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Structure, Metamorphism, and Geochronology Along the Southern Margin of the Breakenridge Orthogneiss, Coast Range, Southern British Columbia

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STRUCTURE, METAMORPHISM, AND GEOCHRONOLOGY ALONG THE SOUTHERN MARGIN OF THE BREAKENRIDGE ORTHOGNEISS, COAST RANGE, SOUTHERN BRITISH COLUMBIA

BY
JOHN A. FELTMAN

Accepted in Partial Completion of the Requirements for the Degree Master of Science

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A Thesis Presented to the Faculty of Western Washington University

In Partial Fulfillment of the Requirements for the Degree Master of Science

by

John A. Feltman
February, 1997
ABSTRACT

The Breakenridge orthogneiss is located at the southern end of the Coast Plutonic Complex, in the southwest Canadian Cordillera. It consists of sheeted orthogneiss sills and metamorphosed country rock folded into a tight, upright antiform. The deformational and metamorphic history along the southern margin of this structure is the focus of this study.

The orthogneiss is in original intrusive contact with enveloping metavolcanic rocks of the Jura-Cretaceous Slollicum Schist. A new U-Pb zircon age of 103.8 ± 0.5 Ma, together with a published age of 96 Ma (Parrish and Monger, 1992), establishes an episode of igneous intrusion and crystallization between 104-96 Ma.

Metamorphic grade ranges from greenschist facies (chlorite zone) on the western side of the area, to amphibolite facies (kyanite zone) in the core of the Breakenridge antiform. Metamorphic pressures range from <5 kb on the western edge of the study area to >8 kb in the gneissic core of the antiform over a horizontal distance of 1.25 km, requiring post-metamorphic truncation of the metamorphic gradient. Isobars are roughly concentric about the antiform, suggesting that the structure exposes a window of high-grade metamorphic rocks. New ⁴⁰Ar/³⁹Ar dates on hornblende and micas indicate that cooling began by 87 Ma, providing a minimum age bracket for high-grade metamorphism.

Two deformation events, D₁ and D₂, are recorded in the study area. D₁ structures consist of penetrative foliation and lineation. Foliation strikes NW and dips steeply to the NE. An abrupt change between down-dip and strike-parallel lineations suggests opposing kinematic regimes of arc-normal shortening and arc-parallel strike-slip deformation. D₂ structures record folding of D₁ fabrics in the Breakenridge antiform. The main intrusive contact of the orthogneiss, marker units of pelitic schist, and D₁ fabrics are folded into a SE-plunging, reclined fold at the south end of the antiform. The doubly-plunging
geometry of the antiform as a whole, accompanied by rare, small scale D₂ folds and down-
dip lineations, suggests post-D₁ shortening and differential upward movement of the high
grade metamorphic core as a regional-scale incipient sheath fold.

A post-metamorphic fault, the Breakenridge fault, is inferred on the western side of
the study area, based on the abrupt change in pressure, truncation of the oligoclase isograd,
different lineation orientations, and the trend of lithologic contacts. The fault, which is best
described as a distributed shear zone, may have been active during or after D₂. The fault
may have formed with the Breakenridge antiform as part of a fault-fold structure.

The narrow temporal window in which down-dip and strike-parallel fabrics appear
to have formed suggests that strain may have been partitioned in a transpressional tectonic
setting. An orogen-scale flower structure model and a detachment partitioning model
are proposed to explain the possible temporal and spatial relationship of orogen-normal
thrusting and orogen-parallel strike-slip kinematics.
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Isotope analysis was conducted by Andy Calvert and Bill McClelland at the University of California, Santa Barbara.

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I. INTRODUCTION

Regional Geology

The study area is located at the western edge of the North American Cordillera in southwest British Columbia, Canada. It lies at the southern end of the Coast Plutonic Complex, a suite of Middle Jurassic to Early Tertiary plutons and associated metamorphic rocks. This large magmatic and metamorphic belt comprises the core of a major orogen extending from southeast Alaska to Washington state along the boundary between the Insular and Intermontane superterranes (Figure 1.1). The Harrison Lake area (Figure 1.2) contains an enclave of wall rocks within the network of plutons, in which a complex history of deformation, metamorphism, and magmatism is recorded. The structural and metamorphic evolution of the Breakenridge orthogneiss, located on the eastern shore of Harrison Lake, is the focus of this study. The study area presents several critical problems characteristic of orogenic belts, including the complex interaction between contractional and strike-slip deformation, high pressure metamorphism, and magmatism.

In southwestern British Columbia, the orogenic belt is composed of three major tectonic blocks, which are from west to east, Wrangellia, the Coast Plutonic Complex, and the Tyaughton-Bridge River terranes (Figure 1.1). In northwestern Washington state, a west-to-east arrangement consists of the Northwest Cascades System, the Crystalline Core, and the Methow-Hozameen terrane. These two regions formed a continuous orogenic belt prior to Tertiary dextral slip on the Fraser River-Straight Creek Fault system (Misch, 1977; Vance and Miller, 1981; Tabor et al., 1989). The Coast Plutonic Complex - Crystalline Core comprises an uplifted, high-grade metamorphic and plutonic core of the orogen. The Tyaughton-Bridge River terrane and the Methow-Hozameen terrane form a belt of syn-orogenic marine and non-marine basin sediments on the eastern flank of the orogen. The
Figure 1.1 Regional geologic setting of the field area and index to detailed map of Figure 1.2. FRF, Fraser River fault; IM, Intermontane superterrane; MT, Methow-Tyaughton terrane; RLF, Ross Lake Fault; SCF, Straight Creek fault; WR, Wrangellia terrane. Modified from Brown and Talbot (1989).
Northwest Cascades System of Washington State, which contains blueschists and
greenschists of oceanic affinity, has been thrust over Wrangellia. The source of thrust
sheets in this tectonic block is not well understood.

The Coast Plutonic Complex - Crystalline Core block consists of metavolcanic and
metasedimentary schists and gneisses intruded by pre- to syn-metamorphic orthogneisses
and post metamorphic plutons. Metamorphic grade ranges from sub-greenschist facies on
the flanks of the orogen to amphibolite facies metamorphism and migmatite formation in the
central core. The high-grade metamorphic core exhibits steep pressure gradients, in which
central areas were metamorphosed under a load of 8-9 kb, compared to 3-4 kb on the
flanks (Brown and Walker, 1993). Isotopic dating of plutons, meta-volcanic protoliths,
and metamorphic index minerals indicate that metamorphism and magmatism were
contemporaneous between 100-80 Ma (Mattinson, 1972).

Many studies cite evidence for Late-Cretaceous orogen-normal shortening.
Sedimentary rocks of the Tyaughton-Methow terranes are deformed in a NE-vergent fold
and thrust belt (McGroder, 1989). The Northwest Cascades System consists of complexly
imbricated thrusts that are interpreted to be southwest-vergent based on regional structural
relationships (Brandon et al., 1988). However, stretching lineations in the Northwest
Cascades System are oriented NW-SE, suggesting orogen-parallel kinematics (Brown,
1987). The presence of contractional belts flanking the high grade core of the orogen has
led to collision-related models, whereby accretion of the Insular terrane to North America
resulted in crustal thickening and subsequent core metamorphism (Misch, 1966; Monger et
The widespread magmatism present in the orogen is not easily explained in this model,
however. Magmatism has been suggested to be the result of anatexis of deep-seated rocks
(Monger et al., 1982; Zen, 1988).
Further studies suggest a post-collision intra-arc shortening model. Journeay and Friedman (1993) cite evidence for the west-vergent Coast Belt Thrust System east of Wrangellia in wall rock septa of the southeast Coast Plutonic Complex. Intra-arc, west-vergent thrust systems were also documented in northwest British Columbia and southeast Alaska (Rubin et al., 1990; McClelland et al., 1990). In this model, contemporaneous arc-related magmatism and west-vergent thrust systems developed across the boundary of the previously accreted Insular and Intermontane superterranes.

Not in agreement with the thrust model, syn-metamorphic fabric patterns in both the paragneisses and orthogneisses of the high-grade core indicate orogen-parallel extension. Foliations strike NW and dip steeply to the NE. Stretching and mineral lineations maintain a strike-parallel orientation throughout many of the high-pressure rocks (Brown and Talbot, 1989). These structures are interpreted to be the result of strike-slip deformation in a transpressional tectonic setting (Brown and Talbot, 1989). In this model, Andean arc-type magmatism developed across previously accreted terranes (van der Heyden, 1992). Transpression allows for the interaction of contractional and strike-slip structures analogous to those documented in Sumatra (Fitch, 1972). An adaptation of the transpressional magmatic arc model invokes an orogen-scale positive flower structure, in which transpressional shear is accommodated in ductile rocks at depth, with the shortening component accommodated by thrusting at shallower structural levels (J.L. Talbot, pers. comm. 1996). The cause of crustal loading and subsequent metamorphism is not easily explained by a transpressional Andean arc type model, however. Metamorphism may have occurred as the arc-related batholith ballooned, loading the rocks below it (Brown and Walker, 1993).
Geology of Harrison Lake

The study area is located along the east shore of Harrison Lake, on the western flank of the high grade orogen. The geology along the east side of the lake consists of three major country rock packages and the Jurassic-Tertiary plutons that intrude them (Figure 1.2). Country rock units are, from west to east, the Slollicum Schist, the Cogburn Creek Group, and the Settler Schist. These metasedimentary and metavolcanic microterranes are juxtaposed along steep, folded thrusts in an imbricate stack, where the Slollicum Schist makes up the basal unit (Lowes, 1972; Monger, 1986).


The Cogburn Creek Group consists of a melange of metacherts, meta-argillites, amphibolite, and marble, and is imbricated with ultramafic lenses.

The Settler Schist consists of high-grade pelitic schists and paragneisses, amphibolite gneisses, quartzite, and marble. This unit is correlative with the Chiwaukum Schist of the Cascade Crystalline Core in Washington state (Misch, 1966; Duggan and Brown, 1994). In an alternative model, the Settler Schist is correlated with blueschist rocks of the Northwest Cascades System (Monger, 1991).

The Spuzzum pluton, Breakenridge orthogneiss, Scuzzy batholith, and Urquhart pluton constitute the major intrusive complexes in the area. Tonalites and diorites of the 96 Ma Spuzzum pluton cross-cut the imbricate thrusts of the country rock (U-Pb, Brown and Walker, 1993). Fabrics within the Spuzzum pluton are dominantly magmatic. The 96 Ma (U-Pb, Parrish and Monger, 1992) Breakenridge orthogneiss exhibits a well-developed.
Figure 1.2 Regional geology of east side of Harrison Lake. HLSZ, Harrison Lake Shear Zone; Msl, Slolicum Schist; Mss, Settler Schist; Pmc, Cogburn Unit. Urquhart pluton is former SW Scuzzy pluton of Brown and Walker (1993). Modified from Monger (1989) and Journey and Friedman (1993).
solid-state metamorphic fabric. Its intrusive and structural relationships with the surrounding Slollicum Schist are the focus of this study. The 90-84 Ma Scuzzy batholith (U-Pb, Parrish and Monger, 1992; Brown and Walker, 1993) and the 91 Ma Urquhart pluton (U-Pb, Brown and Walker, 1993) consist of several diorite, quartz-diorite, and tonalite plutonic phases that intrude the Settler Schist. Fabrics in these plutons are dominantly magmatic. Two large-scale orthogneiss sills, one dated at 91 Ma and the other at 157 Ma (U-Pb, E.H. Brown, unpublished data), are sandwiched between the Scuzzy batholith and the Breakenridge orthogneiss.

