permian to Cretaceous marine sedimentary and volcanic rocks and metamorphic rocks of uncertain age. Jurassic and Paleozoic granitoid plutons in Omineca and Intermontane belts, and Cretaceous and Neogene volcanic rocks in Intermontane belt, are not shown (see Fig. 5.1, Chapter 5). Bv = Beaven fault, M = Malton complex, Mo = Monashee complex, V = Valhalla complex.

Figure 2.2: Geological map of the Shuswap Lake area (compare with central part of Fig. 2.1 for location). Shuswap Lake transfer zone is left step between the en échelon Okanagan - Eagle River and Adams - North Thompson segments of the Okanagan Valley fault system. Circled dots represent localities where reliable indicators of west-directed shear sense have been documented in mylonitic rocks. Locations of cross sections shown by grey lines labelled A-A' (Plate 2a), C-C' (Plate 2c and Fig. 2.4), D-D' (Plate 2d) and E-E' (Plate 2e and Fig. 2.5). C: Crowfoot Mountain; CrC: Craigellachie Creek; CRF: Columbia River fault; CsC: Cross Creek; F: Mount Fowler; FmC: Four Mile Creek; HR: Hunters Range; KwC: Kwikwakaikwet Creek; L: Lichen Mountain; MD: Monashee décollement; P: Pukeashun Mountain; PeR: Perry River; Sil: sillimanite isograd (coincides with Okanagan - Eagle River and Adams - North Thompson faults where not shown); SR: Shuswap Range; TvL: Three Valley Lake; ViC: Victor Creek. Maple leaf symbol designates Trans-Canada Highway. Geology compiled from this study and from Johnson (1990), Okulitch (1979), Schiarizza and Preto (1987), Bardoux (1993), Wheeler (1965), Höy and Brown (1981), Journeay (1986), Journeay and Brown (1986), Bosdachin (1989), Harrap (1990), Coleman (1989), and Crowley (1992).

Figure 2.3: Lower hemisphere equal area stereograms of stretching lineations in mylonitic rocks of the footwall of the Okanagan Valley fault system. Data are from the Oliver (Parkinson 1985), Kelowna (Bardoux 1993), Eagle River and Pukeashun areas (this study).

Figure 2.4: Cross section through the Mara Lake area (location, shown as C-C' in Figure 2.2, is same as that of Plate 2c). Okanagan - Eagle River detachment (O-ER) is inferred as being subparallel to the west-dipping mylonitic foliation in the immediate footwall; double half-arrows indicate apparent dips of foliation at localities of documented west-directed shear sense. Lower plate (Shuswap complex; unpatterned except for marble-calcsilicate marker unit) consists of migmatitic sillimanite-bearing metasediments, amphibolite, and pegmatite. Silver Creek assemblage in upper plate comprises garnet-biotite-muscovite schists, micaceous quartzite and minor carbonate.

Figure 2.5: Kinematic models for the geometric relationships between the Okanagan - Eagle River and Adams - North Thompson (A-NT) segments of the Okanagan Valley fault system and the Pukeashun leucogranite mylonites (Pukeashun shear zone), based on a composite geological section across the Shuswap Lake transfer zone constructed from sections shown in Plates 2e and 2f. Representative apparent dips are shown with short lines; double half-arrows indicate northwest-directed shear sense. High-angle brittle faults cut the mylonites and either feed into or cut the brittle detachment in *a*, *d*, and *e*. (*a*) Pukeashun mylonites are product of northwest-directed backthrusting in the hanging wall of the Monashee décollement (MD); (*b*) O-ER, the Pukeashun granite, and A-NT are part of a contiguous ductile-brittle normal fault system; (*c*) Pukeashun granite mylonites are a distal part of the O-ER that has been subsequently cut by the A-NT; (*d*) Pukeashun mylonites are the excised updip part of the A-NT, folded into a rollover antiform in the hanging wall of O-ER; (*e*) the Pukeashun granite is an abandoned part of the evolving system which has been excised by the A-NT and incised by the O-ER; this is the favoured model (see text). If a steep strike-slip transfer fault links the O-ER and A-NT (see text), then the Pukeashun granite would lie deeper relative to the topographic surface and to the Mount Fowler gneiss than shown, and in models *a*, *d*, and *e* it may lie further west than shown.

WNW

ESE

a

b

c

d

e

Plate 4: Fault rocks and mylonites of the Okanagan - Eagle River fault zone. Cut rock specimens *e*, *f* and *g*, and photomicrograph *h* are oriented so that west is to the left.

(*a*) Southwest-dipping clay gouge zones in fractured footwall gneiss, ~80 m below detachment near Four Mile Creek. Hatchet handle is about 30 cm long. [Station 397].

(*b*) Cut specimen of cataclastic breccia derived from silicified migmatite or granitoid mylonite, Okanagan - Eagle River fault zone near Four Mile Creek. Zones of intense cataclasis appear dark. Scale bar 1 cm. [Sample J88-397E].

(*c*) Photomicrograph of cataclastic breccia shown in (*a*). Silicified and sericitized lenses of mylonite are separated by anastomosing seams of graphite and hematite (dark). Plane-polarized light; scale bar 300 μm .

(*d*) Photomicrograph of crush zone in granodiorite gneiss, from ~10 m above detachment at Four Mile Creek. Biotite is mostly replaced by chlorite (Chl). Crossed nicols, $1/4 \lambda$ plate inserted; scale bar 300 μm . [Sample J88-398A].

(*e*) Migmatitic semipelitic mylonite from peninsula on west side of Mara Lake. Feldspar porphyroclasts (white), ribbon quartz layers (light grey), and matrix are cut by veins of quartz (Qtz) and quartz-hematite (QHem). Scale bar 1 cm. [Sample J88-510].

(*f*) Migmatitic pelitic mylonite showing asymmetrical extensional shear bands (C'), σ -type feldspar porphyroclasts (arrows), and asymmetric west-verging microfolds. Scale bar 1 cm. From northwest flank of Hunters Range [Sample J88-240].

(*g*) Mylonitic pelitic schist from east side of Mara Lake. Sillimanite forms coarse mats (Sil) adjacent to pressure shadows of quartz + plagioclase + chlorite on garnet porphyroblast, and also forms finer-grained blades (light-coloured in matrix) that along with biotite define the foliation. Scale bar 1 cm. [Sample LJ89-19A].

(*h*) Prismatic sillimanite (Sil) in pressure fringe adjacent to K-feldspar porphyroclast (Kfs), in migmatitic pelitic mylonite from the Eagle River valley. Fine-grained recrystallized feldspar (f) forms thin mantle around porphyroclast and forms a "wing" adjacent to its right margin, and is intergrown with fine-grained white mica (m) in pull-apart region between porphyroclast and fringing sillimanite. White mica also occupies transverse fractures between segments of sillimanite bundle. Matrix consists of light-coloured quartzofeldspathic layers and darker-coloured layers of intergrown fibrous sillimanite, biotite and opaque minerals. Grey transverse line across central part of photo is a scratch in the glass slide. Plane-polarized light; scale bar 300 μm . [Sample J88-385].

Plate 5: Field photos (*c* and *f*) and photomicrographs showing mylonitic fabrics. West is to the right in *c* and *f*; all photomicrographs are oriented so that west is to the left. In each photo, the shear plane (C) is subhorizontal and the flattening foliation (S) dips eastward, as labelled in *a* and *f*.

(*a*) C/S fabric in migmatitic pelite mylonite from Eagle River valley near Cambie. C surfaces defined by discrete layers of intergrown biotite and sillimanite (dark) parallel to length of photo; S defined by σ -type feldspar porphyroclasts (F) and by quartz ribbons (Qtz) separated by intergrown biotite and sillimanite. Plane-polarized light; scale bar 600 μm . [Sample J89-938].

(*b*) C/S fabric in migmatitic pelite mylonite from northwest flank of Hunters Range. S defined by long axes of muscovite fish (Ms), σ -type feldspar porphyroclasts (F), and quartz ribbons (Qtz); C surfaces are parallel to length of photo. Plane-polarized light; scale bar 600 μm . [Sample J88-240].

(*c*) Migmatitic pelite mylonite, east side of Mara Lake. Asymmetrical extensional shear bands cut across layering from left centre (parallel to white line) to lower right. Pen is 15 cm long. [Station 232/22].

(*d*) Oblique strain-insensitive foliation in dynamically recrystallized quartz, muscovite granite mylonite from west of Seymour Arm. C surfaces defined by thin sericite residues (ser) are parallel to length of photo. Crossed nicols; scale bar 150 μm . [Sample J90-1097-2].

(*e*) Delta(δ)-type quartz porphyroclast in migmatitic pelite mylonite, from northwest flank of Hunters Range. Plane-polarized light; scale bar 300 μm . [Sample J88-291].

(*f*) Feldspar porphyroclast system of δ type that has been modified such that σ -type tails have grown parallel to the flattening plane of finite strain (S). Coin is 19 mm in diameter. Trans-Canada Highway near Sicamous. [Station 174].

(*g*) Asymmetric west-verging microfold in quartz ribbon, migmatitic pelite mylonite from Eagle River valley near Cambie. Long axis of σ -type feldspar porphyroclast (F) parallels S fabric. Plane-polarized light; scale bar 600 μm . [Sample J89-938].

(*h*) Fabric in migmatitic pelite mylonite from east side of Adams Lake. Sigma(σ)-type feldspars (F), oblique quartz foliation (Qtz), mica fish (Bt = biotite fish), and incipient asymmetrical extensional shear bands (C') indicate west-directed shear sense. Crossed nicols, $1/4 \lambda$ plate inserted; scale bar 600 μm . [Sample J88-373A].

Plate 6: Strain in Pukeashun granite and Mount Fowler orthogneiss.

(a) Sigma(σ)-type plagioclase porphyroclast in biotite granite mylonite from east side of Adams Lake. Adjacent quartz ribbons (Qtz) display oblique, east-dipping, strain-insensitive foliations. Crossed nicols, $1/4 \lambda$ plate inserted; scale bar 300 μm ; west is to left. [Sample J88-373G].

(b) Polished slab of muscovite granite mylonite, from the Pukeashun suite near Adams Lake, showing well-developed C/S fabric. C defined by discrete shears of white mica parallel to length of specimen; S defined by muscovite fish (Ms) and σ -type feldspar porphyroclasts (white) separated by quartz ribbons (light grey). Scale bar 1 cm; NW is to left. [Sample J89-725].

(c) Photomicrograph showing C/S fabric in specimen shown in (b). C (parallel to length of photo) is defined by very fine-grained white mica and dynamically recrystallized quartz, S by long axes of K-feldspars. Recrystallized quartz domains (Qtz) between C planes exhibit strain-insensitive foliation subparallel to S. Crossed nicols; scale bar 600 μm ; NW is to left.

(d) Sigma-type K-feldspar porphyroclast between C planes defined by finely recrystallized white mica and ribbon quartz, in biotite-muscovite granite mylonite from west of Pukeashun Mountain. Crossed nicols, $1/4 \lambda$ plate inserted; scale bar 600 μm ; NW is to left. [Sample J89-719].

(e) 12-cm-wide dike of muscovite granite cuts pegmatite in central part of photo (long arrow); both are moderately foliated. Thin tails of the dike are transposed into parallelism with foliation in migmatite mylonites in lower left (short arrow). Pukeashun Mountain.

(f) Detail of tails of the dike in (e), showing C/S fabric as in (b); however, NW is to right in this photo. Coin is 19 mm in diameter. [Station 764].

(g) Discrete shear defined by recrystallized white mica and chlorite, in muscovite granite mylonite of Pukeashun suite from near Adams Lake. Crossed nicols; scale bar 150 μm ; NW is to left. [Sample J89-723].

(h) Brittle shears in Mount Fowler orthogneiss near Kwikoit Creek. SE-side-down normal slip is indicated by oblique biotite foliation parallel to knifeblade. Knife handle is 8 mm wide. [Station 628].

CHAPTER 3

Stratigraphy and structure of the west-central margin of the Shuswap complex near Shuswap Lake

INTRODUCTION

This chapter presents the stratigraphic and structural results of a 1:50 000 scale mapping program that focused initially on a transect near the Trans-Canada Highway (Johnson 1988), through the part of the Shuswap complex that lies structurally above the Monashee décollement and beneath the Okanagan - Eagle River fault (Fig. 3.1). Further mapping followed the boundary between high-grade and low-grade rocks northwestward to Adams Lake (Johnson 1989, 1990, 1991). This work, compiled at 1:100 000 scale in Plate 1, has enabled characterization of stratigraphic successions and structural style in both the footwall and the hanging wall of the Okanagan Valley fault system, which in turn allows for comparison and preliminary correlations with strata in other parts of the southern Omineca belt.

GEOLOGICAL SETTING

The complex tectonic history of the southern Omineca belt, which involved Jurassic accretion of outboard terranes, continued Cretaceous to Paleocene thrusting

and crustal thickening, early Tertiary extensional denudation of the Shuswap complex, and several episodes of Paleozoic to Tertiary magmatism, is summarized in Chapter 1. The Monashee complex (Fig. 3.1), which comprises Paleoproterozoic basement gneisses and a thin Neoproterozoic to possibly mid-Paleozoic metasedimentary cover succession (Chapter 5 and references therein), occupies the structurally deepest part of the Shuswap core complex exposed in this region. Rocks in the Shuswap Lake area that lie within the Selkirk allochthon, structurally above and to the west of the Monashee complex, have been assigned to eight tectonostratigraphic assemblages and two plutonic suites that are described herein.

The assemblages are distinguished on the basis of lithologic, metamorphic, structural or isotopic characteristics and by structural position relative to tectonic boundaries (Plate 1). Three of these assemblages (*Three Valley*, *Hunters Range* and *Momich*) consist of amphibolite-facies rocks within the Shuswap complex and are defined herein. A fourth newly defined assemblage (*Lichen*) consists of amphibolite-facies rocks that structurally overly the *Pukeashun granite suite*, which occupies the ductile shear zone of the Okanagan Valley fault system in the Shuswap Lake transfer zone; these rocks are metamorphically distinct from the structurally higher *Eagle Bay* assemblage (Schiarizza and Preto 1987), from which they are separated by the *Mount Fowler plutonic suite* (Okulitch et al. 1975). Finally, three units (*Silver Creek*, *Tsalkom* and *Sicamous* formations of Jones 1959) that structurally underly the Eagle Bay in the southern part of the area are considered as separate assemblages because of suspect tectonic contacts between them. These eight assemblages constitute part of the pericratonic Kootenay terrane (Wheeler et al. 1991), to which they have been assigned on the basis of assumed kinship with miogeoclinal strata of ancestral North America. Such kinship has only been demonstrated for the Eagle Bay assemblage, which

contains an archaeocyathid fauna characteristic of Lower Cambrian units of the miogeocline (Schiarizza and Preto 1987).

LITHOLOGY

Tectonostratigraphic assemblages and formations are described below in order of structural stacking above the cover assemblage of the Monashee complex. No complete sections of any of the assemblages are exposed, and contacts between different units are rarely exposed. Therefore, thicknesses of all but the thinnest units are estimated from cross sections (Plates 2a-g). Map units are identified by abbreviations that correspond to those on Plate 1. In most cases these are lithological rather than stratigraphic units.

Three Valley assemblage

Characterized by migmatitic, semipelitic to psammitic paragneisses with boudins of gneissic clinopyroxene- and garnet-bearing amphibolites, the Three Valley assemblage (**TVsa**) is spectacularly exposed along the Trans-Canada Highway at Three Valley Gap (Plate 1). Paragneiss and amphibolite, along with abundant pegmatite and minor diopside-bearing psammite, are exposed continuously for 700 m vertically in cliffs above the highway. Two km east of Three Valley Lake and structurally beneath the amphibolite-bearing unit are migmatitic pelites that contain coarse sillimanite and garnet.

All of the rocks within the assemblage are highly strained. The contact between the Three Valley assemblage and the underlying Monashee complex is a

zone, up to 30 m thick, in which mylonitic rocks of both units are chaotically intermixed and cut by anastomosing shears (Bosdachin and Harrap 1988, Bosdachin 1989, Harrap 1990). On the Trans-Canada Highway, contact relations are obscured by a high-angle brittle normal fault at Victor Creek (Plate 1) which juxtaposes fine-grained mylonitic migmatites of the Three Valley assemblage against Monashee quartzites and calcsilicate gneisses (Brown et al. 1993b). The zone of high strain in which mylonites are developed and units are chaotically sheared is the Monashee décollement.

The top of the Three Valley assemblage is difficult to define because distinction between this assemblage and the overlying Hunters Range assemblage is largely based on isotopic differences. Lithologically, the Hunters Range assemblage is similarly dominated by semipelitic paragneisses, but amphibolite seldom forms the large boudins characteristic of Three Valley. On Eagle Pass Ridge, 3 km north of Three Valley Gap, a zone of high strain up to 300 m thick marks the transition from paragneisses with abundant boudins of amphibolite to those with relatively thin mafic layers typical of the Hunters Range. Within the high-strain zone, the most abundant rock is pegmatite, which exhibits fabrics indicative of top-to-the-northeast and minor superimposed top-to-the-southwest shearing. Screens of white to brown diopsidic marble and green calcsilicate gneiss occur within the pegmatite. Bosdachin (1989) noted that a thin feldspathic quartzite generally underlies the marble, and that this association persists for several km to the north. On Mount English, 7 km southeast of Three Valley Gap, Harrap (1990) observed a similar zone of sheared pegmatite and marble at about the same structural level. This shear zone is provisionally designated as the boundary between the Three Valley and Hunters Range assemblages.

Hunters Range assemblage

Metasedimentary successions in the Hunters Range are dominated by migmatitic garnet-biotite semipelite and feldspathic psammite, with locally abundant pelitic garnet-sillimanite-biotite schist (unit **Hs**). The pelites are black to light purple-grey and commonly weather light orange-brown. They form mm- to m-scale interlayers in psammites and semipelites and typically contain abundant cm-scale leucosome layers (Plate 7 *a-c*). The white to light grey psammites and light grey to light rusty brown semipelites are commonly intergradational. Calcsilicate and marble units (**Hc** and **Hm**) are well exposed along Mara Lake, in the Mount Mara area, and on ridges east of Kingfisher Creek and are associated with distinctive layered amphibolites (unit **Ha** and in unit **Huc**). The northeastern Hunters Range is dominated by grey, medium-grained, garnet-hornblende-quartz-biotite-plagioclase gneiss and biotite-hornblende dioritic to granodioritic gneiss (collectively called the Griffin gneisses; **Hh** and **Hdh**), which are layered but are quite homogeneous (Plate 7 *h*, Plate 8 *a*). They are predominantly orthogneisses, but some of the garnetiferous layers of unit **Hh** could be calcareous metasediments. All units of the Hunters Range assemblage are extensively intruded by sheets and dikes of pegmatite, which typically form over 50% of exposure.

Partial stratigraphic successions have been mapped on the highest ridges of the Hunters Range (Plate 1). Correlation between ridges is facilitated by structural cross sections (Plates 2a, 2b) and by comparison of similar successions on different ridges. Exposures at lower elevations were also mapped but were less useful for stratigraphic purposes.

In the Mount Mara area, a section about 1500 m thick consists of psammitic to pelitic paragneisses of unit **Hs** punctuated by a 300-m thick amphibolite unit (**Ha**) and

a calcareous unit (**Hc**, **Hm**) up to 25 m thick. Paragneisses of unit **Hs** here, and in many other parts of the Hunters Range, contain dm-scale layers and boudins of garnet amphibolite and garnet-biotite-hornblende schist (Plate 7 *a, d, e*). The amphibolites of unit **Ha** are characterized by black hornblende-biotite(-garnet) layers and green diopside-plagioclase layers, which alternate on a scale of a few mm to a few cm (Plate 7 *f, g*). On outcrop scale they are interlayered with semipelite, psammite and minor pelite like those of unit **Hs**. The structural top of unit **Ha** is marked by a thin, discontinuous layer of brown diopsidic marble and calcsilicate. About 50-100 m above unit **Ha** is a discontinuous calcareous unit (**Hc**) which consists of rusty semipelitic paragneiss, amphibolite, and diopsidic calcsilicate gneiss interlayered with and laterally gradational into impure brown marble. A marble/calcsilicate layer up to 5 m thick marks the top of this unit and is shown as **Hm** on the map (Plate 1) and cross sections.

The Mount Mara section is traceable discontinuously as far east as Kingfisher Creek (Plate 1). Similar lithological assemblages occur on the west flank of the Hunters Range and along Mara Lake. Structural projections suggest that marble and calcsilicate exposed in many scattered outcrops near Mara Lake are along strike from one another. These units, including marble and amphibolite at Black Point on the west shore of Mara Lake, may be correlative with units at Mount Mara (Plates 2c and 2b).

East of Kingfisher Creek, remnants of brown to light grey marble and calcsilicate are engulfed in pegmatite. These may be correlative with the calcareous unit of the Mount Mara section or may be structurally higher. They are associated with psammitic to pelitic paragneisses, biotite-hornblende amphibolite, garnet-quartz-biotite-hornblende-plagioclase gneisses like unit **Hh**, and rare quartzite layers a few

cm thick. These units are poorly exposed between Kingfisher Creek and Blue Lake, 5 km to the east. At Blue Lake, a structurally higher succession over 800 m thick consists of psammitic to pelitic paragneiss, an 85-m-thick layer of garnet-quartz-biotite-hornblende-plagioclase gneiss (**Hh**), and several marble units interlayered with green diopsidic calcsilicate gneisses, rusty semipelite, mafic biotite-hornblende-plagioclase gneiss, layered diopside-hornblende amphibolite, and rare hornblendite and quartzite. The marble units are 1-10 m thick and are mainly impure brown marbles (Plate 8 *b*) containing diopside, graphite, quartz and biotite; white to light grey calcite marbles are less common. At the top of the section, amphibolite is overlain by a highly strained and isoclinally folded light brown to grey marble more than 100 m in structural thickness. This is the structurally highest unit documented in the Hunters Range assemblage.

On a ridge 5 km northeast of Blue Lake are calcareous units (including marble, calcsilicate and amphibolite collectively mapped as **Huc**) and psammitic to pelitic paragneisses (**Hs**) that are inferred to be correlative with some of the Blue Lake strata. Capping the ridge is a layer of garnet-quartz-biotite-hornblende-plagioclase gneiss (**Hh**) at least 30 m thick. A similar and probably correlative succession occurs 9 km to the north across Yard Creek, on a ridge that extends westward from Mount Griffin. There the calcsilicate/amphibolite unit is underlain by more paragneiss (**Hs**) and dioritic gneisses of unit **Hdh**.

Momich assemblage

High-grade rocks exposed along the northeast side of Adams Lake and along the Adams River are provisionally named Momich assemblage after the Momich River (unit **M**). These include hornblende-biotite quartz diorite gneiss, pelitic to semipelitic

muscovite-sillimanite-biotite schist, quartzofeldspathic paragneiss, and diopside-bearing amphibolite. They are much like rocks of the Hunters Range and may be correlative, but are classified as a separate assemblage because a 50-km wide zone between the two areas remains to be mapped in detail.

The Momich assemblage is intruded by voluminous pegmatite and by leucogranites of the Pukeashun suite. The contact with the main body of the overlying sheetlike Pukeashun granite (unit **KTg**) is gradational where it has been mapped, marked by a transition into granite with screens of metamorphic rock.

Lichen assemblage

Metasedimentary rocks that lie structurally above the Pukeashun granite and that are isolated from the Eagle Bay assemblage by the intervening Mount Fowler orthogneiss (**Dg**) are here called the Lichen assemblage after Lichen Mountain. The assemblage consists mainly of semipelitic to pelitic (sillimanite-garnet-)biotite-muscovite schists with thin interbeds of feldspathic psammite and quartzite (**Ls**). Graded bedding was observed at three localities but the rocks are tightly folded and no regional sense of stratigraphic facing could be determined. Light bluish grey to light brown calcsilicate gneiss and impure marble (**Lc**) are additional components of the assemblage.

Eagle Bay assemblage

In the Shuswap Range north of Sicamous, the Eagle Bay assemblage consists of a tripartite stratigraphic succession (units **Ec**, **Em** and **Eq**) that is structurally repeated by folds and possibly by thrusts, and within which stratigraphic tops have yet to be identified. The most distinctive unit is **Em**, a clean white to grey marble that

ranges from 0-200 m in structural thickness.

Unit **Ec** is a heterogeneous assemblage of thinly interbedded greyish brown to light greyish green epidote-bearing calcsilicate schist, fine-grained grey and brown marble, quartz-sericite-chlorite phyllite, grey fine-grained quartzite and metasiltite, semipelitic to pelitic biotite-muscovite schists, and minor white quartzite. Epidote-biotite-chlorite-amphibole schists and massive green metabasites are abundant and in some places dominant (unit **Eca**). These locally crosscut calcareous strata and are interpreted as mafic metavolcanic rocks and related shallow intrusions. Structurally below the Mount Fowler orthogneiss in the central and northwestern Shuswap Range is a poorly exposed heterogeneous assemblage designated as unit **Ecq**, which comprises white micaceous and feldspathic quartzites, pelitic to semipelitic (garnet-) biotite-muscovite schists, calcsilicate schists, siliceous and calcareous grey metasiltites, and orange-pink felsic schists that probably are metavolcanic.

Unit **Eq** is a thick succession of light brown to greenish grey, fine to medium-grained, micaceous quartzites interbedded with pelitic to semipelitic biotite-quartz-muscovite schist and minor feldspathic granule conglomerate. In intervals where pelite is abundant, the quartzite layers weather white to light grey. Where composition and metamorphic grade are appropriate, the schists contain porphyroblasts of garnet and staurolite. Mafic metavolcanic rocks like those of unit **Ec** occur within the quartzite succession, especially on the flanks of North Queest Mountain and near the contact with unit **Em**. Rusty yellow and dark greenish grey-weathering sericitic phyllites also are common near this contact. Because similar rocks occur in unit **Ec**, in areas of poor exposure some rocks mapped as **Eq** could be **Ec**. The stratigraphic thickness of unit **Eq** is unknown. Structural thicknesses of more than 300 metres with no obvious repetitions are well exposed east of Queest Mountain and on the southeast

flank of North Queest Mountain; the total thickness of the unit is likely at least twice this estimate.

In the Kwikoit Creek drainage east of Adams Lake, unit **Ec** has been subdivided into sericite-quartz and pyrite-graphite-sericite phyllites (**Ecp**), grey marble and phyllitic marble (**Ecm**), and chlorite schists that probably are mafic metavolcanic rocks (**Eca**). Unit **Ec** is structurally overlain in the Kwikoit Creek area by quartzites, metasiltites, quartz-sericite phyllites, and feldspathic granule conglomerate mapped as **Eq**, which are not necessarily correlative with unit **Eq** in the Shuswap Range.

Silver Creek assemblage

The Silver Creek assemblage is the structurally lowest of three units in the Larch Hills west of Mara Lake, in the hanging wall of the Okanagan - Eagle River fault. It consists mainly of semipelitic to pelitic quartz-muscovite and garnet-biotite-quartz-muscovite schists, interlayered with micaceous and feldspathic quartzites (unit **SCq**). The quartzites contain micaceous partings spaced a few mm to several cm apart. Minor biotite-amphibole schist (**SCa**) and sericite-epidote-quartz-calcite paragneiss interlayered with greyish pink marble (**SCc**) occur near the contact with the overlying Sicamous Formation. Impure greyish brown marble locally occurs within unit **SCq** and was not mapped separately.

Near the southwest end of Mara Lake, schists like those in unit **SCq** are interlayered with calcsilicate, marble and biotite-hornblende amphibolite. The calcareous rocks contain biotite (or phlogopite), tremolite, and locally diopside as a replacement product of tremolite. On Plate 1, the units with abundant calcsilicate and marble and with abundant amphibolite are designated as units **SCc?** and **SCa?**, respectively. Nielsen (1978, 1982) interpreted these units as being correlative with

other Silver Creek rocks to the north, but displaced relative to them along a low-angle fault. He also proposed that they were correlative with calcareous units and amphibolites in the Hunters Range, across Mara Lake to the east. At an early stage in the present investigation, they were interpreted as part of the footwall of the Okanagan - Eagle River fault, exposed in a small antiformal culmination (Johnson 1989).

However, these rocks do not exhibit the gently dipping mylonitic foliation characteristic of rocks in the immediate footwall of the fault, but rather are folded by steep-limbed upright structures. Moreover, the pelitic schists contain abundant muscovite and lack sillimanite, in contrast to the sillimanite-K-feldspar migmatites of the western Hunters Range. A preferable alternative interpretation is that units **SCa?** and **SCc?** constitute a slice that is bounded below by the Okanagan - Eagle River fault and above by a low-angle splay (the low-angle fault mapped by Nielsen). Correlation of these rocks with the Silver Creek assemblage seems reasonable, whereas correlation with the Hunters Range assemblage immediately east of Mara Lake is not demonstrable and probably is incorrect.

Sicamous and Tsalkom formations

The Sicamous Formation (**Sc**) consists of grey, fine-grained, phyllitic marble and calcareous and carbonaceous phyllites. Typically the marbles contain lenses and discontinuous layers of coarse white calcite. Mafic metavolcanic rocks and chlorite-sericite phyllites of the Tsalkom Formation (**Ta**), which discontinuously and probably unconformably underly the Sicamous Formation regionally (Campbell and Okulitch 1973, Okulitch 1979), were observed only in a few isolated outcrops in this study.

Nielsen (1978) mapped an angular discordance between the Sicamous Formation and the Silver Creek assemblage which he interpreted as evidence for a

tectonic contact. From the present study, further evidence for a fault contact is that the uppermost part of the Silver Creek, on the Trans-Canada Highway 2.5 km west of the town of Sicamous, is cut by zones of graphitic clay gouge several metres thick that are subparallel to layering. Both the Silver Creek and the overlying Sicamous Formation are cut by steeply northeast-dipping brittle faults. Whether this contact is everywhere a fault and whether it is a thrust or a normal fault is yet unresolved.

The contact between the Sicamous Formation and chloritic phyllites and metavolcanic rocks of the Eagle Bay assemblage is marked by gradational interlayering, which has been taken as evidence that these units are conformable (Okulitch 1974, K. L. Daughtry pers. comm. 1993).

Mount Fowler orthogneiss suite

Granitoid gneiss of the Mount Fowler suite (unit **Dg**) forms extensive discordant to subconcordant sheets within the Eagle Bay assemblage and between the Eagle Bay and Lichen assemblages. The gneiss ranges in composition from quartz diorite to monzogranite and is typified by medium to coarse-grained biotite granodiorite to tonalite gneiss. Biotite (plus locally hornblende and epidote) constitutes up to 25% of the rock and commonly forms thin lenticular aggregates. Highly sheared gneiss in the southern Shuswap Range contains muscovite and chlorite as well as biotite.

Rare intrusive relationships have been observed between the Mount Fowler gneiss and both the Eagle Bay and Lichen assemblages, generally consisting of screens of country rock in the gneiss or sills of orthogneiss in country rock. Calcsilicate rocks of unit **Ec** near covered contacts with the gneiss commonly contain quartz-calcite-actinolite-epidote skarn zones interpreted as resulting from contact metamorphism. To

what degree the relatively straight map-scale contacts of the gneiss have been modified by shearing is uncertain.

The Mount Fowler gneiss is considered to be of Devonian age, based on a U-Pb concordia upper-intercept date obtained by regression of data from four zircon fractions collected from different localities (Okulitch et al. 1975). This technique is not considered satisfactory for precise age determination, but the close petrological resemblance of the Mount Fowler to other Devonian gneiss suites of the Cordillera argues in favour of a Devonian age. Biotite granodiorite gneiss at Celista Mountain in the southern Seymour Range looks identical to the Mount Fowler gneiss, and a similar granodiorite gneiss further north in the Seymour Range has yielded a Devonian-Mississippian U-Pb zircon date (Parrish 1992).

Pukeashun granite suite

The Pukeashun granite suite, unit **KTg**, consists of multiple generations of medium-grained leucogranite and pegmatite that intrude, and engulf screens of, metamorphic rocks of the Momich and Lichen assemblages (Chapter 2). The contacts shown on the map are approximate, marking the transition from mainly granite to mainly metamorphic rocks with 50% or less granite sheets and dikes. Monzogranite of unit **KTgm**, well exposed at Pukeashun Mountain, contains 10-15% muscovite and generally less than 5% biotite and garnet. Medium to coarse-grained biotite monzogranites of unit **KTgb** form plutons and sheets north of Four Mile Creek. An age of 56 ± 1 Ma has been indicated for a granite dike at Pukeashun Mountain from U-Pb zircon and monazite geochronology (sample J89-762, Chapter 4). The dated dike is one of the younger components of the suite at the sample locality, and the ages of older components are uncertain. Petrologically similar granites at deeper structural

levels to the north range in age from Early Cretaceous to Eocene (Sevigny et al. 1989; Scammell 1993; R. R. Parrish and J. M. Journeay unpublished data 1989), and hence the Pukeashun suite could include components as old as Cretaceous.

Other intrusive rocks

Dikes and sills of medium-grained biotite leucogranite like **KTgb** intrude quartz-biotite-hornblende-plagioclase gneisses of unit **Hh** at Mount Griffin in the Hunters Range. Small plutons of fine to medium-grained leucogranite, mapped as **T(?)g**, intrude the Eagle Bay assemblage and the Mount Fowler gneiss in the Shuswap Range. These contain virtually no varietal minerals, except for small weathered pyrites, and are unstrained.

The Hunters Range, Momich, and Lichen assemblages are extensively intruded by sheets, dikes and stocks of pegmatite that commonly form over 50% of the exposure and that vary in texture from undeformed to mylonitic. Pegmatite and aplite also are common in the Silver Creek assemblage.

Medium to coarse-grained diabase (**Ti**) forms small stocks in the Eagle Bay and Lichen assemblages east of Adams Lake. These rocks are highly altered (Johnson 1990) but are not strained. Medium-grained biotite quartz diorites, also undeformed, intrude the Lichen assemblage north of Kwikoit Mountain and were not mapped separately. They may be related to swarms of lamprophyre and K-feldspar porphyry dikes that intrude other rocks of the Shuswap complex and, at Mount Fowler, orthogneiss of unit **Dg**. The dikes strike north-northwest in the Hunters Range and northeast in the Mount Fowler and Pukeashun Mountain areas.

REGIONAL STRATIGRAPHIC CORRELATIONS

Rocks of the Kootenay terrane are presumably, and to some extent demonstrably, equivalent to strata in the eastern part of the Omineca belt that belong to the ancestral North American miogeoclinal prism (Monger and Berg 1984, Wheeler et al. 1991, Fritz et al. 1991, Colpron and Price 1994 in review). In rocks of low metamorphic grade, rare fossil occurrences have made some correlations possible. In high-grade rocks of the Shuswap complex, direct correlation is possible only where continuity with known stratigraphy is not disrupted by high strain, discontinuous exposure, and Mesozoic and Tertiary intrusions. Most correlations in the Kootenay terrane and particularly in the Shuswap complex have therefore relied on comparison of lithology and stratigraphic successions with known pericratonic sequences.

Footwall assemblages in the Shuswap complex

The Neoproterozoic Windermere Supergroup in the Selkirk, Monashee and Cariboo Mountains consists mainly of poorly sorted feldspathic clastic units and semipelitic to pelitic schists, with minor carbonate and locally abundant amphibolite (Sutherland Brown 1963, Brown et al. 1978, Poulton and Simony 1980, Pell and Simony 1987). Windermere strata have been traced through the northern Selkirk Mountains and into the Shuswap complex north of the Monashee basement culmination (Ghent et al. 1977b, Simony et al. 1980) where they occupy the upper plate of the Monashee décollement. From there, rocks lithologically similar to Windermere strata have been observed along the western flank of the Monashee complex for over half the distance to its southern end (Read and Brown 1981, Journeay and Brown 1986). At the southern end of the Monashee complex, strata that

have been tentatively correlated with the Windermere structurally overly the complex and are overlain by rocks inferred to correlate with Lower Cambrian to Triassic units (Carr 1989, 1990, 1991). Because the Hunters Range and Three Valley assemblages structurally overly the Monashee complex and bear lithological resemblance to the Windermere Supergroup, the possibility that they are correlative with Windermere strata is investigated below.

In the northern Monashee Mountains the Windermere Supergroup is represented by a lower pelite division, a lower marble unit, a semipelite-amphibolite division, a middle marble unit, and an upper clastic division dominated by semipelitic units (Raeside and Simony 1983). Tentative correlation of Hunters Range strata with part of this succession can be made as shown in Figure 3.2. Lithologically, much of the Hunters Range assemblage resembles semipelite-dominated sections of the Windermere, such as the lower pelite and semipelite-amphibolite divisions and the lower part of the upper clastic division. The presence of amphibolite and absence of thickly layered coarse clastic or pelitic metasediments compares favourably with the semipelite-amphibolite (SPA) division (Ghent et al. 1977b, Simony et al. 1980, Poulton and Simony 1980, Raeside and Simony 1983, Pell and Simony 1987). In particular, units similar to the thin layers and boudins of garnet amphibolite and garnet-biotite-hornblende schist in unit **Hs** at Mount Mara and in the lower part of the section east of Kingfisher Creek (Plate 7 *a, d, e*) are reportedly common in the SPA, and have been interpreted as metasediments (Brown et al. 1978) or metatuffs (Sevigny 1987). Layered diopsidic amphibolites like those of unit **Ha** occur in the upper part of the SPA and within the overlying middle marble division in the northern Monashee Mountains and have been interpreted as being calcareous metabasalt flows (Sevigny 1987).

The thick marble unit exposed east of Kingfisher Creek near Blue Lake, which is the structurally highest unit documented within the Hunters Range assemblage (Plate 2b), could be correlative with the middle marble division. The presence of a diopsidic amphibolite unit beneath it is consistent with the association reported by Sevigny (1987). The heterogeneous succession of psammitic to pelitic paragneiss, calcsilicate, amphibolite, and numerous thin marble layers that underly this thick marble could be correlative with zones of alternating carbonate and siliciclastic units in the middle marble (e.g., Brown et al. 1978; Pell and Simony 1982, 1987) or with the carbonate/calcsilicate and upper pelitic zones of the SPA described by Pell and Simony (1981). Marble and calcsilicate units occur in the lower part of the SPA as well (Poulton and Simony 1980) and could be equivalent to calcsilicates lower in the Hunters Range succession. The calcareous unit (**Hc**) and laterally equivalent thin marble (**Hm**) at Mount Mara could be part of the SPA or could be correlative with the underlying lower marble division. If the latter, then the position of marble above amphibolite and rusty semipelite suggests that the sequence is right-way up when compared to the sequence described by Raeside and Simony (1983), and the underlying pelitic to psammitic units of **Hs** and amphibolite of **Ha** should belong to the lower pelite division. On Mount Mara, pelites of unit **Hs** above the marble are commonly rusty and somewhat less sillimanite-rich than those below it, a relationship which also is characteristic of the transition from lower pelite through lower marble to SPA (Raeside and Simony 1983).