Metamorphic grade increases from sub-greenschist facies along the shore of Harrison Lake to amphibolite facies adjacent to the Spuzzum and Scuzzy plutons (Lowes, 1974; Hettinga, 1989). An anomalous 9 kb high is centered on the Breakenridge orthogneiss, disturbing the regional baric pattern (E.H. Brown, pers. comm. 1996). At least three metamorphic events are recorded in the area. The oldest event is recorded in greenschist facies rocks of the Slollicum Schist. Metamorphic fabrics with down-dip stretching lineations parallel thrusts of the imbricated country rock units, and have therefore been associated with a regional thrust event (Bennett, 1989). The second episode is recorded in the contact aureole of the Spuzzum pluton, where fabrics parallel the pluton boundary and overprint thrust-related fabrics. The third event is associated with the aureole of the Scuzzy batholith, which overprints the aureole fabrics of the Spuzzum pluton (E.H. Brown, pers. comm. 1996).

A distinct fabric orientation occurs in the Breakenridge orthogneiss, where stretching and mineral lineations exhibit a shallow-plunge and strike-parallel orientation in a sub-vertical, NW-SE striking foliation. Strike-parallel lineations are also prevalent in the Harrison Lake Shear Zone, which separates high-grade metamorphic rocks on the east side of Harrison Lake from low-grade arc-volcanic rocks and plutons on the western side of the lake.
Geology of the Breakenridge Orthogneiss and Adjacent Rocks

The Breakenridge orthogneiss is an elongate metamorphosed plutonic body exposed over a distance of 20 km, from the lower reaches of Big Silver Creek to Stokke Creek, on the east side of Harrison Lake (Figure 1.3). It is enveloped by metavolcanic and pelitic rocks of the Slollicum Schist. Previous work in the northern part of the Breakenridge orthogneiss by Reamsbottom (1974) shows that the intrusive contact of the orthogneiss, primary lithologic layering in country rock, and overprinting metamorphic foliation are parallel, and are folded together in a broad, upright, north-plunging antiform. Structural and metamorphic relationships at the southern end of the Breakenridge orthogneiss are the focus of this study. Reconnaissance mapping of metamorphic index minerals and thermobarometry, conducted by workers at Western Washington University, has shown that isobars are roughly concentric about the Breakenridge orthogneiss (E.H. Brown, pers. comm. 1996). Metamorphic pressure increases from 4 kb along the shore of Harrison Lake to 9 kb in the core of the orthogneiss, thus defining an unusually steep metamorphic gradient on the western side of the area. Reconnaissance mapping also shows that NW-striking foliation is subvertical in the orthogneiss, and dips moderately to the NE in the Slollicum Schist. Stretching lineations in the orthogneiss and immediately adjacent wall rocks exhibit a strike-parallel orientation, whereas those in the surrounding Slollicum Schist maintain a down-dip orientation.

The high-grade metamorphic rocks on the east side of Harrison Lake are interpreted to represent part of a west-vergent thrust belt described as the Coast Belt Thrust System (Joumeay and Friedman, 1993). In this model, the Breakenridge orthogneiss lies in the core of an allochthonous, antiformal stack of folded igneous sheets and thrust nappes described as the Breakenridge antiform. The antiform is interpreted to be bounded on the
Figure 1.3 Compilation of previously mapped regional geology in the vicinity of the Breakenridge orthogneiss. BF, Breakenridge Fault; CCBD, Central Coast Belt Detachment (Journeay and Friedman, 1993); HLSZ, Harrison Lake Shear Zone; Mbh, Brokenback Hill Formation; Msl, Slollicum Schist; Mss, Settler Schist; Mti, Twin Island Schist; Pmc, Cogburn Creek Group. Adapted from Monger (1989), Journeay and Friedman (1993), and Brown and Walker (1993).
west and east by out-of-sequence, mylonitic reverse faults which truncate the older thrust nappes (Figure 1.4). The Breakenridge Fault, on the west, is cited as the major structure along which uplift of high-pressure rocks in the core of the antiform occurred, and is therefore responsible for the steep metamorphic gradient on the western side of the area. It is mapped as the western contact of the orthogneiss. The Central Coast Belt Detachment, on the east, is interpreted to thrust high grade metamorphic rocks over the Breakenridge antiform.

Objectives

The purpose of this study is to provide a detailed analysis of the structural geology, metamorphism, and uplift of the southern margin of the Breakenridge orthogneiss and the surrounding Slollicum Schist. Several key problems are addressed in this study.

(1) The structural relationship between the Breakenridge orthogneiss and Slollicum Schist is a focus of this study, specifically whether the contact between the two is intrusive, or a fault.

(2) The possible existence of the Breakenridge fault is evaluated by mapping of lithostratigraphic units in country rocks, fabric patterns, and metamorphic zones.

(3) The relative timing and kinematic significance of strike-parallel and down-dip lineations is critical in piecing together the tectonic history of the area. Map patterns of deformational fabrics, and petrographic analysis of the relationship between metamorphic index minerals and deformational fabrics, are used to determine the spatial and temporal relationship of deformation and metamorphism.

(4) The cause of the steep metamorphic gradient on the western side of the area is investigated. Two models are tested, in which the steep gradient may be the result of either
Figure 1.4 Cross section model of Coast Belt Thrust System, showing location of Breakenridge orthogneiss (shaded areas). Adapted from Journeay and Friedman, 1993.
a discrete fault or distributed shear. Thermobarometry and mapping of metamorphic isograds are used to characterize the metamorphic gradient.

(5) Isotope geochronology using the U-Pb zircon and \(^{40}\text{Ar}/^{39}\text{Ar}\) methods provides further constraints on the timing of arc-volcanism, plutonism, metamorphism, and uplift in the vicinity of the Breakenridge orthogneiss.
II. LITHOLOGIC UNITS

Introduction

Major lithologic units exposed in the study area are the Slollicum Schist, a Jurassic orthogneiss sill, and the Breakenridge orthogneiss (Figure 2.1, Plate 1).

The Slollicum Schist is an arc-related sequence of metavolcanic flows, tuffs, volcaniclastic rocks, and interlayered metasedimentary units. Metavolcanic rock types consist of chlorite schist, plagioclase-hornblende schist, and plagioclase-biotite schist. Metasedimentary rocks include pelitic phyllites and schists, metaconglomerates, and metasandstones.

The Breakenridge orthogneiss is a metamorphosed tonalitic pluton intrusive into adjacent metavolcanic and pelitic rocks of the Slollicum Schist. It consists of a central elongate body flanked by sills which parallel its margin.

Specific site localities in the text are shown on Plate 2.

Slollicum Schist

The Slollicum Schist is an arc-related metavolcanic and metasedimentary assemblage exposed along much of the eastern shore of Harrison Lake (Figure 1.2). Previous work by Lowes (1972), Bennett (1989), and Hettinga (1989) divides the Slollicum Schist into a lower metasedimentary unit and an upper metavolcanic unit. Metavolcanic protoliths consist of felsic to intermediate flows and tuffs and associated volcaniclastic rocks. Metasedimentary protolith lithologies consist of pelites, siltstones, sandstones, and conglomerates. Rocks of the Slollicum unit exhibit metamorphic mineral
Figure 2.1 Detailed map of lithologic units in the study area. Section line A-A' corresponds to the cross-section in Figure 2.11.
assemblages ranging from greenschist (chlorite zone) to amphibolite (sillimanite zone) metamorphic facies.

U-Pb zircon ages of 145 Ma (Bennett, 1989) and 102 Ma (Parrish and Monger, 1992) on volcanic layers give the Slollicum a Late Jurassic to Middle Cretaceous age. These ages are consistent with Buchia fossils of Late Berriasian to Albian age in the Peninsula and Brokenback Hill Formations of the Fire Lake Group (Jeletzky, 1965; Arthur, 1986; Lynch, 1990), suggesting a probable correlation across the Harrison Lake Shear Zone with these largely unmetamorphosed units on the western side of Harrison Lake.

The Slollicum Schist as exposed in the study area consists primarily of rhyolitic to andesitic flows and tuffs, and probably correlates with the upper metavolcanic sequence documented by Bennett (1989) further to the south. Metasedimentary units consist of pelitic and conglomerate interlayers within the volcanic stratigraphy.

Metavolcanic Protoliths

The volcanic protoliths within the Slollicum Schist can be separated into three volcano-stratigraphic members based on their bulk rock composition. Multiple SW-NE traverses across the western slope of the study area define a contact between chlorite schists and plagioclase-hornblende schists (Figure 2.1), marked by the abrupt appearance of metamorphic hornblende. This lithologic boundary is proposed to be a fault, as discussed in the Structural Geology chapter. Probable dacite protoliths have chlorite as the predominant mafic mineral, whereas andesitic protoliths contain hornblende. The change between dacitic and andesitic protolith bulk rock composition is inferred based on the absence of actinolite in the dacitic unit, as described in the Metamorphism chapter.
The third unit, a plagioclase-biotite schist, is located in the septum between orthogneiss sills and the west side of the main pluton. This unit probably represents a rhyolitic volcanic sequence. Subtle compositional banding within the schist shows hornblende-rich interlayers on the outcrop scale. These observations serve to illustrate the dependence of hornblende growth on bulk rock composition.

Based on the preceding criteria, chlorite schist, plagioclase-hornblende schist, and plagioclase-biotite schist are mapped and discussed as separate volcano-stratigraphic units (Figure 2.1).

**Chlorite Schist**

The chlorite schist unit consists of massive green to gray schist exposed as blocky outcrops on the western side of the field area. Limited exposure makes stratigraphy within the unit difficult to map. Well-preserved phenocrysts of quartz and feldspar suggest that the chlorite schists are metadacites. Relict layering is poorly preserved in the unit, except where thin pelitic or felsic tuffaceous interlayers are observed. Foliation development in these rocks ranges from good to non-existent, and is probably controlled by lithology. Samples exhibiting good foliation are interpreted to be weak tuffaceous or epiclastic layers. Massive outcrops of poorly-foliated rock are probably resistant lava flows, some of which appear to be very weakly strained (Figure 2.2). Foliation is defined by aligned chlorite and muscovite grains, and flattened porphyroclasts of relict igneous quartz and feldspar (Figure 2.2). The common mineral assemblages are quartz, plagioclase, epidote/clinozoisite, chlorite, and muscovite, with the addition of minor biotite and garnet at the highest metamorphic grades. A U-Pb zircon age of $102 \pm 1 \text{ Ma}$ (Parrish and Monger, 1992) was determined from metadacites on the shore of Harrison Lake west of the Big Silver Creek delta (Figure 1.2).
Figure 2.2 Chlorite schist. A. Sample photo of meta-dacite from site 180-122. Note weakly-deformed relic igneous feldspar phenocrysts. B. Photomicrograph (in plane-polarized light) from site NB-72.
**Plagioclase-Hornblende Schist**

Plagioclase-hornblende schists are the predominant metavolcanic unit in the field area, and are well exposed on both sides of Big Silver Creek. Most outcrops are massive, homogeneous, gray-green schist. Relict tuffaceous layering is very well preserved in a few outcrops, giving the rock a banded appearance (Figure 2.3). Metamorphic hornblende is pervasive throughout the unit.

Foliation is defined by white mica and flattened aggregates of plagioclase. Foliation development in the unit varies depending on the matrix mineral assemblage. In western exposures of the unit, foliation tends to be more schistose than in eastern exposures, due to a higher concentration of micaceous minerals. Eastern exposures have a higher concentration of plagioclase and quartz, which have a polygonal to sub-polygonal texture.

Mineral assemblages consist of plagioclase, quartz, epidote, hornblende, and garnet. Garnet and hornblende commonly occur as large porphyroblasts that cut across the foliation (Figure 2.4). Garnets range from 1-5 mm in diameter, and hornblende prisms range from 1 mm to 5 cm in length. Hornblende prisms are randomly oriented and clearly cross-cut foliation. A few outcrops show some weak alignment of hornblende (Figure 2.3).

**Plagioclase-Biotite Schist**

The plagioclase-biotite schist unit is located in the northwest part of the field area in the septum between the northwest orthogneiss sill and the main orthogneiss body (Figure 2.1). The unit is exposed as outcrops of well foliated, light colored gneiss and schist. Relict features are poorly preserved in the unit. However, rare clastic layers exhibit stretched clasts of uniform felsic composition, indicating that they are probably volcanic
Figure 2.3 Relict tuffaceous layering ($S_0$) in plagioclase-hornblende schist from site 180-24. Note aligned hornblende defining a foliation ($S_1$) parallel to the long axis of photo, and perpendicular to $S_0$. Coin is 2.8 cm in diameter.
Figure 2.4 Plagioclase-hornblende schist. A. Sample photo from site 180-86, showing randomly-oriented post-kinematic hornblende and garnet cutting $S_1$ foliation. B. Photomicrograph (in cross-polarized light) from site 180-177.
agglomerates. Hornblende-rich layers are observed to interfinger with the biotite gneiss on the outcrop scale, suggesting the interlayering of tuffs of andesitic composition. Metaconglomerates and pelitic layers, normally present in a metasedimentary sequence, are absent from the unit. The protolith of the plagioclase-biotite schist unit is therefore interpreted to be a metamorphosed rhyolitic volcanic sequence.

Foliation in the plagioclase-biotite schist unit is defined by aligned biotite grains in a fine-grained matrix of equigranular quartz and feldspar. Biotite is evenly distributed throughout the rock. Isolated porphyroclasts of igneous feldspar are common, some of which exhibit subgrain development. Quartz and feldspar matrix grains exhibit polygonal texture. Elongate relict igneous grains of feldspar and quartz are larger than those in the matrix material. Evidence of post-kinematic recrystallization is the same as in the Breakenridge orthogneiss. Metamorphic minerals consist of plagioclase, quartz, and biotite, with minor garnet (Figure 2.5).

Metasedimentary Protoliths

Pelitic Units

Pelitic rock units occur as three distinct schist belts, and as thin beds within the volcanic units. Two major pelitic schist units, ranging from 200-450m in thickness, wrap around the nose of the Breakenridge antiform parallel to the intrusive contact of the orthogneiss. These units serve as stratigraphic markers within the metavolcanic plagioclase-hornblende schist unit, and help define the Breakenridge antiform. A third unit flanks the entire eastern side of the field area, and is bounded by a N-S linear contact with metavolcanic schists and the Jurassic orthogneiss (Figure 2.1).

Pelitic units are typically exposed as dark, highly weathered, rust-colored outcrops. Fresh exposures of pelitic schist reveal a dark, blue-gray, well-foliated rock, that is
Figure 2.5 Plagioclase-biotite schist. A. Sample photo from site JT-95-9a. B. Photomicrograph (in cross-polarized light) from site 180-83.
typically porphyroblastic at higher metamorphic grades (Figure 2.6). Most outcrops contain thin light-colored interlayers, some of which exhibit randomly oriented hornblende prisms. Foliation is defined by a graphitic parting, elongate quartz grains, and aligned biotite, muscovite, and chlorite. Metamorphic minerals are quartz, plagioclase, epidote/clinozoisite, biotite, muscovite, garnet, and chlorite. Kyanite is also present in samples from the eastern schist belt. Rare staurolite, margarite, and chloritoid were also observed.

Minor pelitic interlayers also occur within the chlorite schist unit on the western side of the field area. Because metamorphic grades are not as high within this unit, pelitic lithologies are predominantly phyllites. Foliation is defined by a graphitic parting, stretched quartz grains, and aligned muscovite and chlorite.

**Metaconglomerates**

Metamorphosed conglomerate layers occur interbedded in the chlorite schist, plagioclase-hornblende schist, and pelitic schist belts. Clasts range from pebble to cobble size and exhibit a range of different lithologies. Clast lithologies observed include pelites, quartzites, volcanic rocks, and granitic rocks (Figure 2.7a). Pelitic clasts may have been shale rip-ups. Quartzites contain microcrystalline quartz, suggesting chert as a protolith. Volcanic clasts are typically rhyolitic in composition, although some were observed to contain metamorphic hornblende. Granitic clasts exhibit good relict igneous texture with varying composition. Matrix material is pelitic or psammitic, and shows preferential growth of biotite and garnet at higher metamorphic grades.

Clasts are typically stretched and exhibit a distinct rheology contrast between lithologies. In a given sample, pelitic, quartzose, and volcanic clasts are highly deformed, whereas granitic clasts may exhibit little or no deformation (Figure 2.7a; see also Reamsbottom, 1974).
Figure 2.6 Pelitic schist. A. Sample photo from site 180-13. Layering is relict bedding. B. Photomicrograph (in cross-polarized light) from site 180-141. Note zoned clinozoisite prisms (blue birefringence).
Figure 2.7 A. Photo of metaconglomerate from boulder located approximately 200m east of site 180-75. Note weak deformation of granitic clasts in contrast to highly strained pelitic clasts. Coin is 2.8 cm in diameter. B. Outcrop photo of deformed meta-agglomerate in plagioclase-hornblende schist unit at site 180-25a.
Based on clast compositions, metaconglomerates are interpreted to be reworked volcanic arc deposits. The presence of granitic clasts suggests that erosion reached older intrusive levels of the volcanic arc.

Volcanic agglomerates are also common in metavolcanic units. They are distinguished from metaconglomerates by their uniform volcanic clast composition and volcanic matrix (Figure 2.7b).

**Jurassic Orthogneiss**

The southern end of a regionally extensive orthogneiss sill, approximately 750 m wide, occurs in the northeastern quarter of the study area (Figure 2.1). The sill parallels the margin of the Breakenridge orthogneiss on a NNW trend, ultimately wrapping around through the north-closing nose of the Breakenridge antiform (Figure 1.2). The sill is intrusive into metavolcanic schists on its western side. Its eastern contact is not well understood, however, and may be tectonically juxtaposed against pelitic schist (E.H. Brown pers. com., 1996). A U-Pb zircon age of 157 Ma was obtained for the sill (McClelland and Brown, unpublished data).

The sill consists of a metamorphic mineral assemblage of plagioclase, hornblende, quartz, and epidote, with varying amounts of biotite, muscovite, and chlorite. Plagioclase is distinctly twinned, and is commonly altered to sericite. Hornblende and plagioclase are by far the most abundant phases in the rock, suggesting an original dioritic composition. Foliation is defined by aligned hornblende and biotite, and elongate clusters of quartz (Figure 2.8).
Figure 2.8 Jurassic orthogneiss. A. Sample photo from site JT-88-39, showing relict igneous plagioclase. B. Photomicrograph (in cross-polarized light) from site 164-509.
Breakenridge Orthogneiss

The Breakenridge orthogneiss is an elongate metaplutonic body that extends NNW from the lower reaches of Big Silver Creek to the northern slopes of Mt. Breakenridge. It was described by Reamshottom (1974), who identified a less deformed core zone and a well-foliated outer rim. Parrish and Monger (1992) obtained a U-Pb zircon igneous-crystallization age of 96 Ma for the Breakenridge orthogneiss. It lies in the middle of a NW-SE trending belt of 96 Ma plutons, which include the Ascent Creek Pluton to the northwest, and the Spuzzum Pluton to the southeast.

Modal analysis, accomplished by point-counting of six samples taken from both the main body and satellite sills, shows an average tonalitic composition, as portrayed in the ternary plot of Figure 2.9. Biotite is the dominant mafic constituent, and typically comprises 20% of the rock. Exceptions to the average tonalitic composition occur in sample JT-93-08 and 180-187, which lie in the granite and granodiorite compositional fields, respectively. These samples suggest that the pluton is composed of multiple igneous phases of differing composition.

This study encompasses the southernmost part of the pluton, where all of the orthogneiss exhibits penetrative, solid-state fabric. Foliation in the Breakenridge orthogneiss is defined by spaced, flattened biotite clusters separating aggregates of flattened and attenuated porphyroclasts of magmatic quartz and feldspar. Porphyroclasts exhibit well-developed, polygonal subgrains surrounded by a fine-grained matrix of equigranular, polygonal quartz and feldspar. Biotite aggregates and elongate plagioclase porphyroclasts define a well-developed mineral lineation. An absence of undulose extinction in quartz and
Figure 2.9 Ternary classification diagram illustrating the composition of the Breakenridge orthogneiss. Plagioclase, alkali feldspar, and quartz are normalized to 100%. Six samples were analyzed, but only five show here due to nearly identical composition of two samples in the tonalite field.
Figure 2.10 Brokenridge orthogneiss. A. Sample photo from site 180-151, showing flattened biotite clusters. B. Photomicrograph (in cross-polarized light) from site JT-88-16c.
feldspar of the matrix, and straight grain boundaries in biotite, indicate post-kinematic static recrystallization (Figure 2.10).