The Hunters Range succession exposed on ridges has a minimum aggregate structural thickness of about 2500 m, within which there are no apparent stratigraphic repetitions. Another 1500 m of poorly exposed section, much of which is quartz dioritic orthogneiss, underlies the amphibolite at Mount Mara. This 4 km thickness is

reasonable for a partial section of the Windermere, which is 6 km thick in the northern Monashee Mountains (Raeside and Simony 1983).

Relatively clean marble units south of the map area near Kingfisher Creek have been tentatively correlated with Lower Cambrian strata by K. L. Daughtry (pers. comm. 1993). The southern vergence of F_2 folds in the Hunters Range suggests that higher structural levels should be exposed toward the south, consistent with southward younging of strata if they are right-way up as proposed. This lends support to the correlation of the Hunters Range assemblage with the Windermere.

Isotopic data provide additional evidence in support of this correlation and are the main basis upon which the Hunters Range and Three Valley assemblages are considered to be different from one another. Sm-Nd and Sr data from Hunters Range amphibolites compare well with data from Windermere volcanic rocks (Parkinson 1992). In contrast, amphibolite boudins and paragneiss from Three Valley Gap have isotopic signatures similar to those of the Monashee basement gneisses, reflecting crustal contamination or a crustal source and strongly suggesting a pre-Windermere age (Armstrong et al. 1991, Parkinson 1992). U-Pb zircon data from the amphibolite boudins at Three Valley Gap suggest an age of at least 1.5 Ga (Parkinson 1992). Along with the Sm-Nd data, this date suggests that the boudins are equivalent to or older than the Moyie sills, which intrude the pre-Windermere (Mesoproterozoic) Belt-Purcell Supergroup.

The garnet-biotite-hornblende quartz dioritic gneisses of units **Hh** and **Hdh** are probably mostly post-Windermere intrusions. A preliminary U-Pb age of $366 \pm 10/-3$ Ma, based on the upper intercept of two single-grain and one multigrain zircon fractions, has been obtained from a gneiss of this type from near Mara Lake (H.A. Wasteneys, unpublished data 1993). This date suggests affinity with Devono-

Mississippian plutonic suites such as the Mount Fowler (Okulitch et al. 1975) and Clachnacudainn (Parrish 1992) orthogneisses, which intrude rocks that lie structurally above major Eocene normal faults. A pre-Windermere crustal source is indicated by a 2.0 Ga Nd model age from a sample of unit **Hdh** from the Trans-Canada Highway (Parkinson 1992).

In summary, lithostratigraphic comparison of the Hunters Range assemblage with a reference section of the Windermere Supergroup shows many similarities, and correlation is tentatively proposed. Although some parts of the Hunters Range could be correlative with Paleozoic strata, thick quartzite and marble units that are characteristic of the Lower Cambrian succession of the miogeocline (see the following section) are apparently absent. Isotopic data are consistent with the proposed correlation and imply that the Three Valley assemblage is most likely of Mesoproterozoic (Belt-Purcell) age. Orthogneisses in the Hunters Range are apparently in part Late Devonian intrusions.

Hangingwall assemblages

The Eagle Bay assemblage regionally ranges in age from Lower Cambrian to Mississippian and may include rocks as old as uppermost Proterozoic (Schiarizza and Preto 1987) and as young as Triassic (Okulitch and Cameron 1976). On the basis of lithological similarity and because it is cut by the Devonian Mount Fowler orthogneiss, Okulitch (1979) concluded that the assemblage is correlative with the lower to middle Paleozoic Lardeau Group and with underlying Lower Cambrian units in the Kootenay Arc. Schiarizza and Preto (1987) strengthened this argument by finding Lower Cambrian archaeocyathids in a marble unit northwest of Adams Lake, enabling correlation to be made with the Badshot Formation which contains an identical fauna

(Wheeler 1963). They concluded that this correlation could be applied to the Tshinakin marble, a regionally widespread unit that occurs within a lower and more southerly thrust sheet than the archaeocyathid-bearing marble but in a stratigraphically similar succession. They also concluded that the younger (Mississippian) parts of the Eagle Bay are not correlative with strata in the Kootenay Arc, but are similar to other pericratonic strata (Earn Group) in the Cassiar terrane to the north.

Unit **Em**, the distinctive white to grey marble in the Shuswap Range, is inferred to correlate with the Tshinakin marble and hence may correlate with the Lower Cambrian Badshot Formation. Based on their lithology and stratigraphic positions relative to **Em**, units **Eq** and **Ec** are tentatively correlated with the Hamill and Lardeau groups, which lie respectively below and above the Badshot in the Kootenay Arc and Selkirk Mountains (Fyles and Eastwood 1962; Wheeler 1963, 1965). If these correlations are correct, then the Eagle Bay section is mostly inverted between Sicamous and Queest Mountain, is repeated by tight folds and possible thrusts in the Queest Mountain area, and is mostly right-way-up at North Queest Mountain.

The Hamill Group consists mainly of quartzite with subordinate pelite and locally abundant mafic volcanic units (Wheeler 1965, Devlin 1989), much like **Eq**. The uppermost unit within the Hamill (Mohican Formation) consists of thin-bedded, commonly brown-weathering calcareous units and phyllites with a locally mappable marble at the base (Fyles and Eastwood 1962, Read and Wheeler 1976). Much of unit **Ec** adjacent to the **Em** marble is like this, especially on North Queest Mountain; however, correlation of units **Ec** and **Em** with the Mohican is not favoured because there is no other prominent clean marble in the Eagle Bay succession with which to correlate the Badshot. Rusty, platy phyllites and calcareous units also occur within unit **Eq** near the contact with **Em** and are more likely correlative with the Mohican.

The Lardeau Group, described in detail by Fyles and Eastwood (1962), includes formations dominated by phyllite (Index, Triune and Sharon Creek), mafic volcanic flows and tuffs (Jowett), and poorly sorted siliciclastic grits (Broadview). The lower part of the Index Formation, at the base of the Lardeau, generally consists of green to dark grey graphitic phyllite or argillite with less abundant platy grey limestone, calcareous mica schist or phyllite and brown or light grey quartzite (Fyles and Eastwood 1962, Wheeler 1963, Fyles 1964, Thompson 1978, Brown 1991, Colpron and Price 1993). The upper part of the Index in the eastern Kootenay Arc is mostly green calcareous phyllite, chlorite schist and massive to pillowed mafic metavolcanic rocks, and is very similar to the Jowett Formation (Fyles and Eastwood 1962, Fyles 1964, Read and Wheeler 1976). Phyllite and phyllitic limestone units of the Index undergo a southwestward facies change to siliceous grit and gritty phyllite (Read 1975, Read and Wheeler 1976). Unit **Ec** in the Shuswap Range contains a similar suite of rocks to that in the Index and Jowett formations, and in approximately the same stratigraphic sequence. A weakness in correlation of these units is that only thin intervals of dark grey graphitic phyllite are exposed in unit **Ec** in the Shuswap Range, whereas they typify the lower Index Formation.

In the Kwikoit Creek area east of Adams Lake, **Ec** is readily correlative with unit EBG of Schiarizza and Preto (1987) and probably with the Lardeau Group. Sericite-quartz and pyrite-graphite-sericite phyllites (**Ecp**) are probably Index Formation. Grey impure marbles of unit **Ecm** probably also correlate with the Index, although they could correlate with the Tshinakin (and Badshot?) marble as mapped by Okulitch (1979). Mafic metavolcanic rocks (**Eca**) are presumably correlative with the upper Index or the Jowett Formation. Quartzites, metasiltites, quartz-sericite phyllites, and feldspathic granule conglomerate mapped as **Eq** could either correlate with the

Hamill Group or with part of the Lardeau (Broadview Formation), and probably correlate with unit EBQ of Schiarizza and Preto.

There is considerable uncertainty about the tectonostratigraphic significance of the Sicamous and Tsalkom formations and the Silver Creek assemblage. Campbell and Okulitch (1973) inferred the Sicamous Formation to be correlative with the Slocan Group of Triassic age, which is part of the Quesnellia terrane. However, similarity of the Sicamous to calcareous and graphitic phyllites of the Index Formation of the Lardeau Group (Fyles and Eastwood 1962, Fyles 1964) and its striking resemblance to grey marble and phyllite of the Eagle Bay assemblage (e.g. units **Ecm** and **Ecp** of this study and unit EBL of Schiarizza and Preto 1987) has prompted several workers to suggest a lower Paleozoic age (Preto and Schiarizza, 1985; A. V. Okulitch, pers. comm. 1986; Schiarizza and Preto 1987; Johnson 1990). Rocks in the Larch Hills that were mapped by Jones (1959) as Permian Cache Creek Group and from which he reported to have collected crinoid stems have been reexamined by K. L. Daughtry (pers. comm. 1993) and reinterpreted as part of the Sicamous Formation. Lower Paleozoic and Triassic ages both remain permissible, as the exact age of the fossil fragments is uncertain.

The Tsalkom Formation has been regarded as being equivalent to the upper Paleozoic Fennell Group of the Slide Mountain terrane (Okulitch 1979, Okulitch and Cameron 1976), but Wheeler and McFeely (1991) have more recently included it in the Kootenay terrane. Either interpretation is admissible at present. The Tsalkom resembles metavolcanic rocks in the Eagle Bay assemblage and could be correlative with the Jowett Formation of the Lardeau Group. A comparative geochemical and isotopic investigation of metavolcanic rocks of the Tsalkom, Eagle Bay, Fennell, and Lardeau is strongly warranted.

The Silver Creek assemblage could be correlative with the Lardeau Group (Broadview Formation) or with Neoproterozoic to Lower Cambrian strata of the Windermere and Hamill (Okulitch 1979, Johnson 1990). The latter is favoured because palinspastic restoration of the Okanagan - Eagle River fault (Chapter 5; Plate 3) juxtaposes the lower part of the Silver Creek near the upper part of the Hunters Range assemblage, which is inferred to be mostly right-way-up Windermere. The Silver Creek bears resemblance to parts of units **Eq** and **Ecq** of the Eagle Bay and to unit **Ls** of the Lichen assemblage. At present the only constraint on its age is that it is intruded by the Little Shuswap orthogneiss, which has been interpreted as being Ordovician in age on the basis of a U-Pb zircon upper intercept (Okulitch 1985). The zircons analyzed were reported to have contained inclusions and large dark cores (possibly inherited components), and as no other granitoid plutons in the southern Omineca belt are known to be of Ordovician age (Wheeler and McFeely 1991, Woodsworth et al. 1991), this interpretation is regarded as suspect.

In light of the possibility that the Sicamous, Tsalkom, and Silver Creek assemblages are entirely or partly unrelated to one another and tectonically juxtaposed, it is suggested that the name "Mount Ida Group," which was defined by Jones (1959) to include most of the low-grade metamorphic rocks in the northwestern part of the Vernon map area (Eagle Bay, Silver Creek, Tsalkom and Sicamous among them), be abandoned.

Summary of stratigraphic conclusions

The footwall of the Okanagan - Eagle River fault contains rocks that are tentatively correlated with the semipelite-amphibolite and middle marble divisions (and possibly also the lower pelite and lower marble divisions) of the Neoproterozoic

Windermere Supergroup. These are structurally underlain by pre-Windermere rocks that may be Mesoproterozoic in age.

The hanging wall of the fault contains quartzite and clean marble units that are tentatively correlated with the Lower Cambrian Hamill Group and Badshot Formation, and heterogeneous metasedimentary and mafic metavolcanic units that are inferred to correlate with the lower Paleozoic Lardeau Group. Some Windermere equivalents may occur in the Silver Creek assemblage. Correlation proposed by previous workers of the Sicamous and Tsalkom formations with rocks of the Quesnellia and Slide Mountain terranes is permissible but is thought to be less likely than correlation with pericratonic Lardeau strata. This will be further investigated in Chapter 5 through palinspastic restoration of the Okanagan - Eagle River fault.

Apparently, both hanging wall and footwall assemblages are intruded by Paleozoic (Late Devonian and possibly older) granitoid plutonic suites.

STRUCTURE AND METAMORPHISM

The Okanagan Valley fault system is a major tectonic boundary across which rocks of contrasting metamorphic grades have been juxtaposed. Footwall assemblages (Three Valley, Hunters Range and Momich) have been metamorphosed to upper amphibolite facies. Assemblages in the hanging walls of the two en échelon detachment segments (Eagle Bay, Silver Creek, Sicamous and Tsalkom) are greenschist to lower amphibolite facies.

Five assemblages (Hunters Range, Eagle Bay, Lichen, and the Pukeashun and Mount Fowler plutonic suites) were mapped in sufficient detail for structural analysis.

Structural details of the Pukeashun granite suite, which has been heterogeneously mylonitized and has excellent west-northwest-directed kinematic indicators, are omitted from this section as they have been described in Chapter 2.

Fabrics, folds, and corresponding periods of deformation are assigned alphanumeric symbols to indicate their relative ages within a given assemblage. For example, S_1 is the oldest foliation preserved and is said to have been produced by D_1 deformation. In some cases more than one generation of structures may have developed during a protracted episode of deformation, and any given phase of deformation may have been regionally diachronous.

Hunters Range and other footwall assemblages

The oldest recognizable deformation fabric in the Hunters Range assemblage, termed S_1 , is a ubiquitous gneissic foliation that in most places is gently dipping to subhorizontal (Fig. 3.3). Compositional layering, defined by contacts between lithological units, by mineralogical variation within units, and by leucosome layers and seams, is parallel to S_1 except in rare instances where thin quartzose layers define rootless intrafolial F_1 isoclines.

Compositional layering and parallel S_1 foliation (together called S_0S_1) are deformed by recumbent to gently inclined folds (F_2) that typically are tight (Plate 8 c) but that range from isoclinal to close. These folds plunge gently west-southwest or east-northeast and have subhorizontal to moderately dipping S_2 axial surfaces that make angles of 0-20° with the enveloping S_0S_1 foliation. In hinge areas of some F_2 folds, layers of biotite that define S_1 are deformed by axial-plane S_2 crenulation cleavage; in others, recrystallization of biotite parallel to S_2 has obliterated most evidence of S_1 . On the long limbs of F_2 folds, S_2 is parallel to compositional layering

and S_1 . This composite gneissic foliation is therefore a product of cumulative transposition and metamorphic recrystallization during protracted or episodic deformation. Long axes of sillimanite and hornblende define an east-northeasterly trending L_2 mineral lineation (mean = $260^\circ-02^\circ$; Fig. 3.3), oriented parallel to F_2 hingelines and to a stretching lineation defined by mineral aggregates (e.g., quartz ribbons and quartz-biotite pressure shadows on garnets).

As noted by Jones (1959) and by Fyson (1970), most of the F_2 folds in the Hunters Range verge southward, although some near Mara lookout verge northward. Only mesoscopic F_2 , typically of the order of a few decimetres to a few metres in amplitude, have been documented. Some south-verging folds near Mount Griffin have shortlimb lengths of the order of 100 m, and although no evidence for closure of a larger structure has been documented, the stratigraphic section on the north flank of Mount Griffin could be repeated by a southward-closing recumbent F_2 antiform cored by unit Hc. Type II interference patterns (Ramsay 1967) are displayed by a few F_2 folds. They are symmetrically disposed about the L_2 mineral stretching lineation and are interpreted as sheath folds (cf. Cobbold and Quinquis 1980) produced by progressive D_2 deformation at high shear strain.

D_2 structures and metamorphic minerals are deformed by moderately to steeply inclined, commonly overturned, west-southwesterly verging F_3 folds (Plate 8 *d-g*). These folds typically are open to close, although minor folds in pelites are tight and have angular hinges. Their axial surfaces dip subvertically to moderately east-northeastward, making moderate to high angles ($30-60^\circ$) with respect to enveloping foliation. Hingelines are subhorizontal to gently plunging, with a mean orientation of $162^\circ-01^\circ$ (Fig. 3.3). Sillimanite wraps around the hinges of F_3 folds but is not dynamically recrystallized (Plate 8 *f*), implying that it grew before, but was stable

during, D_3 . Rarely, metapelites exhibit an axial-plane S_3 cleavage defined by parallel sillimanite and biotite.

The largest F_3 folds in the Hunters Range, notably at Mount Mara, have subvertical to steeply overturned shortlimbs well over 100 metres long. Mesoscopic F_3 folds in pelite-dominated units commonly are tight and angular, and have thickened hinges in which parasitic crenulations have formed. In contrast, folds in psammites are more open and concentric, with curved hinges. Décollements have developed between some interlayered thick pelites and psammites, perhaps so that both of these fold styles could occur without strain incompatibility (Plate 8 g). F_3 folds in pegmatite and migmatitic pelite near the Mara fire lookout are tight to isoclinal, with axial surfaces inclined at a low angle ($10-30^\circ$) to the enveloping foliation. The atypical tightness of these folds is interpreted as having resulted from local high shear strain.

Stereograms of poles to gneissic foliations from the Eagle River area and from near Kingfisher Creek between the ecological reserve and Blue Lake are spread moderately about a great-circle girdle distribution, reflecting F_3 folding (Fig. 3.3). Greater scatter in poles to foliations from the Mount Mara area is due in part to interference between F_2 and F_3 . In contrast, foliations in the Mount Griffin area are typically subhorizontal, deformed only by broad warps and some F_2 folds. The southwest dip of foliations in the Four Mile Creek area is interpreted as a manifestation of the ramp geometry of the underlying Monashee décollement; foliations in the Three Valley assemblage and in the western Monashee complex have similar orientations.

Mylonitic fabrics

Two spatially distinct and kinematically opposed zones of mylonitic fabric are

developed within the footwall panel. Both zones formed under conditions in which the peak metamorphic mineral assemblage was stable and are characterized by intense transposition of compositional layering parallel to mylonitic foliation. The rock is referred to as *mylonitic* if it shows evidence for strain-induced grain-size reduction without loss of cohesion. On outcrop and hand-specimen scales, such evidence typically consists of the presence of (i) irregularly shaped feldspar porphyroclasts with fine-grained recrystallized tails, (ii) elongate ribbon aggregates of quartz, and/or (iii) strain gradients across the foliation that show variability in grain size of an order of magnitude or more within rock of approximately uniform composition. The term *mylonite* is reserved here for rocks that preserve evidence of dynamic grain-size reduction at the thin-section scale; that is, rocks that have nonannealed mylonitic texture.

The first mylonitic zone is most strongly developed near the base of the panel in the Monashee décollement zone. Mylonitic gneisses in this zone have been extensively studied and a consistent relative east-northeastward motion of the upper plate (hereafter called "northeast-directed" sense of shear) has been demonstrated (Read and Brown 1981; Journeay 1986; Journeay and Brown 1986; Scammell 1986; Coleman 1987, 1989, 1990; Bosdachin and Harrap 1988; Bosdachin 1989; Harrap 1990; Carr 1990, 1992; McNicoll and Brown 1994). Mylonitic fabrics with northeast-directed kinematic indicators persist into the lower part of the Three Valley assemblage and occur within the sheared pegmatite that marks the boundary between the Three Valley and Hunters Range assemblages, as described previously. Pegmatites in the northeastern part of the Hunters Range exhibit heterogeneous mylonitic fabric with asymmetrically oriented feldspar porphyroclasts and extensional shear bands that indicate northeast-directed shear (Plate 8 *h*). Elsewhere in the Hunters Range

assemblage, northeast-directed mylonitic fabrics are generally not evident, although pressure shadows on garnets in some paragneisses are asymmetrically disposed and consistent with (but not diagnostic of) northeasterly shear sense. In all thin sections of rocks in which northeast-directed kinematic indicators were observed in the field, quartz grains exhibit annealed texture.

In contrast, mylonites in the immediate footwall of the Okanagan Valley fault system display a variety of kinematic indicators that consistently indicate west-directed shear sense (Bardoux 1985, 1993; Tempelman-Kluit and Parkinson 1986; Journeay and Brown 1986; Johnson 1988, 1989, 1990; Chapter 2; Appendix A). These mylonites are nonannealed and are heterogeneously developed in the uppermost few hundred metres to about 1 km of the footwall, with strain being highest in migmatitic rocks of pelitic to semipelitic paragneiss protolith. Quartz-feldspar aggregates, sillimanite, and pressure shadows marginal to garnet and K-feldspar porphyroclasts define a strong west-trending stretching lineation, which differs in mean orientation from the mineralogically identical L_2 lineation deeper in the footwall and in northeast-directed mylonites ($275^\circ\text{-}08^\circ$ vs. $065^\circ\text{-}03^\circ$; Fig. 3.4). Biotite in discrete west-directed shears typically has been partly replaced by chlorite, and local cataclastic zones display slickensides with west-directed chlorite fibre lineations subparallel to the stretching lineation. West-directed shearing therefore outlasted peak metamorphic conditions.

The distribution of outcrops in which northeast- and west-directed kinematic indicators were observed is shown in Plate 1 and summarized along with orientation data in Figure 3.4. Mylonites with westerly to northwesterly directed kinematic indicators (mean direction 301° based on mean lineation $121^\circ\text{-}13^\circ$) also occur in the Momich assemblage along Adams Lake and in the Pukeashun granite (Plate 1; Fig. 3.4; Chapter 2).

Metamorphic grade in the Hunters Range assemblage is upper amphibolite facies. Metapelites in most of the Hunters Range contain the assemblage garnet + biotite + plagioclase + quartz + sillimanite + K-feldspar and are migmatitic (this study and Bender 1990). Coarse muscovite coexists with sillimanite in pelitic migmatites on ridge tops near Mount Griffin, in the footwall of the Okanagan - Eagle River fault along Craigellachie Creek, and in the Momich assemblage. Muscovite lies parallel to the foliation in these rocks and is assumed to be part of the thermal-peak equilibrium assemblage. Randomly oriented muscovite occurs in small quantity (<2%) in rocks of the sillimanite-K-feldspar zone and is inferred to be a product of retrogression.

Eagle Bay assemblage

The Eagle Bay and other hangingwall assemblages have been deformed by three phases of penetrative fabric-forming structures. A fourth phase characterized by north-trending upright crenulations and brittle fractures (Fyson 1970, Nielsen 1982) is relatively minor.

Primary layering (S_0) and a parallel foliation (S_1) constitute a composite S_0S_1 fabric that is deformed by F_2 folds. A pervasive S_2 schistosity (cleavage in lower-grade phyllites) is defined by alignment of micas and deformed quartz parallel to axial planes of F_2 folds, and is the strongest penetrative fabric in the Eagle Bay rocks. The S_1 foliation is only locally preserved in the hinges of mesoscopic F_2 folds because growth and recrystallization of metamorphic minerals outlasted its formation. For this reason it is generally not possible to distinguish between F_1 and F_2 folds in the field. In some places where S_1 is preserved, S_2 is a crenulation cleavage that makes a low angle to the enveloping S_0S_1 . Except in the hinge areas of F_2 folds, S_2 is parallel to S_0S_1 or makes a very low angle to it. Figure 3.5 shows orientation data for S_0S_1 .

Mesoscopic F_2 folds are most prominent in thinly interlayered quartzites and metapelites of unit Eq (Plate 9 a). They generally are tight, but range from isoclinal to close. They have gently plunging axes (L_2) that regionally exhibit a wide range in trend but that define local southwest and southeast-trending domains (Fig. 3.6). A quartz rodding lineation is commonly developed parallel to F_2 hingelines.

S_2 is folded about gently northwest or southeast-plunging axes by open, upright to steeply inclined F_3 folds and crenulations (Plate 9 a). An axial-plane crenulation cleavage (S_3), defined by foliae of parallel muscovite and biotite (locally retrograded to chlorite) spaced a few mm apart, is developed in phyllites and pelitic schists. The map pattern in the Shuswap Range south of Queest Mountain is dominated by an F_3 synform and complimentary antiform with wavelength of several km (Plates 1 and 2d). These structures are responsible for the great-circle distribution of poles to main foliations in the western and central Shuswap Range in stereographic projections (Fig. 3.5). They apparently die out west of Shuswap Lake where foliations in the Eagle Bay, Mount Fowler and Lichen assemblages dip moderately eastward to southeastward. The Silver Creek and Sicamous formations in the Larch Hills are folded by an F_3 antiform of similar scale, also reflected by girdle distribution of poles to foliations.

Only mesoscopic F_2 have been positively identified, but there probably are important macroscopic folds. The cross section through the central Shuswap Range (Plate 2d) shows limited evidence for a westerly closing isoclinal nappe, cored by unit Eq southwest of Queest Mountain, which has been refolded by the major F_3 synform. Repetition of unit Ec by such a fold is mostly consistent with the geometry of minor folds that were identified as F_2 (but that in some cases could be F_1). If the stratigraphic correlations outlined in the preceding section are correct (i.e. units Eq,

Em and Ec with Hamill, Badshot and Lardeau, respectively), then this structure would be an anticline.

Evidence for this fold is difficult to trace to the northeast. At Queest Mountain, marble of unit Em outlines tight folds with limb lengths of the order of tens or hundreds of metres, and minor structures in interlayered quartzite and schist of unit Eq indicate very tight recumbent folds of a comparable scale. But because of discontinuous exposure it is unclear whether these folds are part of a larger structure. Northeast of Queest Mountain, marble and calcareous schist units mapped as Em and Ec and associated orthogneiss of unit Dg are symmetrically disposed above and below unit Eq. Very few mesoscopic D_2 structures have been observed in this area, but a shallow-dipping crenulation cleavage that makes a low angle with S_0S_1 and that is interpreted as S_2 displays a vergence reversal within unit Eq near Craigellachie Creek that is consistent with closure to the north of a recumbent F_2 fold cored in unit Eq. If this is the same fold as the suspected F_2 anticline southwest of Queest Mountain, or another of the same style, then its hingeline must be curved on the scale of the Shuswap Range; this is permissible in light of the regionally highly variable orientation of L_2 axial lineations. Several alternative explanations for the apparent repetition of map units are equally valid. For example, there could be two marble/calc-schist units in the stratigraphic succession, or section could be repeated by a thrust.

In the southern Shuswap Range, it appears in cross section that orthogneiss of unit Dg closes to the west, outlining the hinge of a recumbent isocline cored by calcareous rocks and chlorite schists of unit Ec (Plate 2a). Some of the orthogneiss displays recrystallized mylonitic texture, and repetition of section could be by low-angle ductile or ductile-brittle faults. The contact between the lower panel of

orthogneiss and unit Ec beneath it is tentatively interpreted as a thrust.

The macroscopic structural geometry in the Adams Lake area is ambiguous. Graphitic and chloritic phyllites (Ecp) and intervening phyllitic marble (Ecm) are exposed in roadcuts along Kwikoit Creek (Plate 1). Marble outcrops on opposite sides of the creek do not line up along strike, and so either there are two belts of marble as shown in Plate 1 or a single belt of marble is cut by a fault in the creek as interpreted by Okulitch (1979). The interpretation shown in Plate 1 is based on the observation that mesoscopic open folds with steep axial surfaces exhibit opposing vergence in the two outcrop areas, which when projected into a vertical cross section suggests duplication of unit Ecm by a synform cored by phyllite of unit Ecp (Plate 2g). Steeply northward-dipping crenulation cleavage is axial planar to mesoscopic folds on the northern limb of the synform and is interpreted as S_3 based on its orientation, implying that the major structures are F_3 folds that are overturned to the southwest. Another possible interpretation is that this marble is the same unit as the marble that is interlayered with mafic metavolcanic rocks and chlorite schists of unit Eca near the Mount Fowler gneiss (Dg), and that it closes in the subsurface as a broad F_3 synform. This would require that the crenulation cleavage and mesoscopic folds in the marble along Kwikoit Creek are D_2 rather than D_3 structures. These ambiguities may be resolved by further mapping.

Metamorphic grade in the Eagle Bay increases to the northeast from biotite to garnet zone of greenschist facies in the Adams Lake area. Staurolite occurs at several localities in metapelites of the Shuswap Range (Plate 1), indicating lower amphibolite facies. In thin sections examined from rocks of the Shuswap Range, garnet and staurolite porphyroblasts contain inclusions of quartz and opaque minerals that define internal fabric (S_i). In some garnets, S_i is approximately perpendicular to external S_2

in the matrix, which wraps around the garnets (Plate 9 *b*). The internal fabric in this case is interpreted as being a pre-D₂ foliation (S₁), though it could be early S₂ that was subsequently rotated during D₂ after garnet growth. In other garnets, S_i inclusion trails are parallel to external S₂, but S₂ is folded by F₃ crenulations whereas S_i is straight (Plate 9 *c*). F₃ crenulations are deflected around all garnets observed. Staurolite porphyroblasts contain S_i inclusion trails that are concordant and continuous with external S₂. In contrast to the straight S_i in garnets, S_i in many idioblastic staurolites outlines crenulations that resemble external F₃ but that have a smaller amplitude and longer wavelength (Plate 9 *d-h*). External S₂ was observed in one instance deflected around a staurolite porphyroblast that had overgrown crenulated S₂ (Plate 9 *g*). Collectively, these relationships imply that some garnet grew before or during the early stages of D₂, garnet growth stopped before D₃, and staurolite growth began late during D₂ and continued during D₃. Metamorphism therefore accompanied D₂ and D₃, and apparently peaked during D₃. These observations and interpretation corroborate those of Fyson (1970).

Metamorphism of the Silver Creek assemblage in the Larch Hills is upper greenschist facies (garnet zone). Nielsen (1978, 1982) noted a southward increase in grade within the Silver Creek, which he attributed to juxtaposition of different structural levels across a low-angle fault that here is interpreted as a splay of the Okanagan - Eagle River fault (see lithological descriptions earlier in this chapter).

Lichen assemblage

Compositional layering (S₀) and parallel foliation (S₁) in the Lichen assemblage are folded by intrafolial F₂ isoclines. The hingelines (L₂) commonly are collinear with a stretching lineation defined by quartz rods and mineral aggregates, and range from

gently southeast plunging at Lichen Mountain to locally subvertical. The variation in orientation of L_2 , for which no regional pattern has been documented, is evident in the scatter on a stereonet (Fig. 3.6).

A crenulation cleavage (S_3) that generally dips steeply to moderately northeast or southwest is well developed in mica schists. Steeply dipping crenulation cleavage in metasediments at Mount Celista is similar in orientation and style to S_3 in the Lichen assemblage, but was mapped as S_2 because only two foliations were recognized in that area.

The structural style of the Lichen assemblage is in most respects like that of the Eagle Bay, although some of the structures are more variable in attitude. The steep west-northwest-striking attitude common of foliations in the Lichen assemblage (Plate 1; Fig. 3.5) may be from rotation caused by strike-slip transfer of displacement between the en échelon segments of the Okanagan Valley fault system (Chapter 2).

Metamorphic grade is amphibolite facies and apparently increases toward the Pukeashun granite. The first appearance of sillimanite generally is fibrolite that overgrows S_2 . Closer to the granite, sillimanite lies within S_2 and parallels L_2 . Muscovite is everywhere part of the metamorphic mineral assemblage.

Mount Fowler orthogneiss

The Mount Fowler orthogneiss displays a foliation that parallels compositional layering in the Eagle Bay and that, along with S_1 in the Eagle Bay, is folded by F_2 folds. This foliation bears a variably developed lineation, defined by aggregates of biotite ($\pm$ epidote $\pm$ hornblende) and/or by quartz ribbons, which trends northeasterly to southeasterly (Fig. 3.6), generally parallel to L_2 lineations in the Eagle Bay and Lichen assemblages. Strongly lineated orthogneiss commonly exhibits ribbon-quartz

aggregates, porphyroclasts of K-feldspar, and strain gradients across the foliation. This association of features suggests mylonitic fabric, but all samples observed in thin section display annealed (granoblastic) textures and lack indicators of noncoaxial flow. Nonannealed mylonitic fabric in the orthogneiss was only observed in discrete west-directed shears near Mount Fowler (Chapter 2).

REGIONAL STRUCTURAL AND METAMORPHIC RELATIONS

Many previous workers have contended that high-grade and low-grade rocks in the Shuswap Lake area have experienced the same deformation history and that structures in each can be correlated (Jones 1959, Fyson 1970, Ross 1981, Neilsen 1982). Comparison of the results of this study with other recent studies cited in the following pages suggests that, despite geometrical similarities, deformation fabrics and metamorphism in high-grade rocks of the Shuswap complex are Cretaceous to early Tertiary in age, whereas those in the lower-grade rocks that lie in the hanging wall of the Okanagan Valley fault system are mostly Early Cretaceous and older.

Footwall structures

The structural style of the Hunters Range is comparable to that of other rocks at the same regional structural position (i.e. above the Monashee décollement and below the Okanagan Valley fault system). Working respectively near the southern and northern ends of the Monashee complex, Carr (1990, 1991) and Scammell (1993) have recognized that, as in the Hunters Range, the main fabric defined by peak metamorphic minerals is a composite foliation into which compositional layering and

parallel foliations have been variably transposed. Results from U-Pb geochronological studies of leucogranites indicate that development of the composite transposition fabric occurred from at least Early to Late Cretaceous (pre-135 Ma to 97 Ma) in the northern Monashee Mountains (Scammell 1993) and that northeast-directed shearing continued throughout the region from Late Cretaceous to Late Paleocene time (Carr 1990, 1991, 1992; Scammell 1993; Chapter 4).

In the Hunters Range assemblage, the S_1 foliation preserved in F_2 fold hinges could have formed earlier in the progressive history of fabric transposition or could have formed during a distinct D_1 event. The peak metamorphic mineral assemblage for the most part grew during D_2 . F_2 folds, L_2 mineral/stretching lineations and northeast-directed mylonitic fabrics formed during D_2 under peak temperatures, and leucosome seams folded by F_2 folds indicate that some partial melting occurred before the final stages of D_2 . D_2 strain is partitioned so that the lower part of the Three Valley assemblage is strongly mylonitic, whereas northeast-directed mylonitic fabrics and strain gradients related to heterogeneous noncoaxial flow in the Hunters Range rocks are most conspicuous in pegmatites in the northeastern part of the range. These observations imply that (i) at least the lower part of the panel of rock that includes the Three Valley and Hunters Range assemblages was subjected to northeast-directed noncoaxial flow at upper amphibolite facies, (ii) this deformation was linked to thrust displacement on the Monashee décollement and related shear zones, and (iii) much of the pegmatite was emplaced syntectonically.

D_3 occurred after crystallization of sillimanite but under conditions favouring sillimanite stability. F_3 folds are consistently west-southwesterly verging, they deform the sillimanite lineation in rocks that locally contain indicators of west-directed shear, and they are cut by discrete west-directed shears that generally dip easterly. They

therefore either formed synchronously with, or were amplified during, west-directed shearing. The atypically tight F_3 folds near Mara lookout were certainly modified (if not generated) by shearing, which probably was facilitated by softening due to partial melting and pegmatite emplacement. U-Pb zircon analyses from folded leucosome at Mara lookout indicate a Late Cretaceous or younger age (~70 Ma; sample J90-1059, Chapter 4), and accordingly this is considered an upper age limit for cessation of D_3 .

Previous interpretations have presumed that the west-directed shears associated with F_3 structures in the Hunters Range are related to the Okanagan Valley normal fault system (Johnson 1988, 1989). However, the moderately to steeply inclined style of most F_3 folds suggests a component of layer-parallel shortening that is inconsistent with the extensional strain implied by the west-trending stretching lineation ubiquitous in mylonites of the normal fault system. This suggests that F_3 folds and the discrete west-directed shears in the central Hunters Range may be backfolds and minor backthrusts, which perhaps formed either *(i)* to maintain critical taper as the rocks were thrust over a ramp in the Monashee décollement or *(ii)* as passive roof structures above a wedge within the allochthon that was being driven eastward above the décollement. Carr (1991) presented an interpretation similar to the second of these scenarios to explain a vergence reversal in F_3 folds south of the Monashee complex, which occurs above the tip point of a blind structure in the hanging wall of the Monashee décollement imaged in a Lithoprobe reflection profile.

High-grade mylonites related to the northeast-directed Monashee décollement and the west-directed Okanagan Valley fault system are only pervasive in, respectively, the lowermost and uppermost parts of the Three Valley / Hunters Range panel. There is little evidence for noncoaxial flow of either sense in the middle part of the panel. Although extension is known to have outlasted thrusting, there is no

evidence from the present study to contradict the contention of Scammell (1993) that extensional crustal denudation began in the Cretaceous and was concurrent with thrusting.

Hangingwall structures

Structures in the hangingwall rocks can be broadly correlated with structures in equivalent rocks elsewhere in Kootenay terrane and in the more inboard pericratonic assemblages of the Selkirk Mountains. In general, polyphase structural history in these areas has been worked out with respect to the timing of regional metamorphism, which is known to have climaxed in many parts of the Selkirk and Kootenay Mountains in the Middle Jurassic (Brown and Tippet 1978, Shaw 1980, Archibald et al. 1983, Brown et al. 1992b, Colpron and Price 1993).

Although the possibilities of regionally diachronous metamorphism and complicated local structural geometry are recognized, the structures of the Eagle Bay assemblage in the Shuswap Range are considered likely to be equivalent to structures in the Selkirk Mountains at the same latitude, because restoration of the Okanagan - Eagle River and Columbia River faults along the approximately east-west directions of extension would result in near juxtaposition of the Shuswap Range and the western Selkirks (Chapter 5). Furthermore, the Eagle Bay in the Shuswap Range exhibits similar structural sequence and style, metamorphic grade, and stratigraphy to the Clachnacudainn terrane in the western Selkirks (Fig. 3.1), and Devonian granitoid plutons are widespread in both of these areas yet, except within the Clachnacudainn terrane, are not known to occur in the Omineca belt east of the Columbia River.

Regionally, the earliest penetrative deformation in the Kootenay terrane is of mid-Paleozoic age, documented by the occurrence of clasts of foliated Lardeau Group

rocks within the Upper Mississippian Milford Group in the Kootenay Arc (Read 1975, Read and Wheeler 1976). The earliest large-scale regional structures in the Selkirk and northern Monashee Mountains are recumbent southwest-vergent nappes that predate the Middle Jurassic metamorphic peak (Ghent et al. 1977b, Simony et al. 1980, Raeside and Simony 1983, Brown and Lane 1988). Southwest-verging folds in the Cariboo Mountains, which are tight and near-recumbent at deep structural levels (Murphy 1985, 1987), deform the Early to Middle Jurassic accretion boundary between the Intermontane superterrane and Kootenay terrane (Struik 1981, Rees and Ferri 1983, Ross et al. 1985, Brown et al. 1986, Rees 1987, Murphy 1987). Since these structures predate Middle Jurassic equilibration of metamorphic isograds, they are at least in part of Early to Middle Jurassic age (see Brown et al. 1986, Brown and Lane 1988 for discussions).

The oldest deformation fabric preserved in the Eagle Bay and other rocks of the hanging wall of the Okanagan - Eagle River fault, S_1 , is a bedding-parallel schistosity in metasedimentary rocks that predates the metamorphic peak and is a typically strongly developed gneissosity in the Mount Fowler orthogneiss. The high degree of strain in the orthogneiss suggests that this unit may have been the locus for large-scale ductile flow. The Devonian age of the gneiss places a maximum limit on the age of D_1 deformation.

Synmetamorphic structures in the Selkirk Mountains include tight to isoclinal, overturned to recumbent southwest-verging F_2 folds (Brown and Tippet 1978, Höy 1979, Crowley 1992, Colpron and Price 1993, Crowley and Brown 1994) which are highly attenuated (Thompson 1978, Crowley 1992) and in places show highly contorted hingelines resulting from complex progressive D_2 deformation in the Middle Jurassic (Colpron and Price 1993). Peak metamorphic conditions were reached

variably during and after D_2 and before D_3 (Brown and Tippet 1978, Shaw 1980, Perkins 1983, Crowley 1992).

D_2 structures in the Eagle Bay assemblage are also synmetamorphic and regionally southwest-verging. The metamorphic peak was syn- D_2 in the Adams Lake area (Schiarizza and Preto 1987), but porphyroblast growth in the Shuswap Range continued beyond D_2 and into D_3 (Fyson 1970; this study). As in the Selkirks, D_2 involved high strain; this is implied by the observations that (i) F_2 hingelines are parallel to the stretching lineation in the Mount Fowler gneiss and to a locally well-developed stretching lineation in the Eagle Bay, and (ii) S_0S_1 generally has been transposed into parallelism with S_2 . There is some evidence for the existence of large recumbent F_2 isoclinal folds in the Shuswap Range, and Okulitch (1979) noted that the Tshinakin marble is repeated by tight southwesterly overturned folds a few km southwest of this area. Southwest-verging folds and thrusts dominate the structural style in the Adams Lake area and are inferred to be of Middle Jurassic to Early Cretaceous age (Schiarizza and Preto 1987).

As in this study, both Fyson (1970) and Okulitch (1979) observed that L_2 axial/stretching lineations in the Eagle Bay exhibit widely variable attitudes, with northerly to northeasterly trends being most common. This variability cannot be accounted for by refolding about the subhorizontal northwest-trending axes of F_3 folds. Differential rotation during Tertiary crustal extension can also be dismissed as a significant factor because L_2 stretching lineations in the immediate hanging wall of the Okanagan - Eagle River fault along the Eagle River valley (where strain might be expected to be highest and such rotation should be greatest) have southerly trends, whereas stretching lineations in the footwall mylonites trend east and west. Domains with different L_2 attitudes could be bounded by splays of the Tertiary fault system, but

this has yet to be demonstrated. The most likely cause for the wide range in L_2 attitude is a complex D_2 history, perhaps involving nonuniform rotation of mesoscale structures. In favour of the latter, L_2 lineations in the Adams Lake area show wide variability but macroscopic F_2 folds are uniformly southwest-verging (Schiarizza and Preto 1987).

D_3 structures in the Eagle Bay, including northwest-trending folds and steeply dipping axial-plane crenulation cleavage, are similar in style and orientation to D_3 structures in the Selkirks (e.g. Brown and Tippet 1978, Crowley 1992, Colpron and Price 1993) and to "postmetamorphic" (D_3) crenulations in the Eagle Bay assemblage west of Adams Lake described by Schiarizza and Preto (1987). These structures predate emplacement of the Albert stock in the Selkirks (Sears 1979, Crowley 1992) and of the Baldy batholith west of Adams Lake (Schiarizza and Preto 1987), which have been dated by U-Pb zircon methods at 104 ± 1 Ma (Crowley 1992) and 116 ± 5 Ma (Calderwood et al. 1990), respectively.

Growth of porphyroblasts in the Shuswap Lake area occurred from early D_2 to syn- D_3 . K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dates from hornblende of the Shuswap Range (Chapter 4 and Okulitch 1979) indicate that these rocks were at temperatures of about 450-500°C in the Late Cretaceous. Since this is within the temperature range of staurolite stability in pelites at low to moderate pressures (Spear and Cheney 1989), it is possible that staurolite growth and D_3 occurred in Late Cretaceous or later time. However, considering that the style and orientation of D_3 structures in the Eagle Bay and in the Selkirks is regionally consistent, and that rocks of the Shuswap Range and western Selkirks would have been close together and at an equivalent structural level before Tertiary extensional faulting, D_3 could be of Early Cretaceous age in both areas.

Summary: contrasts between footwall and hanging wall

The main fabric in footwall rocks of the Three Valley and Hunters Range assemblages is a composite gneissic foliation into which compositional layering has been variably transposed. The foliation and associated tight folds formed in upper amphibolite facies under peak metamorphic conditions; hinges of these folds display a variably recrystallized earlier foliation parallel to the folded layering, and hence the folds are termed F_2 . Regional arguments suggest that the main fabric development and metamorphism were Cretaceous or younger and probably were diachronous. Northeast-directed mylonitic fabric in the lower part of these footwall assemblages, which was related to large-scale thrusting and which appears to be a manifestation of heterogeneous D_2 strain, is at least in part latest Cretaceous to Paleocene in age. West-verging D_3 folds postdate most metamorphic crystallization and are latest Cretaceous (~70 Ma) or younger. West-directed mylonitic fabric in the upper part of the Hunters Range assemblage, related to normal-sense shearing of the Okanagan Valley fault system, is expressed by identical mineralogy to D_2 structures but is at least in part of Early to Middle Eocene age.

Like the footwall rocks, those in the hanging wall of the Okanagan Valley fault system display folds (F_2) that deform compositional layering and a locally preserved parallel S_1 foliation. The S_2 axial-plane schistosity, into which S_0 and S_1 have been variably transposed, is in turn folded by steeply inclined F_3 folds and crenulations. Current metamorphic and thermochronological data allow that D_3 in the Shuswap Range could have occurred in Late Cretaceous or later time, and therefore that it was at least in part synchronous with D_3 in the footwall. However, regional arguments suggest that D_2 and D_3 structures in the hanging wall are most likely of Middle Jurassic and Early Cretaceous age, respectively, and hence significantly older than D_2

and D₃ structures in the footwall. The implications are that structures in the hanging wall and the footwall are not correlative, and any earlier Mesozoic or older strain that may have occurred in the footwall rocks has likely been obliterated by Cretaceous and early Tertiary transposition and metamorphic recrystallization.

CONCLUSIONS

Tectonostratigraphic assemblages in the hanging wall and in the footwall of the Okanagan - Eagle River fault are, according to most evidence, not correlative with each other. The footwall assemblages are dominated by semipelitic paragneisses and probably are correlative with Neoproterozoic and older strata of the North American miogeocline; hangingwall assemblages include abundant micaceous quartzite, a distinctive clean marble, and a thick succession of grey calcareous phyllite that resemble lower Paleozoic units of the miogeocline and that, outside of this study area, contain Paleozoic fossils. Penetrative deformation and metamorphism of the hangingwall rocks probably occurred prior to Late Cretaceous time, whereas the deformation and metamorphism preserved in the footwall rocks is of Cretaceous to Eocene age. Within the footwall, northeast-directed mylonitic fabrics related to thrusting and west-directed mylonitic fabrics related to extensional faulting both formed in upper amphibolite facies. No crosscutting relations were observed between these two sets of fabrics, and it remains permissible that they formed in part concurrently. However, the west-directed extensional fabrics are nonannealed and are overprinted by kinematically congruent brittle structures and retrograde metamorphism, implying that their development outlasted that of the thrust-related fabrics.

Figure 3.1: Tectonic assemblage map of the southeastern Canadian Cordillera, simplified from Wheeler and McFeely (1991) and modified based on Carr (1991) and this study. Shuswap core complex as defined by Brown and Carr (1990) is mostly bounded by Eocene normal faults (thick black lines with barbs on hanging wall side), including the Okanagan Valley (OV), Columbia River (CR), and Slocan Lake - Champion Lake (SL) fault systems. Okanagan - Eagle River and Adams - North Thompson segments of OV step en échelon across a transfer zone west of Shuswap Lake. The OV system transects the sillimanite isograd (Sil; from Read et al. 1991) north of Adams Lake. Black lines of medium weight represent mainly northeast-verging thrusts, except along boundary between the Coast and Intermontane belts where the Fraser - Straight Creek strike-slip fault system is designated by opposing arrows. Southern Rocky Mountain Trench (SRMT) is a morphogeological lineament that defines the boundary between Omineca and Foreland belts. Unpatterned units are (in the Foreland belt) Lower Cambrian to early Tertiary miogeoclinal and platformal strata, and (in the Coast belt) Permian to Cretaceous marine sedimentary and volcanic rocks and metamorphic rocks of uncertain age. Jurassic and Paleozoic granitoid plutons in Omineca and Intermontane belts, and Cretaceous and Neogene volcanic rocks in Intermontane belt, are not shown (see Fig. 5.1, Chapter 5). Bv = Beaven fault, C = Clachnacuda inn terrane, M = Malton complex, MD = Monashee décollement, Mo = Monashee complex, V = Valhalla complex. Inset maps show morphogeological belts (IS: Insular; C: Coast; IM: Intermontane; O: Omineca; F: Foreland) and composite superterrane of the Canadian Cordillera. AB: Alberta; AK: Alaska; BC: British Columbia; WA: Washington; YT: Yukon Territory.

MORPHOGEOLOGICAL BELTS

SUPERTERRANES

Figure 3.2: Schematic stratigraphic sections showing proposed correlation between Hunters Range assemblage (Mount Mara and Blue Lake sections) and the Windermere Supergroup in the northern Monashee Mountains. Windermere section modified from Raeside and Simony (1983).

Figure 3.3: Lower hemisphere equal area projections of poles to main composite gneissic foliation ($S_0S_1 \pm S_2$) in rocks of the footwall of the Okanagan Valley fault system, including six domains within the Hunters Range assemblage and combined data from the Momich assemblage and Pukeashun granite. Best-fit great circle girdles calculated by computer eigenvector analysis are indicated for the Mara Lake - Eagle River, Ecological Reserve and Kingfisher Creek - Blue Lake domains, where the calculated poles (β) to the girdles are similar in orientation to F_3 fold axes. Combined data from the Ecological Reserve and Blue Lake domains yield a girdle with corresponding β of 169° - 16° (not shown), in close agreement with the girdles calculated for the separate domains. Inset shows L_2 and L_3 axial lineations (F_2 and F_3 fold axes, respectively) and poles to S_2 (axial planes of F_2 folds) from Hunters Range assemblage.

Figure 3.4: Lower hemisphere equal area projections of stretching lineations in the footwall of the Okanagan Valley fault system. Note the contrast between the approximately east-west trend of lineations in mylonites of the immediate footwall of the Okanagan - Eagle River fault and the more northeasterly trends deeper in the footwall. Lineations in mylonitic granite of the Pukeashun suite and mylonitic migmatite of the Monnich assemblage trend west-northwest and east-southeast.

Figure 3.5: Lower hemisphere equal area projections of poles to composite $S_0S_1 \pm S_2$ foliation in hangingwall assemblages of the Okanagan Valley fault system. Great-circle girdles, calculated by computer eigenvector analysis, are shown for domains in which the data show a reasonably good girdle distribution.

Figure 3.6: Lower hemisphere equal area projections of D_2 and D_3 elements in hangingwall assemblages. L_2 lineations (F_2 axes and intersections of S_0S_1 on S_2) in the Eagle Bay assemblage define southwesterly and southeasterly trending domains. L_2 axial lineations in the Lichen assemblage show a wide range in orientation. Stretching lineations in the Mount Fowler orthogneiss (unit Dg) trend northeasterly to southeasterly west of Shuswap Lake and trend north-south in the southwestern Shuswap Range; on local scale, these lineations parallel L_2 axial lineations in both the Eagle Bay and Lichen assemblages. L_3 lineations (F_3 crenulation axes) in the Eagle Bay display relatively consistent shallow northwest and southeast plunges; S_3 typically is steeply dipping.

Plate 7: Lithology of the Hunters Range assemblage.

(a) Interlayered pelite (p) and psammite-semipelite (s), typical of unit Hs, containing boudined thin amphibolite layer (a). Bright white layers are migmatite leucosome (leuc). Shoes are 30 cm long.

(b) Migmatitic pelite of unit Hs near Mara lookout. Note garnet porphyroblasts (arrows) with quartz pressure shadows, especially in dark layer above large pod of pegmatitic leucosome. Pen is 15 cm long.

(c) Migmatitic semipelite near Mara Lookout. Small diameter of hand lens is 25 mm.

(d) Semipelite with interlayered garnet amphibolite (dark coloured), unit Hs near Mara lookout. Garnet porphyroblasts appear light grey against darker matrix. Diameter of hand lens is 25 mm.

(e) Semipelite (s) with boudin of garnet amphibolite (dark coloured, below hammer), intruded by pegmatite (peg). Unit Hs near Mara lookout. Hammer shaft is 33 cm long.

(f) Layered amphibolite of unit Ha (dark) cut by pegmatite dike (light coloured, upper right portion of photo). Mara lookout area. Pack in lower right corner is 50 cm high.

(g) Closeup of amphibolite outcrop shown in (f). Dark layers are rich in hornblende; light-coloured layers contain alternating zones rich in plagioclase (Pl) and diopside (Di). Knife is about 8 cm long.

(h) Biotite-hornblende quartz dioritic gneiss of unit Hdh, Craigellachie station. Coin is 19 mm in diameter.

Plate 8: Lithology and structural style of the Hunters Range assemblage.

(a) Garnet-quartz-hornblende-biotite-plagioclase gneiss of unit Hh on ridge west of Mount Griffin. White lenticular pods are pegmatite.

(b) Tightly folded impure marble (unit Hm) near Blue Lake. Boudined calcsilicate layers stand out in relief. Hammer shaft is about 35 cm long.

(c) Interlayered semipelite (s), pelite (dark layer at arrow), psammite (ps) and garnet amphibolite (a) deformed by tight F_2 fold near Mara lookout. Biotite defines axial planar S_2 foliation in pelitic layer (arrow). Hammer shaft is 33 cm long; south is to left.

(d) Migmatitic pelite folded by steeply inclined F_3 folds near Blue Lake. Pen is 15 cm long; west is to right.

(e) Cut surface of specimen of migmatitic pelite from Blue Lake area showing angular F_3 folds. Quartzofeldspathic layers, biotite (dark coloured) and prismatic sillimanite are folded. Scale bar 1 cm.

(f) Photomicrograph of specimen shown in (e), showing prismatic sillimanite (Sil) folded around F_3 . Fine-grained quartz and feldspar (QF) and small blades of sillimanite (distinguished by higher relief) occur between the larger sillimanite blades; dark mineral is biotite. Plane-polarized light; scale bar 0.5 mm.

(g) F_3 folds in pelite (dark grey) with thin semipelite interlayers (light grey). Folded pelites are detached from underlying psammities along décollement near bottom of photo (between arrows). Unit Hs, along boundary of ecological reserve west of Kingfisher Creek. Hammer left of centre is 33 cm long.

(h) Protomylonitic pegmatite on ridge northeast of Blue Lake. Asymmetrical extensional shears (left side of photo between arrows) dip easterly, consistent with east-directed shear sense. Hammer shaft is about 35 cm long. [Station 132].

Plate 9: Structural and metamorphic relations in the Eagle Bay assemblage.

(a) Interlayered pelitic and semipelitic schists (grey) and quartzite (thin white layers), Eagle Bay assemblage near Queest Mountain. Quartzite layers outline tight F_2 folds, to which main penetrative foliation (S_2) is axial planar. Pelite layers show well developed S_3 crenulation cleavage. Coin in centre of photo is 2.5 cm in diameter.

(b) Opaque inclusions (arrow) in garnet porphyroblast (Grt) define an internal foliation (S_i) that makes a high angle with external S_2 defined by muscovite and biotite. S_i is interpreted as a D_1 fabric and the garnet is interpreted as having grown after D_1 and before, or early during, D_2 . Plane-polarized light; scale bar 1 mm. [Sample J88-464, from North Queest Mountain].

(c) Garnet porphyroblast (Grt) with straight trails of quartz inclusions (S_i) that are subparallel to the form surface of external S_2 . External S_2 , defined by matrix muscovite (Ms) and quartz, is deformed by F_3 crenulations and S_3 crenulation cleavage that do not deform S_i ; hence, S_i in this case is a D_2 fabric and the garnet is late or post- D_2 and pre- D_3 . Garnet is rimmed with biotite (Bt; dark grey) and quartz (Qtz). Crossed nicols; scale bar 1 mm. [Sample LJ89-853, from Queest Mountain].

(d) Staurolite porphyroblasts (St), with inclusions of opaque minerals and quartz defining S_i that is parallel and continuous with S_2 in matrix muscovite and biotite. Matrix S_2 is folded by F_3 crenulations, and the staurolites are interpreted as having grown during D_3 . Plane-polarized light; scale bar 1 mm. [Sample J88-464].

(e) Staurolite porphyroblast with trails of opaque inclusions (S_i) that outline a gentle fold; F_3 crenulations that deform external S_2 are, in contrast, much tighter, and therefore staurolite is interpreted as having grown early during D_3 . Matrix is intergrown muscovite and biotite with minor opaques. Plane-polarized light; scale bar 1 mm. [Sample J88-464].

(f) Same as (e), with crossed nicols.

(g) Staurolite porphyroblast with S_i defined by opaque and quartz inclusions. S_i is parallel and continuous with matrix S_2 and, apparently, with F_3 crenulations on left side of porphyroblast; but it makes a high angle with S_2 on the right. Porphyroblast is interpreted as having grown before, or early during, D_3 . Plane-polarized light; scale bar 1 mm. [Sample J88-464].

(h) Staurolite porphyroblasts that overgrow gently warped S_p defined by quartz and opaque inclusions. S_i is continuous with external S_2 in the area between the porphyroblasts but makes a high angle with S_2 along the upper right edge of the staurolite on the right. S_2 in the matrix is folded by F_3 crenulations that are tighter than the warps which deform S_p ; staurolites are interpreted as having grown early during D_3 . Plane-polarized light; scale bar 1 mm. [Sample J88-464].

CHAPTER 4

$^{40}\text{Ar}/^{39}\text{Ar}$ and U-Pb geochronology and implications for Cretaceous to Middle Eocene thermotectonic history of the Omineca belt between Three Valley Gap and Adams Lake

INTRODUCTION

Geochronological studies have highlighted the importance of early Tertiary thermotectonic processes in the evolution of the Omineca belt of southern British Columbia and northern Washington. U-Pb studies in conjunction with careful field mapping have led to the recognition of widespread Paleocene-Eocene mid-crustal granite magmatism (Carr et al. 1987; Parrish et al. 1988; Sevigny et al. 1989; Carr 1990, 1992; Whitehouse et al. 1992; Parkinson 1992; Scammell 1993; Bardoux 1993) and have demonstrated that regional crustal extension in the southern Omineca belt began at most 2 million years after (Carr 1992), and more likely during the final stages of (Carr 1990, Scammell 1993), contractional faulting. K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dating has shown that the high-grade metamorphic rocks of the Shuswap complex, a metamorphic core complex within the southern Omineca belt (Chapter 1, Fig. 1.1), were rapidly cooled from high temperatures during the Early and Middle Eocene (Medford 1975, Ewing 1980, Mathews 1981, Armstrong 1982, Parrish et al. 1988, Bardoux 1993, Scammell 1993). The rapid cooling has been attributed to tectonic removal of the upper crust along low to moderate-angle normal faults and associated

shear zones that bound the complex (see Parrish et al. 1988 for a summary) and that were in part localized by the granite melts (Carr 1992; Chapter 2). Rocks in the hanging walls of these normal faults, in contrast, generally have Cretaceous and older K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages.

This chapter presents the results of $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende and U-Pb zircon and monazite geochronological work conducted as part of a structural and tectonic study of the west-central margin of the Shuswap complex near Shuswap Lake, British Columbia (Fig. 4.1). The western margin of the complex is delineated by the west-side-down Okanagan Valley normal fault system (Chapter 2), which is known to have been active in the Middle Eocene (Parkinson 1985, Tempelman-Kluit and Parkinson 1986, Parrish et al. 1988, Bardoux 1993) but along which the time of first displacement is uncertain. The main objectives of this work were to gain information on the timing of shearing from exhumed mylonites of the Okanagan Valley fault system, to compare cooling histories of rocks in the hanging wall and the footwall of the system, and to see if any additional information could be gained regarding the timing of the transition between thrusting and extension. A summary of the geochronological results is given with sample locations in Figure 4.1.

AMPHIBOLE $^{40}\text{Ar}/^{39}\text{Ar}$ GEOCHRONOLOGY

Amphiboles from twelve samples of amphibolite and hornblende gneiss were dated using the $^{40}\text{Ar}/^{39}\text{Ar}$ technique. Of these, seven were from upper amphibolite-facies rocks of the Shuswap complex, in the footwall of the Okanagan - Eagle River fault (the main segment of the Okanagan Valley fault system, see Fig. 4.1 and Chapter

2). These were selected in order to obtain dates from an east-west transect, subparallel to the extension direction and adjacent to the Trans-Canada Highway, through the tectonically denuded part of the thrust sheet above the Monashee décollement, a crustal-scale Mesozoic-Paleocene thrust zone (Brown et al. 1992a). The other five samples were from structurally higher and further northwest, between Shuswap Lake and Adams Lake. The sampling strategy for these was to obtain dates from an east-west transect and to test the hypothesis that a belt of Devonian orthogneiss (the Mount Fowler gneiss; Okulitch et al. 1975) was detached from underlying high-grade metamorphic rocks along a splay of the Okanagan Valley fault system (Johnson 1990). Recent mapping has since indicated that the region between Shuswap Lake and Adams Lake is a transfer zone between two en échelon segments of the Okanagan Valley fault system (Shuswap Lake transfer zone) within which the limited evidence for such a detachment is restricted to an area west of Lichen Mountain (Fig. 4.1; Chapter 2). Sample locations are shown in Figure 4.1 and are described in Appendix B.

Analytical methods

Amphibole separates greater than 99% pure were obtained using conventional crushing, grinding, sieving, magnetic separation and heavy liquid techniques. 250-mg samples of 80-100 mesh (177-149 μm) fraction were wrapped in Al foil and irradiated at the MacMaster University Reactor in Hamilton, Ontario, for 33.6 h. Neutron flux was monitored with biotite standards LP-6 (127.9 ± 1.1 Ma; Odin et al. 1982) and SP-85 (18.65 ± 0.05 Ma; Queen's University intralaboratory standard, D. A. Archibald pers. comm. 1993). Irradiated samples were loaded into crimped Nb-sheet crucibles, heated in a quartz tube using a Lindberg resistance furnace, and analyzed using an ultrahigh-vacuum extraction system online with a highly modified AEI MS-10 mass

spectrometer run in static mode at Queen's University in Kingston, Ontario. Each sample was heated in 10-12 steps of approximately 30 minutes each, from 850-1200°C (heating schedules are given in Figs. 4.2-4.13). Isotopic ratios were corrected by online computer for mass fractionation, mass-spectrometer sensitivity, atmospheric contamination, interference reactions from K and Ca, and time elapsed since irradiation. Apparent $^{40}\text{Ar}/^{39}\text{Ar}$ ages for each step were calculated using the decay constants of Steiger and Jäger (1977). Analytical uncertainties, reported at the 2σ level, were derived using the equations of Dalrymple et al. (1981) and do not include uncertainty in the irradiation parameter J . Addition of $\pm 0.5\%$ to the quoted errors (± 0.3 Ma for samples with plateau ages in the range 50-60 Ma) could reasonably account for uncertainty in J .

Five points on one grain from each sample were analyzed by electron microprobe at the Geological Survey of Canada in Ottawa (Appendix C). All of the grains proved to be of the calcic amphibole group, most of them being tschermakitic or ferroan pargasitic hornblendes (terminology recommended by the International Mineralogical Association; Leake 1978). J89-642H (ferro-tschermakite) and J89-788H (magnesian hastingsite) were lower in Si than true hornblendes.

Following criteria recommended by Fleck et al. (1977), plateau ages were calculated wherever at least three consecutive steps, each consisting of at least 3% and together consisting of at least 50% of the total ^{39}Ar evolved from the sample, give the same apparent age within 2σ uncertainty. In every plateau interval, Ca/K ratios calculated from $^{37}\text{Ar}/^{39}\text{Ar}$ isotopic ratios for each step have a mean deviation of no more than 12% ($\leq 7\%$ in footwall rocks) (Appendix D). The mean ratio generally agrees to $\pm 9\%$ with the Ca/K ratio calculated from the average of the 5 microprobe analyses, and in every case except sample J89-3H is within the range of values

determined from these analyses.

Results and interpretation

Amphiboles from the footwall of the Okanagan - Eagle River fault gave Late Paleocene to earliest Middle Eocene apparent ages. Five of the seven samples yielded plateaus, over which generally >80% of the ^{39}Ar was released. The plateau ages are considered to represent the time of closure (Dodson 1973) of the amphiboles to release of radiogenic ^{40}Ar . The lowest-temperature step (and in some cases the second step) of most samples gave an apparent age much older than the plateau age. During these anomalous steps, less than 3% of the ^{39}Ar was released and the Ca/K ratio differed markedly from that of the amphibole. Harrison and McDougall (1980b) have made a strong case, through high-resolution analyses, for initial steps with unreasonably high apparent ages such as these being caused by small amounts of excess ^{40}Ar near the outer surfaces of hornblendes.

J89-18 gave the youngest apparent age, a plateau of 50.5 ± 1.7 Ma (Fig. 4.2). This hornblende was from the westernmost of two amphibolites collected from the immediate footwall of the fault. The other is J88-514H, which gave a plateau age of 56.8 ± 3.3 Ma (Fig. 4.3). This age was derived from three steps (1075, 1095, and 1110°C) that constituted only 58% of the analyzed ^{39}Ar , because operator errors during gas extraction rendered results from the 1050°C step uninterpretable and resulted in loss of the 1130°C step. Nevertheless, apparent ages from the 940, 985, 1030, and 1150°C steps (during which collectively another 38% of the ^{39}Ar was released) are within error of the plateau age, and this age is therefore thought to be the best estimate of the closure age for this sample.

A well-defined plateau age of 56.5 ± 3.2 Ma was derived from J89-36 (Fig.

4.4), a hornblende quartz diorite from the Trans-Canada Highway. Two amphibolites from ridges in the Hunters Range, J88-351A (Fig. 4.5) and J89-905 (Fig. 4.6), yielded similar plateau ages of 58.3 ± 3.8 Ma and 54.0 ± 3.1 Ma, respectively.

Only the easternmost of the footwall samples, J89-3H (Fig. 4.7) and J89-788H (Fig. 4.8), did not yield plateaus. J89-3H is from an amphibolite at Three Valley Gap, the structurally deepest of all the rocks sampled. J89-788 is from a mafic gneiss at Eagle Pass Ridge, just north of Three Valley Gap. Age spectra from both of these samples show minima in the intermediate steps, with older apparent ages from the low and high-temperature steps. Similar "saddle-shaped" spectra are characteristic of materials that contain excess ^{40}Ar (Lanphere and Dalrymple 1976), including well-documented cases of metamorphic hornblendes (Harrison and McDougall 1981). However, release patterns that show an increase in apparent age toward the high-temperature end of the spectrum are also characteristic of materials that either have lost ^{40}Ar by diffusion in response to episodic reheating (Turner 1968) or that have cooled slowly through their closure temperatures (McDougall and Harrison 1988); if excess ^{40}Ar is subsequently absorbed by such a material, the intermediate-temperature steps would define minima in the resulting spectrum (e.g. Harrison and McDougall 1980b). The resolution of the analyses in the present study is below that required for positive identification of diffusional loss or slow cooling profiles. In any case, the apparent-age minimum of the youngest step can be considered to be the maximum age of final closure of the mineral to loss of ^{40}Ar (e.g. Copeland et al. 1991). Thus, maximum closure ages of 58.3 ± 2.5 Ma and 53.8 ± 1.4 Ma can be inferred for J89-3H and J89-788H, respectively.

Amphiboles from the hanging wall of the Okanagan - Eagle River fault in the Shuswap Lake transfer zone yielded generally more complicated age spectra that

indicate Mesozoic cooling histories. J89-970, from a granodiorite gneiss of the Devonian Mount Fowler suite, yielded a plateau over the 940-1060°C steps with a corresponding age of 74.2 ± 3.6 Ma (Fig. 4.9). In detail, the apparent age gradually increases with increasing temperature from about 70 Ma to about 80 Ma, but the anomalously low Ca/K ratio of step 1 probably reflects contamination by a phase other than amphibole. The 70 Ma age minimum therefore is probably meaningless, and the plateau age is considered to be the best estimate of the closure age of the sample.

J89-662H, from a screen of foliated mafic quartz diorite in the Mount Fowler granodiorite gneiss, yielded a plateau age of 110.6 ± 5.6 Ma over the 940-1115°C interval (Fig. 4.10). Apparent ages of the two higher-temperature steps are older by about 15 m.y. The lowest-temperature step gave a considerably older apparent age of about 290 Ma that, as in the footwall samples, is probably an artifact of excess ^{40}Ar . A similar release pattern is exhibited by J90-1092, from a quartzofeldspathic gneiss structurally beneath the Mount Fowler orthogneiss (Fig. 4.11). The first step, with an anomalously high apparent age of about 230 Ma likely owing to excess ^{40}Ar , is followed by a plateau that corresponds to an age of 87.1 ± 2.6 Ma. The final steps have apparent ages several m.y. older than the plateau age. The plateau ages of J89-662H and J90-1092 are interpreted as closure ages.

The most complex data are from two samples from near Adams Lake. J89-417H is from an amphibolite layer in fibrolite-bearing quartz-mica schists and psammites, structurally beneath the Mount Fowler orthogneiss. The age spectrum (Fig. 4.12) exhibits an anomalously high apparent age of ~500 Ma for the initial step, a young (~100 Ma) second step, and an anomalous age peak over the third and fourth steps of ~285 Ma. A minimum of 99 ± 20 Ma in step 5 (1060°C) is followed by a gradual increase to a short plateau over steps 8-9 corresponding to an age of 158 ± 9

Ma. J89-642H (ferro-tschermakite) is from an amphibole-chlorite schist in upper greenschist-facies rocks (garnet zone) of the Eagle Bay assemblage, structurally above the Mount Fowler orthogneiss. The release pattern (Fig. 4.13) is grossly similar to that of J89-417H, but the apparent ages are older. A peak of ~200 Ma occurs in step 3 between minima of about 160 Ma in steps 2, 4 and 5. From step 5 the apparent age gradually increases to 202 ± 17 Ma over four steps.

Slopes in the high-temperature portions of the age spectra for both J89-417H and J89-642H approximate episodic loss or slow cooling profiles. Slow cooling cannot be ruled out, but it is considered less likely than episodic reheating in the case of J89-417H because this sample was from a structural level in which granites and pegmatites, some of which are known to be of Cretaceous to early Tertiary age, are widespread (Chapters 2 and 3). The apparent-age minimum of 99.1 ± 20 Ma is interpreted as the maximum age of resetting and final closure (e.g. Harrison and McDougall 1980b). This compares well with plateau ages of the samples from further east, and is consistent with heating related to mid- to Late Cretaceous plutonism. The maximum closure age for J89-642H based on its apparent-age minimum is 159 ± 17 Ma, significantly older than closure ages inferred from the other amphiboles of this study; implications of this will be discussed later.

According to the Turner (1968) model for episodic ^{40}Ar loss, the apparent ages at the high-temperature ends of the spectra of J89-417H and J89-642H would represent minimum ages of crystallization or previous closure to radiogenic ^{40}Ar loss at, respectively, about 160 and 200 Ma. However, as mentioned earlier, without very high resolution it can be difficult to discriminate between patterns of episodic loss or slow cooling and those produced as a result of excess ^{40}Ar alone. For example, Heizler and Harrison (1988) have shown that multiple components of trapped excess

^{40}Ar can lead to complex release patterns. This is because age spectrum diagrams have the intrinsic assumption that the $^{40}\text{Ar}/^{36}\text{Ar}$ composition of the argon trapped in the system is uniform and equal to the atmospheric value (Merrihue and Turner 1966, McDougall and Harrison 1988). If this assumption is incorrect, anomalous peaks will appear in the spectrum. The isotopic composition of trapped argon can often be resolved through the use of isotope correlation diagrams, especially inverse isochron diagrams wherein $^{36}\text{Ar}/^{40}\text{Ar}$ is plotted against $^{39}\text{Ar}/^{40}\text{Ar}$ (Roddick 1978, Heizler and Harrison 1988). In a concurrent study of amphiboles that were analyzed consecutively with those reported here, Scammell (1993) concluded that high system blanks and low volumes of non-blank ^{36}Ar rendered such correlation plots meaningless. Since the sample sizes, the quantities of gas evolved during step heating, and the resultant age spectra of these amphiboles were comparable to those of the present study, attempts to use such diagrams in this study were abandoned. Consequently, whether apparent-age gradients are a function of excess ^{40}Ar , slow cooling, or episodic diffusional loss is unresolved.

U-Pb ZIRCON AND MONAZITE GEOCHRONOLOGY

Zircon and monazite obtained from deformed pegmatites and granite from four localities were dated by U-Pb methods in order to put limits on the timing of (i) northeast-directed ductile thrusting above the Monashee décollement, (ii) west and northwest-directed normal-sense ductile shearing associated with the Okanagan Valley fault system, and (iii) west-vergent folding during D_3 deformation in the part of the Shuswap complex that lies structurally beneath the Okanagan Valley fault system and

above the Monashee décollement.

Analytical procedures

Heavy mineral separates were obtained by conventional crushing, grinding, Rodgers gold table, and heavy liquid techniques at Carleton University. All other procedures were carried out at the Geological Survey of Canada in Ottawa. Zircon and monazite were recovered by Frantz magnetic separation followed by sieving and handpicking. Crystals were selected for clarity, euhedral morphology and, where possible, lack of inclusions and cracks. All zircon fractions were air-abraded following the method of Krogh (1982) to reduce the effects of low-temperature Pb loss near crystal surfaces. Sample dissolution and column chemistry were performed using methods and materials modified from those of Krogh (1973) as outlined by Parrish et al. (1987), Parrish (1987), and Roddick et al. (1987). Samples were spiked with a mixed ^{205}Pb - ^{233}U - ^{235}U isotopic tracer (Parrish and Krogh 1987, Parrish et al. 1987) prior to dissolution. Isotopic compositions of Pb and U were measured using a Finnigan-MAT 261 variable multicollector mass spectrometer equipped with a secondary electron multiplier, as described by Roddick et al. (1987). Measured isotopic ratios, reported in Table 4.1, have been corrected for mass fractionation, common Pb, blank Pb and U, and spike. For the common Pb correction, a model Pb isotopic composition derived from the growth curve of Stacey and Kramers (1975) for the $^{207}\text{Pb}/^{206}\text{Pb}$ age of the sample was assumed. Decay constants used for derivation of ages were those recommended by Steiger and Jäger (1977). Total procedural Pb blanks were generally in the range of 10-35 pg, although a few analyses had higher blanks (Table 4.1); U blanks were 0-5 pg. Analytical uncertainties, quoted in Table 4.1 as one standard error of the mean (1σ) for isotopic ratios and as 2σ for the

corresponding ages, include uncertainty in all relevant components of the analysis (Roddick 1987). The $^{206}\text{Pb}/^{238}\text{U}$ and $^{207}\text{Pb}/^{206}\text{Pb}$ isotopic ratios and dates were corrected, as outlined by Schärer (1984) and Parrish (1990), to compensate for excess or deficient ^{206}Pb resulting from Th fractionation during crystallization of monazite and zircon, respectively (Schärer 1984). These corrected ratios and ages are reported in Table 4.2.

Sample descriptions, results and interpretation

J89-788P

Sample J89-788P is from a sheet of sheared pegmatite and pelitic migmatite on Eagle Pass Ridge, north of Three Valley Gap and about 1.3 km structurally above the Monashee décollement (Fig. 4.1). The pegmatite sheet, mapped by Bosdachin (1989), is 300 m thick and marks a zone of high strain between structurally overlying pelitic to psammitic paragneisses of the Hunters Range assemblage (Chapter 3) and underlying biotite-hornblende-plagioclase gneisses and semipelitic paragneisses that contain large boudins of amphibolite (Three Valley assemblage; Chapter 3). Screens of marble and calcsilicate gneiss are common within the zone.

The pegmatite (Plate 10 *a, b*) is everywhere foliated and is well-lineated in zones of highest strain. In coarser-grained layers of pegmatite, feldspar porphyroclast systems (Passchier and Simpson 1986) indicate predominantly NE-directed (i.e. upper member to the NE) shear sense. Mylonitic shears, ~10-15 cm wide, have gradational contacts with the coarser-grained pegmatite. Most of these shears have retrograde chlorite after biotite on shear planes (C-surfaces of Berthé et al. 1979a) and chlorite pressure shadows on feldspar porphyroclasts indicate SW-directed shear. Other shears

have black matrix rich in nonretrograded biotite, and have feldspar porphyroclast systems that indicate NE-directed shear (Plate 10 *c, d*). Although crosscutting relations were not observed between these two sets of shears, lack of retrogression in the NE-directed ones suggests that they are older. In the regional context, the NE-directed shears are interpreted as having accommodated thrust displacement, whereas the SW-directed ones are most likely related to Eocene normal faulting and tectonic denudation. The pegmatite is therefore regarded as having been emplaced before or during the last stages of thrusting.

Sample J89-788P was collected from part of the zone in which NE-directed kinematic indicators were displayed to ensure that the strain was not younger than thrusting. The sample was a mylonitic pegmatite containing disseminated fine-grained biotite and thin biotite seams interpreted as being relict metapelite.

Four multigrain fractions of zircon and one fraction comprised of three monazite crystals were analyzed (Table 4.1, Fig. 4.14). The zircon fractions represent three different morphologies: subequant multifaceted crystals, elongate bipyramidal prisms with aspect ratios of 1:3 to 1:5, and stubby bipyramidal prisms with multifaceted terminations. All of the zircons analyzed are interpreted as being of magmatic origin, because prismatic and sharply faceted forms are characteristic of magmatic zircons (van Breemen et al. 1987).

The stubby prisms (fraction 3) are highly discordant and have a corrected $^{207}\text{Pb}/^{206}\text{Pb}$ age of 333.0 ± 3 Ma, reflecting inheritance of Mississippian or older age. Fraction 2 (subequant multifaceted zircons) is barely concordant before correction for Th deficit. After correction, the error ellipse overlaps concordia at 66.3 ± 0.1 Ma (Fig. 4.15). A second fraction of the same morphology (fraction 1), and elongate prisms (fraction 4), plot below concordia. Analyses 1, 2, and 4 fit a chord through the

origin with an upper intercept of 67.2 ± 2.2 Ma and MSWD of 2.19, calculated by regression using the method of York (1969) modified as outlined by Roddick (1987). The Th-corrected $^{207}\text{Pb}/^{206}\text{Pb}$ ages of fractions 1 and 4 are slightly older than the concordant date from fraction 2 but are within error of the upper intercept date. Discordance of fractions 1 and 4 is therefore attributed to recent Pb loss, an interpretation further supported by the correlation of younger U/Pb ages with greater U concentration. The concordant date, considered together with the upper intercept of the chord through fractions 1, 2 and 4, are consistent with a crystallization age of $67.2 \pm 2.2/-1.0$ Ma for the pegmatite.