Field mapping resulted in the identification of three discontinuous orthogneiss sills, each approximately 200-300 meters wide, which flank the main body of orthogneiss parallel to its contact with the Slollicum Schist on the east and west sides (Figure 2.1). The sill adjacent to the SW flank of the orthogneiss appears to have been connected directly to the main body prior to offset along a brittle fault, suggesting that the sills are part of the same igneous suite as the central pluton. The sills are folded in the Brekenridge antiform, and may have been continuous across the hinge of the antiform prior to erosional dissection. These observations suggest that the pluton was intruded as a horizontal sheeted-sill complex, in which the main body of orthogneiss as exposed today represents a thick stack of igneous sheets (Figure 2.11). Individual sills within the main body of orthogneiss, as portrayed on the diagram of Figure 2.11, are schematic. A small number of outcrop-scale, country rock lenses were identified in the main body of orthogneiss near the nose of the antiform, but were not large enough to map. These lenses exhibit mixing zone features, and are probably screens sandwiched between sheets of orthogneiss. Apparent thickening by folding of the sheets would explain the absence of major windows into underlying country rock in the core of the main body of orthogneiss.

Intrusive contacts are well exposed in road outcrops and stream channels at several localities. The most definitive outcrop occurs at the southern tip of the Brekenridge orthogneiss in the polished bedrock of Big Silver Creek (site 180-105, Plate 2). This outcrop exposes a complex mixing zone between orthogneiss and pelitic-volcanic schists. Xenoliths of country rock in the orthogneiss, and apophyses of orthogneiss in the wall rocks define a 20 m wide injection zone that clearly illustrates the nearly concordant intrusive emplacement of the Brekenridge orthogneiss. Mixing zone features are flattened and overprinted by solid-state fabric. Mixing zones were also observed in stream-polished
Figure 2.11 Cross section of field area along line A-A' of Figure 2.1, showing sheeted sills of Breakenridge orthogneiss, satellite sills, and Jurassic sill folded in the Breakenridge antiform. No vertical exaggeration. BF indicates proposed Breakenridge Fault. For key to units, see Figure 2.1.
bedrock at the sharp bend in Clear Creek and at two road cuts on the western contact of the orthogneiss. Photographs and descriptions of the structural relationships at these sites are presented in the Structural Geology chapter.
III. METAMORPHISM

Introduction

The rocks of the study area record a regional metamorphic gradient, varying from greenschist facies (chlorite zone) assemblages in the west to amphibolite facies assemblages in the east. The gradient may be interrupted by the Breakenridge fault, as discussed in the next chapter.

Isograds of metamorphic index minerals were mapped by field study and thin-section petrography. Mineral assemblages are referenced by sample number in Appendix A. Sample localities are shown on Plate 2.

New thermobarometric analysis was conducted on two samples from the plagioclase-hornblende schist unit. Analysis of anorthite content in plagioclase was conducted on samples from the chlorite schist unit in an effort to map the albite-oligoclase transition in that unit. These results complement pre-existing thermobarometry data collected during unpublished reconnaissance work by Dr. E.H. Brown and Dr. J.L. Talbot at Western Washington University.

Isograds

Regional metamorphic isograds were mapped for biotite, garnet, and hornblende. Isograd map patterns indicate an increase in metamorphic grade from west to east.

Sample traverses eastward from Harrison Lake define a NW-SE trending biotite isograd, sub-parallel to, and close to the shoreline of Harrison Lake (Figure 3.1). In the southern part of the area, the isograd is constrained by four samples in the chlorite schist unit that contain post-kinematic biotite. In the northern part of the area, the isograd swings
Figure 3.1 The distribution of biotite and the biotite isograd in the field area.
westward into Harrison Lake. Biotite in this area is syn-kinematic, and occurs mostly in layers of pelitic and psammitic schist.

The first occurrence of garnet defines an isograd that lies east of, and sub-parallel to, the biotite isograd (Figure 3.2). The southern segment of the isograd is constrained by sample 180-205 (Plate 2), which contains post-kinematic garnet. The middle and northern segments of the isograd are constrained by syn-kinematic garnets. In higher grade rocks of the eastern half of the field area, garnets are post-kinematic, with rare exceptions. Post-kinematic garnets are typically euhedral crystals that truncate foliation, and may exhibit compositional zoning.

Definition of the biotite and garnet isograds by both syn- and post-kinematic minerals is somewhat problematic, because minerals of different ages are used. However, post-kinematic metamorphic index minerals were not observed to overprint syn-kinematic forms of the same mineral, suggesting that the two generations may be part of a continuous metamorphic event. This interpretation would imply that in higher grade areas post-kinematic mineral growth occurred during prolonged heating after deformation had ceased.

The rare occurrence of biotite and garnet in the southern part of the chlorite schist unit is probably controlled by the bulk rock composition of the protolith. The occurrence of fresh, euhedral, post-kinematic biotite and garnet in a few compositionally favorable, outlying samples of chlorite schist requires that isograds be drawn further west than the majority of the data would suggest. Therefore, much of the chlorite schist unit experienced biotite-garnet grade metamorphic conditions.

Hornblende appears abruptly in metavolcanic lithologies along a NW-SE trending line (Figure 3.3), and is part of the evidence for the Breakenridge fault discussed in the Structural Geology chapter. In the southeast, the hornblende isograd swings eastward around the large eastern unit of pelitic schist. Several lines of evidence support the interpretation that the first occurrence of hornblende is controlled by bulk rock composition
Figure 3.2 The distribution of garnet and the garnet isograd in the field area.
Figure 3.3 The distribution of hornblende and the hornblende isograd in the field area.
instead of progressive metamorphism. (1) Chlorite schists west of this line are devoid of actinolite, thus indicating that the appearance of hornblende is due to a lithologic change, and does not represent a metamorphic change in amphibole composition within a single lithologic unit. (2) The first occurrence of hornblende lies well to the east of the albite-oligoclase transition line (discussed below), at a higher metamorphic grade than is typical for the greenschist-amphibolite transition at medium pressures (e.g. Miyashiro, 1975). This change in bulk rock composition and lithology could be stratigraphically or structurally controlled.

Hornblende occurs predominantly as post-kinematic prisms associated with post-kinematic garnets and static recrystallization textures in quartz and plagioclase, indicating high grade recrystallization after deformation (Figure 2.8).

Other metamorphic index minerals observed include chloritoid, margarite, and kyanite. These mineral occurrences are noted on Figure 3.1. Kyanite occurs at the extreme eastern edge of the field area, and is pervasive in pelitic rocks further to the east (Brown and Walker, 1993).

Albite-Oligoclase Transition

Traverses across the metamorphic gradient were conducted in an effort to map the albite-oligoclase transition within the chlorite schist unit. Eleven samples were chosen for microprobe analysis to compliment pre-existing data, and are listed in Table 1.

Results define a distinct jump in anorthite content of plagioclase from albite (An$_{0.2}$) to oligoclase (An$_{17.22}$). Composition values, when plotted on the map, define a boundary maintaining a NNW-SSE trend within the chlorite schist unit (Figure 3.4). Further south, the transition line bends SE and is covered in the Big Silver Creek delta.
Table 1. Thermobarometry results. Sample locations are shown on Plate 2. Site letters are keyed to Figures 3.4, 3.6, and 3.7. Samples in bold are original to this study. Microprobe data for samples 180-191 and 180-192 are listed in Appendix B.

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*GABI = garnet-biotite Fe-Mg exchange (Berman, 1991)
GAHB = garnet-hornblende Fe-Mg exchange (Graham and Powell, 1984)
GAHP = garnet-hornblende-plagioclase (Kohn and Spear, 1984)
GAMI = garnet-biotite-muscovite-plagioclase (Berman, 1991)
GASP = garnet-alumino-silicate-plagioclase-quartz (Berman, 1991)
Figure 3.4 The distribution of metamorphic temperatures, anorthite content in plagioclase coexisting with clinzoisite, and the albite-oligoclase isograd in the field area. Mineral compositions were determined from microprobe analysis. Site letters are keyed to Table 1 and Figure 3.7.
The abrupt change in composition, known as the peristerite gap, has been described in many metamorphic belts, and is interpreted to be a miscibility gap between albite and anorthite plagioclase components in progressive metamorphism (Crawford, 1966; Maruyama et al., 1982). The compositional variation of plagioclase with increasing temperature is highly dependent on bulk rock composition, as portrayed in the T-X_an and compositional phase diagrams of Figure 3.5. Greenschists containing amphibole can develop two coexisting plagioclases. Rocks with clinozoisite, as in the area of this study, show a jump in plagioclase composition.

The presence of the albite-oligoclase transition within the field area, combined with the position of the biotite and garnet isograds, suggests a metamorphic gradient that increases from west to east within the chlorite schist unit. The temperature and pressure conditions at which the transition occurs can be estimated as follows. Sample JT-89-19B (Plate 2; Figure 3.4, site H), located very close to the transition at its northern end, is the only sample with appropriate mineralogy for thermobarometry. Using garnet-mica (GAMI/GABI) thermobarometry, pressure-temperature conditions of 5.40 kb and 405°C were obtained with a plagioclase composition of An_{17} (Figures 3.4 and 3.6; Table 1). The temperature at which albite and oligoclase coexist has been shown to be pressure-dependent, based on a survey of studies in low- and high-pressure metamorphic belts by Maruyama and others (1982). The authors estimate that the transition temperature increases by 20°C/kb. The P-T gradient of this study area is similar to that of the Appalachian metamorphic belt. Albite of An_{2} was shown to coexist with anorthite of An_{22} in Appalachian rocks (Crawford, 1966). Various methods of thermobarometry suggested a temperature for the peristerite solvus in Vermont between 450-525°C (Crawford, 1966; Nord et al., 1978; Spear, 1980; Laird and Albee, 1981). Pressure was estimated at 3.5-5.7 kb (Laird and Albee, 1981). Considering the metamorphic conditions obtained for
Figure 3.5  A.  T-$X_{An}$ graph drawn along the albite-anorthite join of figure 3.5b above.  
B. Compositional phase diagram showing plagioclase stability with various phases.
"X" marks the likely composition of the chlorite schist unit of this study. Ab = albite, Act = actinolite, An = anorthite, Cz = clinzoisite, Hb = hornblende, Marg = margarite, Pg = paragonite (Modified from Maruyama et. al., 1982).
Figure 3.6 The distribution of metamorphic pressures in the field area. Site letters are keyed to Table 1 and Figure 3.7.
sample JT-89-19B, and pressure and temperature estimated for similar rocks in Vermont, a temperature of 400-500°C is suggested for the oligoclase isograd in this study area. This value fits well with the P-T curve determined by Brown (1978), based on $^{18}_{}$O isotope fractionation and thermodynamic calculations, in which the first occurrence of oligoclase at 5 kb corresponds to a temperature of 450°C. It also fits well with the P-T curve determined by Begin (1992), using thermobarometry, in which the first occurrence of oligoclase at 5 kb corresponds to a temperature of 475°C.

In the southern part of the area, eastward continuity of the albite-oligoclase transition across the Big Silver Creek delta is constrained by sample JT-88-7f (Figure 3.4, site A). Because this sample records An$_{27}$, the transition must continue south or southeast of this site. Therefore, the albite-oligoclase transition is not folded in the Breakenridge antiform, and either traverses southward into Harrison Lake or is truncated by a fault. The biotite and garnet isograds do not appear to be folded either. It is therefore probable that the metamorphic gradient in the chlorite schist unit is discontinuous with that in higher grade rocks to the east. The discontinuity of metamorphic isograds between the western and eastern halves of the study area is used as evidence for the Breakenridge fault in the Structural Geology Chapter.