Monazite fraction M-5 shows reverse discordance, which typically is attributed to excess ^{206}Pb resulting from initial incorporation and subsequent decay of ^{230}Th in monazite (Schärer 1984). Correction for Th/U fractionation by the method of Schärer (1984) caused a shift in the position of the error ellipse to well below concordia (Fig. 4.14). This dramatic overcorrection implies that the intrinsic assumptions of this method are incorrect. A likely reason is that the Th/U ratio of the whole rock used to calculate the fractionation factor $f = (\text{Th}/\text{U})_{\text{mineral}}/(\text{Th}/\text{U})_{\text{magma}}$ (Schärer 1984) does not reflect the Th/U that was available at the time of monazite growth. In the case of monazite M-5, f (Table 4.2) has apparently been overestimated, suggesting that Th/U available during monazite growth was higher than that in the whole rock. Parrish (1990) has pointed out that sequential crystallization of accessory minerals can lead to fractionation of U or Th in magmatic systems. In the case of J89-788P, earlier crystallization of zircon may have tied up uranium and increased the effective Th/U available for subsequent monazite crystallization.

Because $^{207}\text{Pb}/^{235}\text{U}$ ages are unaffected by Th/U disequilibrium, for monazite they generally are still meaningful even when the corrections do not lead to

concordant dates (Parrish 1990). The $^{207}\text{Pb}/^{235}\text{U}$ age of monazite M-5 is 58.2 ± 0.2 Ma, 8 m.y. younger than the crystallization age of the pegmatite inferred from zircon data. This date could represent the age of monazite growth subsequent to crystallization of the pegmatite or it could be a time of Pb loss from initially older magmatic monazite. The euhedral morphology of the crystals suggests a magmatic origin (see Parrish 1990), and thus if they grew after the pegmatite was emplaced they likely did so at supersolidus temperature in local anatectic melts. If the monazites grew during crystallization of the pegmatite, then the relatively young monazite date likely reflects post-crystallization Pb loss.

In summary, the zircon data are consistent with a crystallization age of $67.2 \pm 2.2/-1.0$ for the pegmatite. The younger monazite date of 58.2 ± 0.2 Ma is interpreted as representing a time of either subsequent partial melting or high-temperature Pb loss.

J90-1059

Sample J90-1059 is a pegmatitic migmatite from near the Mara fire lookout in the central Hunters Range (Fig. 4.1) that has been deformed by an F_3 fold. F_3 folds in the Mount Mara area are overturned toward the west, have moderately east to northeast-dipping axial planes, and crenulate the peak-metamorphic minerals such as biotite and sillimanite (Chapter 3). At the sample site, migmatitic sillimanite-garnet-biotite semipelite and garnet-biotite psammite are intruded by pegmatite sheets that nearly everywhere lie subparallel to compositional layering. About 70% of the rock is pegmatite and leucosome; the rock sampled for dating was mostly leucosome. The leucosome, which is coarse-grained to pegmatitic and locally has a weak layer-parallel foliation defined by quartz aggregates, clearly is folded by F_3 folds (Plate 10 *e*); the

thicker pegmatite sheets locally were observed to also follow layering around fold hinges. The pegmatite and leucosome are therefore interpreted to be as old as, or older than, D_3 deformation. A few hundred metres west of the sample locality, discrete, gently east-dipping, west-directed shears locally cut F_3 folds in similar rocks (Plate 10 f), and at the sample locality the limbs of some folds have been sheared out (Plate 10 g). The shears are up to 30 cm wide and are defined by preferred orientation of biotite, sillimanite, and dynamically recrystallized quartz and feldspar. They are either backrotated extensional structures related to the Okanagan Valley fault system (Johnson 1989, 1990) or minor backthrusts related to ramping on the Monashee décollement (Chapter 3).

Three multigrain zircon fractions and one fraction comprising a single crystal tip gave variably discordant analyses (Fig. 4.16; Table 4.1). Regression of the zircon data resulted in a chord with intercepts at 2345 ± 56 Ma and 69 ± 1 Ma (MSWD = 45.9). Elongate prismatic zircons (4A) and the tip of a prismatic crystal (4C) plot close to concordia near 72 Ma. Discordance is interpreted as being due mainly to inheritance rather than Pb loss, because the crystals were strongly abraded before analysis and some crystals contained very small inclusions that morphologically resembled zircon. The lower intercept date is therefore interpreted as approximating the crystallization age of the leucosome. Scatter in the data indicated by the high MSWD is probably a result of multiple ages of inheritance, and therefore the upper intercept age may have little geological meaning other than indicating Precambrian inheritance (in part at least 1495 Ma based on the $^{207}\text{Pb}/^{206}\text{Pb}$ age of fraction 6A).

Two single-crystal monazite analyses plot above concordia (Fig. 4.16). M-3, a clear yellow, euhedral, tabular crystal of probable magmatic origin, yielded a $^{207}\text{Pb}/^{235}\text{U}$ age of 53.8 ± 0.2 Ma. M-2 was a pale yellow, very clear, subhedral, tabular crystal

with slightly rounded facets and with a thin, pale orange core zone approximately 1/3 the length of the crystal; M-2 plots closer to concordia than M-3 and has an older $^{207}\text{Pb}/^{235}\text{U}$ age of 60.6 ± 0.4 Ma.

The faceted morphology of the monazites would tend to favour a magmatic origin. If the leucosome is about 70 Ma old as suggested by the zircon data, then the younger monazite dates can be attributed most readily to high-temperature partial Pb loss from the monazites after crystallization. A $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende plateau age of 58.3 ± 3.8 Ma from J88-351A (Figs. 4.1, 4.5) indicates that rocks presently about 1 km away from Mara lookout had cooled well below the nominal closure temperature of monazite (ca. $725 \pm 25^\circ\text{C}$; Parrish 1990) by 54 Ma, the $^{207}\text{Pb}/^{235}\text{U}$ age of monazite M-3. Therefore, unless the migmatite at Mara lookout was reheated to $\geq 700^\circ\text{C}$ in the Eocene by a local source (the pegmatite?), it would appear that the monazites (in particular M-3) lost Pb at a temperature of perhaps $\sim 500^\circ\text{C}$. Since both monazites M-3 and M-2 had very high U concentrations (> 5000 ppm; Table 4.1), this interpretation is consistent with recent conclusions of Parrish and Carr (1994) that monazite with U concentrations greater than ~ 4000 ppm can lose Pb at temperatures as low as 400°C .

Correction for Th/U fractionation did not cause a significant change in the position of the error ellipses of either the monazites or zircons (Fig. 4.17). Discordance of the zircons is attributed to inheritance, as discussed above. However, the fact that the monazites remain reversely discordant after correction suggests that the calculated fractionation factors (Table 4.2) are too low, for which the simplest explanation is that the Th/U ratio of the whole rock overestimates the Th/U that was available for monazite growth. Possible reasons for this include the presence of inherited monazite (for example restitic crystals in the metapelite component of the migmatite) or punctuated growth of monazite during progressive metamorphism and

partial melting.

In summary, the zircon data from J90-1059 suggest that the leucosome is Late Cretaceous, approximately 70 Ma old. The younger monazite dates probably reflect Pb loss due to open-system behaviour correlated with very high U content; an alternative possibility, that Pb loss was caused by local reheating in the Eocene, could be tested by dated pegmatites from the same locality, as these are the only likely sources of local heat. In the regional context, considering that the leucosome and pegmatite have been folded by F_3 folds, the U-Pb data from this rock indicate that D_3 deformation in the Hunters Range occurred, or at least ended, no earlier than about 70 Ma.

J87-174

Migmatitic pelite and semipelite mylonites in the footwall of the Okanagan - Eagle River fault are well exposed in a long roadcut on the Trans-Canada Highway 3 km northeast of Sicamous. These rocks contain the synkinematic mineral assemblage garnet-sillimanite-biotite-plagioclase-K-feldspar-quartz and have asymmetrical fabrics indicative of west-directed shear. Sample J87-174 is from a several-metre-thick sheet of garnet-sillimanite-biotite granite pegmatite that is sheared and transposed into parallelism with the mylonitic foliation, and is therefore clearly older than at least the last stages of ductile shearing.

Three discordant zircon fractions fit a chord that has upper and lower intercepts of $150.0 \pm 5.6/-5.0$ Ma and $60.0 \pm 5.6/-6.3$ Ma, respectively (MSWD = 0.71; Fig. 4.18). The chord could be a mixing line reflecting a combination of Jurassic and Paleocene components; however, zircons within each fraction were physically similar, and as such there is no reason to strongly suspect that individual zircons in a given fraction

are of different ages. No cores were visible in any of the zircons, although several contained small unidentified inclusions. Some crystals in each fraction exhibited zones of fracturing and moderate radiation damage around their central portions that were not completely removed by abrasion. All of the zircon fractions had very high U concentrations, and those with the higher U concentrations yielded younger U-Pb ages (Table 4.1). This pattern is consistent with discordance due to radiation damage and loss of radiogenic Pb, where more Pb loss was incurred by more U-rich zircons. In light of these observations, the discordia is most reasonably interpreted as reflecting crystallization of the pegmatite in the Late Jurassic and subsequent partial loss of radiogenic Pb. The Paleocene - Early Eocene lower intercept date suggests that Pb loss occurred at a time when these rocks are known to have been at high temperature, and within uncertainty of a time when shearing along the Okanagan Valley fault system is known to have been occurring (see discussion). Lattice damage enhanced by a combination of radiation and shearing deformation probably enabled Pb loss at high temperature.

Two multigrain monazite analyses yield Late Cretaceous $^{207}\text{Pb}/^{235}\text{Pb}$ ages of 81.8 ± 0.2 Ma and 71.4 ± 0.2 Ma. These dates may also be interpreted as representing variable amounts of high-temperature Pb loss during deformation, or they could represent different times of growth. Fraction M-4 consisted of two clear yellow, euhedral, highly faceted crystals of probable magmatic origin, whereas M-5 included rounded crystals of probable metamorphic origin that could have grown at different times. M-4 had a much higher U concentration than M-5 (Table 4.1) and yielded a younger apparent age, consistent with a greater degree of Pb loss. Failure of the Th/U fractionation correction to return the error ellipses to concordia in this case could be explained by either punctuated monazite crystallization (see J90-1059 above) or Pb

loss; because the concentration of Th in the monazite is calculated from ^{208}Pb , loss of ^{208}Pb would result in an underestimate of $(\text{Th}/\text{U})_{\text{mineral}}$ and an underestimate of f . Leaching of U from the rock after crystallization, for example by fluids in the shear zone, would lead to an overestimate of $(\text{Th}/\text{U})_{\text{rock}}$ and thus also to an underestimate of f .

Pegmatite J87-174 is interpreted as most likely having crystallized in the latest Jurassic (between 156 and 145 Ma). Zircons apparently lost variable amounts of radiogenic Pb at high temperatures during early Tertiary shearing. Late Cretaceous monazite dates are inferred to reflect some combination of punctuated metamorphic growth and high-temperature Pb loss from older monazite.

J89-762

In the Shuswap Lake transfer zone (Fig. 4.1), ductile strain related to westerly directed extension is distributed heterogeneously through two-mica leucogranites of the Pukeashun suite (Johnson 1990; Chapter 2). These granites form about 80% of the rock within a zone of high strain called the Pukeashun shear zone, which is interpreted as having been a major channel of displacement in the transfer zone. The Pukeashun suite consists of multiple intrusions of muscovite granite pegmatite and medium-grained muscovite and garnet-biotite-muscovite monzogranites. Most of the granite is strongly to mildly foliated and well-lineated. Mylonitic fabrics in the zones of highest strain consistently indicate top to the west-northwest shear sense. Weakly foliated leucogranite dikes have thin apophyses that are sheared and flattened parallel to the mylonitic foliation, indicating that intrusion was coeval with NW-directed shearing (Plate 6e in Chapter 2). At Pukeashun Mountain, the mylonitic foliation dips southeastward and is interpreted as having been produced by normal-sense shearing

along a mid-crustal portion of the Okanagan Valley fault system that later was antiformally warped and abandoned during progressive crustal extension and exhumation (Chapter 2).

Sample J89-762 is from a 70-cm-wide dike, at Pukeashun Mountain (Fig. 4.1), of medium-grained garnet-biotite-muscovite monzogranite of the Pukeashun suite that cuts migmatitic semipelite gneiss and pegmatite with low-angle discordance (Plate 10 *h*). The dike is less highly strained than the migmatite country rock, but both are mylonitic and both exhibit microstructures that reliably indicate west-northwesterly directed ductile shear. The dike is therefore interpreted as a late synextensional intrusion.

Three multigrain fractions of prismatic zircon each gave discordant analyses, clustering near concordia between 55 and 57 Ma (Fig. 4.19; Table 4.1). Two single-crystal monazite analyses gave $^{207}\text{Pb}/^{235}\text{U}$ ages of 55.0 ± 0.7 Ma and 55.9 ± 1.5 Ma. Th/U corrections moved the error ellipses so that monazite M-3 overlaps concordia near 55 Ma, monazite M-2 lies below concordia, and zircons 3 and 4 intersect concordia at about 56.0 and 56.9 Ma respectively. Slight overcorrection of M-2 reflects probable overestimation of f , which might be explained by fractionation of U into earlier-formed zircon; U concentrations of the zircon fractions are very high (1400-2300 ppm; Table 4.1).

The analytical uncertainties in corrected $^{207}\text{Pb}/^{206}\text{Pb}$ ages of the three zircon analyses overlap in the range 60-74 Ma, and thus one possible interpretation is that the dike crystallized in the latest Cretaceous or Paleocene and the zircons subsequently lost radiogenic Pb. A preferable interpretation is that discordance in the zircon analyses results from a combination of different, but small, amounts of inheritance and Pb loss. The zircons of fraction 3 were of the best quality, displaying good igneous

zoning and no visible inclusions or cores. A few zircons in the other fractions contained inclusions and possible cores. The corrected age of fraction 3 barely intersects concordia at 56.0 ± 0.1 Ma (Table 4.2; Fig. 4.19); this is interpreted as a close approximation to the crystallization age of the rock, but the analytical uncertainty underestimates uncertainty in the age. The slightly older age of fraction 4 is inferred to reflect an inherited component (Late Cretaceous or older). Discordance of fraction 5 is attributed to recent low-temperature Pb loss, an interpretation consistent with its higher U content and smaller size of the zircons relative to those of fraction 3 (Table 4.1). Collectively, the zircon and monazite data from J89-762 are consistent with crystallization of the granite between 55 and 57 Ma, and accordingly an age of 56 ± 1 Ma is assigned.

THERMOCHRONOMETRY AND TECTONIC IMPLICATIONS

Anatectic melting

Of the four granitic rocks dated by U-Pb methods in this study, three have inferred crystallization ages in the range 55-72 Ma. The lower intercept of the zircon discordia from the fourth rock, pegmatite J87-174, is also within this range (60 ± 6 Ma). Many leucogranites and pegmatites from the same structural level elsewhere in the southern Omineca belt (i.e. above the Monashee décollement and beneath the Okanagan Valley and related normal fault systems) are 52-71 Ma in age (Carr 1986, 1989, 1990, 1992; Carr et al. 1987; Parrish et al. 1988; Sevigny et al. 1989; Parkinson 1992; Bardoux 1993; Scammell 1993); they have generally been interpreted as having been derived from crustal anatexis (Sevigny et al. 1989, Carr 1992, Scammell 1993).

Generation of anatectic melts commenced during mid-Cretaceous to Paleocene metamorphism and continued into the Eocene with production of voluminous leucogranites (Carr 1990, 1991, 1992; Scammell 1993; Parrish 1994 in review).

Paleocene-Eocene magmatism is thought to have been facilitated by decompression, as a result of thrusting of mid-crustal rocks up a crustal scale ramp in the Monashee décollement coupled with or closely followed by tectonic denudation (Carr 1992, Scammell and Dixon 1993, Scammell 1993). The U-Pb results of this study supplement the results of other workers.

Cooling history

Results of the $^{40}\text{Ar}/^{39}\text{Ar}$ dating point clearly to Paleocene-Eocene cooling of rocks in the footwall of the Okanagan - Eagle River fault below ^{40}Ar closure temperatures for amphibole. As will be shown below, the footwall rocks were cooled rapidly during this time from high temperatures. Plateaus in the amphibole age spectra are therefore considered to signify a homogeneous distribution of ^{40}Ar preserved in the crystals rather than homogenization due to phase changes during laboratory heating (e.g. Lee et al. 1991), and the plateau ages are thus interpreted as representing times of closure to ^{40}Ar loss. For amphiboles that did not yield plateaus, minima in the age spectra are interpreted as maximum ages of final closure.

For reasonable rates of cooling of metamorphic terranes (1-100°C/m.y.), closure temperatures for loss of radiogenic ^{40}Ar range from about 450-550°C for hornblende (Harrison 1981) and from 280-345°C for biotite (Harrison et al. 1985). The closure temperature of monazite to loss of radiogenic Pb is in the range 675-750°C (Copeland et al. 1988, Parrish 1990) provided that the U concentration in the monazite is not excessively high (Parrish and Carr 1994). With geochronological data from this and

previous studies, these figures can be used to place limits on the cooling history of footwall rocks. Three localities will be considered: (1) Three Valley Gap, the deepest structural level studied, (2) Eagle Pass Ridge, north of and ~2000 m higher than Three Valley Gap, and (3) the immediate footwall of the Okanagan - Eagle River fault south of Sicamous.

Data from Three Valley Gap include a U-Pb zircon lower intercept of 73.4 ± 1.7 Ma (Parkinson 1992), a $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende date of 58.3 ± 2.5 Ma (sample J89-3H of this study), and a K-Ar biotite date of 49.5 ± 1.7 Ma (Armstrong et al. 1991). The zircon and hornblende were from large amphibolite boudins in semipelitic paragneiss; the biotite was from the paragneiss. Parkinson interpreted the zircons as being of metamorphic origin with Proterozoic inheritance, and the lower intercept date as approximating the time at which peak metamorphic temperatures were reached. Equilibration temperature of the amphibolite based on the garnet-clinopyroxene Fe/Mg exchange geothermometer has been estimated at 710°C (Nicholls et al. 1988, 1991); an uncertainty of $\pm 30^\circ\text{C}$ is assumed. Considering these constraints and assuming that the hornblende and biotite dates are closure ages, the estimated cooling rate averaged over the time interval between metamorphic zircon growth and hornblende closure is $7\text{--}27^\circ\text{C}/\text{m.y.}$, and the average cooling rate between closure of hornblende and biotite is estimated at $8\text{--}59^\circ\text{C}/\text{m.y.}$ (Fig. 4.20).

Monazite U-Pb and amphibole $^{40}\text{Ar}/^{39}\text{Ar}$ dates from Eagle Pass Ridge are 58.2 ± 0.2 Ma and 53.8 ± 1.4 Ma, respectively (samples J89-788P and J89-788H of this study). The amphibole date is interpreted as a maximum closure age. The monazite is regarded as most likely dating the time of high-temperature Pb loss, though it could have grown at ~58 Ma at supersolidus temperature of 640°C or more (cf. Thompson and Tracy 1979). Since the U concentration of the monazite fraction was not

unusually high, there is little reason to suspect Pb loss at lower temperatures than this; hence the monazite date is inferred to document ambient temperature in the range 640-750°C. An average cooling rate in the range 15-107°C/m.y. is calculated for the period between 58 and ~53.

Cooling ages from the immediate footwall of the Okanagan - Eagle River fault are inferred from the $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende plateau age of 50.5 ± 1.7 Ma from Mara Lake (J89-18 of this study) and a K-Ar biotite date of 44.5 ± 1.4 Ma from near Enderby (Mathews 1981). These dates yield cooling rates of 13-90°C/m.y., similar to those calculated above. Because the dates are from 20 km apart, the assumption that there were no marked cooling-rate gradients along strike in that part of the shear zone during this early Middle Eocene time interval is implied.

The cooling rates calculated above are consistent with the considered ranges for closure temperatures. Since closure temperatures are dependent on cooling rate (Dodson 1973), the extreme values in the ranges of calculated rates can be ignored. Closure temperatures of 500°C and 300°C are appropriate for hornblende and biotite, respectively, for cooling rates of the order of 10°C/m.y. The relative difference in the expected closure temperatures is changed by less than 2% for higher rates (compare Harrison 1981 and Harrison et al. 1985). Cooling rates recalculated using these values are in the range 15-71°C/m.y. for the Late Paleocene to Middle Eocene (Fig. 4.20).

In contrast to the rapid early Tertiary cooling history of footwall rocks, the amphibole $^{40}\text{Ar}/^{39}\text{Ar}$ results from hangingwall rocks of the Shuswap Lake transfer zone indicate cooling below about 500°C in the Mesozoic. From east to west across this zone there is a general increase in the inferred hornblende closure ages from Late Cretaceous (74 ± 4 Ma) to Early Cretaceous and possibly Jurassic (159 ± 17 Ma) (Fig. 4.1). This pattern could reflect a lateral thermal gradient or differential erosion (see

Sonder and Chamberlain 1992), but it is interpreted as being primarily a function of relative structural depth and of the distribution of late granitic intrusions. For example, J90-1092 and J89-417, with indistinguishable dates of 87 ± 3 Ma and 99 ± 20 Ma respectively, were from high-grade rocks laced with abundant pegmatites that lie structurally above the Pukeashun granite and beneath the Mount Fowler orthogneiss. Pegmatite intrusions also are abundant in the Mount Fowler gneiss in the Shuswap Range, but are uncommon in the gneiss west of Shuswap Lake and are very rare in the superjacent medium to low-grade rocks of the Eagle Bay assemblage. This could account for the young cooling age of J89-970 relative to that of J89-662H (both from the orthogneiss) and for the pattern of increasing cooling ages up structural section west of Shuswap Lake. In addition, the younger dates came from more easterly structural blocks, bounded by predominantly west-side-down high-angle normal faults, that have been uplifted relative to adjacent blocks and hence expose somewhat deeper structural levels (Fig. 4.1).

J89-642H, from the Eagle Bay assemblage above the Mount Fowler gneiss, yielded a much older date (159 ± 17 Ma) than any of the other amphiboles. This Late Jurassic date is interpreted as a maximum cooling age because it is the age minimum in the release spectrum of the sample. By comparison, a short plateau at the high-temperature end of the release spectrum of J89-417H, from below the Mount Fowler gneiss, yields an apparent age of 158 ± 9 Ma (Fig. 4.12); this date can be interpreted as the minimum age of metamorphic crystallization and is indistinguishable from the cooling age inferred for J89-642H. The maximum cooling age of J89-417H, in contrast, is mid-Cretaceous (99 ± 20 Ma). These data are consistent with two tectonic hypotheses presented earlier (Chapter 2): (1) structural omission across a possible low-angle normal fault at the base of the orthogneiss near Adams Lake, and (2) the role of

the Mount Fowler orthogneiss as a mechanical barrier or "lid" to leucogranite and pegmatites of probable Cretaceous to early Tertiary age emplaced from below, that effectively insulated the Eagle Bay rocks above from associated thermal effects.

Northeasterly thrusting

The zircon date from J89-788P indicates that northeasterly ductile thrusting in a zone similar to the Monashee décollement was active at $67 \pm 2/-1$ Ma or later. Although this is the first attempt to date such a zone at the latitude of the Trans-Canada Highway, previous work at the southern end of the Monashee complex has shown that the Monashee décollement was active at 59 Ma and inactive by 58 Ma (Carr 1990, 1992; Brown et al. 1992a).

Extensional shearing on the Okanagan Valley fault system

Cooling ages of hornblende and biotite in the footwall shear zone of the Okanagan - Eagle River fault near Sicamous are several m.y. younger than those from deeper in the footwall to the east (Fig. 4.1). There is no distinguishable difference in hornblende ages from the deeper levels, and thus it appears that the differential cooling was a local phenomenon perhaps related to dissipative heating or hot fluids in the shear zone rather than to a regional cooling gradient. Mylonites in the shear zone formed within the stability field of sillimanite and were subsequently overprinted by chloritic shears and brittle extensional structures (Chapter 2). The shear zone was therefore active at temperatures of at least 500°C (Holdaway 1971, Spear and Cheney 1989) and also at much lower temperatures as it was exhumed. Accordingly, since closure of hornblende to ^{40}Ar loss occurs at about 500°C, the plateau age of 50.5 ± 1.7 Ma from hornblende J89-18 is considered to date synextensional cooling.

In the part of the footwall shear zone currently exposed near Kelowna, ductile shearing is inferred to have stopped by ~46 Ma, the approximate K-Ar age of volcanic rocks that crosscut and unconformably overly mylonites (Bardoux 1993). This does not discount the possibility of continued near-surface brittle faulting and shearing at depth. In fact, hangingwall volcanic rocks at Enderby dated by K-Ar at 42.5 ± 1.7 Ma (Mathews 1981) appear to have been rotated toward the detachment (Bardoux 1993), and so some extension may have continued beyond 43-44 Ma.

The time of first motion has been more difficult to resolve. Parrish et al. (1988) argued that displacement on the Okanagan Valley system and other west-side-down normal faults in the Omineca belt began at 52 Ma. Their argument for the Okanagan Valley system was based on U-Pb ages of sheared intrusions of 50 ± 2 Ma from near Vernon (Solberg 1976) and 51 ± 1 Ma from Okanagan Falls (Parkinson 1985, Parrish et al. 1988). However, because these intrusions are not demonstrably synkinematic, these dates place upper limits on the time of cessation of ductile shearing but provide no conclusive evidence for the time of initiation of extension. In a recent study, Bardoux (1993) has concluded that extension must have commenced by the time of the first eruptions of Marron Formation lavas in the Kelowna area at ~53 Ma. Results from the present study indicate the oldest age yet documented for extensional shearing on the Okanagan Valley fault system. U-Pb data from J89-762, a late synextensional granite of the Pukeashun suite, imply that west-northwesterly directed extensional shearing was underway in the Shuswap Lake area by 56 ± 1 Ma.

CONCLUSIONS

1. Rocks of the Shuswap complex, in the footwall of the Okanagan - Eagle River fault, cooled rapidly from temperatures in excess of 500°C in Late Paleocene to Middle Eocene time. Average cooling rates during this period were at least 15°C/m.y. and probably $\geq 30^\circ\text{C/m.y.}$ In contrast, rocks in the hanging wall cooled below ~500°C in the Mesozoic.
2. Ductile shearing in the Pukeashun shear zone of the Okanagan Valley fault system was underway by 56 ± 1 Ma. This timing is consistent with the contention that rapid cooling of the Shuswap complex was caused, or at least enhanced, by extensional tectonic denudation.
3. Mylonitic rocks in the Okanagan - Eagle River fault zone at Mara Lake were at ~500°C in the Middle Eocene (50.5 ± 1.7 Ma). Extensional shearing occurred beyond this time as the rocks passed through the brittle upper crust.
4. West-southwesterly verging D_3 folding in the Hunters Range occurred at ~70 Ma or later.
5. Northeasterly directed shearing within the hanging wall of the Monashee décollement occurred beyond $67 \pm 2/-1$ Ma at the latitude of the Trans-Canada Highway.

Table 4.1: U-Pb Analytical Data

sample ^a fraction	wt. ^b (mg)	U (ppm)	Pb ^c (ppm)	$\frac{206\text{Pb}}{204\text{Pb}}$ ^d	Pb _{com.} ^e (pg)	$\frac{208\text{Pb}}{206\text{Pb}}$ ^c	$\frac{206\text{Pb}}{238\text{U}}$ ^f (Ma)	$\frac{207\text{Pb}}{235\text{U}}$ ^f (Ma)	corr. coef.	$\frac{207\text{Pb}}{206\text{Pb}}$ ^f	$\frac{207\text{Pb}}{206\text{Pb}}$ ^g (Ma)
J89-788P pegmatite, Eagle Pass Ridge 399300E, 5646400N^h											
1, 105-149,e	0.156	414.4	3.90	4615	9	0.036	0.01010 ± 0.09%	0.06603 ± 0.10%	0.93	0.04742 ± 0.04%	70.5 ± 1.8
2, 105-149,e	0.104	406.9	3.94	2074	13	0.043	0.01032 ± 0.09	0.06746 ± 0.12	0.85	0.04740 ± 0.06	69.2 ± 2.9
3, 105-149,s	0.107	403	4.26	2360	13	0.043	0.01121 ± 0.09	0.08220 ± 0.12	0.86	0.05316 ± 0.06	335.8 ± 2.7
4, 105-149,p	0.090	613.6	5.28	1456	22	0.033	0.00926 ± 0.12	0.06074 ± 0.17	0.66	0.04755 ± 0.13	76.9 ± 6.2
M-5	0.036	1634.7	188.1	710	51	13.014	0.00944 ± 0.11	0.05902 ± 0.20	0.70	0.04532 ± 0.15	-38.2 ± 7.1
J87-174 pegmatite, Sicamous 363600E, 5635600N											
1, +149,p	0.062	2528.2	35.32	9666	16	0.004	0.01544 ± 0.11	0.10311 ± 0.12	0.93	0.04842 ± 0.04	119.9 ± 2.1
2, +149,p	0.039	1641.4	31.54	6080	14	0.005	0.02122 ± 0.09	0.14313 ± 0.11	0.86	0.04892 ± 0.06	144.0 ± 2.7
3, +105,e	0.046	2009.2	30.55	9363	10	0.004	0.01681 ± 0.09	0.11254 ± 0.11	0.95	0.04855 ± 0.04	126.2 ± 1.7
M-4	0.021	18103	521.1	8486	32	1.925	0.01120 ± 0.13	0.07290 ± 0.14	0.97	0.04720 ± 0.03	59.1 ± 1.6
M-5	0.035	3748.3	189.5	4573	24	3.480	0.01290 ± 0.10	0.08395 ± 0.11	0.92	0.04718 ± 0.04	58.4 ± 2.0
J89-762 monzogranite, Pukeashun Mountain 342900E, 5674550N											
3, +149,p	0.024	1390.9	11.17	1000	18	0.026	0.00870 ± 0.10	0.05689 ± 0.24	0.62	0.04740 ± 0.19	69.3 ± 9.1
4, 105-149,p	0.025	2279.1	18.76	364	90	0.034	0.00886 ± 0.20	0.05814 ± 0.56	0.63	0.04762 ± 0.46	80.2 ± 21.9
5, 105-149,p,f	0.028	1620.2	12.96	2231	11	0.030	0.00864 ± 0.16	0.05651 ± 0.21	0.64	0.04746 ± 0.17	72.3 ± 8.0
M-2	0.028	1235.1	138.7	160	14	13.422	0.00896 ± 0.29	0.05664 ± 1.38	0.67	0.04587 ± 1.21	-9.2 ± 59.4
M-3	0.007	3180	139.2	387	33	4.771	0.00869 ± 0.16	0.05569 ± 0.64	0.61	0.04648 ± 0.56	22.7 ± 27.1
J90-1059 migmatite, Mara Lookout 370100E, 5625000N											
4A, +149,p	0.076	2157.9	21.49	11180	10	0.008	0.01096 ± 0.11	0.07405 ± 0.12	0.95	0.04898 ± 0.04	147.1 ± 1.8
4C, +149,p,f	0.012	1569.8	15.82	222	64	0.000	0.01119 ± 0.28	0.07475 ± 1.06	0.63	0.04846 ± 0.91	122.0 ± 43.6
5, +149,t	0.048	2215	26.94	6275	13	0.057	0.01263 ± 0.10	0.11035 ± 0.12	0.94	0.06337 ± 0.04	720.5 ± 1.8
6A, +149,s	0.046	317.3	8.40	862	20	0.500	0.01927 ± 0.14	0.24799 ± 0.20	0.73	0.09332 ± 0.14	1494.4 ± 5.2
M-2	0.028	5248.2	193.6	583	154	3.428	0.00952 ± 0.12	0.06150 ± 0.33	0.65	0.04683 ± 0.27	40.5 ± 13.0
M-3	0.017	5027.2	185.2	933	49	3.940	0.00854 ± 0.11	0.05441 ± 0.23	0.65	0.04624 ± 0.18	10.1 ± 8.8

^a 3-digit numbers (e.g. +149) refer to size of zircons in microns; ^e = subequant, ^p = elongate prisms, ^s = stubby prisms, ^t = tabular, ^f = fragments; ^M = monazite

^b Weighing error = 0.001 mg.

^c Radiogenic Pb.

^d Measured ratio, corrected for spike and Pb fractionation of 0.09 ± 0.03%/AMU.

^e Total common Pb in analysis corrected for fractionation and spike.

^f Corrected for blank Pb and U, and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the rock); errors are 1 sigma in percent for ratios and 2 sigma in Ma for ages.

^g Corrected for blank and common Pb, errors are 2 sigma in Ma.

^h UTM coordinates in grid zone 11.

Table 4.2: U-Pb Data Corrected for Th-U Disequilibrium

sample, fraction	^{208}Pb ^a (nmole/g)	f ^b	^{206}Pb ^c ^{238}U cor.	^{206}Pb ^c ^{238}U (Ma)	^{207}Pb ^d ^{235}U (Ma)	^{207}Pb ^c ^{206}Pb cor.	^{207}Pb ^c ^{206}Pb (Ma)
J89-788P pegmatite, Eagle Pass Ridge 0.2 ppm Th, 0.3 ppm U ^e							
1, zircon	0.6301	0.1688	0.01011	64.9	64.8	0.04735	67.1
2, zircon	0.7587	0.2028	0.01034	66.3	66.3	0.04734	66.2
3, zircon	0.8051	0.1793	0.01123	72.0	80.2	0.05310	333.0
4, zircon	0.7934	0.1555	0.00928	59.5	59.9	0.04748	73.3
M-5, monazite	837.91	63.435	0.00839	53.9	58.2	0.05100	240.8
J87-174 pegmatite, Sicamous 6.0 ppm Th, 1.4 ppm U ^e							
1, zircon	0.7081	0.00315	0.01546	98.9	99.6	0.04837	117.4
2, zircon	0.7187	0.00361	0.02124	135.5	135.8	0.04888	142.2
3, zircon	0.5697	0.00293	0.01683	107.6	108.3	0.04850	123.8
M-4, monazite	1628.0	1.4109	0.01120	71.8	71.4	0.04722	60.6
M-5, monazite	702.01	2.5644	0.01288	82.5	81.8	0.04728	63.3
J89-762 monzogranite, Pukeashun Mountain 8.5 ppm Th, 5.9 ppm U ^e							
3, zircon	1.3240	0.05646	0.00872	56.0	56.2	0.04731	65.1
4, zircon	2.8389	0.07233	0.00887	56.9	57.4	0.04753	76.0
5, zircon	1.7434	0.06427	0.00865	55.5	55.8	0.04737	68.1
M-2, monazite	619.26	29.893	0.00847	54.3	55.9	0.04850	123.8
M-3, monazite	549.87	10.482	0.00853	54.8	55.0	0.04735	66.9
J90-1059 migmatite, Mara Lookout 14.0 ppm Th, 2.7 ppm U ^e							
4A, zircon	0.7836	0.00464	0.01098	70.4	72.5	0.04891	143.6
4C, zircon	0.0	0.000	0.01120	71.8	73.2	0.04839	118.4
5, zircon	6.6221	0.02601	0.01265	81.0	106.3	0.06329	717.9
6A, zircon	12.741	0.16472	0.01929	123.1	224.9	0.09325	1493.1
M-2, monazite	714.63	2.0807	0.00951	61.0	60.6	0.04692	45.0
M-3, monazite	705.09	2.4143	0.00851	54.6	53.8	0.04637	16.7

^a Radiogenic ^{208}Pb .^b Fractionation factor $f = (\text{Th}/\text{U})_{\text{mineral}}/(\text{Th}/\text{U})_{\text{rock}}$; $\text{Th}_{\text{mineral}}$ calculated assuming that all ^{208}Pb in analysis is a radiogenic decay product of ^{232}Th (Schärer 1984).^c Corrected using equations given by Schärer (1984) and Parrish (1990); uncertainty given in Table 4.1.^d From Table 4.1; uncertainty given in Table 4.1.^e Measured by ICPMS courtesy of the Geological Survey of Canada; detection limit 0.1 ppm.

Figure 4.1: Geological map of the Shuswap Lake area showing sample locations and a summary of geochronological data. Station numbers indicated by bold numerals correspond to the last digits of sample numbers (e.g. samples J89-788H and J89-788P are from station 788). All dates are in Ma and are identified by type as follows: Hbl (p) = hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ plateau, Hbl (min) = minimum in hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ age spectrum, Zrc (c) = concordant U-Pb zircon, Zrc (lw) = U-Pb zircon lower intercept, Zrc (u) = U-Pb zircon upper intercept, Mzt = U-Pb monazite. CrC: Craigellachie Creek; CRF: Columbia River fault; F: Mount Fowler; FmC: Four Mile Creek; HR: Hunters Range; L: Lichen Mountain; MD: Monashee décollement; P: Pukeashun Mountain; PeR: Perry River; Sil: sillimanite isograd; SR: Shuswap Range; TvL: Three Valley Lake; ViC: Victor Creek. Maple leaf symbol designates Trans-Canada Highway.

J89-18 HORNBLENDE 80/100

Run: RS-29, C79/P23

Mass: 250 mg

Date: APRIL 18/91

J Value: 0.00501

Total ^{39}Ar : 6.886 E-8 cm³ NTP

Approx. 0.78% K

Integrated Age: 51.1 $\pm$ 8.5 MaPlateau Age: 50.5 $\pm$ 1.7 Ma (83.9% of ^{39}Ar , steps marked by *)

Temp C	40/39	36/39	37/39	Vol ^{39}Ar E-8 cm ³	Fraction ^{39}Ar	% ^{40}Ar Rad.	Age Ma	+/-	Error 2 sigma
850	23.045	0.0807	2.074	0.099	0.014	2.89	-6.0	+/-	47.9
940	21.214	0.0485	7.179	0.115	0.017	34.84	65.9	+/-	33.3
1000	8.244	0.0101	5.445	0.384	0.056	68.44	50.4	+/-	11.2
* 1035	6.439	0.0037	4.902	2.623	0.381	88.46	50.9	+/-	0.9
* 1065	6.099	0.0028	4.844	1.544	0.224	91.67	50.0	+/-	1.0
* 1095	6.592	0.0043	4.629	0.726	0.105	85.50	50.4	+/-	5.1
* 1120	6.973	0.0056	4.696	0.883	0.128	80.99	50.5	+/-	2.3
1120	8.001	0.0080	4.612	0.229	0.033	74.46	53.2	+/-	7.9
1145	8.119	0.0089	4.581	0.239	0.035	71.35	51.7	+/-	9.8
1170	21.208	0.0506	4.567	0.038	0.006	30.93	58.5	+/-	54.9
1200	198.764	0.0619	4.787	0.006	0.001	90.96	1165.6	+/-	5126.6

Figure 4.2: $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data and age spectra for J89-18 hornblende.

J88-514H HORNBLENDE 80/100

Run: RS-12, C79/P18 Mass: 250 mg
Date: MARCH 10/91 J Value: 0.00499

Total ^{39}Ar : 4.636 E-8 cm³ NTP Approx. 0.52% K
Integrated Age: 64.4 $\pm$ 8.6 Ma
Plateau Age: 56.8 $\pm$ 3.3 Ma (58.4% of ^{39}Ar , steps marked by *)

Temp C	40/39	36/39	37/39	Vol ^{39}Ar E-8 cm ³	Fraction ^{39}Ar	%40Ar Rad.	Age Ma	+/-	Error 2 sigma
850	48.971	0.0855	3.799	0.114	0.025	48.91	204.2	+/-	20.7
940	12.485	0.0176	4.234	0.189	0.041	60.66	67.1	+/-	16.9
965	0.923	0.0070	5.085	0.243	0.052	81.41	66.6	+/-	11.1
1030	7.896	0.0055	5.059	1.059	0.228	83.66	58.7	+/-	3.5
1050	22.559	0.0219	0.696	0.044	0.010	71.38	139.5	+/-	48.9
* 1075	9.090	0.0098	5.081	0.289	0.062	72.05	58.2	+/-	12.5
* 1095	7.370	0.0048	5.382	1.734	0.374	85.98	56.4	+/-	1.3
* 1110	7.771	0.0057	5.080	0.684	0.148	82.88	57.3	+/-	4.5
1130									
1150	10.416	0.0144	4.991	0.253	0.055	62.52	57.9	+/-	11.7
1170	49.402	0.1131	4.383	0.021	0.004	32.94	141.5	+/-	129.3
1200	612.567	1.5044	2.708	0.006	0.001	27.46	1100.8	+/-	1759.1

Figure 4.3: $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data and age spectra for J89-514H hornblende.

J89-36 HORNBLLENDE 80/100

Run: RS-15, C79/P19 Mass: 250 mg
Date: MARCH 12/91 J Value: 0.00496

Total ^{39}Ar : 9.050 E-8 cm³ NTP Approx. 1.0% K
Integrated Age: 58.6 $\pm$ 5.3 Ma
Plateau Age: 56.5 $\pm$ 3.2 Ma (92.5% of ^{39}Ar , steps marked by *)

Temp C	40/39	36/39	37/39	Vol ^{39}Ar E-8 cm ³	Fraction ^{39}Ar	% ^{40}Ar Rad.	Age Ma	+/-	Error 2 sigma
850	63.769	0.1496	1.147	0.114	0.013	30.75	167.5	+/-	97.0
940	12.856	0.0175	3.648	0.226	0.025	61.65	69.7	+/-	14.5
* 1000	7.615	0.0046	3.981	0.998	0.110	85.71	57.6	+/-	2.5
* 1030	7.149	0.0031	3.797	1.921	0.212	90.81	57.3	+/-	3.6
* 1060	6.791	0.0025	3.623	2.248	0.248	92.60	55.5	+/-	1.5
* 1090	6.742	0.0024	3.609	1.444	0.160	92.98	55.4	+/-	4.0
* 1115	7.185	0.0036	3.666	0.943	0.104	88.50	56.2	+/-	4.4
* 1140	7.519	0.0038	3.630	0.817	0.090	88.26	56.6	+/-	4.4
1170	8.394	0.0055	3.540	0.328	0.036	83.42	61.7	+/-	15.4
1200	84.157	0.1987	2.816	0.010	0.001	30.43	216.1	+/-	226.0

Figure 4.4: $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data and age spectra for J89-36 hornblende.

J88-351A HORNBLENDE 80/100

Run: RS-32, C79/P22 Mass: 250 mg
Date: APRIL 20/91 J Value: 0.00501

Total ^{39}Ar : 5.031 E-8 cm³ NTP Approx. 0.57% K
Integrated Age: 65.3 $\pm$ 6.0 Ma
Plateau Age: 58.3 $\pm$ 3.8 Ma (87.6% of ^{39}Ar , steps marked by *)

Temp C	40/39	36/39	37/39	Vol ^{39}Ar E-8 cm ³	Fraction ^{39}Ar	%40Ar Rad.	Age Ma	+/-	Error 2 sigma
850	120.405	0.2951	1.392	0.128	0.026	27.65	278.2	+/-	31.7
940	31.850	0.0672	8.066	0.072	0.014	39.45	110.6	+/-	44.3
1010	13.753	0.0218	8.251	0.351	0.070	57.54	70.4	+/-	7.2
* 1040	8.535	0.0082	6.889	0.770	0.153	77.44	59.0	+/-	2.4
* 1065	8.375	0.0082	7.413	0.597	0.119	77.40	57.9	+/-	4.7
* 1090	8.410	0.0086	6.768	0.340	0.068	75.48	56.7	+/-	11.0
* 1115	8.280	0.0084	6.195	0.707	0.141	75.28	55.6	+/-	4.2
* 1140	8.538	0.0075	6.178	1.179	0.234	79.05	60.2	+/-	1.9
* 1170	8.495	0.0081	5.901	0.814	0.162	76.74	58.1	+/-	3.7
1200	19.208	0.0494	5.873	0.073	0.014	26.10	44.9	+/-	50.7

Figure 4.5: $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data and age spectra for J88-351A hornblende.

J89-905 HORNBLENDE 80/100

Run: RS-25, C79/P16 Mass: 252 mg
Date: APRIL 16/91 J Value: 0.00499

Total ^{39}Ar : 8.076 E-8 cm³ NTP Approx. 0.91% K
Integrated Age: 54.9 $\pm$ 5.2 Ma
Plateau Age: 54.0 $\pm$ 3.1 Ma (94.6% of ^{39}Ar , steps marked by *)

Temp C	40/39	36/39	37/39	Vol ^{39}Ar E-8 cm ³	Fraction ^{39}Ar	%40Ar Rad.	Age Ma	+/-	Error 2 sigma
850	18.026	0.0328	1.595	0.062	0.008	46.79	74.4	+/-	51.3
940	18.221	0.0374	5.056	0.205	0.025	41.26	66.6	+/-	14.8
* 1000	7.456	0.0055	3.311	0.805	0.100	81.07	53.7	+/-	5.9
* 1030	7.145	0.0046	3.218	2.172	0.269	83.75	53.2	+/-	1.3
* 1060	7.344	0.0056	3.200	1.710	0.212	80.38	52.5	+/-	1.3
* 1090	7.510	0.0050	3.318	1.550	0.192	83.36	55.6	+/-	5.9
* 1115	8.724	0.0092	3.572	0.830	0.103	71.46	55.4	+/-	1.8
* 1140	8.164	0.0072	3.359	0.575	0.071	76.57	55.5	+/-	6.4
1165	11.492	0.0188	3.226	0.160	0.020	53.56	54.7	+/-	11.6
1200	96.696	0.1294	0.575	0.008	0.001	60.46	461.7	+/-	1234.1

Figure 4.6: $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data and age spectra for J89-905 hornblende.

J89-3H HORNBLENDE 80/100

Run: RS-33, C79/P17 Mass: 251 mg
Date: APRIL 20/91 J Value: 0.00499

Total ^{39}Ar : 5.380 E-8 cm³ NTP Approx. 0.61% K
Integrated Age: 63.9 $\pm$ 5.6 Ma

Temp C	40/39	36/39	37/39	Vol ^{39}Ar E-8 cm ³	Fraction ^{39}Ar	% ^{40}Ar Rad.	Age Ma	+/-	Error 2 sigma
850	63.658	0.1408	2.287	0.057	0.011	34.85	189.7	+/-	174.9
940	20.746	0.0420	6.421	0.086	0.016	42.29	77.6	+/-	31.3
1010	10.610	0.0133	6.115	0.490	0.091	66.97	63.1	+/-	4.2
1040	9.204	0.0095	5.834	0.389	0.072	73.85	60.4	+/-	5.0
1065	8.930	0.0066	5.493	0.793	0.147	82.49	65.4	+/-	2.5
1090	8.022	0.0057	5.334	0.648	0.120	83.75	59.7	+/-	2.0
1115	9.342	0.0107	5.561	0.580	0.108	70.23	58.3	+/-	2.5
1140	8.700	0.0070	5.443	1.188	0.221	80.55	62.2	+/-	2.0
1170	8.817	0.0070	5.447	1.133	0.211	80.93	63.3	+/-	1.6
1200	115.430	0.3476	5.092	0.017	0.003	11.33	114.4	+/-	253.2

Figure 4.7: $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data and age spectra for J89-3H hornblende.

J89-788H AMPHIBOLE 80/100

Run: RS-31, C79/P14 Mass: 251 mg
Date: APRIL 19/91 J Value: 0.00498

Total ^{39}Ar : 10.591 E-8 cm³ NTP Approx. 1.2% K
Integrated Age: 60.8 $\pm$ 2.5 Ma

Temp C	40/39	36/39	37/39	Vol ^{39}Ar E-8 cm ³	Fraction ^{39}Ar	%40Ar Rad.	Age Ma	+/-	Error 2 sigma
850	90.188	0.1486	1.525	0.110	0.010	51.41	375.2	+/-	17.8
940	19.968	0.0276	3.485	0.121	0.011	60.37	105.4	+/-	18.8
1000	7.496	0.0039	3.294	1.752	0.165	87.40	58.0	+/-	0.8
1035	7.391	0.0035	3.275	1.492	0.141	88.89	58.2	+/-	1.6
1065	7.100	0.0031	3.247	1.323	0.125	90.14	56.7	+/-	2.5
1095	6.568	0.0024	3.214	2.181	0.206	92.43	53.8	+/-	1.4
1120	6.994	0.0025	3.251	1.896	0.179	92.31	57.2	+/-	0.8
1145	6.900	0.0026	3.235	1.620	0.153	91.91	56.2	+/-	1.7
1170	16.625	0.0357	3.166	0.086	0.008	37.77	55.7	+/-	39.6
1200	204.895	0.4684	3.395	0.009	0.001	32.55	518.4	+/-	567.1

Figure 4.8: $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data and age spectra for J89-788H amphibole.

J89-970 HORNBLLENDE 80/100

Run: RS-19, C79/P12 Mass: 250 mg
Date: MARCH 20/91 J Value: 0.00498

Total ^{39}Ar : 9.687 E-8 cm³ NTP Approx. 1.1% K
Integrated Age: 75.8 $\pm$ 3.9 Ma
Plateau Age: 74.2 $\pm$ 3.6 Ma (53.8% of ^{39}Ar , steps marked by *)

Temp C	40/39	36/39	37/39	Vol ^{39}Ar E-8 cm ³	Fraction ^{39}Ar	% ^{40}Ar Rad.	Age Ma	+/-	Error 2 sigma
850	12.782	0.0163	0.327	0.610	0.063	62.26	70.1	+/-	4.0
* 940	9.060	0.0029	3.692	1.486	0.153	93.43	74.6	+/-	1.8
* 1000	9.086	0.0029	3.447	1.357	0.140	93.05	74.5	+/-	3.3
* 1030	9.060	0.0033	3.692	1.482	0.153	92.08	73.5	+/-	6.2
* 1060	9.026	0.0030	3.774	0.890	0.092	93.09	74.1	+/-	2.7
1090	9.628	0.0032	3.719	0.917	0.095	98.79	78.6	+/-	3.9
1115	9.214	0.0027	3.789	1.688	0.174	95.49	77.5	+/-	1.6
1140	9.370	0.0023	3.820	1.217	0.126	95.80	78.8	+/-	2.0
1170	18.815	0.0319	3.477	0.031	0.003	45.30	67.2	+/-	56.3
1200	76.499	0.1068	2.776	0.009	0.001	58.97	365.9	+/-	641.2

Figure 4.9: $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data and age spectra for J89-970 hornblende.

J89-662H HORNBLENDE 80/100

Run: RS-27, C79/P20 Mass: 252 mg
 Date: APRIL 17/91 J Value: 0.00500

Total ^{39}Ar : 3.583 E-8 cm³ NTP Approx. 0.40% K
 Integrated Age: 120.7 $\pm$ 6.6 Ma
 Plateau Age: 110.6 $\pm$ 5.6 Ma (71.7% of ^{39}Ar , steps marked by *)

Temp C	40/39	36/39	37/39	Vol ^{39}Ar E-8 cm ³	Fraction ^{39}Ar	% ^{40}Ar Rad.	Age Ma	+/-	Error 2 sigma
850	105.036	0.2391	1.514	0.134	0.037	32.81	287.0	+/-	24.8
* 940	19.772	0.0299	9.162	0.200	0.056	58.63	102.2	+/-	12.4
* 1000	15.976	0.0146	8.311	0.501	0.140	76.66	107.8	+/-	6.5
* 1030	15.498	0.0104	7.967	0.628	0.175	83.87	114.2	+/-	4.9
* 1060	16.160	0.0097	8.035	0.151	0.042	85.83	121.5	+/-	14.8
* 1090	15.414	0.0158	8.050	0.124	0.035	73.40	99.8	+/-	10.2
* 1115	14.147	0.0069	7.920	0.965	0.269	89.53	111.3	+/-	2.1
1140	16.655	0.0103	8.050	0.639	0.178	85.11	124.1	+/-	4.4
1165	19.096	0.0177	7.650	0.242	0.067	75.53	126.2	+/-	12.8

Figure 4.10: $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data and age spectra for J89-662H hornblende.

J90-1092 HORNBLLENDE 80/100

Run: RS-23, C79/P15 Mass: 253 mg
Date: MARCH 22/91 J Value: 0.00499

Total ^{39}Ar : 7.181 E-8 cm³ NTP Approx. 0.80% K
Integrated Age: 96.0 +/- 3.7 Ma
Plateau Age: 87.1 +/- 2.6 Ma (64.6% of ^{39}Ar , steps marked by *)

Temp C	40/39	36/39	37/39	Vol ^{39}Ar E-8 cm ³	Fraction ^{39}Ar	%40Ar Rad.	Age Ma	+/-	Error 2 sigma
850	69.108	0.1416	1.033	0.350	0.049	39.53	230.5	+/-	8.3
940	15.601	0.0227	6.075	0.377	0.052	59.76	82.3	+/-	8.3
* 1000	11.955	0.0083	5.258	1.142	0.159	82.62	87.0	+/-	3.4
* 1030	11.117	0.0055	4.691	1.919	0.267	88.41	86.6	+/-	1.0
* 1060	10.928	0.0051	4.573	0.775	0.108	89.10	85.8	+/-	3.9
* 1090	11.565	0.0057	4.656	0.805	0.112	88.10	89.6	+/-	4.1
1115	12.338	0.0063	4.794	0.870	0.121	87.66	95.0	+/-	2.1
1140	12.219	0.0065	4.625	0.784	0.109	86.76	93.2	+/-	3.4
1165	14.503	0.0114	4.614	0.143	0.020	79.02	100.5	+/-	11.0
1200	52.397	0.0863	4.107	0.016	0.002	51.84	229.7	+/-	160.3

Figure 4.11: $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data and age spectra for J90-1092 hornblende.

J89-417H HORNBLENDE 80/100

Run: RS-17, C79/P13 Mass: 251 mg
Date: MARCH 13/91 J Value: 0.00498

Total ^{39}Ar : 2.662 E-8 cm³ NTP Approx. 0.30% K
Integrated Age: 181.4 $\pm$ 14.7 Ma

Temp C	40/39	36/39	37/39	Vol ^{39}Ar E-8 cm ³	Fraction ^{39}Ar	% ^{40}Ar Rad.	Age Ma	+/-	Error 2 sigma
850	294.851	0.7433	3.763	0.135	0.051	22.98	510.0	+/-	69.3
940	53.286	0.1443	10.569	0.159	0.060	21.40	100.3	+/-	31.6
1000	45.385	0.0409	11.163	0.143	0.054	75.17	284.9	+/-	12.8
1030	35.154	0.0227	11.842	0.444	0.167	83.35	247.4	+/-	8.2
1060	16.615	0.0207	10.698	0.115	0.043	67.78	99.1	+/-	20.2
1090	16.217	0.0133	12.766	0.347	0.131	81.37	115.7	+/-	10.9
1115	18.797	0.0126	12.622	0.478	0.180	85.02	139.2	+/-	6.3
1140	21.624	0.0130	12.157	0.478	0.180	86.27	161.4	+/-	8.0
1170	20.403	0.0124	11.493	0.322	0.121	86.07	152.3	+/-	10.3
1200	27.928	0.0690	11.174	0.040	0.015	29.84	73.9	+/-	75.6

Figure 4.12: $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data and age spectra for J89-417H hornblende.

J89-642H AMPHIBOLE 80/100

Run: RS-21, C79/P21

Mass: 250 mg

Date: MARCH 21/91

J Value: 0.00500

Total ^{39}Ar : 2.868 E-8 cm³ NTP

Approx. 0.32% K

Integrated Age: 182.4 $\pm$ 7.9 Ma

Temp C	40/39	36/39	37/39	Vol ^{39}Ar E-8 cm ³	Fraction ^{39}Ar	%40Ar Rad.	Age Ma	+/-	Error 2 sigma
850	80.612	-0.0069	3.331	0.002	0.001	102.81	626.6	+/-	947.2
940	22.082	0.0144	8.416	0.340	0.119	83.43	159.4	+/-	3.9
1000	24.765	0.0071	10.825	0.876	0.305	94.69	201.4	+/-	1.7
1030	20.180	0.0085	9.521	0.112	0.039	90.97	159.4	+/-	16.8
1060	19.936	0.0070	10.507	0.560	0.195	93.47	161.8	+/-	4.5
1090	21.971	0.0077	11.282	0.491	0.171	93.34	177.3	+/-	5.8
1115	23.795	0.0072	11.691	0.338	0.118	94.61	193.8	+/-	12.8
1140	39.525	0.0567	10.572	0.080	0.028	59.54	202.0	+/-	17.1
1185	30.687	0.0291	10.168	0.055	0.019	74.33	196.1	+/-	42.7
1200	59.725	0.1211	9.547	0.014	0.005	41.22	210.7	+/-	165.9

Figure 4.13: $^{40}\text{Ar}/^{39}\text{Ar}$ analytical data and age spectra for J89-642H amphibole.

Figure 4.14: U-Pb concordia diagram for J89-788P, pegmatite from Eagle Pass Ridge. Error ellipse labelled M-5' represents analysis of fraction M-5 corrected for Th/U fractionation.

Figure 4.15: U-Pb concordia diagram for J89-788P, showing detail of Figure 4.14. Shaded ellipses are corrected for Th/U fractionation.

Figure 4.16: U-Pb concordia diagram for J90-1059, a migmatite with pre- or syn-D₃ leucosome from the Hunters Range near Mara fire lookout.

Figure 4.17: U-Pb concordia diagram for J90-1059, showing detail of Figure 4.16. Shaded ellipses are corrected for Th/U fractionation.

Figure 4.18: U-Pb concordia diagram for J87-174, pegmatite from a roadcut on the Trans-Canada Highway near Sicamous.

Figure 4.19: U-Pb concordia diagram for J89-762, a mylonitic garnet-biotite-muscovite monzogranite of the Pukeashun suite. Shaded ellipses are corrected for Th/U fractionation.

Figure 4.20: Temperature vs. time plot for Three Valley Gap, Eagle Pass Ridge and the footwall shear zone of the Okanagan - Eagle River fault south of Sicamous. Boxes represent uncertainty in cooling ages and closure temperatures for the minerals indicated. Thick black lines represent geologically reasonable cooling curves. Numbers on curve segments are ranges of possible average cooling rates in °C/m.y., assuming closure temperatures for monazite (640-700°C), hornblende (500°C) and biotite (300°C). See text for discussion.

Plate 10: Field relationships of rocks dated by U-Pb methods.

(a) Outcrop of deformed pegmatite at Eagle Pass Ridge from which J89-788P was collected. Pen for scale is 15 cm long. Southwest is to the left.

(b) Pegmatite at Eagle Pass Ridge showing strong foliation. Strain is particularly high in migmatitic layers in lower half of photo, where feldspar porphyroclasts are isolated in dark biotite-rich matrix. Pen for scale. Southwest is to left.

(c) Discrete 10-cm thick mylonite zone in migmatitic pegmatite at Eagle Pass Ridge, dark in colour relative to light-coloured, less-strained pegmatite below. Obliquity of flattening fabric in the mylonite relative to boundaries of the zone indicate NE-directed shear sense (to left). Pen for scale.

(d) Rock slab cut from a shear zone similar to the one shown in (c), showing plagioclase porphyroclast systems with asymmetry indicative of NE-directed shear (to right). Scale bar near top of photo 1 cm.

(e) Outcrop near Mara fire lookout from which J90-1059 was collected, showing pegmatitic migmatite deformed by F_3 fold. Lipbalm tube for scale in fold hinge is 50 mm long.

(f) F_3 fold cut by 15-cm wide W-directed (to left) shear near Mara lookout. Nearly vertical foliation in bottom centre of photo is deflected into the shear zone. Pen along upper contact of shear near centre of photo is 15 cm long.

(g) Outcrop near Mara lookout from which J90-1059 was collected. Migmatitic metapelites and semipelites are polydeformed. F_3 folds (e.g. to left of pack in upper centre of photo) here are tight W-verging structures that commonly have sheared-out limbs. Note folded pegmatites in left foreground and in background. West is to left.

(h) Monzogranite dike of the Pukeashun suite from which J89-762 was collected. Dike cuts older pegmatite and granite phases of the suite and darker-coloured migmatites. Note low-angle discordance between basal contact of dike (marked by hammer head) and foliation in the older granite and migmatite. Shaft of hammer is about 70 cm long.

CHAPTER 5

Crustal structure and early Tertiary extensional tectonics of the Omineca belt at 51°N latitude

INTRODUCTION

A crustal section through the southern Omineca belt in the region traversed by the Trans-Canada Highway incorporates results from a study of the early Tertiary Okanagan Valley fault system on the western margin of the Shuswap core complex (Johnson 1990; Chapters 2, 3) and seismic reflection results from the Lithoprobe Southern Canadian Cordillera transect (Cook et al. 1992). The section illustrates the structural style resulting from a long and complex history of contractional orogeny followed by extensional tectonic exhumation of the Shuswap complex, and serves as the basis for a palinspastic restoration of the extensional structures that considers regional metamorphic, thermochronological and structural data. Results of the restoration are compared against previous estimates of the magnitude of extension across the Omineca belt and of pre-extension crustal geometry, and are used to evaluate hypotheses regarding processes of low-angle normal faulting and crustal extension.

TRANS-CANADA HIGHWAY TRANSECT, SURFACE GEOLOGY

Plate 3a shows an interpretation of the present structure of the Omineca belt in a vertical cross section traversed by the Trans-Canada Highway (see Fig. 5.1 for location and Table 5.1 for sources of data). The section is a composite of two segments: "W-E" extends along latitude 50°55'N from the eastern Intermontane belt to the Selkirk Mountains, crossing the western Omineca belt and the Shuswap complex subparallel to the direction of early Tertiary extension; "WSW-ENE" crosses the northern Selkirk and Purcell Mountains perpendicular to older compressional structures, where the effects of extension have been negligible. Only major structures such as the Moho, the Monashee décollement and the Columbia River fault are correlative in the subsurface across the splice between these two segments.

The Shuswap complex is bounded on the east at Revelstoke by the east-dipping Columbia River fault and on the west at Sicamous by the west-dipping Okanagan - Eagle River fault, both being early Tertiary normal faults. The deepest exposed structural level within the Shuswap complex is the Monashee complex, a basement-cored tectonic window bounded by the Monashee décollement and truncated on its east side by the Columbia River fault. High-grade metamorphic rocks of the Three Valley and Hunters Range assemblages lie above the Monashee décollement and below the Okanagan - Eagle River fault in the western part of the Shuswap complex. Low to medium-grade rocks are juxtaposed against the Shuswap complex on both its eastern and western flanks. On the east side, miogeoclinal assemblages of the Selkirk and northern Purcell Mountains lie in the hanging wall of the Columbia River fault. On the west side, above the Okanagan - Eagle River fault, are assemblages that have been assigned to the Kootenay terrane in a recent compilation by Wheeler and

McFeely (1991) and that are at least in part correlative with miogeoclinal strata. Still further west are accreted arc and oceanic rocks of the Quesnellia terrane (Intermontane superterrane). Details of the cross section are described below from east to west.

Northern Purcell and Selkirk Mountains

Rocks known to be part of the ancestral North American miogeoclinal succession occur in the eastern part of the southern Omineca belt in the Cariboo, northern Monashee, Selkirk and Purcell mountains and in the Kootenay Arc, an arcuate fold belt south of the Trans-Canada Highway (Fig. 5.1). Of importance to this investigation are the dominantly coarse clastic and pelitic units of the Neoproterozoic Windermere Supergroup, the overlying quartzite-dominated Hamill Group, Lower Cambrian marble of the Badshot Formation, and the Cambro-Ordovician(?) Lardeau Group which consists of graphitic phyllite, chlorite schist, calcareous phyllite, marble and impure quartzite (Fyles and Eastwood 1962, Wheeler 1963, Read and Wheeler 1976).

In the Dogtooth Range of the northern Purcell Mountains, southwest-dipping panels of Windermere and Hamill Group are imbricated by northeasterly directed thrusts carried in the hanging wall of the Purcell thrust (Simony and Wind 1970, Kubli and Simony 1994). These structures steepen westward to the axis of the Selkirk fan structure, beyond which they give way to southwesterly overturned folds and northeast-dipping faults (Wheeler 1963, Brown and Tippet 1978, Brown et al. 1993a). The western Selkirk Mountains are dominated by the Illecillewaet synclinorium, a major southwesterly overturned structure cored by Lardeau Group (Wheeler 1963, Zwanzig 1973, Colpron and Price 1993). The Clachnacudainn terrane in the westernmost Selkirks consists of Hamill, Badshot and Lardeau strata (Read and

Wheeler 1976; Thompson 1972, 1978; Read and Thompson 1980; Crowley 1992; Crowley and Brown 1994) intruded by Devonian-Mississippian, Mesozoic and Tertiary plutons (Parrish 1992, Crowley 1992), folded by tight, recumbent, southwest-verging nappes and metamorphosed to garnet zone and higher grade (Crowley 1992, Crowley and Brown 1994). The terrane is bounded on the east by the Standfast Creek fault (Thompson 1972, 1978), a pre-Late Cretaceous low-angle normal fault interpreted as having accommodated little displacement (Crowley 1992, Crowley and Brown 1994). Southwest-vergent D_2 folding, thrusting and metamorphism in the Selkirk Mountains occurred in Middle Jurassic time (Brown and Tippet 1978, Colpron and Price 1993). Upright to northeast-vergent D_3 folds and crenulation cleavage were superimposed on these structures before emplacement of mid-Cretaceous granitic plutons (Sears 1979, Crowley 1992).

Columbia River fault

The Columbia River fault forms the northeastern boundary of the Shuswap complex (Fig. 5.1). Mylonitic foliation and discrete ductile and brittle shears in the fault zone dip 20-40° eastward (Read and Brown 1981, Lane 1984a). Asymmetrical shear fabrics consistently indicate east-directed shear sense (Brown and Murphy 1982, Lane 1984a, Parrish et al. 1988, Lane et al. 1989, Parkinson 1992), and slickenfibres on brittle shears indicate a major component of east-side-down normal-sense slip and minor, mainly dextral, strike-slip reactivation (Lane 1984a, 1984b). The fault coincides with the arched Monashee décollement along the northeast flank of the Monashee complex (Read and Brown 1981), and therefore it is difficult to discern how much of the ductile strain is related to the Columbia River fault (Lane et al. 1989); older annealed mylonite has been interpreted as forming during early stages of

thrusting on the décollement (Lane et al. 1989), whereas younger nonannealed fabrics defined by greenschist-facies mineral assemblages are considered to be related to normal-sense displacement on the Columbia River fault (Parrish et al. 1988, Coleman 1990, Carr 1992, Parkinson 1992, McNicoll and Brown 1994). The fault dips 20-30° eastward at the surface and is inferred to have steepened above the present erosion surface (Read and Brown 1981). Subsurface seismic reflectors correlated with the fault dip ~35° and become listric at depth (Cook 1986).

Monashee complex and Monashee décollement

Early Proterozoic basement gneisses in the Monashee complex are exposed to the south and north of the highway in, respectively, Thor-Odin and Frenchman Cap culminations (Fig. 5.1). These gneisses, inferred to be parautochthonous North American basement, have yielded U-Pb zircon dates in the range 1.85-2.1 Ga (Wanless and Reesor 1975, Parrish and Armstrong 1983, Parrish and Ross 1990, Parkinson 1991, Armstrong et al. 1991). Unconformably overlying the basement is a cover sequence of quartzite, pelite, calcsilicate gneiss, marble and amphibolite that represents a continental platform and rift succession (McMillan 1973, Scammell and Brown 1990 and references therein) of Neoproterozoic to possibly mid-Paleozoic age (Parrish and Scammell 1988, Höy and Godwin 1988, Scammell and Parrish 1993). The cover rocks together with the basement gneisses have been deformed by large-scale northerly to easterly verging folds (Fyles 1970; McMillan 1970, 1973; Reesor 1970; Reesor and Moore 1971; Read and Klepacki 1981; Höy and Brown 1981; Duncan 1984; Brown et al. 1986; Journeay 1986; Scammell 1986).

The top of the Monashee complex is marked by the Monashee décollement, a zone of annealed mylonitic gneisses up to 1 km thick (Read and Brown 1981,

Journey and Brown 1986, Brown et al. 1992a). Mylonitic fabrics within the zone have been studied extensively and consistently indicate a northeast-directed thrust sense of displacement (Journey 1986; Journey and Brown 1986; Scammell 1986; Coleman 1987, 1989, 1990; Bosdachin 1989; Lane et al. 1989; Harrap 1990; Carr 1990, 1992; McNicoll and Brown 1994). Lithoprobe reflection profiles indicate that the décollement roots to the west in the lower crust beneath accreted terranes of the Intermontane belt (Carr 1991, Cook et al. 1992, Brown et al. 1992a), and Brown et al. (1992a) have argued from the basis of Lithoprobe profiles and timing of motion that it correlates with the sole thrust of the Rocky Mountain foreland belt.

Three Valley and Hunters Range assemblages

The structural panel between the Monashee décollement and the Okanagan - Eagle River fault (Fig. 5.1; Plate 3a) consists mainly of migmatitic gneisses and pegmatite that have been thrust northeastward over the Monashee complex. Brown and Carr (1990; Carr 1991, 1992) have referred to rocks at this regional structural level as the "middle crustal zone" in reference to the crustal depth at which they were deformed and metamorphosed. Within this zone, voluminous leucogranites and pegmatites were emplaced during the transition between compressional orogeny and extensional collapse of the southern Omineca belt (see Carr 1992, Scammell 1993). Along the Trans-Canada Highway transect, rocks in this zone preserve excellent structural evidence for both northeast-directed compressional and west-directed extensional shearing, both at upper amphibolite facies. The gneisses have been assigned to two assemblages, between which tectonostratigraphic relationships have yet to be worked out in detail (Chapter 3).

Three Valley assemblage

The Three Valley assemblage is characterized by semipelitic paragneisses with large boudins of amphibolite, and less abundant pelitic schist and calcareous psammite. These rocks generally have been regarded as being equivalent to the Neoproterozoic Windermere Supergroup, but recent isotopic studies have shed doubt on this correlation. In particular, U-Pb zircon data of Parkinson (1992) from the amphibolite boudins suggest an age of at least 1.5 Ga, much older than Windermere rocks but close to the age of the Moyie sills which intrude the Mesoproterozoic Belt-Purcell Supergroup. In addition, a Sm-Nd model age of 2.0 Ga for the amphibolite (Parkinson 1992) and extremely high initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios from the paragneiss and amphibolite (Armstrong et al. 1991) reflect crustal contamination or a crustal source and support a pre-Windermere age.

Hunters Range assemblage

A succession of migmatitic garnet-sillimanite-biotite schists, quartzofeldspathic paragneiss, amphibolite, marble, calcsilicate gneiss and biotite-hornblende granodioritic to dioritic gneiss occurs in the Hunters Range, structurally above the Three Valley assemblage (Plate 1). This succession, which has been tentatively correlated with Windermere strata (Chapter 3), forms a structural pile about 4 km thick characterized by regionally subhorizontal foliation and apparently lacking any major internal structural breaks (Chapter 3; Plates 2a, 2b, 3a).

Two mylonitic zones with opposing shear sense

The main fabric elements of the Three Valley assemblage are an annealed mylonitic foliation, into which layering has been transposed, and a west-southwest-

trending mineral stretching lineation. Asymmetrical shear structures indicate relative northeastward movement of the upper plate, consistent with thrusting associated with the décollement. Along the provisional contact between the Three Valley and Hunters Range assemblages, a sheet of pegmatite 300 m thick has been heterogeneously sheared in a northeast-directed sense (Chapter 3). This contact probably is not a major tectonic boundary if the proposed correlations of the Three Valley and Hunters Range assemblages with, respectively, pre-Windermere and Windermere strata are correct, but it is nonetheless a zone of high strain.

Up structural section in the Hunters Range, the foliation persists; the lineation (mean orientation $065^{\circ}-02^{\circ}$) is well developed in relatively weak rocks such as migmatitic schist and paragneiss, but is only weakly to moderately developed in stronger rocks such as quartz dioritic orthogneisses. Noncoaxial shear fabrics are rare except in pegmatites in the northeastern part of the range, which commonly are heterogeneously mylonitic and locally display winged feldspar porphyroclasts and asymmetrical extensional shear bands indicative of northeast-directed shear (Plate 1; Appendix A). Throughout the Hunters Range, tight to isoclinal F_2 folds have hingelines that lie parallel to the stretching lineation. Superimposed F_3 folds are steeply to moderately inclined and verge west-southwestward. Heterogeneous development of a stretching lineation and mylonitic fabric in the Hunters Range, and incomplete fabric transposition indicated by preservation of folds, contrast with the relatively uniform mylonitic foliation in the Three Valley assemblage and implies that strain increases toward the Monashee décollement at the base of the "middle crustal" panel.

At the top of the panel, in the immediate footwall of the Okanagan - Eagle River fault, mylonitic paragneisses exhibit nonannealed C/S fabrics, winged feldspar

porphyroclasts, asymmetrical extensional shear bands, and oblique foliations in dynamically recrystallized quartz aggregates that consistently indicate westward movement of the upper plate (Chapter 2; Appendix A). A strong stretching lineation defined by sillimanite and quartz-feldspar aggregates trends westward (mean azimuth 275°), in contrast to the northeast-trending lineation at deeper levels. Close to the brittle detachment, local cataclastic zones display slickensides with west-directed chlorite fibre lineations subparallel to the stretching lineation, and are taken as evidence that normal-sense shearing continued as the rocks reached brittle upper-crustal levels.

Okanagan - Eagle River fault

The Okanagan Valley fault system consists of two west-dipping, en échelon, ductile-brittle normal fault segments and an intervening transfer zone (Fig. 5.1). The Okanagan - Eagle River segment (Tempelman-Kluit 1984; Parkinson 1985; Bardoux 1985; Journeay and Brown 1986) is delineated by what probably for the most part is a single continuous low-angle detachment fault (Parrish et al. 1985; Journeay and Brown 1986; Tempelman-Kluit and Parkinson 1986; Johnson 1989, 1990). The Adams - North Thompson segment (Campbell 1968; Pell and Simony 1981; Johnson 1990) may itself consist of en échelon segments. Mylonites exhibiting west-directed kinematic indicators (described in the preceding section) occur within the uppermost few hundred metres to 2 km of the footwall of the system (Bardoux 1985, 1993; Parkinson 1985; Tempelman-Kluit and Parkinson 1986; Journeay and Brown 1986; Johnson 1989, 1990; Chapter 2), and are interpreted as exhumed former parts or splays of a subhorizontal mid-crustal root zone inferred from Lithoprobe reflection profiles.

Between the major en échelon fault segments is the Shuswap Lake transfer

zone, a 45-km-wide left step in the system within which there is abundant evidence for low-angle ductile displacement and limited evidence for high-angle brittle, or ductile-brittle, strike-slip transfer (Chapter 2). West-northwest-directed (mean azimuth 301°) mylonitic strain in this zone occurs in a belt of two-mica leucogranites of the Pukeashun suite (Pukeashun shear zone), which were at least partly syntectonic. The mylonitic foliation in the granite has been warped into a broad antiform that is interpreted as having resulted from isostatic rebound during tectonic denudation.

Hangingwall rocks in the western Omineca belt

In the hanging wall of the Okanagan - Eagle River fault are garnet-mica schists, impure quartzites, and minor carbonate and mafic schist of the Silver Creek assemblage, which are thought to correlate with either Neoproterozoic or lower Paleozoic strata (Windermere/Hamill or Lardeau) (Okulitch 1979; Johnson 1989, 1990).

Above and in probable faulted contact with the Silver Creek are discontinuous metabasites of the Tsalkom Formation and a thick succession of phyllitic marble and calcareous phyllite of the Sicamous Formation. There is considerable uncertainty about the tectonostratigraphic significance of these rocks; correlations with the Triassic Slocan Group (Campbell and Okulitch 1973, Okulitch and Cameron 1976) and with lower Paleozoic strata of the Lardeau Group and Eagle Bay assemblage (Schiarizza and Preto 1987, Johnson 1990) have been suggested. They could therefore be part of either the pericratonic Kootenay terrane (Lardeau/Eagle Bay) or the Slide Mountain terrane (Slocan; inboard edge of the Intermontane superterrane).

Northeast of and structurally above the Sicamous Formation is the Eagle Bay assemblage, comprised of quartzite, mica schist, marble, calcsilicate schist, phyllite

and metabasite. These strata range in age from Lower Cambrian or older to Mississippian and are equivalent to the Hamill(?), Badshot and Lardeau succession of the miogeocline (Okulitch 1979, Schiarizza and Preto 1987; Chapter 3).

The Eagle Bay is intruded by granodioritic gneiss of the Mount Fowler suite, of probable Devonian age (Okulitch et al. 1975). The Little Shuswap orthogneiss, which intrudes the Silver Creek assemblage near the southwest end of Shuswap Lake and is petrologically similar to the Mount Fowler gneiss, has been inferred as being of Ordovician age (Okulitch 1985) but this is poorly constrained. It is important to note that, east of the Columbia River at this latitude, Paleozoic granitoid rocks intrude miogeoclinal strata within the Omineca belt only in the northern part of the Clachnacudainn terrane (Wheeler and McFeely 1991, Woodsworth et al. 1991), where the Devono-Mississippian Clachnacudainn gneiss (Parrish 1992) intrudes the Lardeau Group.

Compositional layering and subparallel S_1 foliation in these rocks are deformed by tight, southeast to northeast-trending, synmetamorphic D_2 folds and associated penetrative foliation of probable Middle Jurassic to Early Cretaceous age (Schiarizza and Preto 1987; Chapter 3). South-southwesterly overturned D_2 folds and associated thrusts dominate the map pattern west of Shuswap Lake (Okulitch 1979, Schiarizza and Preto 1987) and are highly oblique to the cross section (Plate 3). Metamorphic grade ranges from middle greenschist facies in the southwest to staurolite zone in the Shuswap Range north of Sicamous (Plate 1; Fyson 1970, Johnson 1990). The metamorphic peak predated formation of open, steeply inclined, D_3 folds and crenulation cleavage, but metamorphism outlasted D_3 . As in the Selkirk Mountains, mid-Cretaceous granite plutons postdate all major penetrative strain.