**Thermobarometry**

Thermobarometric analysis of the area is based largely on a pre-existing database assembled by workers at Western Washington University. Thermobarometry results from this database are summarized in Table 1 and keyed to Figures 3.4, 3.6, and 3.7. Two additional samples, 180-191 (CC) and 180-192 (DD), were chosen to complement pre-existing data. Both new samples are from the sharp bend in Clear Creek on the eastern side.
of the area (Figures 3.4, 3.6, 3.7, Plate 2). Results from these samples are discussed below, and are included in Table 1.

Microprobe analysis, for both the pre-existing database, and samples collected in this study, was conducted by Dr. E.H. Brown using the JOEL electron microprobe at the University of Washington. Averages of from two to ten spots per mineral were used to obtain a representative composition. Rim sites on garnets were used exclusively to represent equilibrium between phases in each rock. Mineral compositions for the two samples analyzed in this study are listed in Appendix B. It should be noted that garnet rims were assumed to be in equilibrium with adjacent matrix minerals.

Two thermometers were used to determine metamorphic temperatures for the two samples analyzed in this study. Pelitic samples were analyzed using the garnet-biotite Fe-Mg exchange (GABI) thermometer (Berman, 1991). Garnet-hornblende schists were analyzed using the garnet-hornblende Fe-Mg exchange (GAHB) thermometer (Graham and Powell, 1984).

Metamorphic pressures were determined using three barometers. Pelitic samples were analyzed using the garnet-biotite-muscovite-plagioclase barometer (GAMI) of Berman (1991). Hornblende-bearing schists were analyzed using the garnet-hornblende-plagioclase barometer (GAHP) of Kohn and Spear (1990).

Results

Sample 180-191 (CC) contains both pelitic and basic mineral assemblages, and was analyzed using two thermobarometers. The GABI/GAMI method gave a pressure of 8.2 kb and a temperature of 660°C. The GAHB/GAHP method gave a pressure of 5.4 kb and a temperature of 620°C. The pressure differs by 2.8 kb between the two methods. Based on work in equivalent rocks in the Cascades of Washington, Brown and Burmester (1991)
estimate an error range of ± 1.0 kb for pressures, and ± 50°C for temperatures, determined using different equilibria. The pressure discrepancy in sample 180-191 is clearly a large one. A similar result was obtained for sample NB23 (O), with a pressure difference of 1.94 kb (Table 1).

Sample 180-192 (DD), located within 50 m of 180-191, records a temperature and pressure of 575°C and 4.0 kb using the GAHB/GAHP thermobarometer.

In each of these samples, pressures, and rarely temperatures, obtained with the GAHB/GAHP thermobarometer are too low when compared with the majority of P-T data in the field area (Figure 3.4 and Figure 3.6; Table 1). Garnets in sample 180-192 contain Ca-rich rims, suggesting higher pressures than were determined. This may be due to error in analysis, or calibration of the barometer. Alternately, the low pressures obtained may be the result of disequilibrium within the mineral assemblage of specific sample localities. Because garnet rim sites were assumed to be in equilibrium with matrix minerals, disequilibrium is possible if the porphyroblast and matrix phases actually formed at different times. Disequilibrium at localized sites may also explain why GAHB/GAHP data at other sites are clearly concordant with surrounding data.

**Discussion of Thermal Gradient**

Thermal data, plotted on the map of Figure 3.4, portray a clear increase in metamorphic temperatures between the shore of Harrison Lake and the eastern side of the field area. Figure 3.7a is a temperature-distance plot based on the projection of thermobarometry sites onto line A-A' of Figure 3.4, which traverses the metamorphic gradient in the northern part of the study area. Site letters are keyed to Table 1 and Figure 3.4. Error bars of ± 50°C are based on estimates from previous studies in the Cascades (Brown and Burmester, 1991). In the eastern half of the study area, metamorphic
Figure 3.7 Plots of thermobarometric data projected onto a line (A-A', Figures 3.4 and 3.6) traversing the metamorphic gradient in the NW part of the study area. Site letters are keyed to Table 1 and Figures 3.4 and 3.6. Error bars are adapted from Brown and Burmester (1991). A. Temperature vs. distance plot. B. Pressure vs. distance plot.
temperatures range from 550-675°C. The difference in temperature between any two adjacent sites is within ± 50°C, indicating that there is no statistical change in temperature in the eastern half of the area.

The major change in temperature occurs on the western side of the study area, where the albite-oligoclase transition is the best evidence for increasing temperature. The transition occurs within one lithology, suggesting that it defines a true metamorphic isograd. It is assigned here as a 450°C isotherm, based on discussion earlier in the text. The west-to-east increase in temperature is also reflected between samples JT-89-19b (H) and JT-89-30e (J), which exhibit temperatures of 405°C and 588°C, respectively (Figures 3.4 and 3.7a).

Discussion of Pressure Gradient

Pressure data are plotted on the map of Figure 3.6. Figure 3.7b shows a pressure-distance plot based on the projection of thermobarometry sites onto a line traversing the metamorphic gradient in the northern part of the study area. Site letters are keyed to Table 1 and Figure 3.6.

In the eastern part of the study area, metamorphic pressures range from 6.00-9.20 kb. Pressure values between any two adjacent samples are within ± 1 kb of each other, suggesting that there is no statistical difference in pressure east of the chlorite schist unit (Figure 3.7b). However, large scale mapping of isobars around the entire Breakenridge antiform do appear to be concentric about a 9 kb high in the Breakenridge orthogneiss (E.H. Brown, pers. comm. 1996). This pattern is consistent with a dome structure, where deeper rocks are exposed in the core. Garnets within this eastern area are in places compositionally zoned with Ca-rich rims, suggesting a jump in metamorphic pressures during crystallization of the garnets (Figure 3.8).
Figure 3.8 Back-scatter electron image of zoned garnet from site 180-192. The transition from red/orange to green/blue indicates an increase in calcium composition during garnet growth. Image was taken using the JOEL electron microprobe at the University of Washington. $X_{\alpha} \text{ core} = 5.55\%$ ; $X_{\alpha} \text{ rim} = 19.16\%$. 
Within the chlorite schist unit, two samples were analyzed for P and T. Sample JT-89-19b (H), on the shore of Harrison Lake, records a pressure of 5.4 kb (Figure 3.6; Table 1). It lies just east of the albite-oligoclase transition, and was used earlier to estimate pressure conditions at which the transition occurs. Sample CB-11B (R) records a pressure of 7.05 kb (Figure 3.6 and 3.7b). It lies 1.25 km east of the albite-oligoclase transition. When allowable error is taken into account, no significant jump in pressure occurs between these two samples.

Sample JT-89-30E (J), at 8.32 kb, lies 1.25 km east of JT-89-19B (5.4 kb). These two samples define an unusually large pressure change, with a difference of almost 3 kb between them (Figure 3.6 and 3.7b). A maximum observable pressure gradient would occur if a normal gradient, assumed to be 0.32 kb/km (Yardley, 1989), were turned on end, resulting in vertically-oriented isobaric surfaces. Although there is no evidence for a tilted gradient, the maximum possible pressure gradient is clearly exceeded by the two samples mentioned above, which record 9 km of relative vertical offset between them (Figure 3.7b). When an error of ± 1.0 kb is taken into account, 3-15 km of offset has occurred across the zone. The abrupt change in pressure is used as evidence for the Breakenridge fault, discussed in the Structural Geology chapter.
IV. STRUCTURAL GEOLOGY

Introduction

Structures within the field area consist of deformational fabrics and map scale structures. Deformational fabrics include foliation, mineral lineations, and stretching lineations. Small-scale folds are rare. Map scale structures consist of the Breakenridge antiform and the Breakenridge fault.

Two deformation events, $D_1$ and $D_2$, are recorded in the area. $D_1$ structures are well-developed throughout the entire study area. $D_2$ is associated with development of the Breakenridge antiform. Small-scale $D_2$ structures are weakly-developed, and occur only in localized outcrops of higher grade rocks in the central and eastern parts of the study area.

Microscopic fabric descriptions are included in the Lithologic Units chapter.

Deformational Fabrics

$D_1$ Foliation

A NW-striking foliation is the predominant solid-state fabric in the study area (Figure 4.1). Cross-cutting of lithologic layering by foliation, and excellent preservation of relict protolith textures, suggest that the foliation is the primary deformational fabric ($S_1$) in the region. Evidence of fabrics pre-dating $S_1$ is restricted to a limited number of ambiguous samples in which linear inclusion trails in the cores of garnets are oblique to surrounding foliation.

In metavolcanic lithologies, $S_1$ is defined by aligned white mica and chlorite, and flattened clusters of quartz and feldspar. In the Breakenridge orthogneiss, $S_1$ is defined by
Figure 4.1 Foliation ($S_2$) attitudes in the field area. Stereonet contour plots of poles to foliation at bottom and top of page correspond to rocks west and east of the proposed Breakenridge fault, respectively. Contour interval = 2.0%/1% area.
aligned biotite clusters, and flattened aggregates of plagioclase and quartz. Foliation in pelitic schist is defined by a graphitic parting, quartz ribbons, and, at higher metamorphic grades, aligned biotite.

The relationship of $S_1$ development to metamorphism is defined by metamorphic index minerals. Biotite is predominantly syn-kinematic within the Breakenridge orthogneiss and the plagioclase-biotite schist unit. Biotite is also syn-kinematic in northern sections of the chlorite schist unit, where it is accompanied by syn-kinematic garnet. Hornblende and garnet are predominantly post-kinematic to $S_1$ throughout all lithologies east of the chlorite schist unit. Garnet is also post-kinematic in the southern part of the chlorite schist unit (Figure 4.2). Hornblende occurs as prisms that are randomly oriented in the $S_1$ foliation plane. In some samples hornblende and garnet cut across the foliation (Figure 2.4). These post-kinematic high-grade minerals indicate that peak metamorphic conditions occurred after deformation ceased.

In the chlorite schist unit, on the western side of the study area, foliation maintains a NW strike and a constant dip of 40-50° E, as shown by the tight cluster of data on inset A of Figure 4.1. In the higher grade rocks to the east of the chlorite schist unit, $S_1$ is deformed by the Breakenridge antiform, and consequently displays a considerable variation in strike. In the northern part of this high grade area, $S_1$ maintains the dominant NW-SE strike. In the western limb of the Breakenridge antiform, foliation is sub-vertical. In the eastern limb, foliation dips approximately 60°E. Along strike to the southeast, however, $S_1$ folds around the south-closing nose of the antiform while maintaining an east dip on both limbs of the fold. This pattern is reflected in inset B of Figure 4.1, which shows a spread of data points that define a SE-plunging fold axis. A cylindrical best fit method, using poles to foliation, was used to calculate a fold axis plunge and trend of 62°, $S79^\circ$E.