Eastern Intermontane belt

Rocks of the eastern Intermontane belt belong to the Quesnellia terrane of the Intermontane superterrane. They include Devonian to Permian argillite, volcanoclastic rocks and limestone of the Harper Ranch Group, and overlying Upper Triassic arc-related volcanic and sedimentary units of the Nicola Group (Monger 1985, Monger and McMillan 1989). These are intruded by Late Triassic to Early Jurassic granodiorite and quartz monzonite plutons, and are deformed by southwest-verging overturned folds and possible thrusts of probable Middle Jurassic age that are similar to structures in the western Omineca belt (Monger 1985, 1989). Metamorphic grade decreases southwestward from upper greenschist to subgreenschist facies (Monger 1989).

The northeast-verging thrust boundary along which the Intermontane superterrane was accreted to North America (Rees 1981, 1987; Struik 1981; Rees and Ferri 1983) is obscured at this latitude by younger displacement on the Louis Creek fault, interpreted as a Jurassic southwest-verging thrust (Monger et al. 1985) that may have had an Eocene strike-slip history as well (Schiariizza and Preto 1987).

SUBSURFACE STRUCTURE

The Lithoprobe Southern Canadian Cordillera transect resulted in the acquisition of high-quality seismic reflection profiles across the Cordilleran orogen from the Southern Rocky Mountain Trench (eastern edge of the Omineca belt) to the west coast of Vancouver Island. The results are detailed in several publications (Clowes et al. 1987; Cook et al. 1988, 1992; Carr 1991; Brown et al. 1992a; Varsek et

al. 1993) and those data considered to be important to the interpretation of the early Tertiary extensional tectonic history of the Omineca belt at the latitude of the Trans-Canada Highway are analyzed further here. Special consideration is given to line 19 because it crosses a region mapped during the present investigation that is isolated from the main Lithoprobe transect (Fig. 5.1).

The crustal section shown in Plate 3 is 60-75 km north of the main Lithoprobe transect (lines 7-10); it crosses the southwest end of Lithoprobe line 19 and crosses the Akolkolex River near the northeast end of an independent seismic transect of Cook (1986).

Lithoprobe lines 7-10

High-grade rocks above the Monashee décollement are traversed by Lithoprobe profiles 7, 8 and 9 (Fig. 5.2). Line 10 crosses the Quesnellia terrane in the hanging wall of the Okanagan - Eagle River fault; the surface trace of the fault lies between lines 9 and 10. Details of the reflection geometry have been presented by Cook et al. (1992) and by Brown et al. (1992a).

In these profiles the Moho corresponds to a thin zone of subhorizontal reflections that changes very little in depth over the 300 km transect, from about 11.8 s two-way travel time (~35 km assuming P-wave velocity of 6.0 km/s) in the east to about 10.3 s (~31 km) in the west. The Monashee décollement projects southwestward from the surface to a zone of strong reflections up to 5.0 s (15 km) thick near the surface at the eastern edge of profile 7. This zone, interpreted as corresponding to sheared and transposed layering of the décollement and of related northeast-directed thrust zones in its footwall, narrows as it continues westward and becomes listric into the lower crust in profile 9, outlining a crustal-scale thrust ramp.

Data from a north-south profile (line 6; see Carr 1991, Cook et al. 1992, Brown et al. 1992a) confirm that the décollement, the ramp, and associated fabric in the intervening zone are southwest-dipping structures.

The Okanagan - Eagle River fault projects toward a zone of subhorizontal to gently west-dipping reflections in profile 10 that is traceable into the middle crust at about 5.0 s (15 km) depth. Reflections in both hanging wall and footwall appear to be cut off against this zone. To the west in profile 11 the relationship of the Okanagan - Eagle River fault to the steeply dipping Quilchena and Coldwater normal faults is not clear (Cook et al. 1992). These faults die out to the north and are not crossed by the section in Plate 3.

Lithoprobe line 19

Line 19 crosses the Monashee décollement and the Shuswap Lake transfer zone of the Okanagan Valley fault system north of the Trans-Canada Highway (Fig. 5.1). The reflection profile from Cook et al. (1992) is shown in Figure 5.3. The overall pattern of reflections within the crust defines a broad synform, with the crest of a complimentary antiform apparent near the southwest edge of the profile. The Moho is imaged as a narrow band of coherent subhorizontal reflectors that dips very gently eastward from about 11.0 s (33 km) to 11.5 s (35 km). The uppermost mantle is seismically transparent, with only a few coherent reflections. The middle and upper crust are characterized by numerous strongly layered reflections. At the northeast end of the profile, these dip southwestward and persist to depths of almost 9.0 s (27 km). The lower crust, beneath ~5.0 s in the southwestern part of the profile and beneath ~9.0 s in the northeastern part, is dominated by gently to moderately northeast-dipping reflections that become listric into the Moho.

A highly reflective band can be traced to the surface at the northeast end of the profile where it correlates with strongly layered, moderately southwest-dipping, mylonitic gneisses in the upper part of the Monashee complex. The thickness of this band is 1.0-2.0 s (3-6 km), approximately the same as the thickness of Monashee cover rocks between the basement and the Monashee décollement in this part of the Frenchman Cap culmination (Journey 1986). To the southwest, these reflections outline the trough and part of the northeast-dipping limb of the broad synform, and then become indistinct.

The line along which these data were acquired is not very straight (for example, between shotpoints 501 and 731 the direction of the transect varies within a 270° range), and geology is only sparsely exposed along it. For these reasons, the data are also presented in the form of a line drawing wherein the predominant reflections have been projected, horizontally and approximately parallel to orogenic strike, into a vertical section that crosses areas of good surface exposure and that is oriented subparallel to most of profile 19 (Fig. 5.4). The section passes through the intersection of line 19 with the surface trace of the Monashee décollement and also through Pukeashun Mountain (Fig. 5.1). Data from segments of the seismic profile that are oriented at high angles to this section have been omitted in order to filter out potentially misleading apparent dips; the surface geology is shown across the top of the section and is shown in detail in Plate 2g.

In surface outcrop at Pukeashun Mountain, early Tertiary mylonitic foliation of west-northwest-directed, presumably extensional, origin in the Pukeashun granite dips southeastward (Plate 1). The foliation persists in this orientation for about 15 km eastward to Seymour Arm (near shotpoint 801 on line 19, Fig. 5.3). Corresponding apparent dips shown in Figure 5.4 are subhorizontal to gentle northeasterly. Between

Pukeashun Mountain and Adams Lake, the mylonitic foliation defines a broad antiformal warp (Plate 2f) that is interpreted as having resulted from isostatic rebound during tectonic denudation (Chapter 2). To illustrate this relationship, surface data from the west-dipping limb of this antiform, in the immediate footwall of the Adams - North Thompson segment of the Okanagan Valley fault system, have been projected down an estimated shallow south-southwest plunge to 4-7 km depth near the southwestern edge of the section in Figure 5.4. In the interpretation shown in Figure 5.5, the Pukeashun shear zone is truncated in the subsurface against, or merges with, a high-angle transfer fault that connects the Okanagan - Eagle River and Adams - North Thompson normal faults and that merges with the Okanagan - Eagle River fault in the upper-middle crust. This interpretation is consistent with geological data (Chapter 2) and leads to a viable palinspastic restoration as discussed in a later section.

It is worth noting that the overall broad synform-antiform geometry outlined by seismic reflections throughout the crust is approximately concordant with apparent dips of mylonitic foliation related to the Monashee décollement and Pukeashun shear zone, the two major tectonic features exposed at the surface. The Pukeashun shear zone is considered to have been warped during Eocene extension, and the Monashee décollement and related pre-Eocene fabric also appear to be warped; hence, the northeast-dipping orientation of reflectors in the southwestern part of profile 19 may also be a function of Eocene extension. Based on this reasoning, it has been suggested that the domain of northeast-dipping reflections in the middle and lower crust represent a zone into which remobilized lower crust or magma flowed (e.g., Gans 1987, Block and Royden 1990, Wernicke 1990) to compensate for Eocene crustal stretching (Johnson 1991); however, this interpretation leads to an unlikely pre-extension restored crustal geometry, with anomalously thin crust in this part of the

section. A more reasonable interpretation is that the lower crust at the northeast end of the profile, where the northeast-dipping reflections become listric, was significantly stretched in the Eocene but that to the southwest it was not. This mode of lower-crustal strain partitioning is discussed further in the section on tectonic implications at the end of this chapter. Other interpretations that have been offered for the northeast-dipping reflectors are that they represent the root of southwest-verging Middle Jurassic thrusts in the Kootenay terrane and Selkirk Mountains (Johnson and Brown 1990), or that they are either inherited Precambrian structures or structures related to basement imbrication during northeast-verging contraction (Cook et al. 1992). These are permissible but are largely untestable.

Akolkolex transect

Cook (1986) conducted an independent seismic reflection study along the Akolkolex River, east of the Columbia River near Revelstoke. He concluded that the Columbia River fault, which was clearly imaged, is listric into the middle or lower crust. The Monashee décollement was not clearly imaged and could either coincide with the Columbia River fault zone into the middle crust or could be cut off in its hanging wall at about 5 km depth.

Interpretation of crustal structure along Trans-Canada Highway transect

The crustal-scale structural elements that are most important to the interpretation of early Tertiary tectonic history in this region are the Okanagan Valley fault system, the Columbia River fault, the Moho and the Monashee décollement. In the eastern part of the crustal section shown in Plate 3a (segment "WSW-ENE"), these features are adapted with slight modification from a section by Monger et al. (1985).

In the western segment of the section ("W-E"), the geometry is largely extrapolated from seismic reflection and surface data, and is in part constrained by requirements for a retrodeformable cross section.

The depth of the Moho is extrapolated from Lithoprobe reflection data and, east of the Columbia River, from seismic refraction data (Chandra and Cumming 1972, Cumming et al. 1979). The crust tapers gradually from about 32 km thick in the eastern Intermontane belt to about 40 km thick beneath the Selkirk Mountains, then thickens more abruptly to about 50 km beneath the Rocky Mountain Trench.

At the surface, the Monashee décollement dips about 25° southwest on the southwest flank of the Monashee complex (Harrap 1990). The Okanagan - Eagle River fault is inferred to dip about 15° westward at Sicamous (Chapter 2). A steeper fault north of Sicamous (Solsqua fault) is interpreted as a splay. In the Eagle River valley east of the Solsqua fault, the detachment is subhorizontal in the plane of section and is warped by corrugations that plunge gently northward (Chapter 2; Plate 2a).

Lithoprobe line 19 intersects the cross section (Plate 3a) at the southwest end of Shuswap Lake. This part of line 19 follows the trace of a high-angle fault for 10 km along the lower reaches of Scotch Creek (Okulitch 1979) and so the depths of structures are not tightly constrained. The Monashee décollement is projected to between 12-15 km depth at the southwest end of the line (Fig. 5.5). The Okanagan - Eagle River fault is drawn in Plate 3a with a uniform 15° dip west of Sicamous, so that it is 8-9 km deep where the section intersects line 19. This interpretation is admissible given the reflection data (Fig. 5.5) and, upon restoration, it leads to an adequate thickness of cover on the Hunters Range assemblage and to a realistic initial fault trajectory and topographic slope. The high angle fault at Scotch Creek is depicted as cutting both the Okanagan - Eagle River fault and the Monashee

décollement; the 2 km of east-side-down separation shown is consistent with surface geology.

Profile 19 indicates that the Monashee décollement and associated transposed fabric flatten to subhorizontal in the upper middle crust before steepening again to the west. In Plate 3a the décollement is drawn subparallel to the Okanagan - Eagle River fault between line 19 and where it is exposed at the surface, faithful to this arched or ramp-flat geometry and to the constraint that it should not dip eastward in the restored section. Profiles 9 and 10 show that the Okanagan - Eagle River fault does not cut the Monashee décollement, and there is no evidence from profile 19 to the contrary.

Therefore, in keeping with profiles 9 and 10, the décollement is interpreted as flattening to subhorizontal at about 20 km depth beneath the eastern Intermontane belt, and the Okanagan - Eagle River fault may become subhorizontal in the middle crust west of the section. Within the basement beneath the Monashee décollement, a thrust is inferred as ramping up to the east from the lower crust and shallowing to subhorizontal in the middle crust beneath the Monashee complex. This interpretation is consistent with Lithoprobe data (e.g. Fig. 5.2) and, implicitly, is in agreement with previous interpretations that arching of the Monashee complex is due in part to thrusting over a basement ramp (Brown et al. 1986; Journeay 1986; Brown and Journeay 1987).

PRE-EXTENSION PALINSPASTIC RESTORATION

Plate 3b shows a crustal section across the Omineca Belt with the Tertiary normal faults restored. The most important problems that this exercise attempts to

resolve are the initial geometry of the normal faults, the amount of displacement that has occurred along them, and the crustal thickness before extension. Constraints and assumptions are outlined and followed by a discussion of tectonic implications.

Relative timing of extension and contraction

The restoration assumes that displacement on the normal faults postdates final thrust motion on the Monashee décollement and related structures. This assumption is consistent with current knowledge of the timing of the major structures. For example, the Monashee décollement at the southern end of Thor-Odin culmination (Fig. 5.1) is known to have been active at 59 Ma and to have become inactive by 58 Ma (Carr 1990, 1992); thrust displacement along the contact between the Three Valley and Hunters Range assemblages occurred at or possibly after ~67 Ma (Chapter 4); normal-sense displacement on the Columbia River fault was predominantly Early Eocene but ductile shearing related to extension may have been underway as early as 60 Ma (Parrish et al. 1988); much of the displacement on the Okanagan Valley fault system is Middle Eocene (Parrish et al. 1988) but some ductile shearing occurred at, and possibly before, about 56 Ma (Chapter 4). Evidence for earlier extension has been presented by Scammell (1993; Scammell and Dixon 1993), who proposed that thermobarometric and geochronological data from north of the Monashee complex could be explained by periods of extensional tectonic exhumation at 100-94 Ma and 52-41 Ma; the later period is correlated with west-side-down shearing but the earlier period has yet to be correlated with specific structures. Extensional shearing may have therefore overlapped with the last stages of thrusting, but extension accommodated by major normal fault systems began at about the same time as thrusting stopped.

Metamorphic and thermochronological constraints on pre-extension crustal depths and exhumation history

Footwall rocks

Journeay (1986) and Scammell (1986) documented metamorphic and structural evidence for a two-stage thrusting history for the Monashee décollement at the north end of Frenchman Cap culmination (Fig. 5.1). The first stage occurred at pressures of 6.4-7.1 kbar (640-710 MPa) sometime between late Middle Jurassic (Journeay and Brown 1986) and latest Cretaceous (Lane et al. 1989). The second stage of thrusting involved syntectonic growth of low-pressure (2-3.4 kbar) metamorphic assemblages that indicate depth of no more than 12 km. A Late Cretaceous to Paleocene age has been inferred for this low-pressure episode (Journeay 1986). Preservation of an inverted low-pressure metamorphic gradient in the footwall of the décollement is attributed to exhumation rates of the order of 2 mm/yr (Journeay 1986).

At the Trans-Canada Highway, the mineral assemblage that defines fabrics related to the final stage of thrusting includes sillimanite + K-feldspar $\pm$ muscovite (Bosdachin 1989, Harrap 1990), implying temperature above $\sim 600^{\circ}\text{C}$ within the décollement zone and pressure greater than 2 kbar (Carmichael 1978, Spear and Cheney 1989). No evidence for the low-pressure metamorphism has been documented either here or further south in the Thor-Odin culmination. The tentative assumption is made here that the décollement at the highway was at about 4 kbar pressure (~ 14 km depth) during the Late Paleocene.

In the Hunters Range assemblage, the peak metamorphic assemblage grew during Cretaceous to Paleocene thrusting and fabric transposition and was stable during Late Cretaceous or younger west-vergent D_3 folding (Chapter 3). Temperature

in the central Hunters Range was high enough for partial melting to occur ($>640^{\circ}\text{C}$ in the second sillimanite field; Thompson and Tracy 1979) at least as late as ~ 70 Ma, the approximate age of a pre- or syn- D_3 migmatite leucosome based on discordant U-Pb zircon analyses (sample J90-1059, Chapter 4). Retrograde muscovite is rare in rocks within the second sillimanite zone, suggesting that the rocks were cooled rapidly from temperatures well above 600°C . Hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ dates from the Hunters Range and Three Valley assemblages are Late Paleocene to Early Eocene and are interpreted as corresponding to time of cooling below $\sim 500^{\circ}\text{C}$ (Chapter 4).

Several thermobarometric studies have been conducted on the high-grade rocks between the Monashee décollement and the Okanagan Valley fault system. Ghent et al. (1977a) and Nicholls et al. (1988, 1991) have estimated the temperature and pressure conditions of metamorphism from the rocks at Three Valley Gap; Nicholls et al. (1991) concluded that mineral assemblages in the amphibolites and enclosing paragneisses equilibrated at about 700°C and 8 kbar. These conditions were most likely reached before the last thrusting on the Monashee décollement because (i) the Three Valley assemblage lies in the immediate hanging wall of the décollement, which as noted above was probably at a lower pressure during late thrusting, and (ii) a $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende date from Three Valley Gap of 58 ± 3 Ma is interpreted as representing Late Paleocene or Early Eocene cooling through temperatures much lower than the equilibration temperature (Chapter 4). A Late Cretaceous age for equilibration is inferred, consistent with a U-Pb zircon lower intercept date of ~ 73 Ma from the amphibolites interpreted as the age of metamorphic zircon growth at high temperature (Parkinson 1992).

Other studies have indicated similar metamorphic temperatures and somewhat lower pressures. Scammell (1993) calculated peak conditions of $\sim 720^{\circ}\text{C}$ and 7.3 kbar

and retrograde conditions of $\sim 610^{\circ}\text{C}$ and 4.6 kbar from rocks north of Frenchman Cap. He concluded on the basis of U-Pb monazite and titanite dates that these conditions were reached at 100 Ma and 94 Ma respectively. Bardoux (1993) estimated average peak and retrograde conditions of 700°C / 6.0 kbar and 550°C / 4.5 kbar from the Kelowna area (~ 45 km south of Vernon, Fig. 5.1; see Fig. 1.1). He inferred peak conditions to have been reached during late stages of Late Cretaceous-Paleocene thrusting and retrograde conditions to have resulted from tectonic exhumation in the Eocene, though these inferences are not well constrained.

Average geothermal gradients calculated from the above thermobarometric results are in the range $25\text{-}40^{\circ}\text{C}/\text{km}$. The lower values ($<35^{\circ}\text{C}/\text{km}$) are derived from thermal peak data, whereas higher values correspond to post-peak pressures and temperatures that are correlated by both Scammell (1993) and Bardoux (1993) with tectonic exhumation and crustal thinning. No pretence is made here to assume a linear geothermal gradient throughout the crustal column. Instead, the point is made that the average geothermal gradient in the upper crust during the last stages of thrusting should have been lower than that calculated from pressures and temperatures of tectonically thinned middle crust. In the following section, therefore, a geothermal gradient of $35^{\circ}\text{C}/\text{km}$ is assumed to be reasonable for upper crustal rocks in the Late Cretaceous to Paleocene and is used to estimate Late Paleocene burial depths of those rocks.

Hangingwall rocks

Rocks in the hanging walls of the Columbia River and Okanagan - Eagle River faults were metamorphosed in the Mesozoic (Read and Wheeler 1976, Brown and Tippet 1978, Archibald et al. 1983, Schiarizza and Preto 1987, Parrish et al. 1988 and

references therein). Two independent lines of evidence suggest that the Clachnacudainn terrane was at a depth of about 7 km in the Late Paleocene. First, geobarometry based on aluminum in hornblende from the Albert stock indicates a pressure of 3.6 kbar (Crowley et al. 1994 in review). Assuming that this pressure corresponds to emplacement depth of 12-12.5 km, an average exhumation rate of 0.14 mm/yr is calculated for the time between intrusion of the stock (104 ± 1 Ma; Crowley 1992) and the present. At this rate, 5 km of cover would have been removed from the Clachnacudainn terrane by 58-60 Ma. Secondly, biotite $^{40}\text{Ar}/^{39}\text{Ar}$ ages from the Lardeau Group in the southern Clachnacudainn terrane are ~65 Ma and are slightly older to the east (M. Colpron unpublished data 1994). Assuming a constant cooling rate to the present ($\sim 4^\circ\text{C}/\text{m.y.}$), an argon closure temperature for biotite of $280\text{-}300^\circ\text{C}$ (Harrison et al. 1985), and a geothermal gradient of $35^\circ\text{C}/\text{km}$, a burial depth of 6.5-7.5 km is calculated for 58 Ma.

A similar depth in Late Paleocene time is likely for the Eagle Bay assemblage in the Shuswap Range north of Sicamous. Hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ and K-Ar dates are Late Cretaceous (Chapter 4; Geological Survey of Canada internal reports cited in Okulitch 1979) and indicate cooling below $\sim 500^\circ\text{C}$ before the Paleocene. Biotite and muscovite K-Ar dates are Middle Eocene (Wanless et al. 1978, Geological Survey of Canada internal reports cited in Okulitch 1979); these have been interpreted as either cooling ages or ages of resetting, and suggest equivocally that the rocks were above $280\text{-}300^\circ\text{C}$ before motion on the Okanagan - Eagle River fault. A zircon fission-track date of 51 ± 8 Ma (C. Naeser cited in Okulitch 1979) suggests a minimum pre-extension temperature of $\sim 175^\circ\text{C}$, the approximate closure temperature for annealing of fission tracks in zircon (Harrison et al. 1979). For a geothermal gradient of $35^\circ\text{C}/\text{km}$, temperatures in the range $175\text{-}400^\circ\text{C}$ correspond to burial depths of 5-11.5 km, and a

reasonable temperature of 300°C corresponds to 8.5 km depth.

Biotite and muscovite K-Ar dates from the Little Shuswap orthogneiss, near the southwest end of Shuswap Lake, are Early Cretaceous (147-122 Ma; Lowden et al. 1963, Wanless et al. 1978, Geological Survey of Canada internal reports cited in Okulitch 1979, Stevens et al. 1982), implying that the temperature there was probably much less than 300°C by the Paleocene. Apatite fission-track dates from hangingwall rocks west of the Okanagan Valley range from latest Cretaceous to Late Eocene (Medford 1975), suggesting that some hangingwall rocks cooled below about 105°C (closure temperature for apatite fission tracks; Harrison and McDougall 1980a, Parrish 1983) before motion on the Okanagan - Eagle River fault but that others were warmer or were reheated in the Eocene. In keeping with these data, the Little Shuswap orthogneiss is assumed to have been buried by less than 8.5 km, and probably closer to about 3 km, of cover in the Paleocene.

Finally, in the Intermontane belt, Triassic and Early Jurassic plutons have yielded dates by K-Ar and Rb-Sr methods that are close to their crystallization ages determined by U-Pb methods (Monger 1989 and references therein). Country rocks of the Quesnellia terrane were therefore below argon retention temperatures (~300°C for biotite) in the Jurassic and have not since been reheated. By the Paleocene, they likely were at no more than 100°C, and hence buried by no more than 3-4 km of cover.

Relative depths of footwall and hangingwall rocks

From the foregoing analysis, it is apparent that the Monashee complex was about 7 km deeper than the Clachnacudainn terrane (~14 km vs. ~7 km depth) before movement on the Columbia River fault. The relative depths of footwall and

hangingwall rocks of the Okanagan - Eagle River fault are more difficult to constrain. However, prior to extension the Eagle Bay assemblage of the Shuswap Range was probably at about the same structural depth as the Lardeau Group of the Clachnacudainn terrane. This is suggested by similar stratigraphy, metamorphic grade and structural style, and by the unique association of widespread Devonian orthogneisses with both assemblages. The estimated pre-extension depth of hangingwall rocks that presently are exposed west of the Shuswap Range decreases westward to ≤ 4 km in the eastern Intermontane belt.

Restoration of the Shuswap Range and Clachnacudainn terrane to about the same depth places geometric constraints on vertical displacement (throw) across the major normal faults and limits the amount of net hangingwall extension across the Shuswap culmination to a maximum of about 60 km, the distance between the surface traces of the Columbia River and Okanagan - Eagle River faults.

Geometric assumptions

All faults are modelled in Plate 3 as discrete detachments and, for lack of adequate documentation to the contrary, their hanging walls are restored as coherent blocks with only minor internal strain to resolve local room problems. This method is not rigorous but it serves to illustrate a reasonable pre-extension structure of the Shuswap complex and gives an approximation of the total amount of extension across the Omineca belt. In doing this, it is recognized that a problem encountered in many attempts to restore extensional fault and shear zone systems is that such systems commonly exhibit complex crosscutting relationships that result from abandonment of splays during progressive displacement and uplift (e.g. Davis and Lister 1988, Lister and Davis 1989). The Okanagan Valley fault system shows such complexity

regionally (Chapter 2) but the effects in the plane of this section are thought to be only locally significant. If major abandoned splays intersect the plane of section, then restoration would have to involve multiple steps.

Estimates of displacement across the Columbia River fault range from 1-10 km (Lane 1984a, 1984b; Harrap 1990) to 40 km or more (Parkinson 1992). In Plate 3a, the Monashee décollement in the footwall of the Columbia River fault is extrapolated above the present erosion surface parallel to the foliation in the Monashee complex from a section by Read and Klepacki (1981), and its position in the hanging wall is modified from one of the interpretations of seismic data presented by Cook (1986) as per Brown et al. (1993a). The distance between the hangingwall and footwall cutoffs of the décollement is 15 km. This is the minimum displacement inferred by Read (1979a, 1979b; Read and Brown 1981) and upon restoration puts cover on the Monashee complex of a thickness that is consistent with the assumed 4 kbar pressure (14-15 km; Plate 3b).

The restored Okanagan - Eagle River fault in Plate 3b dips westward at an average of $\sim 14^\circ$ before flattening to horizontal in the middle crust, and is shown with 32 km of dip slip restored. Rationale behind this geometry is as follows. First, the dip of the fault was drawn so that the paleotopographic surface has, on average, a subhorizontal or very low easterly dip and so that the Monashee décollement has an overall westerly dip; these constraints satisfy criteria for an east-directed tapered thrust wedge (Davis et al. 1983). Next, the Shuswap Range was restored to approximately the same crustal level as the Clachnacudainn terrane and to 6 km higher in the crust than the Three Valley assemblage at Three Valley Gap. This satisfies metamorphic constraints and puts the presently exposed parts of the Eagle Bay and Hunters Range assemblages far enough apart that different pre-extension deformation histories

(Chapter 3) are conceivable. Finally, the Sicamous Formation is restored to well above the Hunters Range assemblage because there is no possibility of correlation between them. The hangingwall cutoff of the Sicamous is well constrained south of the section and its projected depth in the plane of section is a minimum based on surface geology. The apparent thickness of the Sicamous Formation (~1.5 km) is based on minimum thickness established from surface geology (Plate 2d; Okulitch 1979).

The restoration in Plate 3b shows two different configurations for the footwall domain. In one, the footwall is treated as though it were not stretched in the Eocene. In the other, hangingwall and footwall extension are balanced within the limits of the Omineca belt and it is assumed that no volume has been added to the plane of section (e.g. by Early and Middle Eocene calc-alkaline magmatic underplating) or subtracted from it except by erosion; unless it is demonstrated in subsequent work that material has moved into or out of the plane of this section since the Late Paleocene, these assumptions are permissible. Implications of the two alternative configurations are discussed later.

TECTONIC IMPLICATIONS

Initial geometry of the Okanagan - Eagle River fault

Hangingwall domains of low-angle normal fault systems typically display internal extension and thinning accommodated by slip and rotation of subsidiary block faults (Wernicke and Burchfiel 1982 and many other studies; Fig. 5.6). Consequently, the master detachment is rotated to a shallower dip isostatically. The Okanagan -

Eagle River fault is shown with an average initial dip of $\sim 14^\circ$ in the pre-extension restoration. Because the hanging wall was modelled as a coherent block without any internal extension, the actual trajectory was probably somewhat steeper than this.

Northerly striking faults that cut the Eagle Bay and other hangingwall assemblages (Okulitch 1979 and pers. comm. 1989; Schiarizza and Preto 1987; Plate 1; Chapter 2) are interpreted as high-angle normal faults that accommodated only minor extension, perhaps collectively 4-5 km ($<10\%$ extension across the western Omineca belt). Similarly, in the southern Okanagan Valley where the hanging wall is better exposed, only minor internal extension has been documented (Fig. 2 of Tempelman-Kluit and Parkinson 1986). In the Intermontane belt, the total extension that can be attributed to mapped faults is only about 5% across the 140 km width of the belt (Monger et al. 1985).

Comparison of the present and restored cross sections shows that relative uplift and minor arching of the Okanagan - Eagle River fault has occurred as a result of tectonic unloading; former flat mid-crustal parts now dip gently westward and part of the fault that originally dipped gently westward has been backrotated to a subhorizontal dip (in the plane of section) along the Eagle River valley (Plate 3). Spencer (1984) has shown that normal faults with initially uniform low dips should develop a similarly warped post-extension geometry by isostatic rebound if the hanging wall is extended nonuniformly and if this extension decreases in the downdip direction. Rebound should be greatest below the most highly extended part of the hanging wall, and a pronounced antiformal arch will develop if the fault has a steep listric breakaway (Fig. 5.7). By analogy, the Shuswap Range and the now-eroded hanging wall to the east should be the most highly extended parts of the hanging wall of the Okanagan - Eagle River fault.

Figure 5.8 shows that when a wedge-shaped hangingwall of length x and maximum thickness h above a low-angle fault dipping θ degrees is stretched uniformly by a factor β , the resulting dip of the fault θ' can be calculated from the expression

$$\tan\theta = \beta^2 \tan\theta' \quad [1]$$

assuming complete isostatic rebound. From this equation, restoration of 10% ($\beta=1.1$) internal hangingwall extension in Plate 3b would only change the average dip of the fault from 14° to 17° . Since the hanging wall was probably extended nonuniformly, with greatest extension east of the present limit of exposure, the fault could have had a steeper breakaway in the upper crust to the east. It is concluded, however, that below ~ 8 km depth it had an average initial westward dip of less than 20° . This concurs with a growing number of geological, thermochronological and paleomagnetic studies that have concluded that normal faults can initiate at low angles (Howard and John 1987, John 1987, Burchfiel et al. 1987, Davis and Lister 1988, Yin and Dunn 1992, Holm et al. 1992, Scott and Lister 1992, Livaccari et al. 1993, Dokka 1993, Axen 1993, John and Foster 1993), but neither supports nor conflicts with arguments that low-angle normal faults in the upper crust accommodate seismogenic slip.

Relationships between fault dip, hangingwall extension, and isostatic rebound can have other regional implications. For example, a flat segment of the fault east of the Okanagan Valley further south is up to 70 km wide (Carr 1991, Cook et al. 1992), reflecting either greater net extension or perhaps a different distribution of hangingwall strain than at the Trans-Canada Highway. To the north, antiformal arching of the Pukeashun shear zone could have resulted from greater lateral strain gradients in the

hanging wall above an initially more gently dipping or subhorizontal detachment zone (Chapter 2). Rebound of the footwall along the Trans-Canada transect could have been accommodated by displacement on steep west-side-down normal faults that cut the footwall, for example at Victor Creek and Three Valley Lake. If so, these faults should be younger from east to west, a prediction that may be testable through $^{40}\text{Ar}/^{39}\text{Ar}$ dating of micas and feldspars and fission track dating of zircon and apatite across these faults.

Magnitude of extension and crustal strain partitioning

The total extension across the Shuswap core complex accommodated by the Okanagan Valley and Columbia River fault systems is shown in Plate 3 as being 44 km, or 22% of the restored distance from the Rocky Mountain Trench to the North Thompson River. Somewhat less than this is possible if stratigraphic omission and metamorphic mismatches across the Okanagan - Eagle River fault are partly accommodated by displacement on yet unknown high-angle block faults within the hanging wall. Conversely, since the hanging wall west of the Eagle River has been internally extended by such faults as discussed in the previous section, this figure could be an underestimate. Restoration of 10% internal extension would increase overall extension across the Omineca belt to 28%, whereas the combined effects of internal extension and a more conservative heave on the detachment would nearly cancel each other out. Further detailed mapping may lead to improvements upon these estimates; at this point it is concluded that net extension of at least 20% and less than 30% occurred across the Omineca belt at this latitude, leading to its present width of about 220 km. This is about three times the extension that was previously estimated by Brown and Journeay (1987). Parrish et al. (1988) estimated that the Omineca belt

175 km to the south was extended by about 75 km to its present 300 km width at that latitude, a considerably greater magnitude of extension than that estimated for the Trans-Canada Highway section.

Two configurations for the lower crust are considered in the restored section (Plate 3b). The first, wherein the lower crust has not been involved in extension, is unlikely. Given that the Okanagan - Eagle River fault does not penetrate the entire crust, the excess lower crust in the restored section would have to be balanced by shortening of upper crust to the west of the present section. However, the only early Tertiary structures to the west are strike-slip and normal faults in the Intermontane and Coast belts, many of which similarly feed into the middle crust (Varsek et al. 1993).

The preferred solution is that extension of the upper crust in the hanging walls of normal faults has been balanced by reciprocal ductile stretching of the lower crust (Price 1992, 1993; Cook et al. 1992; Varsek et al. 1993); hangingwall extension is greatest between the Okanagan - Eagle River and Columbia River faults as discussed above, whereas the footwall has been most highly attenuated beneath the eastern Intermontane belt. This interpretation is consistent with seismic reflection data which show that the lower crust beneath the Monashee complex contains dipping reflectors that probably are pre-extensional in origin, whereas the lower crust beneath the western Omineca and eastern Intermontane belts is a zone of strongly layered subhorizontal reflections into which both compressional and extensional structures apparently become listric (Cook et al. 1992; Fig. 5.2). It also is applicable to the region transected by Lithoprobe line 19, where attenuation of upper crust (now eroded away) above the Pukeashun shear zone is inferred to have been balanced by stretching of the lower crust to the northeast. Restored according to this model, the crust in Plate 3b has an approximately uniform thickness of just over 50 km. This agrees well

with restorations of Brown and Journeay (1987) and Cook et al. (1992); Parrish et al. (1988) estimated a 60 km thickness based on restoration of normal faults that penetrate the entire crust, but this geometrical interpretation is now known to be incompatible with seismic reflection data from Lithoprobe.

Importantly, the interpretation outlined above carries the implied assertion that the regionally subhorizontal Moho is a geodynamic equilibrium configuration rather than mere happenstance (Price 1993). A flat Moho may be maintained by lateral flow of weak lower or middle crust in order to relieve lateral pressure gradients that arise where the crust is of variable thickness; the rate of flow is dependent upon viscosity and yield stress and is therefore enhanced by high heat flow (Block and Royden 1990; Wernicke 1990; Buck 1991). Heat flow in the Intermontane and Omineca belts is presently very high ($>80 \text{ mW/m}^2$; Lewis et al. 1992), and probably was as high or higher in the early Tertiary when these belts were magmatically active. It should be emphasized that the model advocated here differs from other models that invoke lower crustal flow (Gans 1987; Block and Royden 1990; Wernicke 1990) in that the lower crust did not flow toward and reflate areas of crustal attenuation, but merely stretched heterogeneously to compliment heterogeneous attenuation of the upper crust. Testing of this hypothesis would require numerical and analogue modelling.

Stratigraphic implications

The Silver Creek assemblage, which is at least 2.5 km thick near Sicamous (Plate 2d), is restored so that its lower part is in strike continuity with the upper part of the Hunters Range assemblage. This lends credence to an assertion of Nielsen (1982) that these units are correlative, even though his conclusion that there is no major fault in the valley south of Sicamous is inconsistent with the structural

interpretation presented here. Nielsen argued that semipelitic paragneiss, marble and hornblende gneiss of the Silver Creek assemblage west of the south end of Mara Lake (units labelled "SCa?" and "SCc?" on Plate 1) are correlative with similar units in the Hunters Range on the east side of the lake. These units have been reinterpreted here as belonging to a small slice (~10 km²) bounded by splays of the Okanagan - Eagle River fault (Plate 1 and Chapter 3), and they need not have travelled farther than a few km relative to the footwall.

The Sicamous and Tsalkom formations project into the space above the Hunters Range and Three Valley assemblages, from which their footwall equivalents have been eroded. The problem of whether these units are correlative with the Lardeau Group or the Slide Mountain terrane remains unresolved, but the former is more readily acceptable because the only rocks of Slide Mountain terrane east of the Columbia River fault lie far to the south (Wheeler et al. 1991). If it is accepted that the Silver Creek assemblage and the Sicamous Formation are correlative with, respectively, Windermere/Hamill and Lardeau strata, then the inferred fault between them puts younger rocks on older and apparently omits Lower Cambrian marble of the Badshot Formation.

CONCLUSIONS

Metamorphic, stratigraphic, structural and thermochronological arguments suggest that 30 km of normal-sense displacement on the Okanagan - Eagle River fault is geologically reasonable, that the fault initially dipped no more than 20° westward in the upper middle crust, and that a steeper breakaway in the upper crust was possible.

The present geometry of the fault in the Eagle River area and its inferred geometry in the subsurface are consistent with the profile that would be expected to result from isostatic rebound in response to nonuniform stretching within the hanging wall domain, with greatest stretching east of Sicamous in rocks that have since been removed by erosion. The implied nonuniform distribution of hangingwall strain, considered together with Lithoprobe reflection data from the lower crust, geological and geophysical constraints from the Intermontane and Coast belts, and requirements for palinspastic restoration, favour a model of crustal strain partitioning whereby stretching of the upper crust is balanced by reciprocal nonuniform ductile stretching of the lower crust.

Restoration of early Tertiary extensional strain and of an estimated amount of cover inferred to have been removed by erosion leads to an estimated Late Paleocene crustal thickness of over 50 km. Net extension across the Omineca belt at latitude 51°N is estimated to have been 20-25% (~45 km) between the North Thompson River and the Rocky Mountain Trench.

Table 5.1: Sources of data for Plate 3a

Intermontane Belt	Monger and McMillan (1989)
Western Omineca Belt	Okulitch (1979)
Shuswap Range, Hunters Range	this study
Monashee Complex	Read and Klepacki (1981), Harrap (1990)
Western Selkirk Mountains	Crowley and Brown (1994), Sears (1979), Brown et al. (1993a)
Eastern Selkirk Mountains	Monger et al. (1985), Brown et al. (1993a)
Dogtooth Range	Monger et al. (1985)
Subsurface	Cook et al. (1992), Brown et al. (1992a), Cook (1986)

Figure 5.1: Tectonic assemblage map of the Omineca belt and adjacent parts of the Intermontane and Foreland belts between 50° and 52°N latitude. Line of cross section shown in Figure 5.4 is indicated; cross sections "W-E" and "WSW-E" are shown in Plate 3. Seismic reflection profiles include 1-12 and 19 of Lithoprobe (Cook et al. 1992) and the Akolkolex River transect (A) of Cook (1986). Trans-Canada Highway (not shown) passes through Golden, Revelstoke, Sicamous and Kamloops. 1 = Paleoproterozoic basement of Monashee complex; 2 = Proterozoic to Paleozoic cover of Monashee complex; 3 = undivided metamorphic rocks of Shuswap complex above Monashee décollement (includes equivalents of 4-7); 4 = Purcell (Mesoproterozoic) and Windermere (Neoproterozoic) supergroups; 5 = Paleozoic strata in hanging walls of major Eocene normal faults (includes Hamill, Badshot, Lardeau, Eagle Bay, Silver Creek, Tsalkom and Sicamous; some rocks adjacent to west side of Shuswap complex may be of Neoproterozoic age); 6 = Paleozoic and Mesozoic strata of Foreland belt (east of SRMT) and undifferentiated rocks of Coast belt (west of FF); 7 = Intermontane superterrane (Devonian-Jurassic; includes Slide Mountain, Quesnellia, and Cache Creek terranes, Jurassic overlap assemblages and Cretaceous volcanic assemblages); 8 = Paleozoic granodioritic to dioritic gneiss; 9 = Jurassic granodiorite, quartz monzonite, quartz diorite; 10 = Cretaceous granite, granodiorite; 11 = Paleocene-Early Eocene granite; 12 = Middle Eocene volcanic and sedimentary rocks (Kamloops Group); 13 = Neogene volcanic rocks. A-NT = Adams - North Thompson segment of Okanagan Valley fault system; Bv = Beaven fault; CR = Columbia River fault; CT = Clachnacudainn terrane; DR = Dogtooth Range; FC = Frenchman Cap culmination; FF = Fraser fault system; HR = Hunters Range; MD = Monashee décollement; O-ER = Okanagan - Eagle River segment of Okanagan Valley fault system; P = Pukeashun Mountain; PT = Purcell thrust; SR = Shuswap Range; SRMT = Southern Rocky Mountain Trench; TO = Thor-Odin culmination.

Figure 5.2: (*Top panel*) Lithoprobe seismic reflection data from the central and western Omineca belt (lines 7-9) and the eastern Intermontane belt (line 10). Data are migrated and coherency filtered and are shown with no vertical exaggeration for a velocity of 6.0 km/s. (*Bottom panel*) Structural interpretation. Shaded pattern represents crystalline basement interpreted as being of ancestral North American cratonic origin. Cover rocks have been stripped from, and thrust northeastward over, the basement along the Monashee décollement (MD). Eocene normal faults are shown as alternating long and short dashes: BF = Beaven fault, CF = Cherryville fault, CRF = Columbia River fault, OVF = Okanagan - Eagle River fault (Okanagan Valley fault system). Location of seismic lines is shown in Figure 5.1. From Brown et al. (1992a).

OKANAGAN LAKE

TWO-WAY TIME (s)

TWO-WAY TIME (s)

LITHOPROBETRANSECT - SOUTHERN CORDILLERA LINE 19 MIGRATED

Figure 5.3: Migrated and coherency filtered seismic reflection data from Lithoprobe profile 19. MD indicates surface trace of Monashee décollement. Approximate locations of the villages of Scotch Creek (~4 km off profile line) and Seymour Arm (2 km off line) are indicated. From Cook et al. (1992).

Figure 5.4: Scaled line drawing of Lithoprobe reflection data from profile 19 projected, horizontally and approximately parallel to strike of compressional structures (azimuth = 340°), into a vertical plane striking 060° (see Fig. 5.1 for location). Uniform seismic velocity of 6.0 km/s is assumed. Geology shown across the top: plus pattern = early Tertiary Pukeashun granite, with arrows and short line segments indicating apparent dips of mylonitic foliation and relative westerly sense of upper-member shear; arrows and short line segments near southwest edge of section are apparent dips and shear sense of mylonitic fabric projected from near Adams Lake; hachured pattern to left of Pukeashun Mountain is Devonian Mount Fowler orthogneiss; fold in marble unit of Eagle Bay assemblage shown between orthogneiss and shotpoint 2101; MD = Monashee décollement. Points labelled 101 and 2101 are projected initial and final shotpoints of the seismic transect.

Figure 5.5: As Figure 5.4, with geological interpretation. Pukeashun shear zone (PSZ) either merges with or is truncated against the steeply dipping "Grizzly Lake transfer fault" (GLF), which may have transferred displacement between the Okanagan - Eagle River fault (O-ERF) and the Adams - North Thompson segment of the Okanagan Valley fault system (see Fig. 5.1 for locations and text for discussion). Section is highly oblique to the west-northwest slip directions on the O-ERF, PSZ and GLF. Monashee décollement (MD) dips southwestward at the surface and flattens to a subhorizontal dip at about 5 km depth in central part of profile. Highly reflective band beneath MD, labelled *cover?*, is interpreted as correlating with strongly transposed layering in cover rocks infolded, and possibly imbricated, with basement slices. Reflectors beneath this zone are inferred as correlating with basement rocks. The domain that lies below ~15 km depth at the southwest end of the section and that tapers northeasterly to below ~28 km depth is interpreted as middle and lower crust that was differentially stretched during early Tertiary extension.

Figure 5.6: Diagram illustrating the different deformation styles exhibited by the hanging wall and the footwall of a low-angle normal fault as a result of crustal extension. (a) Configuration of the crust prior to extension. (b) Configuration of distended crust after about 20% extension of the hanging wall, accommodated by slip and rotation of subsidiary high-angle block faults within the hangingwall domain. Arching and isostatic rebound of footwall, in response to tectonic unloading, is greatest just downdip from where hanging wall has been most highly attenuated. Modified from Davis et al. (1986).

Figure 5.7: Isostatic rebound and warping of low-angle normal faults assuming complete isostatic rebound and flat initial surface topography. Six different patterns of uplift and warping result from three different arbitrary distributions of internal extension within hangingwall domains of two different forms (planar and listric). For example, dotted line at top, representing one distribution of extension, corresponds to dotted lines in two cross sections, representing final shape of fault after denudation and uplift. All cases are based on 20 km net extension; for example, point A is displaced 20 km to point A'. Part of hanging wall initially to right of point A is uniformly displaced 20 km and undergoes no internal extension. Part of hanging wall initially to left of point A is internally extended and is displaced less than 20 km. Adapted from Spencer (1984).

Figure 5.8: Cross-sectional representation of a wedge-shaped hanging wall domain, with no internal extension (*left*) and uniformly stretched internally by a factor β (*right*), illustrating mathematical relationship between extensional attenuation and fault dip. Initial dips (θ) of 17°, 20°, 29° and 45° result in a final dip of $\theta'=14^\circ$ (restored dip of Okanagan - Eagle River fault in Plate 3b) for β values of 1.1, 1.2, 1.5 and 2.0, respectively.

SUMMARY OF SCIENTIFIC CONTRIBUTIONS MADE BY THIS STUDY AND SUGGESTIONS FOR FUTURE WORK

Mylonitic rocks in the footwall of the Okanagan - Eagle River fault have been shown to consistently display nonannealed microfabric and west-directed kinematic indicators, and their distribution has been shown to be limited to within 1 or 2 km of the hanging wall. Northeast-directed mylonitic fabrics related to thrusting deeper in the footwall are defined by the same metamorphic mineral assemblage as, and are not overprinted by, the west-directed fabrics related to extension; these observations allow that thrusting and extensional faulting could have overlapped in time.

The Shuswap Lake transfer zone, a step between major en échelon normal fault segments, is an important and unexpected discovery. Rather than a discrete detachment between high-grade and low-grade rocks, the transfer zone is characterized by (i) an antiformally arched composite sheet of mylonitic leucogranite intrusions, (ii) minor high-angle normal faults, and (iii) evidence for a high-angle strike-slip transfer fault. These components are considered to have accommodated displacement during different stages of the evolution of the zone. En échelon steps in low-angle normal fault systems have been mapped in other parts of the North American Cordillera, but the deep level of exhumation in the Shuswap area provides excellent opportunities to investigate the processes involved, and in particular the role of granite melts, in the evolution of such zones. Future work should focus on (i) delineation and characterization of the northern boundary of the Pukeashun shear zone, (ii) detailed mapping (perhaps at 1:10 000 scale) of selected areas within the zone to document strain variation and the geometry and distribution of granite bodies, (iii) U-Pb geochronology of intrusions of the Pukeashun suite, representing various strain states and different structural levels, (iv) comparative $^{40}\text{Ar}/^{39}\text{Ar}$ mineral dating of the Momich, Lichen, and Eagle Bay assemblages, the Pukeashun granite, and the Mount Fowler orthogneiss in NNE-SSW transects across the transfer zone, and (v) efforts to further document and delineate a transfer fault or fault zone through mapping, kinematic studies and remote sensing.

New information on lithology, stratigraphic relationships and structure of various tectonostratigraphic assemblages in the Shuswap area have been gained through detailed mapping. The preliminary regional correlations proposed herein may be testable by detailed and specifically focused studies. In particular, the Eagle Bay assemblage could be mapped over a broader extent, with detailed stratigraphic work for comparison with sections of Paleozoic strata in the Selkirk Mountains. Isotopic work should be done on the metabasites and carbonates of the Eagle Bay, Tsalkom, Lardeau and Hamill for the sake of comparison; further isotopic work on amphibolites of the Hunters Range assemblage could serve as a test for the proposed correlation with Windermere rocks. In the Eagle Bay assemblage, a more comprehensive analysis of fabric and metamorphic relations may help in testing the inferred correlation of the

deformation history with that in the Selkirk Mountains, and a $^{40}\text{Ar}/^{39}\text{Ar}$ mineral dating study could be designed to test assumptions made herein about the depth of burial of these rocks in the Paleocene.

From the geochronological work, the most important conclusions are that (i) mylonitic strain in the Pukeashun granite is in part earliest Eocene in age (56 ± 1 Ma), (ii) rocks in the footwall of the Okanagan - Eagle River fault cooled through the closure temperature of amphiboles to loss of ^{40}Ar in the Paleocene to Middle Eocene, whereas hangingwall rocks cooled mainly in the Cretaceous, and (iii) average Paleocene-Eocene cooling rates of the footwall rocks were at least $15^\circ\text{C}/\text{m.y.}$ and probably were much greater. Extensional shearing on the west side of the Shuswap complex (west side down) can therefore no longer be considered to have begun much later than east-side-down extensional shearing on the east flank of the complex, and the rapid cooling rates can be attributed to coeval extensional tectonic denudation.

Through palinspastic restoration of early Tertiary normal faults, quantitative estimates have been made for the amount of displacement on the Okanagan - Eagle River fault (~ 30 km), the amount of extension across the Omineca belt at 51°N latitude (~ 45 km, or 20-25%), and the Late Paleocene crustal thickness (> 50 km). Comparison of the present and restored crustal cross sections lends support to the hypothesis that extension was partitioned differentially in the upper and lower plates of the major normal faults, and shows that models wherein the lower crust flows toward areas thinned during extension are inapplicable to the Shuswap complex.

APPENDIX A

Descriptions of mylonitic rocks and other fault rocks

TEMPLATE

Station # Map unit; location* (1:50 000 map sheet; UTM coordinates (grid zone 11))

Sense of shear (azimuth of stretching lineation in parentheses)

Kinematic indicators observed in field

Thin-section descriptions (sample #, rock type, mineralogy, microfabrics and kinematic indicators)

*locations shown in Plate 1

MINERAL SYMBOLS used in appendix A (from Kretz 1983)

Act actinolite
Bt biotite
Cal calcite
Chl chlorite
Ep epidote
Grt garnet
Gr graphite
Hbl hornblende
Hem hematite

Kfs K-feldspar
Mag magnetite
Ms muscovite
Pl plagioclase
Py pyrite
Qtz quartz
Rt rutile
Sil sillimanite

Other mineral abbreviations:

carb = carbonate (Cal?)
feld = Pl ± Kfs; undifferentiated feldspar
QF = Qtz + feld; quartzofeldspathic
ser = sericite
wh mica = Ms +/- ser

6 Hu; N of Yard Ck. (Malakwa 82L/15; 373800E, 563850N)

Top-to-W (271)

C/S fabric, asymmetrical extensional shear bands, rotated feldspar porphyroclasts

19 Hu; Mara Lk. (Sorrento 82L/14; 357500E, 5624600N)

Top-to-WNW (285)

C/S fabric; asymmetric Qtz pressure shadows on Grt porphyroblasts in sample LJ89-19A

J87-19: pelite schist tectonite

Grt-Sil-Bt-Pl-Kfs?-Qtz; Chl after Bt locally abundant; minor crosscutting Ms

C/S fabric (S: flattened felds and lenticular domains of Qtz; C: discrete layers of Qtz $\pm$ feld + Bt + Sil)

26 Hu; SE of Mara Lk. (Sorrento 82L/14; 358300E, 5623900N)

Top-to-W (277)

Asymmetrical extensional shear bands, foliation fish, asymmetrical pressure shadows of Qtz on Grt porphyroblasts in slabbed hand specimen J87-26

27 Hs; Eagle R. valley (Malakwa 82L/15; 363500E, 5635500N)

Top-to-W (270)

C/S fabric, asymmetrical extensional shear bands

28 Hs; Eagle R. valley (Malakwa 82L/15; 364200E, 5636200N)

Top-to-W (262)

C/S fabric

44 Hu; Pinnacle peak near Blue Lk. (Malakwa 82L/15; 380800E, 5630300N)

Top-to-W (255)

C/S fabric, asymmetrical extensional shear bands

J87-44: migmatitic pelite protomylonite

Sil-Bt-Pl-Qtz; intense retrogression of Bt to Chl and random sprays and needles of Rt; random plates of Ms cut Chl and Rt

C/S fabric (S: micas anastomosing around polycrystalline Qtz lenses and local flattened Pl; C: layers of Bt/Chl + coarse Sil)

Asymmetric extensional shear band (Chl + wh mica + fine-grained recrystallized Qtz; Chl pinches and swells into bundles)

65 Hdh; SE of Mt. Griffin (Malakwa 82L/15; 393500E, 5640700N)

Top-to-NE ? (044)

Probable C/S fabric in pegmatite

75 Hu; Wap Lk. (Malakwa 82L/15; 390500E, 5634900N)

Top-to-E ? (094)

Possible C/S fabric and shear bands; thin section shows no obvious asymmetry

98 Hh; Mt. Griffin (Malakwa 82L/15; 389000E, 5643100N)

Top-to-ENE ? (070)

Possible C/S fabric in pegmatite

131 Hh; 7 km W of Wap Lk. (Malakwa 82L/15; 383900E, 5634500N)

Top-to-SW (234)

Asymmetrical extensional shear bands in migmatitic pelite

132 Hh; 7 km W of Wap Lk. (Malakwa 82L/15; 383600E, 5634700N)

Top-to-ENE (060)

C/S fabric, asymmetrical extensional shear bands in pegmatite

134 Hh; 5 km W of Wap Lk. (Malakwa 82L/15; 385200E, 5634200N)

Top-to-NE (050)

C/S fabric, asymmetrical extensional shear bands in pegmatite

171 Hs; 0.75 km W of Mara Lookout (Malakwa 82L/15; 369200E, 5625200N)

Top-to-NE ? (080)

A few asymmetric Qtz pressure shadows on Grt porphyroblasts in outcrop and in slabs cut from hand specimens

173 Hs; 0.5 km W of Mara Lookout (Malakwa 82L/15; 369600E, 5625200N)

Top-to-WSW (250)

Discrete shear zone 5-30 cm wide cuts W-verging F3 fold in migmatite

J87-173: migmatite tectonite

Grt-Sil-Bt-Qtz-feld(Pl?); random sprays of Ms in leucosome

Deformation bands and low-angle subgrains in Qtz, no consistent asymmetry

Myrmekite at feld grain contacts

No kinematic indicators in thin section

174 Hs; Eagle R. valley (Malakwa 82L/15; 363600E 5635600N)

Top-to-W (270)

C/S fabric, feldspar porphyroclast systems (including the complex σ - δ type shown in Plate 5f), asymmetrical extensional shear bands

J88-174A: migmatitic pelite mylonite

Grt-Sil-Bt-Pl-Kfs-Qtz; some late crosscutting wh mica; Chl in pressure fringes on Grt

C/S fabric (S: lenticular QF domains separated by pelitic domains; C: pelitic layers)

240 Hs; ridge N of Yard Ck. (Malakwa 82L/15; 378200E, 5640900N)

Top-to-W (262)

C/S fabric, feldspar porphyroclast systems, asymmetrical extensional shear bands, west-verging isoclinal folds

J88-240: migmatitic pelite mylonite (~20% porphyroclasts)

Sil-Ms-Bt-Pl-Kfs?-Qtz; Bt locally replaced by Chl; matrix is fine-grained Bt + wh mica + Qtz $\pm$ feld $\pm$ Sil

σ -type feldspar porphyroclasts

Ms fish

Local foliation fish

Incipient shear bands

241 Hs; ridge N of Yard Ck. (Malakwa 82L/15; 377700E, 5640700N)

Top-to-W (270)
C/S fabrics

242 Huh; N of Yard Ck. (Malakwa 82L/15; 380000E, 5640200N)

Top-to-(N)E (no lineation here)
Stairstep tails on σ -type feldspar porphyroclasts on outcrop face oriented approximately E-W

285 Hs; Eagle R. valley (Malakwa 82L/15; 365400E, 5639200N)

Top-to-W (262)
C/S fabric, rotated feldspar porphyroclasts

287 Hs; Eagle R. valley (Malakwa 82L/15; 363800E, 5638200N)

Top-to-W (264)
C/S fabric

J88-287: migmatitic pelite protomylonite
Sil-Bt-Pl-Kfs-Qtz; some secondary wh mica and Chl

Deformation bands, some subgrains in Qtz; myrmekite and mantles of recrystallized new grains on Kfs

Oblique flattened feldspars
Myrmekite preferentially developed on long sides of Kfs (quarter structure)

291 Hs; ridge N of Yard Ck. (Malakwa 82L/15; 376200E, 5640000N)

Top-to-W (273) based mainly on thin section
Chlorite and quartz slickenfibres indicating slip direction observed in 2 places in outcrop

J88-291: hosed-over mung; migmatitic pelite mylonite ($\geq 30\%$ porphyroclasts); matrix is sericite and fine-grained recrystallized Bt + Qtz $\pm$ feld
Sil-Bt-Pl-Kfs?-Qtz

δ -type feld and Qtz porphyroclast systems
Oblique inclined feld porphyroclasts with asymmetric pressure shadows of Bt + Qtz
West-vergent microfolds

- 318** Hu; Shuswap R. valley 4 km E of Ashton Ck. (Mabel Lake 82L/10; 361500E, 5601000N)

Top-to-WSW (258)

Several rotated feldspar porphyroclast systems in migmatitic biotite schist

- 364** Hs; ridge N of Yard Ck. (Malakwa 82L/15; 382900E, 5642900N)

Top-to-WSW in field (244)

Slickenfibres on discrete, polished, chloritic surfaces indicate slip parallel to stretch in 2 places

J88-364: semipelite schist

Grt-Bt-Ms-Chl-feld-Qtz; most of the Bt is replaced by Chl; Ms occurs as blades and intergrown with Chl

No kinematic indicators in thin section

Discrete shears parallel to foliation (Chl + wh mica + fine-grained Qtz and feld)

Qtz and feld flattened, deformation bands in Qtz but no good subgrains

- 373** M; Momich R. near E shore Adams Lk. (Cayenne Creek 82M/6; 331300E, 5690000N)

Top-to-WNW (283)

C/S fabric, oblique feldspar porphyroclasts

J88-373G: Muscovite-biotite monzogranite mylonite

C/S fabric (S: flattened and σ -type feld porphyroclasts; C: Bt and fine-grained brown material)

Oblique foliation in recrystallized Qtz aggregates

σ -type feld porphyroclast systems

Myrmekite and a Ms mat on quarters of one Kfs

J88-373A: Migmatitic pelite mylonite

Grt-Sil-Ms-Bt-Kfs-Pl-Qtz

σ -type Kfs porphyroclasts with tails of Bt + dynamically recrystallized Ksp

Some quarter mats of Ms and quarter myrmekite on Kfs

Ms fish

Asymmetrical microfolds

Incipient asymmetrical extensional shear bands; one long shear band has Ms + Bt parallel to it and Sil deflected into it

J88-373B: mylonitic pelite

Ms-Grt-Sil-Bt-Kfs-Pl-Qtz

Asymmetrical microfolds

C/S fabric (S: σ -type Pl porphyroclasts; C: pelite layers)

Mica fish (Ms and Bt)

Foliation fish (oblique Bt in pelite layers bounded by discrete C planes)
 Oblique foliation in recrystallized Qtz aggregates within leucosome
 Pressure shadows of Bt + Pl on Grt
 Asymmetrical extensional shear bands in Bt-rich layers

385 Hs; Eagle R. valley (Malakwa 82L/15; 363900E, 5639500N)

Top-to-WNW (295)
 Oblique feldspar porphyroclasts, C/S fabric

J88-385: migmatitic pelite mylonite
 Sil-Bt-Pl-Kfs?-Qtz; abundant secondary ser in pressure shadows, Ms in some pelitic layers; Sil in pressure shadows and along shear bands; pull-apart between one Kfs porphyroclast and Sil in pressure shadow contains ser + QF

 Oblique inclined feld porphyroclasts (most of their wings are in-plane)
 Incipient shear bands
 Backrotated lenticular foliation segments (foliation fish) in at least 3 places in thin section

397 Hu/Dg; Four Mile Ck. (Perry River 82M/2; 371700E, 5658200N)

Top-to-SW (216) based on thin sections J88-397D-1 and J88-397D-2
 No kinematic indicators observed in field
 Samples J88-397-D and J88-397-E described below are from middle of fault zone described in text; sample J88-398A is from ~10 m into hanging wall

J88-397D-1: migmatitic semipelite protomylonite
 Ms-Bt-feld-Qtz; more than half of Bt replaced by Chl

 C/S fabric (S: flattened felds and domains of Qtz polycrystals separated by Bt/Chl, feld porphyroclasts with Chl pressure fringes; C: phyllosilicates, dynamically recrystallized Qtz and feld)
 Ms fish
 2 asymmetrical extensional shear bands (defined by Chl with some relict Bt)

J88-397D-2: migmatitic semipelite mylonite
 Ms-Bt-feld-Qtz; more than half of Bt replaced by Chl

 C/S fabric as above
 Oblique foliation in recrystallized Qtz aggregates
 Ms fish
 Anastomosing shears defined by Chl + Bt + wh mica + opaque

J88-397E: quartzofeldspathic cataclasite breccia
 Protolith from hand sample looks like granite (pegmatite?), but very difficult to assess
 Silicified (subgrains in Qtz but no distinct grain boundaries) and then sericitized
 Cut by anastomosing seams up to 1 mm wide (2 mm in hand sample) of graphite that internally

contain fragments of wallrock elongate parallel to seam margins. Gr seams are cut locally by thin Hem veinlets and finally by dilational open fractures.

Foliation in rock defined by these Gr seams and by the elongate lenticular domains that they bound. In hand sample it looks as though there is a relic flattening fabric defined by alternate yellowish altered feld and Qtz grains or lenses, between which are thin rims/veinlets of Hem and probably ser.

J88-398A: epidote-biotite granodiorite gneiss

Much of Bt replaced by Chl; large Ep crystals could be primary

Moderate foliation; subgrains and some deformation bands in Qtz, a few bent Pl twins, undulose extinction in Qtz and Pl

Intense comminution mainly is localized around crosscutting fractures (crush zones) up to 1 mm wide. Foliation is discordant on opposite sides of prominent crush zone.

Some carb-Ep veinlets

424

Ls; Adams Lake (Albas 82M/2; 326600E, 5678900N)

No kinematic indicators

Steeply NNE-dipping brittle fault zone >35 m wide. In a transect through the zone from SW to NE, relatively unaltered Sil-Bt-Ms-Qtz schist and psammite give way gradationally to oxidized and retrograded schist containing abundant Chl after Bt and progressively more abundant Ms (5 m wide); black cataclasite containing Ms flakes, in sharp contact with adjacent zones (7 m wide); highly weathered and oxidized quartzite and Ms-Qtz schist (23 m wide); fractured and slightly oxidized pegmatite and weathered Hbl diorite (~100 m wide).

425

Ls; Adams Lake (Albas 82M/2; 327600E, 5680200N)

No kinematic indicators

Zone ~100 m wide of rusty and highly weathered Qtz-Ms and Sil-Bt-Qtz-Ms schists cut by steeply SW-dipping clay gouge zones.

452

Hs; 5 km W of Wap Lk. (Malakwa 82L/15; 385100E, 5634600N)

Top-to-SW (230)

Oblique feldspar porphyroclasts, numerous small shear bands, C/S fabric in a 1 cm-thick pelite layer

510

Hs; Mara Lk. (Sorrento 82L/14; 356700E, 5625300N)

Top-to-W (277)

Feldspar porphyroclast systems, C/S fabric (S: flattened felds and QF lenses; C: Chl shears spaced ~2 mm apart in some layers)

J88-510: migmatitic semipelite mylonite
 Grt-Hem-Chl-feld-Qtz; almost all of what was Bt has been replaced by Chl and Hem; some Ms in pressure fringe on Grt; feldspars (probably Kfs) are wickedly sericitized

Subgrains and dynamically recrystallized new grains of Qtz are abundant

C/S fabric (S: flattened feldspars; C: fine-grained recrystallized QF, concentration of Chl and opaque Hem)

Microfractures in Grt filled with Hem; younger microfractures filled with Qtz cut Grt at higher angle to C planes; abundant microfractures cut feldspars and Qtz-rich matrix, as do veins up to 3 mm wide of Qtz and Qtz-Hem

514 Hu; Four Mile Ck. (Perry River 82M/2; 372300E, 5658800N)

Top-to-WSW (257)

Asymmetrical extensional shear bands and mica fish

J88-514: migmatitic pelite schist tectonite

Sil-Pl-Ms-Qtz-Bt; almost half of Bt has been replaced by Chl, especially in shear bands; large Ms plates contain inclusions of Sil and Bt

Asymmetrical extensional shear bands defined by fine-grained Chl, and by some ser, Ms, Ep, and relic Bt

Ms fish (a few)

612 Dg?; near Kwikoit Ck. (Albas 82M/3; 332300E, 5672200N)

No kinematic indicators

Fractures, slickensides, small Qtz veins

J89-612: Chl-Ms-feld-Qtz schist

Chl is retrograde, some in large plates that crosscut older minerals; Ms is highly altered (ser?); felds altered to clay + ser

Qtz veinlets

613 Dg; near Kwikoit Ck. (Albas 82M/3; 332100E, 5671900N)

No kinematic indicators

Silicified breccia with scattered enclaves of Ms schist, abundant slickensides, abundant anastomosing fractures; grades up into highly fractured and chloritized orthogneiss

J89-613: Breccia

Silicified; protolith not recognizable

Irregular zones of coarse-grained polycrystalline Qtz (with subgrains and undulose extinction) separated by crush zones of fine-grained recrystallized Qtz; patches and veins of carb,

wh mica, Chl

J89-613A: Ep-Bt granitoid orthogneiss

About half of Bt is retrograded to Chl; feld is quite sericitized; Ep looks like it was there before retrogression, but difficult to tell; some crosscutting Ms plates

Annealed fabric; Qtz has undulose extinction, but polygonal to slightly embayed grain contacts

628 Dg; NW of Kwikoit Ck. (Albas 82M/3; 330900E, 5673100N)

Top-to-SE

Discrete brittle shears, 5-75 mm wide and spaced on order of 10-15 cm apart, dip steeply SE subparallel to compositional layering (048/72) in gneiss. Flattening fabric defined by oblique Bt plates within these zones indicates normal-sense slip.

629 Dg; NW of Kwikoit Ck. (Albas 82M/2; 330400E, 5674800N)

Top-to-E (087) based on slab cut from sample J89-629 (biotite granodiorite gneiss)

Asymmetrical extensional shear bands

653 Dg; Mt. Fowler (Albas 82M/3; 347300E, 5661700N)

Top-to-SSW (194)

A few σ -type feld porphyroclasts observed in outcrop and in slab cut from sample J89-653B

654 Dg; Mt. Fowler (Albas 82M/3; 346800E, 5662200N)

Top-to-WNW (302)

Asymmetrical extensional shear bands in orthogneiss adjacent to discrete shear zones up to 2 cm wide; shear sense corroborated by steps on slickensides

J89-654A: Biotite orthogneiss mylonite

Shear bands are defined by finely recrystallized Bt + QF

668 Dg; Cross Ck. (Albas 82M/3; 335800E, 5669100N)

No kinematic indicators

Orthogneiss cut by numerous Qtz and Ep-Qtz veins and by slickensided fractures

J89-668: Ep-Chl granitoid orthogneiss

Bt has all been replaced by Chl + opaque (Mag?); feld replaced by ser $\pm$ clay; crosscutting Ms

plates
Some carb veinlets
Qtz grains and subgrains have embayed contacts

716 KTgm; W of Pukeashun Mtn. (Albas 82M/3; 336800E, 5674000N)

Top-to-NW (314) based on thin section
No kinematic indicators observed in field

J89-716: garnet-muscovite monzogranite mylonite

C/S fabric (S: σ -type feld (mostly Kfs); C: wh mica shears cutting zones of intense dynamic recrystallization)
Ms fish

719 KTgm; W of Pukeashun Mtn. (Albas 82M/3; 338200E, 5675100N)

Top-to-NW (307) based on thin section
No kinematic indicators observed in field

J89-719: biotite-muscovite monzogranite mylonite

C/S fabric (S: σ -type feld (mostly Kfs); C: discrete shears of wh mica (mainly ser) + minor Bt, cutting zones of ribbon Qtz and recrystallized QF)
Ms fish

723 KTgm; Adams Lk. (Cayenne Creek 82M/6; 328800E, 5683700N)

Top-to-WNW (299) based on thin section
No kinematic indicators observed in field

J89-723: muscovite monzogranite mylonite (~50% porphyroclasts)

C/S fabric (S: flattened felds, lenticular QF domains; C: discrete shears of ser + Chl that cut parallel sub-mm-scale zones of dynamically recrystallized Qtz ribbons)
Oblique foliation in recrystallized Qtz aggregates well developed
Some Ms fish

725 KTgm; Adams Lk. (Cayenne Creek 82M/6; 327900E, 5681200N)

Top-to-NNW (328)
Outstanding C/S fabric; granite mylonite breaks into plates as thin as 3 mm along C planes

J89-725: muscovite monzogranite mylonite

C/S fabric (S: σ -type feld porphyroclasts; C: discrete to 100 μ m-wide shears of wh mica)
 Myrmekite quarter structures in some Kfs
 Ms fish
 Oblique foliation in recrystallized Qtz aggregates

746 Ls; 1 km E of Adams Lake (Albas 82M/2; 327000E, 5678600N)

No kinematic indicators
 Rusty Ms-Qtz semipelite, crumbly and cut by clay gouge zones.

752 M; E of Adams Lk. (Cayenne Creek 82M/6; 333400E, 5685600N)

Top-to-WNW (299) based on thin section
 No kinematic indicators observed in field

J89-752: migmatitic pelite mylonite (20-25% porphyroclasts)
 Ms-Bt-Kfs-Pl-Qtz

Ms fish
 Oblique foliation in recrystallized Qtz aggregates
 σ -type feldspars
 Myrmekite quarter structure on one Kfs porphyroclast
 Bt pressure fringes on some feldspars

754 TVsa; Victor Lk. (Revelstoke 82L/16; 301400E, 5644300N)

Top-to-NE (046)
 C/S fabric, σ -type and δ -type feldspar porphyroclast systems in migmatitic pelite mylonite

755 TVsa; 2 km NE of Three Valley Lk. (Revelstoke 82L/16; 300000E, 5644300N)

Top-to-NE (042)
 C/S fabric defined by oblique feldspar porphyroclasts with pressure shadows

LJ89-755-1: migmatitic pelite tectonite
 Kfs-Pl-Grt-Sil-Qtz; minor Chl after Bt

Pressure fringes of Bt + Sil on Grt porphyroclasts

LJ89-755-2: migmatitic pelite tectonite
Kfs-Pl-Grt-Sil-Qtz; minor Chl after Bt

Oblique flattened Grt porphyroclasts with pressure fringes of Bt + Sil
Oblique flattened feld porphyroclasts and Qtz aggregates with pressure shadows of Bt + Qtz
Bt fish (only one observed)

762 KTgm; Pukeashun Mtn. (Albas 82M/3; 342900E, 5674500N)

Top-to-WNW (301)
Feldspar porphyroclasts

J89-762: garnet-biotite-muscovite monzogranite protomylonite

C/S fabric (S: σ -type felds (mainly Kfs); C: Ms + Bt, short Qtz ribbons and trails of recrystallized QF from porphyroclast tails)
Quarter structures (myrmekite and Ms mats on some Kfs)
Asymmetrical extensional shear bands (Ms, Bt)

J89-762A: migmatitic Bt-feld-Qtz semipelite gneiss protomylonite
Bt-Pl-Kfs-Qtz; some Chl after Bt

C/S fabric (S: flattened felds and oblique domains of ribbon Qtz + Bt; C: Qtz ribbons)
Oblique foliation in recrystallized Qtz ribbons

764 KTgm; Pukeashun Mtn. (Albas 82M/3; 343100E, 5674500N)

Top-to-NW (314)
C/S fabric in tongues of granite dikes that are flattened into parallelism with the shear foliation

770 KTgm; W of Pukeashun Mtn. (Albas 82M/3; 340700E, 5676100N)

Top-to-NW (304)
C/S fabric

J89-770: chlorite-muscovite monzogranite mylonite
Chl is retrograde after Bt

C/S fabric (S: σ -type Kfs and Pl porphyroclasts; C: wh mica (mainly ser) and recrystallized QF)
Ms fish
Quarter structure (myrmekite and Ms quarter mat on one Kfs)
Some oblique foliation in recrystallized Qtz aggregates

788

Hu/TVsa; Eagle Pass Ridge N of Three Valley Lk. (Revelstoke 82L/16; 399300E, 5646400N)

Top-to-NE (046) overall, with superimposed top-to-SW in discrete chloritic shears

Sheet of migmatitic pegmatite is everywhere foliated, is well-lineated in zones of highest strain, contains screens of marble and calcsilicate gneiss. In coarser-grained layers of pegmatite, feldspar porphyroclasts indicate ambiguous but predominantly top-to-NE shear sense. Mylonitic shears, ~10-15 cm wide, have gradational contacts with the coarser-grained pegmatite. Most of these zones have retrograde Chl after Bt on C-surfaces, and Chl pressure shadows on very round feldspar porphyroclasts indicate top-to-SW shear. Other zones have black matrix rich in nonretrograded Bt, and have round feldspar porphyroclasts with recrystallized tails (including δ -type in sample J89-788-3) that indicate top-to-NE shear. Although crosscutting relations were not observed between these two sets of shear zones, lack of retrogression in the NE-directed ones suggests that they are older. Structurally beneath this pegmatite sheet are Bt-Hbl-Pl gneisses in which Hbl crystals define an oblique foliation consistent with top-to-NE shear.

J89-788-3: migmatitic pegmatite mylonite

Bt-Kfs?-Qtz-Pl; Chl as random plates and as retrograde product after about 10% of Bt

Primary recrystallization prominent in thin section, confirms mylonitic texture; foliation only is visible at outcrop and hand-sample scales

Pl porphyroclast systems, including δ -type, are best seen in slabs cut from sample

845

E?c/Dg; Cambie cliffs (Malakwa 82L/15; 368400E, 5640700N)

Top-to-W (slickenlines trend 280) based on field observation of foliation (S) in mylonitic marble oblique to sheared contact (C) with overlying silicified and altered hangingwall rocks; no reliable kinematic indicators in thin sections of marble

Above alluvium in creekbed is 1.5 m-thick exposure of mylonitic marble with oblique E-dipping foliation (J89-845-3, 5). A knife-sharp contact separates the marble from overlying silicified orthogneiss (J89-845-6) and amphibolite (J89-845-7), which are cut by anastomosing chloritic shears. Silicification and retrograde alteration decrease upward in the orthogneiss (J89-845-8).

J89-845-3: impure marble mylonite

Chl-Py-Gr-Qtz-feld-Cal

Pressure shadows of Cal on Qtz and feld porphyroclasts, are texturally indistinct from recrystallized Cal of surrounding matrix; larger porphyroclasts (> 500 μ m) are pulled apart (or embayed), and the pull-aparts contain recrystallized Cal as in the matrix
C-planes are Chl + opaque Gr

J89-845-5: mylonitic impure marble

wh mica - opaque (Py $\pm$ Gr) - Qtz - Cal

Compositional layering defined by alternate coarse Cal and fine-grained Qtz + wh mica + Py + Cal

Coarse Cal from secondary recrystallization

J89-845-6: altered orthogneiss

Act-Ep-Pl-Qtz; alternating zones of Qtz, QF-rich, and Ep-rich altered gneiss; Chl-carb-Ep

alteration and veins; intensely fractured

J89-845-7: amphibolite gneiss
Qtz-Ep-Pl-Act; cut by veinlets of Qtz, carb, and Ep-carb; some Pl near veinlets are sericitized

J89-845-8: biotite monzogranite(?) gneiss
About half of the Bt is chloritized; some Ep is associated with Chl, probably is also secondary; much of the feld is untwinned and is highly sericitized; Ep-Qtz veinlets

891 Hu; Pinnacle peak near Blue Lk. (Malakwa 82L/15; 381200E, 5630500N)

Top-to-W (252)

Feldspar porphyroclast systems, Type 2 asymmetrical pull-apart (cf. Hammer 1986) in amphibolite layer

J89-891: migmatitic pelite mylonite
Sil-Bt-Pl-Kfs-Qtz; Chl after Bt, moderate ser alteration of feld; rare Ms plates are parallel to oblique with respect to Bt foliation (probably retrograde)

C/S fabric (S: flattened feld porphyroclasts and lenticular aggregates of recrystallized Qtz; C: discrete to diffuse bands of Bt/Chl + Sil + fine-grained recrystallized Qtz ($\pm$ feld))

A couple of incipient shear bands (Chl-Bt-wh mica)

Dynamic recrystallization and asymmetric myrmekite quarter structure on edges of Kfs

938 Hs; Cambie cliffs (Malakwa 82L/15; 368900E, 5640500N)

Top-to-W (275)

C/S fabric, rotated feldspar porphyroclasts, shear bands

J89-938: migmatitic pelite mylonite
Sil-Bt-Pl-Kfs?-Qtz

Within this thin section are two domains, displaying 16° angular discordance in the dip of their respective C surfaces and characterized by different degrees of finite strain. The older domain dips gently eastward and is the less highly strained of the two. The younger crosscutting shear domain dips gently westward, and is characterized by more intense comminution of grains by dynamic recrystallization. C planes in the younger domain are discrete shears up to 0.5 mm wide, spaced on sub-mm scale; C planes in the older domain are less well-defined, laterally less continuous, more widely spaced, and somewhat coarser-grained.