Because the antiform is not cylindrical, however, this orientation is an approximation of the fold axis.
Figure 4.2 Photomicrograph (in cross-polarized light) of euhedral post-kinematic garnet cutting foliation at site 180-205 in the chlorite schist unit.

Figure 4.3 Photomicrograph (in cross-polarized light) of syn-kinematic, rotated garnet from site JT-89-30b of the chlorite schist unit, showing top-to-the-left sense of shear.
The structural relationship of $S_1$ to lithologic layering, $S_0$, varies in the south-central part of the field area, where $S_1$ is folded around the nose of the Breakenridge antiform. At the intrusive contact in Big Silver Creek, $S_1$, $S_0$, and the intrusive contact are parallel and change strike abruptly through the tight hinge of the antiform. This relationship suggests that the orthogneiss intruded sub-parallel to $S_0$, followed by layer-parallel development of $S_1$, prior to folding. In the vicinity of the major pelitic unit that crosses Big Silver Creek, however, the map-scale foliation trajectory of $S_1$ crosscuts mapped lithologic contacts between pelitic and plagioclase-hornblende schists on the western limb of the antiform (Figure 4.1). This structural relationship was confirmed on the outcrop scale, where $S_1$ foliation was observed to cut across an $S_0$ conglomerate layer with identical orientations to those observed in map scale (E.H. Brown, pers. comm. 1996). This relationship may be explained by the development of $S_1$ at an oblique angle to original stratigraphic layering prior to folding in the antiform. This model could also be explained by the development of tight $F_1$ folds, in which $S_1$ is an axial-planar fabric. The localized occurrence of a large $F_1$ fold in the western limb might explain the absence of cross-cutting foliation in the eastern limb.

**D$_1$ Lineation**

Lineations ($L_1$) consist of well-developed stretching and mineral lineations which lie in the $S_1$ foliation (Figure 4.4). Stretching lineations are defined by elongate clasts and bombs in conglomerates and agglomerates, respectively. Mineral lineations are defined mostly by syn-tectonic biotite, occurring in the Breakenridge orthogneiss and plagioclase-biotite schist. At the northern end of the chlorite schist unit, rotated syn-kinematic garnets associated with down-dip mineral lineations indicate a top-to-the-west sense of
Figure 4.4 Lineation ($L_r$) attitudes in the field area. Stereonet contour plots at bottom and top of page correspond to regions west and east of the proposed Breakenridge fault, respectively. Contour interval = 2.0%/1% area.
shear (Figures 4.3 and 4.4). Rare occurrences of garnet and biotite in the southern part of
the chlorite schist unit are post-kinematic to the foliation, however (Figure 4.2). Some
outcrops within the plagioclase-hornblende schist exhibit a rare occurrence of lineated
hornblende (Figure 2.3). These outcrops probably represent isolated zones where strain
continued during peak metamorphism. Mineral and stretching lineation orientations are
parallel when observed together in outcrop.

Lineation map patterns define two distinct domains of lineation orientations.
Lineations in the chlorite schist unit, which are defined predominantly by stretched clasts in
metaconglomerates, maintain a dominant down-dip orientation in the northeast-dipping
foliation, as shown on inset A of Figure 4.4. Down-dip lineations are attributed to orogen-
normal tectonic transport. Inset A of Figure 4.4 also shows a cluster of north-plunging
lines representing strike-parallel lineations along the shore of Harrison Lake. These
lineations are attributed to strike-slip shear along the Harrison Lake Shear Zone (Monger,
1989; Talbot, unpublished). East of the chlorite schist unit, in the Breakenridge
orthogneiss and the high-grade metavolcanic rocks that envelop it, lineations are
predominantly strike-parallel, except where folded in the Breakenridge antiform (Figure
4.4).

The boundary between down-dip and strike-parallel domains is well defined in the
north, where it is used later in this chapter as evidence for the proposed Breakenridge fault.
In the south, the domain boundary is poorly defined (Figure 4.4). East-plunging, down-
dip lineations in the hinge area of the Breakenridge antiform appear to be aligned with
down-dip lineations in the southern part of the chlorite schist unit to the west. Definition of
the domain boundary in this area is based on the assumption that lineations in the hinge area
of the antiform were folded with S₁, and therefore used to be strike-parallel like those
further to the north. Those in the chlorite schist unit are assumed to be unaffected by the
development of the antiform. This is consistent with the map trend of the eastern contact of
the chlorite schist unit, which does not parallel folded lithologic marker units of pelitic schist, and has therefore not been folded in the antiform either. Juxtaposition of folded and unfolded areas is interpreted to have occurred along the proposed Breakenridge fault, discussed later in this chapter.

Down-dip lineations also occur adjacent to the eastern edge of the field area, where lineations plunge moderately to the NE and E, oblique to strike-parallel lineations along the eastern margin of the Breakenridge orthogneiss.

**D₁ Folds**

Folds (F₁) are uncommon in the field area, due in part to poor exposure. One outcrop in the chlorite schist unit displays a clear relationship between F₁ and S₁, where foliation is axial-planar to east-plunging isoclinal folds in relict tuff layers (Figure 4.5). Many outcrops in the unit do show S₁ cross-cutting S₀, which may indicate the presence of larger scale folds with an axial-planar foliation. Axes of inferred folds at these outcrops, although not measurable directly, were assumed to be parallel to the intersection lineations of S₁ on S₀, which show east-plunging attitudes with considerable variation in trend. The above relationships suggest a simple, single-event deformation history similar to that reported by Lynch (1990) for the Fire Lake Group at the northern end of Harrison Lake.

Folds (F₁) of compositional layering (S₀) were observed at several localities within the plagioclase-hornblende schist unit. Folds in western exposures of the unit are generally open structures in felsic to intermediate tuff layers (Figure 4.6a). An axial-planar relationship of S₁ to F₁ was observed in one northwest-plunging fold at site 180-85 (Plate 2). Folds in eastern exposures of the plagioclase-hornblende schist unit exhibit a clear relationship between F₁ and S₁. Compositional layering is isoclinal folded with an axial planar foliation (S₁). Sample 180-138 (Plate 2) shows tight isoclinal folds of felsic and
Figure 4.5 Outcrop photo of isoclinal $F_1$ fold in relict volcanic layering of the chlorite schist unit. Axial-parallel foliation is parallel to long edge of transit. Looking ENE, down-dip (site 180-36b).
Figure 4.6 F, folds in the plagioclase-hornblende schist unit. A. Open fold in relict tuffaceous layering from site 180-24. B. Isoclinal fold in felsic and mafic layers of site 180-138. Axial-planar foliation is parallel to long-axis of photo. Coin is 2.8 cm in diameter.
intermediate metavolcanic layers, in which $S_1$ is an axial planar fabric (Figure 4.6b). Aligned biotite and hornblende on the foliation surface are perpendicular to fold axes.

$D_2$ Structures

$D_2$ structures consist of folds and S- and L-fabrics that deform $D_1$ fabrics. $F_2$ folds in $S_1$ were observed in the hinge area of the Breakenridge antiform, where they occur in well-layered, fine-grained felsic metavolcanic rocks and well-foliated mylonites (Figures 4.7 and 4.8). $F_2$ folds were also observed in pelitic schist adjacent to the intrusive contact of the Breakenridge orthogneiss in Big Silver Creek (site 180-105, Figure 4.7). Folds exhibit an open geometry with low amplitude and wavelength. Fold axes plunge E to SE, and are sub-parallel to the fold axis calculated for the Breakenridge antiform. Based on this latter observation, and their localized occurrence, $F_2$ folds are interpreted to be second-order folds formed by buckling in the tightly-folded, SE-plunging nose of the Breakenridge antiform. The antiform itself is interpreted to be a large-scale $D_2$ structure.

Small scale secondary folding of $D_1$ fabrics in the Breakenridge orthogneiss was observed at one locality. At site JT-88-16e (Plate 2), located inside the orthogneiss approximately 650 m from its western intrusive contact, strike-parallel $L_1$ lineations are deflected into a down-dip orientation on a sub-vertical $S_1$ surface, similar to sheath fold geometry (Figure 4.9). This observation would suggest that strike-parallel movement during $D_1$ was followed by vertical motion during $D_2$.

Secondary fabrics consist of rare mineral lineations defined by biotite in felsic metavolcanic layers. Field observations of overprinting relationships were observed in four samples. Sites 180-73, 180-74, and 180-145 (Plate 2) occur in felsic, fine-grained, felsic mylonites in the western hinge area of the Breakenridge antiform (Figure 4.7). Elongate clusters of biotite define high-strain stretching lineations ($L_1$) in a sub-vertical
Figure 4.7 Detailed map of hinge area of the Breakenridge antiform, showing attitudes of $D_2$ structures.
Figure 4.8 $F_2$ folds of $S_1$ foliation surface at site 180-145. Folds plunge toward the right hand side of photo. Folded $L_1$ stretching lineations on the $S_1$ surface alternate between an east and a west plunge through the sequence of folds.

Figure 4.9 $F_2$ fold of $L_1$ stretching lineations on sub-vertical $S_1$ foliation surface in Breakenridge orthogneiss at site JT-88-16c. Lineations were originally strike-parallel, and have been deflected into a down-dip orientation, similar to sheath fold geometry.
foliation. In each case, \( L_1 \) is strike-parallel and plunges 20-50° E. Individual biotite grains have been re-oriented within the clusters, defining a foliation-parallel lineation (\( L_2 \)) with a steep, down-dip plunge of 60-80° E (Figure 4.10). At site 180-145, reoriented biotite grains and aligned pressure shadows on pyrite cubes define an \( L_2 \) lineation that is axial-parallel to east-plunging \( F_2 \) folds of mylonitic foliation (Figure 4.10a). Folded stretching lineations (\( L_3 \)) alternate between an east and a west plunge through the sequence of folds, while \( L_2 \) maintains a constant axial-parallel plunge of 70-80° E (Figure 4.8). These observations may indicate that \( L_2 \) is related to \( F_3 \) development.

\( D_2 \) fabric development appears to be absent from the orthogneiss itself. This may be a rheological effect, where the orthogneiss was more competent than the volcanic rocks that envelop it. This interpretation would explain the preservation of syn-kinematic biotite, which defines \( S_1 \), in the orthogneiss.

**Structural Geology of Injection Zones**

Several outcrops expose contact relationships between the Breakenridge orthogneiss and surrounding country rock. These outcrops are critical to understanding the geologic history of the area, because they define the orthogneiss-country rock boundary as an intrusive contact. In addition, they define the relative timing of igneous emplacement, deformation, and metamorphism.

Four outcrops, exposed in road cuts and stream beds, exhibit injection zone features overprinted by solid state foliation. At site 180-20 (Plate 2), orthogneiss intrudes plagioclase-hornblende schist in a 5m-wide mixing zone. Xenoliths of metavolcanic schist in orthogneiss are flattened parallel to foliation (Figure 4.11a). A similar zone was observed at site JT-95-8b (Plate 2), where xenoliths of plagioclase-biotite schist in orthogneiss exhibit reaction rims and show overprinting by solid state foliation (Figure 65).
Figure 4.10  Fabric overprinting relationships.  A. Sample photo of $L_2$ crossing $L_1$ on $S_1$ foliation surface in mylonite of site 180-145. $L_2$ is defined by pressure shadows (white) on pyrite cubes (dark grains), and re-oriented biotite grains (not visible in photo). B. Photomicrograph (in plane-polarized light) of site 180-74. $L_1$ is defined by biotite clusters extending from upper left to lower right of photo. $L_2$ is defined by individual, re-oriented biotite grains lined up parallel to long-axis of photo.
Figure 4.11 Outcrop photos of orthogneiss-country rock injection zones overprinted by S₁ foliation. **A.** Deformed mixing zone textures at site 180-20. **B.** Flattened xenoliths from mixing zone at site JT-95-8b.
4.1 lb). At site 180-193 (Plate 2), in the sharp bend in Clear Creek, apophyses of orthogneiss intrude plagioclase-hornblende schist, and are in turn overprinted by solid-state foliation (Figure 4.12).

The best-exposed outcrop of injection zone textures occurs in the polished bedrock of Big Silver Creek at site 180-105 (Plate 2; Figure 4.13). This site consists of a complex 10m-wide mixing zone of highly deformed xenoliths and apophyses that lies in the hinge zone of the Breakenridge antiform. Structural relationships, portrayed in the schematic diagram of Figure 4.13b, are complex, and appear to be unique within the field area. Foliation, striking approximately N30°E with a 60-70°SE dip, overprints the injection zone parallel to both the dominant intrusive contact and compositional layering in adjacent wall rocks. Foliation is accompanied by a distinct down-dip lineation, defined in orthogneiss by deformed primary igneous minerals and stretched xenoliths, and in pelitic schist by aligned biotite. The lineation is parallel to the axes of isoclinal folds of S₁ in pelitic schist immediately adjacent to the mixing zone. The age of folds relative to lineations is difficult to determine, however. Lineations may not be related to fold development. Because the lineations are a primary deformational feature, they are most likely L₁ structures associated with S₁ foliation. Because S₁ is itself folded, it appears that folding occurred after lineation development. It is therefore probable that fold axes developed parallel to the pronounced L₁ linear fabric, a structural relationship that is common in sheath folds (Watkinson and Cobbold, 1981). Folds at this site are therefore D₂ structures. Fold axes are parallel to the fold axis calculated for the southern nose of the Breakenridge antiform, which supports the interpretation that they are F₂ folds. The presence of tight F₂ folds at this location is to be expected, because the site lies in the hinge zone of the antiform, where deformation is more pronounced. This is similar to the geometry of a developing sheath fold.
Figure 4.12 Apophyses of orthogneiss flattened in $S_1$ at site 180-193.
Figure 4.13 Injection zone at site 180-105. A. Photograph of outcrop in Big Silver Creek (ogn = orthogneiss; ps = pelitic schist). Looking southeast down dip of foliation. D<sub>1</sub> folds are faintly visible in pelitic schist. B. Schematic block diagram showing structural relationships of D<sub>1</sub> and D<sub>2</sub> structures in injection zone. Looking north.
The Breakenridge antiform is an elongate, dome-like fold that extends 25 km NNW from the lower reaches of Big Silver Creek (Figure 1.3). Its core is occupied by the Breakenridge orthogneiss, which is flanked by metavolcanic and metasedimentary schists and gneisses of the Slollicum Schist. Previous work on the antiform concentrated on its north-closing nose, where it is defined by a broad, upright, NNW-plunging fold of the intrusive contact between the Breakenridge orthogneiss and the Slollicum Schist (Reamsbottom, 1974). Foliation ($S_1$) and compositional layering ($S_0$) are parallel to this folded contact. Second order folds associated with the antiform are defined by mutually parallel marker units of pelitic schist and foliation trajectories.

This study focuses on the southern margin of the antiform, where structural relationships are analogous to those in the north. The intrusive contact between the Breakenridge orthogneiss and the Slollicum Schist is folded into a SE-closing nose. Distinct pelitic marker units follow the folded contact (Figure 2.1). Foliation ($S_1$) in both the Breakenridge orthogneiss and the Slollicum Schist wraps around the nose parallel to the intrusive contact (Figure 4.1). When combined with observations from the northern end of the structure, the stratigraphic and structural data presented above define the Breakenridge antiform as a doubly-plunging, mantled gneiss dome.

East-dipping foliation on both limbs of the antiform at its southern end indicate that it is overturned to the west and reclined to the east. A cylindrical best-fit method, using poles to $S_1$ foliation, was used to calculate a fold axis plunge and trend of 62°, S79°E for the southern nose of the Breakenridge antiform. Because the antiform is not cylindrical, this orientation is an approximation of the fold axis. The reclined geometry of the southern end of the dome is easily seen on a small scale at site 180-105 (Plate 2), which lies in the
hinge zone of the antiform, as discussed in the immediately preceding section (Figure 4.13).

The kinematic development of the Breakenridge antiform is constrained by the geometry of the structure as a whole. Both ends of the antiform are deflected downward relative to the central part of the dome (Figure 4.14). At the northern end of the structure, this deflection is gentle, resulting in a broad, upright, north-plunging fold of $S_1$. $L_1$ is NW-trending and remains parallel to the axis of the fold. At the southern end, the structure is a tight fold that is overturned to the SW and reclined to the SE. $L_1$ lineations formed during strike-parallel movement are deflected downward to the southeast. This geometry is analogous to large-scale sheath folds cored by gneissic basement in the Moine thrust zone of northern Scotland (Holdsworth, 1989). Basement-cover relationships in the Moine rocks are slightly different, because mantling metasedimentary rocks were deposited unconformably on gneissic basement prior to sheath folding and associated metamorphism. Nevertheless, the Moine thrusts illustrate that tightly appressed gneiss domes commonly develop as sheath folds in compressional tectonic regimes.

Differential uplift of the Breakenridge antiform as a developing sheath fold, where the middle section of the dome rises rapidly relative to the southern section, would cause the extreme downward warping of the southern end. This geometry is supported by small scale structures. $F_2$ fold axes are parallel to the axis of the reclined southern end of antiform. $L_2$ lineations, which cross-cut $L_1$, are oriented down-dip, and are therefore consistent with vertical motion. Similarly, small scale sheath folds of strike-parallel $L_1$ lineations in orthogneiss of site JT-88-16e suggests upward movement of rocks during $D_2$. Vertical uplift during doming also explains the presence of high-pressure rocks in the core of the antiform.
Figure 4.14 Schematic block diagram of the Breakenridge antiform, showing representative lithostratigraphic layering in country rock, and folded Breakenridge orthogneiss. Arrows represent orientation of folded strike-parallel L\textsubscript{s} stretching lineations. The diagram graphically portrays the open, upright, shallow plunging geometry of the northern end of the antiform, and the tight, steeply plunging geometry of the southern end of the structure. The SE-plunging, reclined geometry of the southern end is not shown, but is portrayed in Figure 4.13. Lithostratigraphic layering is truncated by the Breakenridge fault, but is not intended to represent offset of a single stratigraphic marker in this diagram. View looking NNE.
Kinematics of \( D_1 \) are poorly constrained in the field area, due to a limited amount of reliable shear-sense data. This condition is probably due to regional post-kinematic static recrystallization, during which kinematic indicators were partly erased by grain boundary area reduction in quartz and feldspar, and straight grain-boundary growth of micas. This condition is dominant in higher grade rocks east of the chlorite schist unit, where only two reliable shear-sense indicators were observed.

Map patterns of lineations define two domains, as discussed previously. Down-dip, NE-plunging mineral lineations at the northern end of the western domain are associated with rotated, snowball-type garnets that indicate a top-to-the-southwest sense of shear (Figure 4.3 and 4.4). Identical shear-sense indicators were observed along strike 10 km to the north (T. Lapen, pers. comm. 1996). These observations suggest that down-dip lineations in the chlorite schist unit developed during SW-directed thrusting. \( L_1 \) lineation orientations east of the chlorite schist unit are predominantly strike-parallel, except where folded in the Breakenridge antiform, suggesting orogen-parallel tectonic transport in these higher grade rocks. However, no kinematic indicators were observed in these rocks.

The different \( L_1 \) lineation orientations suggest the juxtaposition of rocks deformed in arc-normal and arc-parallel tectonic regimes, respectively. Attempts at quantitatively restoring strike-parallel lineations to an orogen-normal orientation by unfolding the Breakenridge antiform were not successful. This is to be expected, because the antiform is not a cylindrical fold. Lineations in the central and northern areas of the Breakenridge antiform are strike-parallel, and therefore do not rotate when unfolded about the axis of the structure. Only the southern end of the antiform shows deflection of \( L_1 \) away from the strike-parallel orientation. Therefore, lineation orientations were NW-SE trending prior to
folding in the antiform. The boundary between strike-parallel and down-dip lineation domains is inferred to be along the proposed Breakenridge fault, discussed later in this chapter.

Along the eastern edge of the field area, lineations trend NE at an oblique angle to strike-parallel lineations in the adjacent Breakenridge orthogneiss. This transition is not well-understood. It may result from folding of lineations in a D₂ fold of smaller scale than the Breakenridge antiform itself. Alternatively, the discontinuity may represent a kinematic transition from strike-parallel to down-dip tectonic transport in rocks further to the east.

**Strain Analysis**

Strain data were collected on 17 samples from deformed metaconglomerates, metaagglomerates, and orthogneiss biotite clusters. Two out of three strain ratios were measured on multiple strain markers from each outcrop, where the X/Y plane is assumed to equal foliation, and X equals the lineation direction (Ramsay and Huber, 1983). Results are listed in Table 2. Ellipses representing the X/Z strain ratio are plotted on Figure 4.15. A Flinn diagram is shown on Figure 4.16.

Strain patterns deduced from these seventeen samples are somewhat limited, due to the small number of data points. Three sites within the chlorite schist unit all record relatively low K values, indicating flattening strain. These values appear to be representative when considering that foliation and metamorphic recrystallization are poorly developed in this unit as a whole. Strain in higher grade rocks to the east of the chlorite schist unit appears to be highly variable. Six of fourteen sites record constrictional strain, all of which occur in rocks with high-strain fabrics, such as mylonites. The remaining eight samples record flattening strain.
Table 2. Strain data. Site letters are keyed to Figures 4.15 and 4.16.