C/S fabric (S: flattened felds (including σ -type) and QF lenses surrounded by thin layers of Bt + Sil; C: very finely comminuted Bt + QF)

Oblique foliation in recrystallized Qtz aggregates (occurs within Qtz and QF lenses in both domains, and in ribbon-Qtz layers in younger domain)

West-verging microfolds (in ribbon-Qtz layers of younger domain; Sil appears folded around hinges, and oblique foliation in recrystallized Qtz aggregates is approximately axial planar to these microfolds)

A few shear bands defined by Bt

Brittle fractures cut across felds and matrix at high angles to C surfaces in both domains

940 Hs; Cambie cliffs (Malakwa 82L/15; 369900E, 5640100N)

Top-to-WNW (290)

C/S fabric, asymmetrical extensional shear bands

J89-940: migmatitic pelite mylonite
Sil-Bt-Pl-Kfs?-Qtz

Well-developed subgrains and recrystallized new grains in Qtz and feld; deformation bands and undulose extinction in Qtz

C/S fabric (S: flattened feldspars and lenticular leucosome pods surrounded by Bt and Sil; C: Bt + Sil + fine-grained recrystallized QF)

Oblique foliation very well developed in recrystallized Qtz aggregates

Local lenticular back-rotated foliation segments (foliation fish)

A few shear bands defined by Bt and fine-grained recrystallized Qtz $\pm$ feld

Probable δ and complex σ - δ Pl porphyroclast systems

942 Hs; Cambie cliffs (Malakwa 82L/15; 371400E, 5640600N)

Top-to-WNW (288)

C/S fabric, rotated feld porphyroclasts, asymmetrical tight folds, asymmetrical extensional shear bands

1097 KTgm; near Seymour Arm (Albas 82M/3; 355900E, 5671400N)

Top-to-W (280) based on thin section from 60 cm-thick highly strained leucogranite sill in retrograded Chl-Ms-Qtz semipelitic schists

No kinematic indicators observed in field

J90-1097-1: Grt-Bt-Chl-feld-Qtz psammite
Nearly all Bt has been replaced by Chl
Qtz displays prominent deformation bands, lobate to serrate grain contacts

Chl pressure fringes on felds look E-verging

J90-1097-2: muscovite monzogranite mylonite
Felds are brownish, altered to ser $\pm$ clay

C/S fabric (S: lenticular flattened Qtz and feld domains, some flattened felds; C: mainly ser, also very thin Hem residues)

Oblique foliation in recrystallized Qtz within lenticular Qtz domains

Ms fish

1102 KTgm; E shore Seymour Arm (Perry River 82M/2; 360500E, 5665000N)

No kinematic indicators observed in field or in thin section

Sheets of highly strained medium-grained muscovite leucogranite, several cm to tens of cm thick, interlayered with pegmatite, migmatitic Bt-Qtz-feld gneiss and Bt-Ms schist. Boudinage in a few places. Stretching lineation (280-11) defined by Qtz and mineral aggregates.

232/22 Hs; Mara Lk. (Malakwa 82L/15; 361000E, 5630800N)

Top-to-WNW (285)

Asymmetrical extensional shear bands, foliation fish

J86-2 Huc; Mara Lk. (Malakwa 82L/15; 359600E, 5629600N)

Top-to-WNW (286), based on thin section

No kinematic indicators observed in field

J86-2: migmatitic semipelite mylonite

Sil-Grt-Bt-Pl-Kfs?-Qtz; abundant Chl after Bt, especially in pressure fringes

Subgrains, deformation bands, and mantles of recrystallized new grains in Qtz

Weak C/S fabric (S: alternate domains of QF and Bt, and domains of feld with pressure fringes; C: fine-grained Bt and QF)

Some asymmetrical extensional shear bands defined by Bt

J86-22 Hs; Mara Lk. (Malakwa 82L/15; 359800E, 5628200N)

Top-to-WNW (285)

Good C/S fabric

86-3 Hs; Mara Lk. (Malakwa 82L/15; 360100E, 5628600N)

Top-to-WNW (282)

Rotated feldspar porphyroclasts, west-verging folds, possible C/S

APPENDIX B

Amphibole samples dated by $^{40}\text{Ar}/^{39}\text{Ar}$ methods

Abbreviations: Bt=biotite, Chl= chlorite, Di=diopside, Ep=epidote, feld=feldspar, Grt=garnet, Hbl=hornblende, Ms=moscovite, Pl=plagioclase, Sil=sillimanite, Qtz=quartz.

UTM coordinates are in Grid Zone 11.

J89-18 Amphibolite (unit Hu, E side of Mara Lake; 357800E, 5624700N; elev. 1200'). Long roadcut on Hwy. 17A; layered amphibolite boudins in matrix of quartzite, Qtz-feld-Bt schist, and graphitic calcisilicate gneiss; overlain by succession of interlayered Grt-Sil-Bt schist and Bt-Qtz-feld gneiss.

J88-514H Bt-Pl-Hbl amphibolite (unit Hu, Four Mile Creek; 372300E, 5658800N; elev. 4640'). Amphibolite boudins in migmatitic Qtz-Bt schist; from NE end of long roadcut that exposes strongly foliated granitoid gneisses and less-strained leucogranites, next to road going up off of main logging road to Four Mile Creek.

J89-36 Bt-Hbl quartz diorite gneiss (unit Hdh, Trans-Canada Hwy. 1 km NW of Taft, across from Eagle River Fish Hatchery; 384400E, 5650800N; elev. 1450'). Sampled ~50 cm away from a pegmatite dike.

J88-351A Amphibolite (unit Ha, Mount Mara lookout road; 371200E, 5624400N; elev. 6550'). Outcrops on W side of road; amphibolite (>50% Hbl), and psammitic Qtz-feld and Grt-Bt-Qtz-feld paragneiss. Amphibolite locally contains Di and Bt. Pegmatite forms ~50% of exposure, amphibolite and psammite each ~25%.

J89-905 Bt-Di-Pl-Hbl amphibolite (unit Ha, 2 km E of Blue Lake; 382900E, 5630400N; elev. 6070'). From succession of amphibolite, marble and psammitic paragneiss cut by pegmatites.

J89-3H Di-Hbl-Pl amphibolite (unit TVsa, Three Valley Gap, long roadcut on Trans-Canada Highway; 397950E, 5642850N; elev. 1700'). From one of many large amphibolite boudins in Grt-Sil-Bt-Qtz-Pl-Kfs semipelite paragneiss.

J89-788H Bt-Pl-Hbl gneiss (unit TVsa, Eagle Pass Ridge, Jordan Range N of Three Valley Gap; 399350E, 5646400N; elev. 6420'). Mafic gneiss beneath 300-m-thick sheet of sheared migmatitic pegmatite.

- J89-970** Bt-Hbl granodiorite orthogneiss, containing abundant Ep + titanite (unit Dgh, Shuswap Range N of Queest Creek; 366050E, 5657250N; elev. 2890'). Roadcut ~100 m long of orthogneiss cut by thin ($\leq$ 1 m) pegmatite dikes. Sample locality is 15 m from nearest dike.
- J89-662H** Hbl-Bt quartz diorite gneiss (unit Dg, Scotch Creek; 343800E, 5665550N; elev. 3810'). Cliff beneath powerline along Scotch Creek Road, Bt granodiorite gneiss contains screens and xenoliths of diorite gneiss. Sample is from a screen.
- J90-1092** Bt-Hbl-Qtz-Pl gneiss (unit Lu, 6 km N of Mount Fowler; 348700E, 5668300N; elev. 4260'). From logging-road cuts exposing interlayered Bt-Hbl-Qtz-Pl gneiss, Sil-Bt-Ms schist, siliceous paragneiss and calcsilicate schist.
- J89-417H** Amphibolite (unit Ls, Adams East Road; 325000E, 5674400N; elev. 2450'). From pod of amphibolite in Bt-Ms-Qtz schist (with rare Grt and Sil).
- J89-642H** Hbl-Chl schist (unit Eca, between Adams Lake and Kwikoi Creek; 327750E, 5674500N; elev. 5040'). From succession of buff-brown marble, fine-grained micaceous quartzite, Chl-sericite phyllite, Hbl-Chl schist and Grt-Bt-Ms schist, near contact with granodiorite gneiss of unit Dg.

APPENDIX C: Electron microprobe analyses of amphiboles dated by ⁴⁰Ar/³⁹Ar

J89-18

J88-514H

J89-36

J88-351A

tschermakitic hornblende										tschermakitic hornblende										potassium ferroan pargasitic hornblende										edenitic hornblende									
	1	2	3	4	5	avg.	SD	1	2	3	4	5	avg.	SD	1	2	3	4	5	avg.	SD	1	2	3	4	5	avg.	SD											
SiO ₂	43.30	43.27	42.93	43.02	43.37	43.18	.19	42.98	41.53	41.76	41.54	41.55	41.87	.63	43.20	43.39	43.31	43.33	42.68	43.18	.29	44.03	44.05	43.65	44.99	45.34	44.41	.72											
Al ₂ O ₃	11.15	11.18	11.34	11.12	10.19	11.00	.46	11.22	13.21	13.36	13.18	12.67	12.73	.88	11.93	12.08	11.99	11.98	12.19	12.03	.10	11.03	11.34	11.54	10.93	9.48	10.86	.81											
TiO ₂	1.92	1.94	2.43	2.39	2.32	2.20	.25	.88	.97	1.11	.89	.91	.95	.09	1.17	1.21	1.30	.94	1.13	1.15	.13	1.92	1.90	2.03	1.93	1.70	1.90	.12											
Cr ₂ O ₃	.05	.04	.04	.07	.10	.06	.03	.00	.02	.02	.00	.08	.02	.03	.04	.06	.06	.04	.06	.05	.01	.16	.08	.16	.04	.08	.10	.05											
Fe ₂ O ₃	2.31	3.12	2.79	2.63	3.07	2.71	.33	6.65	6.68	7.10	5.73	7.26	6.69	.60	2.76	2.61	2.20	3.55	4.19	3.08	.80	.67	.22	1.14	.61	.00	.00	.44											
FeO	15.17	14.87	15.10	14.85	14.25	14.91	.36	11.98	11.91	11.85	12.24	12.16	12.02	.17	13.21	13.74	13.90	13.27	12.81	13.38	.44	13.24	14.01	12.93	13.27	10.27	13.22	1.44											
MnO	.24	.36	.26	.28	.33	.29	.05	.32	.31	.38	.30	.32	.32	.03	.45	.44	.36	.35	.40	.40	.05	.29	.20	.18	.20	.31	.24	.06											
MgO	9.51	9.35	9.44	9.68	9.81	9.56	.19	10.16	9.29	9.20	9.25	9.25	9.43	.41	10.41	10.22	10.23	10.33	10.20	10.27	.09	11.62	11.44	11.58	11.86	7.73	10.85	1.75											
CaO	11.49	11.39	11.46	11.51	11.36	11.44	.06	11.95	11.62	11.56	11.51	11.77	11.68	.18	11.99	12.19	12.02	12.12	11.94	12.05	.10	11.95	12.04	11.94	11.96	12.08	11.99	.06											
Na ₂ O	1.48	1.45	1.52	1.48	1.33	1.45	.07	1.07	1.12	1.17	1.20	1.14	1.14	.05	1.08	.95	1.04	.99	1.06	1.02	.05	1.72	1.79	1.72	1.64	.85	1.54	.39											
K ₂ O	1.19	1.21	1.26	1.29	.99	1.19	.12	.88	1.21	1.19	1.21	1.08	1.11	.14	1.6	1.60	1.66	1.62	1.64	1.62	.03	1.14	1.24	1.17	1.16	.59	1.06	.27											
F	.03	.06	.00	.00	.46	.11	.20	.04	.05	.00	.00	.10	.04	.04	.15	.00	.07	.00	.17	.08	.08	.07	.22	.13	.17	.09	.14	.06											
Cl	.02	.01	.00	.00	.00	.01	.01	.07	.09	.05	.08	.06	.07	.02	.02	.02	.05	.05	.05	.04	.02	.01	.04	.04	.01	.05	.03	.02											
total	97.86	98.25	98.57	98.32	97.58	98.11	.39	98.20	98.01	98.75	97.13	98.35	98.07	.60	98.01	98.51	98.19	98.57	98.52	98.35	.25	97.85	98.57	98.21	98.77	88.57	96.34	4.39											

Formula based on 23 oxygens

	Si	Al ^{IV}	Al ^{VI}	Ti	Cr	Fe ⁺³	Fe ⁺²	Mn	Mg	Ca	Na ^{VI}	Na	K	F	Cl
Si	6.491	6.469	6.401	6.424	6.531	6.464	.052	.028	.028	.028	.028	.028	.028	.028	.028
Al ^{IV}	1.509	1.531	1.599	1.576	1.469	1.536	.052	.028	.028	.028	.028	.028	.028	.028	.028
Al ^{VI}	.461	.438	.394	.382	.340	.404	.048	.048	.048	.048	.048	.048	.048	.048	.048
Ti	.217	.218	.273	.268	.262	.248	.028	.028	.028	.028	.028	.028	.028	.028	.028
Cr	.006	.004	.005	.009	.012	.007	.003	.003	.003	.003	.003	.003	.003	.003	.003
Fe ⁺³	.260	.351	.313	.295	.348	.306	.038	.038	.038	.038	.038	.038	.038	.038	.038
Fe ⁺²	1.903	1.859	1.883	1.855	1.795	1.867	.041	.041	.041	.041	.041	.041	.041	.041	.041
Mn	.030	.045	.033	.036	.041	.037	.006	.006	.006	.006	.006	.006	.006	.006	.006
Mg	2.124	2.084	2.099	2.154	2.201	2.133	.047	.047	.047	.047	.047	.047	.047	.047	.047
Mg	5.001	4.999	5.000	4.999	4.999	5.002	.001	.001	.001	.001	.001	.001	.001	.001	.001
Ca	1.845	1.825	1.831	1.842	1.833	1.835	.008	.008	.008	.008	.008	.008	.008	.008	.008
Na ^{VI}	.155	.175	.169	.158	.167	.165	.008	.008	.008	.008	.008	.008	.008	.008	.008
Na	2.000	2.000	2.000	2.000	2.000	2.000	.000	.000	.000	.000	.000	.000	.000	.000	.000
Na	.274	.246	.270	.271	.222	.257	.022	.022	.022	.022	.022	.022	.022	.022	.022
K	.228	.230	.239	.245	.191	.227	.021	.021	.021	.021	.021	.021	.021	.021	.021
F	.502	.476	.509	.516	.413	.484	.042	.042	.042	.042	.042	.042	.042	.042	.042
F	.016	.026	.000	.002	.221	.053	.095	.095	.095	.095	.095	.095	.095	.095	.095
Cl	.005	.003	.000	.000	.001	.002	.002	.002	.002	.002	.002	.002	.002	.002	.002
Cl	.021	.029	.000	.002	.222	.055	.094	.094	.094	.094	.094	.094	.094	.094	.094

J89-905

J89-3H

J89-788H

J89-970

potassian ferroan pargasitic hornblende										tremolitic hornblende										potassian magnesian hastingsite										potassian magnesian hastingsitic hornblende									
	1	2	3	4	5	avg.	SD	1	2	3	4	5	avg.	SD	1	2	3	4	5	avg.	SD	1	2	3	4	5	avg.	SD	1	2	3	4	5	avg.	SD				
SiO2	41.43	40.53	41.20	40.56	45.55	41.85	2.10	54.11	54.46	54.38	54.38	54.08	54.28	.17	40.94	41.34	41.06	40.57	41.06	40.99	.28	42.41	42.89	42.79	43.07	43.00	42.83	42.41	42.89	42.79	43.07	43.00	42.83	.26					
Al2O3	12.84	14.07	13.08	14.05	10.08	12.82	1.63	1.34	1.31	1.46	1.48	1.37	1.39	.07	12.89	12.53	13.04	12.97	12.96	12.88	.20	11.83	11.58	11.46	11.39	11.93	11.64	11.83	11.58	11.46	11.39	11.93	11.64	.23					
TiO2	1.40	1.53	1.52	1.43	.78	1.33	.31	.01	.00	.00	.03	.09	.03	.04	1.77	1.77	1.68	1.71	1.71	1.73	.04	.54	.55	.61	.61	.62	.59	.54	.55	.61	.62	.59	.04						
Cr2O3	.02	.02	.05	.05	.00	.03	.02	.00	.03	.00	.00	.00	.01	.01	.00	.03	.06	.00	.00	.02	.03	.03	.02	.05	.02	.07	.00	.03	.02	.05	.02	.07	.00	.03	.03				
Fe2O3	1.49	1.40	1.09	1.58	1.55	1.40	.20	24.57	24.93	24.55	24.36	24.11	24.50	.30	5.03	4.35	6.11	5.98	5.03	5.31	.74	4.58	5.69	4.58	5.04	4.37	4.87	4.58	5.69	4.58	5.04	4.37	4.87	.53					
FeO	15.46	16.61	15.88	16.02	15.66	15.94	.44	.00	.00	.00	.00	.00	.00	.00	13.86	13.60	12.73	12.87	13.68	13.34	.51	14.35	13.27	14.09	13.77	13.89	13.85	14.35	13.27	14.09	13.77	13.89	13.85	.40					
MnO	.34	.22	.32	.13	.29	.26	.09	.72	.74	.76	.89	.76	.77	.07	.34	.49	.50	.43	.38	.43	.07	.38	.47	.27	.41	.33	.37	.38	.47	.27	.41	.33	.37	.08					
MgO	9.47	8.74	9.35	8.88	10.38	9.36	.65	18.26	18.28	18.01	18.17	18.13	18.17	.11	9.20	9.17	9.14	8.92	8.92	9.07	.14	9.32	9.63	9.62	9.66	9.69	9.58	9.32	9.63	9.62	9.66	9.69	9.58	.15					
CaO	12.02	12.04	11.92	11.81	12.39	12.04	.22	1.05	1.05	1.19	1.12	1.08	1.10	.06	11.67	11.43	11.41	11.21	11.33	11.41	.17	11.93	11.82	11.87	11.90	11.97	11.90	11.82	11.87	11.90	11.97	11.90	11.97	.06					
Na2O	1.72	1.80	1.85	1.76	1.23	1.67	.25	.03	.03	.06	.08	.04	.05	.02	1.48	1.36	1.59	1.41	1.45	1.42	.05	1.26	1.19	1.23	1.20	1.16	1.21	1.26	1.19	1.23	1.20	1.16	1.21	.04					
K2O	1.69	2.09	1.90	.20	.87	1.75	.53	.01	.04	.01	.01	.01	.01	.01	1.72	1.63	1.41	1.69	1.70	1.67	.05	1.51	1.44	1.47	1.41	1.51	1.47	1.51	1.44	1.47	1.41	1.51	1.47	.04					
F	.26	.20	.31	.24	.13	.23	.07	.00	.00	.00	.00	.00	.00	.00	.22	.05	.05	.09	.09	.10	.07	.29	.04	.20	.15	.16	.17	.29	.04	.20	.15	.16	.17	.09					
Cl	.09	.15	.06	.13	.13	.11	.04	.00	.01	.00	.00	.00	.00	.00	.21	.20	.23	.24	.26	.23	.02	.00	.01	.03	.01	.04	.02	.00	.01	.03	.01	.04	.02	.02					
total	98.23	99.40	98.53	98.84	99.04	98.79	.45	100.10	100.88	100.42	100.52	99.67	100.31	.46	99.33	97.95	99.01	98.09	98.57	98.60	.59	98.42	98.63	98.24	98.69	98.67	98.53	98.42	98.63	98.24	98.69	98.67	98.53	.19					

Formula based on 23 oxygens

	Si	Al ^{IV}	Al ^{VI}	Ti	Cr	Fe ⁺³	Fe ⁺²	Mn	Mg	Ca	Na ^{VI}	Na	K	F	Cl	Total
6.256	6.096	6.220	6.119	6.729	6.286	2.58	2.28	2.58	2.58	2.58	2.58	2.58	2.58	2.58	2.58	2.58
1.744	1.904	1.780	1.881	1.271	1.714	.258	.258	.258	.258	.258	.258	.258	.258	.258	.258	.258
8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
.541	.591	.548	.618	.484	.556	.051	.051	.051	.051	.051	.051	.051	.051	.051	.051	.051
.159	.173	.173	.162	.087	.150	.036	.036	.036	.036	.036	.036	.036	.036	.036	.036	.036
.002	.002	.006	.006	.000	.003	.003	.003	.003	.003	.003	.003	.003	.003	.003	.003	.003
.169	.159	.124	.180	.173	.159	.022	.022	.022	.022	.022	.022	.022	.022	.022	.022	.022
1.953	2.089	2.005	2.021	1.935	2.003	.061	.061	.061	.061	.061	.061	.061	.061	.061	.061	.061
.043	.028	.041	.017	.036	.033	.011	.011	.011	.011	.011	.011	.011	.011	.011	.011	.011
2.132	1.958	2.103	1.997	2.287	2.096	.129	.129	.129	.129	.129	.129	.129	.129	.129	.129	.129
4.999	5.000	5.000	5.001	5.002	5.000	.001	.001	.001	.001	.001	.001	.001	.001	.001	.001	.001
1.944	1.941	1.928	1.909	1.962	1.937	.020	.020	.020	.020	.020	.020	.020	.020	.020	.020	.020
.056	.059	.072	.091	.038	.063	.020	.020	.020	.020	.020	.020	.020	.020	.020	.020	.020
2.000	2.000	2.000	2.000	2.000	2.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000	.000
.447	.446	.469	.424	.315	.424	.064	.064	.064	.064	.064	.064	.064	.064	.064	.064	.064
.326	.401	.365	.423	.163	.335	.103	.103	.103	.103	.103	.103	.103	.103	.103	.103	.103
.773	.867	.834	.847	.478	.759	.161	.161	.161	.161	.161	.161	.161	.161	.161	.161	.161
.124	.093	.146	.116	.061	.108	.032	.032	.032	.032	.032	.032	.032	.032	.032	.032	.032
.024	.038	.015	.034	.033	.029	.009	.009	.009	.009	.009	.009	.009	.009	.009	.009	.009
.148	.131	.161	.150	.094	.137	.026	.026	.026	.026	.026	.026	.026	.026	.026	.026	.026

J89-662H

J90-1092

J89-417H

J89-642H

tschermakitic bornblende

magnesian-bornblende

tschermakitic bornblende

ferro-tschermakite

	1	2	3	4	5	avg.	SD	1	2	3	4	5	avg.	SD	1	2	3	4	5	avg.	SD
SiO ₂	44.05	45.01	44.98	43.76	42.75	44.11	.94	48.18	47.79	44.18	44.66	44.32	45.83	1.98	43.99	42.72	42.07	42.72	44.16	43.13	.90
Al ₂ O ₃	12.05	11.38	11.28	12.32	13.74	12.15	.99	7.81	8.45	10.71	10.82	10.90	9.74	1.49	14.53	15.71	16.41	15.91	13.98	15.31	1.01
TiO ₂	.39	.46	.52	.54	.49	.06	.01	.37	.42	.42	.57	.76	.51	.16	.55	.76	.71	.66	.44	.62	.13
Cr ₂ O ₃	.02	.00	.03	.00	.00	.01	.01	.05	.00	.02	.02	.02	.02	.02	.00	.03	.07	.08	.12	.06	.05
Fe ₂ O ₃	4.47	5.69	5.74	5.00	4.13	5.02	.72	3.26	2.33	3.65	4.19	2.54	3.22	.77	5.62	5.57	5.06	5.10	6.28	5.51	.49
FeO	12.66	11.30	11.00	12.11	12.89	11.98	.83	12.50	12.84	13.90	12.90	14.01	13.21	.68	11.56	12.41	12.54	12.21	10.77	11.92	.74
MnO	.39	.30	.34	.26	.33	.32	.05	.38	.24	.26	.34	.42	.33	.08	.19	.31	.26	.28	.28	.27	.05
MgO	10.21	11.00	11.00	10.22	9.37	10.36	.68	12.13	12.05	10.33	10.73	10.37	11.12	.90	9.03	8.06	8.04	8.23	9.42	8.55	.63
CaO	12.28	12.10	12.08	12.06	12.04	12.11	.10	12.26	12.20	12.20	12.04	12.05	12.15	.10	11.33	11.35	11.57	11.38	11.36	11.40	.10
Na ₂ O	1.05	1.02	.96	1.09	1.20	1.06	.09	.64	.77	1.08	1.11	1.09	.94	.22	1.07	1.10	1.13	1.13	1.03	1.09	.04
K ₂ O	.71	.67	.49	.66	.65	.64	.08	.64	.74	1.11	1.14	1.29	.98	.28	.38	.55	.58	.54	.33	.48	.11
F	.00	.04	.00	.00	.00	.01	.02	.17	.16	.08	.15	.00	.11	.07	.07	.07	.05	.00	.15	.07	.05
Cl	.03	.03	.00	.02	.03	.02	.01	.02	.01	.00	.02	.00	.01	.01	.00	.03	.04	.00	.04	.02	.02
total	98.31	99.00	98.42	98.02	97.67	98.28	.49	98.41	98.00	97.94	98.69	97.77	98.17	.38	98.32	98.67	98.53	98.30	98.36	98.43	.16

Formula based on 23 oxygens

	Si	Al ^{IV}	Al ^{VI}	Ti	Cr	Fe ⁺³	Fe ⁺²	Mn	Mg	Ca	Na ^{VI}	Na	K	F	Cl
Si	6.502	6.563	6.578	6.578	6.578	6.578	6.578	6.578	6.578	6.578	6.578	6.578	6.578	6.578	6.578
Al ^{IV}	1.498	1.437	1.422	1.422	1.422	1.422	1.422	1.422	1.422	1.422	1.422	1.422	1.422	1.422	1.422
Al ^{VI}	.599	.518	.523	.523	.523	.523	.523	.523	.523	.523	.523	.523	.523	.523	.523
Ti	.043	.050	.057	.058	.060	.060	.060	.060	.060	.060	.060	.060	.060	.060	.060
Cr	.497	.625	.632	.632	.632	.632	.632	.632	.632	.632	.632	.632	.632	.632	.632
Fe ⁺³	1.563	1.379	1.345	1.496	1.602	1.475	1.475	1.475	1.475	1.475	1.475	1.475	1.475	1.475	1.475
Fe ⁺²	.049	.037	.042	.033	.041	0.40	.006	.006	.006	.006	.006	.006	.006	.006	.006
Mn	2.246	2.391	2.399	2.250	2.076	2.273	1.32	1.32	1.32	1.32	1.32	1.32	1.32	1.32	1.32
Mg	4.999	5.000	5.001	5.002	5.000	4.999	.001	.001	.001	.001	.001	.001	.001	.001	.001
Ca	1.942	1.891	1.893	1.909	1.918	1.910	.021	.021	.021	.021	.021	.021	.021	.021	.021
Na ^{VI}	.058	.109	.107	.091	.082	.090	.021	.021	.021	.021	.021	.021	.021	.021	.021
Na	2.42	1.78	1.65	2.20	.263	.213	.042	.042	.042	.042	.042	.042	.042	.042	.042
K	.376	.302	.257	.345	.387	.333	.054	.054	.054	.054	.054	.054	.054	.054	.054
F	.000	.016	.000	.000	.000	.003	.007	.007	.007	.007	.007	.007	.007	.007	.007
Cl	.007	.007	.000	.005	.008	.005	.003	.003	.003	.003	.003	.003	.003	.003	.003

All amphiboles were analyzed, courtesy of the Geological Survey of Canada in Ottawa, with a Cameca SX-50 electron microprobe using both natural and synthetic standards (15 kV acceleration voltage; 10 nA current and 20 s counting time for Na, K, Fe and Mg; 30 nA current and 10 s counting time for all other elements; ZAF correction procedures). Formulas were calculated using the computer algorithm *Mineral Report* (version 1991-09-03) written by G. Pring. Relative amounts of ferrous and ferric iron were calculated using charge-balance assumptions. Amphibole nomenclature assigned to each sample (Leake 1978) is based on formula from average of five spot analyses.

APPENDIX D

Ca/K ratios in amphiboles determined from Ar isotopic and electron microprobe analyses

Sample no.	J89-18	J88-514H	J89-36	J88-351A	J89-905	J89-3H	J89-788H	J89-970	J89-662H	J90-1092	J89-417H	J89-642H
Heat steps ^a	3.79	6.95	2.10	2.55	2.92	4.18	2.79	0.60	2.77	1.89	6.89	3.98
	13.14	7.75	6.68	14.76	9.25	11.75	6.38	5.64 *	16.77 *	11.12	19.34	15.40
	9.96	9.31	7.28 *	15.10	6.06 *	11.19	6.03	6.31 *	15.21 *	9.62 *	20.43	19.81
	8.97 *	9.26	6.95 *	12.61 *	5.89 *	10.68	5.99	6.76 *	14.58 *	8.58 *	21.67	17.42
	8.86 *	1.27	6.71 *	13.57 *	5.86 *	10.05	5.94	6.91 *	14.70 *	8.37 *	19.58	19.23
	8.47 *	9.30 *	6.60 *	12.39 *	6.07 *	9.76	5.88	6.81	14.73 *	8.52 *	23.60	20.65
	8.59 *	9.85 *	6.71 *	11.34 *	6.54 *	10.18	5.95	6.93	14.49 *	8.77	23.10	21.39
	8.44	9.40 *	6.64 *	11.30 *	6.15 *	9.96	5.92	6.99	14.73	8.46	22.25	19.35
	8.38	---	6.48	10.80 *	5.90	9.97	5.79	6.36	14.00	8.44	21.03	18.61
	8.36	9.13	5.15	10.75	1.05	---	6.21	5.08	4.49	8.69	20.45	17.47
Plateau ^b	8.72 ± 3%	9.52 ± 3%	6.82 ± 7%	12.00 ± 13%	6.10 ± 7%			6.40 ± 12%	15.08 ± 11%	8.77 ± 10%		
Microprobe ^c	8.09	11.34	6.30	8.83	5.96	154	5.71	6.65	14.49	16.13	24.96	28.23
	7.94	8.10	6.40	8.14	4.84	25.5	5.88	6.92	15.25	13.88	17.30	19.86
	7.66	8.16	6.09	8.60	5.28	87.5	6.01	6.77	20.58	9.24	16.71	18.09
	7.52	8.00	6.29	8.69	4.51	164	5.56	7.07	15.27	8.86	17.89	22.26
Avg.	9.60	9.16	6.10	17.16	12.04	79.5	5.60	6.67	15.47	7.86	28.64	24.87
	8.08	8.82	6.23	9.54	5.78	53.7	5.74	6.81	15.92	10.40	20.09	22.14

^a Ca/K determined from measured ³⁷Ar/³⁹Ar isotopic ratios; * denotes plateau step.

^b mean Ca/K averaged over plateau interval ± maximum deviation expressed as percent of mean.

^c Ca/K based on calculated formulas for each of five electron microprobe spot analyses and on average formula (Appendix C).

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LITHOLOGICAL UNITS	
Tertiary:	
TI	altered, nondeformed diabase stocks
Cretaceous(?) to early Tertiary:	
TTG	leucocratic granite stocks
KTg	medium-grained, massive to mylonitic, leucocratic granites and pegmatite of Pukeashun suite, locally containing abundant screens of country rock; KTgb and KTgm are biotite and (garnet-biotite-muscovite monzonites, respectively
Early Cretaceous:	
Kg	biotite monzonite of Baldy batholith
Devonian:	
Dg	medium to coarse-grained (hornblende-biotite granodioritic, monzonitic and quartz dioritic) gneiss of Mount Foster suite; locally includes (chlorite-biotite-muscovite) granitic gneiss and biotite-chlorite-muscovite-quartz-feldspar schist interpreted as being mylonitic granite; Dgb is epidote-hornblende-biotite granodioritic to quartz dioritic gneiss
Paleozoic or Triassic:	
Sc	Sicamous Formation: grey, fine-grained, phyllitic marble and calcareous phyllite
Paleozoic:	
Ta	Taliskum Formation: epidote-chlorite-amphibole schist, massive metabasalt, chloritic phyllite
Proterozoic(?) - Paleozoic:	
Silver Creek assemblage:	
SCq	pelitic to semipelitic (garnet-biotite-quartz-muscovite schist, micaceous quartzite, minor calcareate schist and mafic schist; intruded by locally abundant apitic and pegmatite
SCa	biotite-amphibole schist, semipelitic schist, quartzofeldspathic paragneiss; SCa? includes biotite-hornblende amphibole and rocks like those of unit SCq
SCc	calcareous and quartzofeldspathic paragneisses, greyish pink marble, biotite-amphibole schist, semipelitic schist; SCc? includes marble (m), (diopside-tremolite calcareate gneiss, semipelitic schist and paragneiss
Eagle Bay assemblage:	
Ec	calcareate schist; marble; chloritic, sericitic, calcareous and graphitic phyllites; metabasite; mafic schist; minor quartzite; Ecm dominated by calcareate schist, grey fine-grained marble and phyllitic marble; Ecp dominated by phyllite and metacalcite; Eca dominated by epidote-biotite-amphibole-chlorite schist and quartz-sericite-chlorite phyllite; Ecg includes abundant micaceous quartzite
Em	white to medium grey marble
Eq	light brown, greenish grey, and white quartzite and micaceous quartzite; (staurolite-garnet-biotite-quartz-muscovite schist; minor feldspathic granule conglomerate; locally abundant epidote-biotite-amphibole-chlorite schist and quartz-sericite-chlorite phyllite

Proterozoic and/or Paleozoic:	
Lichen assemblage:	
Lu	undifferentiated pelitic to semipelitic schist, psammitic paragneiss, calcareate gneiss, biotite-hornblende quartz dioritic gneiss, rare amphibolite
La	semipelitic to pelitic (sillimanite-garnet-biotite-muscovite schist with thin interbeds of psammitic paragneiss and quartzite; intruded by locally abundant pegmatite and leucogranite
Lc	light bluish grey to light brown calcareate gneiss and impure marble
Munich assemblage:	
M	pelitic to semipelitic muscovite-sillimanite-biotite schist, psammitic paragneiss, hornblende-biotite quartz dioritic gneiss, diopside-bearing amphibolite, extensively intruded by pegmatite and leucogranite
Hunters Range assemblage:	
Hu	undifferentiated quartzofeldspathic gneiss, psammitic paragneiss, semipelitic and pelitic schist, amphibolite, calcareate gneiss, marble, rare quartzite; extensively intruded by sheets and dikes of pegmatite; Huc, Huc and Huh are dominated by rocks like those of units Hc, Hs and Hdb, respectively
Hh, Hdb	Griffin gneisses: Hh is grey quartzofeldspathic garnet-biotite-hornblende gneiss, at least in part of granuloid protolith, with minor leucocratic (garnet-biotite) granitic gneiss; Hdb is mostly grey dioritic to granodioritic (garnet-biotite-hornblende) gneiss
Hs	semipelitic to pelitic garnet-biotite and garnet-sillimanite-biotite (muscovite) schists, psammitic paragneiss, rare quartzite, local quartzofeldspathic garnet-biotite-hornblende gneiss and amphibolite
Ha	diopside amphibolite, semipelitic, psammitic paragneiss, quartzofeldspathic garnet-biotite-hornblende gneiss, minor pelitic schist and marble
Hc	diopside calcareate gneiss, marble, diopside amphibolite, rusty weathering semipelitic, minor psammite and pelite
Hm	light brown to light grey diopside marble and calcite marble, diopside calcareate gneiss, minor amphibolite and semipelite
Proterozoic:	
Three Valley assemblage:	
TVsa	semipelitic to psammitic paragneisses with boudins of clinopyroxene- and garnet-bearing amphibolites; pelitic garnet-sillimanite-biotite schist; minor diopside-bearing psammite; locally abundant pegmatite
Proterozoic (to Paleozoic?):	
Munich complex:	
MC	garnet-kyanite-sillimanite-biotite schist, quartzite, calcareate gneiss, marble

Plate 1:
Geological map of the western margin of the Shuswap complex, Shuswap Lake area, British Columbia (parts of 82L, 82M)
by B.J. Johnson

GEOLOGICAL SYMBOLS	
	geological contact (defined, approximate; inferred)
	normal fault (barbs on hanging wall)
	thrust fault (teeth on hanging wall)
Fabrics and folds (structures of a given phase are not correlative between all assemblages):	
	strike and dip of S ₁ , unless shown by another symbol, S ₁ (in high-grade rocks, e.g. Hunters Range assemblage, this symbol denotes the main composite gneissic layering/foliation)
	strike and dip of S ₂ , where graded beds indicate right-way-up
	strike and dip of S ₃ , where attitude differs from S ₂ ; trend and plunge of L ₁ intersection location (S ₂ ∩ S ₃) or F ₁ fold axis, vergence unknown
	trend and plunge of F ₁ fold axis; half arrowhead indicates F ₁ fold vergence
	strike and dip of S ₄ crenulation cleavage; trend and plunge of L ₂ intersection location (S ₃ ∩ S ₄) or F ₂ crenulation axis
	trend and plunge of fold axis, phase uncertain (F ₁ or F ₂); half arrowhead indicates fold vergence
	trend and plunge of mineral stretching lineation
	trend and plunge of stretching lineation in mylonitic rocks, upper-plate-to-the-west shear sense indicated
	trend and plunge of stretching lineation in mylonitic rocks, upper-plate-to-the-east shear sense indicated
	axial surface trace of upright or inclined F ₁ (antiform; synform); arrow indicates direction of plunge

Plate 2a: cross section A-A', SHUSWAP LAKE to THREE VALLEY GAP

94d
2pt
1404
1402
1402

Plate 2b: cross section B-B', CENTRAL HUNTERS RANGE

Plate 2c: cross section C-C', MARA LAKE

Plate 2d: cross section D-D', LARCH HILLS to CRAIGELLACHIE CREEK

E
WNW

Plate 2e: cross section E-E', ADAMS LAKE to SHUSWAP RANGE

Adams L.
A-N-T

Kwikoit Crk.

Lichen Mtn.

Scotch Crk.

Mt. Fowler

Seymour Arm
(Shuswap L.)

Anstey Arm
(Shuswap L.)

North Queest Mtn.

Craigellachie Crk.

O-ERF

E'
ESE

Plate 2f: cross section F-F', ADAMS LAKE to SEYMOUR ARM

F
WNW

A-NT

Mu

KTgm

Lu

KTgm

Pukeashun Mtn.

Lu

KTgm

Dg

Dg

Lu

KTgm

Lu

KTgb

F'
ESE

THIS
PH.D.
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Plate 2g: cross section G-G',
ADAMS PLATEAU to PUKEASHUN MTN.

Plate 3: crustal cross section and Paleocene restoration of the OMINECA BELT at latitude 51°N

b LATE PALEOCENE (~58 Ma)