<table>
<thead>
<tr>
<th>Site</th>
<th>Sample number</th>
<th>strain marker</th>
<th>X/Y</th>
<th>Y/Z</th>
<th>X/Z</th>
<th>K*</th>
<th>Lineation Plunge, Trend</th>
</tr>
</thead>
<tbody>
<tr>
<td>A.</td>
<td>180-12</td>
<td>agglomerate</td>
<td>1.97</td>
<td>2.86</td>
<td>5.64</td>
<td>0.52</td>
<td>unrecorded</td>
</tr>
<tr>
<td>B.</td>
<td>180-21</td>
<td></td>
<td>1.45</td>
<td>3.55</td>
<td>5.14</td>
<td>0.18</td>
<td>42, S70E</td>
</tr>
<tr>
<td>C.</td>
<td>180-24</td>
<td></td>
<td>3.82</td>
<td>1.89</td>
<td>7.24</td>
<td>3.17</td>
<td>50, S40E</td>
</tr>
<tr>
<td>D.</td>
<td>180-28</td>
<td>conglomerate</td>
<td>1.46</td>
<td>8.65</td>
<td>12.61</td>
<td>0.06</td>
<td>50, N50E</td>
</tr>
<tr>
<td>E.</td>
<td>180-31</td>
<td></td>
<td>3.20</td>
<td>6.30</td>
<td>9.21</td>
<td>0.42</td>
<td>48, N45E</td>
</tr>
<tr>
<td>F.</td>
<td>180-32</td>
<td></td>
<td>2.76</td>
<td>3.98</td>
<td>11.00</td>
<td>0.59</td>
<td>40, N40E</td>
</tr>
<tr>
<td>G.</td>
<td>180-70</td>
<td></td>
<td>1.77</td>
<td>3.33</td>
<td>5.90</td>
<td>0.33</td>
<td>70, N75E</td>
</tr>
<tr>
<td>H.</td>
<td>180-71</td>
<td></td>
<td>5.40</td>
<td>4.31</td>
<td>23.30</td>
<td>1.33</td>
<td>53, N70E</td>
</tr>
<tr>
<td>I.</td>
<td>180-75</td>
<td></td>
<td>8.80</td>
<td>2.88</td>
<td>25.35</td>
<td>4.14</td>
<td>36, S84E</td>
</tr>
<tr>
<td>J.</td>
<td>180-97</td>
<td>agglomerate</td>
<td>1.79</td>
<td>2.77</td>
<td>4.96</td>
<td>0.45</td>
<td>65, S65E</td>
</tr>
<tr>
<td>K.</td>
<td>180-104</td>
<td>conglomerate</td>
<td>2.60</td>
<td>3.86</td>
<td>10.00</td>
<td>0.56</td>
<td>50, N50E</td>
</tr>
<tr>
<td>L.</td>
<td>180-136</td>
<td>agglomerate</td>
<td>3.55</td>
<td>3.86</td>
<td>13.7</td>
<td>0.89</td>
<td>38, N18E</td>
</tr>
<tr>
<td>M.</td>
<td>180-184</td>
<td></td>
<td>16.00</td>
<td>1.68</td>
<td>26.88</td>
<td>22.10</td>
<td>35, S80E</td>
</tr>
<tr>
<td>N.</td>
<td>JT-88-38d</td>
<td></td>
<td>1.10</td>
<td>11.86</td>
<td>12.89</td>
<td>0.01</td>
<td>48, S38E</td>
</tr>
<tr>
<td>O.</td>
<td>JT-89-12</td>
<td>orthogneiss</td>
<td>2.50</td>
<td>2.00</td>
<td>5.00</td>
<td>1.50</td>
<td>13, N49W</td>
</tr>
<tr>
<td>P.</td>
<td>JT-95-2e</td>
<td>biotite cluster</td>
<td>4.00</td>
<td>5.00</td>
<td>15.00</td>
<td>0.75</td>
<td>69, S42E</td>
</tr>
<tr>
<td>Q.</td>
<td>JT-95-4</td>
<td></td>
<td>4.00</td>
<td>10.00</td>
<td>40.00</td>
<td>0.33</td>
<td>55, S38E</td>
</tr>
</tbody>
</table>

*K = (X/Y - 1)/(Y/Z - 1); K=1 is plane strain. Underlined K values represent apparent constrictional strain; other values represent apparent flattening strain.
Figure 4.15 Distribution of strain data in the field area. Strain ellipses are oriented parallel to elongation direction of stretching lineation. Site letters are keyed to Table 2.
Figure 4.16 Flinn diagram. Letters are keyed to Table 2 and Figure 4.18.
A high-strain shear zone, coincident with the main ridge-top on the western side of the Big Silver Creek drainage, is the best example of concentrated high strain development in the field area. The shear zone consists of a series of apparently discontinuous outcrops of mylonite, which collectively define a 100m-wide, NW-SE linear structure across the western side of the field area (Figure 4.15). Foliation in these zones is conformable to the regional foliation in the surrounding rocks. The shear zone is offset in an apparent sinistral sense across a NNW-trending, brittle high angle fault in the nose of the antiform.

Eastward continuity across Big Silver Creek is not known.

Shear zone outcrops are characterized by mylonitic fabrics exhibiting varying levels of strain intensity. Mylonitic fabric in felsic volcanic rocks shows intense grain-size reduction and tectonic laminar foliation (4.17). These outcrops typically contain stretched biotite clusters which form a distinct streak-like lineation on foliation surfaces (Figure 4.17). Strain ratios on these stretched clusters were difficult to measure, but X/Z ratios are estimated at greater than 50:1. Constrictional strain is common to the zone, as seen in samples 180-75 and 180-184 (Plate 2). Kinematics of the shear zone are not well understood, as no reliable shear-sense indicators were observed. L, lineations are strike-parallel within the shear zone, suggesting that it was simply a higher strain zone associated with strike-parallel motion in the higher grade rocks.

Faults

Breakenridge Fault

The Breakenridge Fault was defined previously as a mylonitic, high-angle reverse fault of the Coast Belt Thrust System, along which uplift of high-grade metamorphic rocks
Figure 4.17 A. Boulders of mylonite from site 180-74. Note stretched biotite clusters on foliation (X-Y) surface, and their elliptical intersections on Y-Z surfaces. Coin is 2.8 cm in diameter.

B. Photomicrograph (in cross-polarized light) of mylonite from site 180-74, showing tectonically-layered foliation, grain-size reduction, and static-recrystallisation textures in quartz and feldspar.
Occurred (Journey and Friedman, 1993; Monger, 1989). It was mapped in the same location as the high-strain shear zone mentioned in the Strain Analysis section of this study, and is assumed to have been based on that structure. Several lines of evidence suggest that the shear zone has limited displacement, and is not an important structure in the uplift history of the high grade rocks. (1) Thermobarometry studies from samples which bracket the zone show no appreciable change in metamorphic grade. (2) The shear zone cuts across a sill associated with the main body of the Breakenridge orthogneiss, thus limiting large-magnitude vertical displacement. Lithologic contacts are not offset along strike, suggesting that the mylonite outcrops are localized high-strain zones. (3) Lineation and foliation attitudes within the shear zone are conformable with adjacent regional fabrics. (4) Lineation orientations are strike-parallel within the zone, thus eliminating the possibility of an orogen-normal, top-to-the-southwest sense of shear as required in the reverse fault model.

The Breakenridge fault is redefined here as a post-metamorphic structure that trends NW-SE, west of its previously mapped location (Figure 4.15). The fault zone itself was not observed during the course of this study. Instead, the fault is inferred based on metamorphic, lithologic, and structural discontinuities that coincide along the west side of the study area, as follows:

(1) In the northern part of the area, a west-to-east increase of at least 3 kb in metamorphic pressure occurs across a zone 1.25 km wide, as discussed in the Metamorphism chapter (Figure 3.7). Within error, this abrupt change in pressure requires a minimum of 3 km relative displacement across the zone, and is therefore significant evidence for a post-metamorphic fault zone.

(2) In the southern part of the area, the albite-oligoclase transition is discontinuous across the Big Silver Creek delta, where it either is truncated by a fault, or bends sharply to the south. Projection of the albite-oligoclase transition across the delta as drawn would
place sample JT-88-7f (Figure 3.4, site A), of An_{27}, on the low-temperature side of the albite-oligoclase transition (An_{17-20}). It is therefore probable that a fault is present in the delta that juxtaposes rocks which record high temperatures against rocks which lie below the 450°C albite-oligoclase transition.

(3) The eastern boundary of the chlorite schist unit does not follow folded marker units of pelitic schist that define the antiform in rocks to the east. Instead, the boundary swings southward, suggesting a possible structural break between folded and unfolded areas. This interpretation is supported by the map trend of the albite-oligoclase transition, which does not continue eastward, and is therefore not folded in the Breakenridge antiform either. The eastern boundary of the chlorite schist unit defines a distinct lithologic change between chlorite schists to the west and hornblende-bearing schists to the east. Because the chlorite schists are devoid of actinolite, the abrupt occurrence of hornblende is caused by a lithologic change, and does not represent a metamorphic change in amphibole composition within a single lithologic unit (see Metamorphism chapter). Therefore, the change may be structurally controlled by the proposed Breakenridge fault.

(4) The boundary between lineation domains, discussed earlier in this chapter, coincides spatially with the lithologic and metamorphic changes discussed above. Because no overprinting was observed, the change between down-dip and strike-parallel lineations is interpreted to be controlled by the proposed Breakenridge fault.

Because it was never seen as a discrete structure in the field, the Breakenridge fault is better described as a zone of distributed shear. Based on the pressure data described in the metamorphism chapter, at least 3 km of relative displacement has occurred across a zone approximately 1.25 km wide (Figure 4.18). If it is assumed that distributed simple shear occurred across the entire width of 1.25 km along discrete shears spaced 1 cm apart, the amount of displacement required of a single discrete shear would be approximately 2.4 cm in order to produce a 3 km relative displacement. This interpretation would allow for
Figure 4.18 Schematic cross-section showing proposed deflection of isobars along the post-metamorphic Breakenridge fault. The true orientation of isobars west of the fault is not known, and is assumed here to be sub-horizontal. Data points at present day erosional surface are keyed to Table 1, Figure 3.6, and Figure 3.7.
the relatively small amount of post-metamorphic deformation observed in the chlorite schist unit. Post-metamorphic shear-bands were observed to wrap metamorphic biotite in some samples within the chlorite schist unit (Figure 4.19), suggesting that some late-stage deformation occurred after D1 in these rocks. A wide shear zone model would require that the discrete line representing the Breakenridge fault as mapped in this study really represents the eastern edge of the shear zone, an interpretation that explains the transition from down-dip to strike-parallel lineation attitudes along this line. The true width of the fault zone is not known, however. It should be noted that a significant part of the uplift may have occurred as high-pressure rocks moved upward during shortening in the core of the Breakenridge antiform, allowing for even less post-metamorphic deformation in the shear zone.

The tectonic significance of the Breakenridge fault zone is not known. The map trace of the proposed fault remains sub-linear across topographic variations, suggesting that it is a steeply dipping structure. The abrupt change in metamorphic grade is explained by the reverse fault model of Joumeay and Friedman (1993), where the Breakenridge fault is responsible for the uplift and westward translation of high grade rocks in the eastern part of the area. It is possible that the Breakenridge fault and the Breakenridge antiform developed together as a fault-fold structure. It is also possible that the Breakenridge fault is a normal fault.

**Harrison Lake Shear Zone**

The Harrison Lake shear zone (Harrison fault) was mapped by Monger (1989) as a major dextral strike-slip fault zone that extends the full length of Harrison Lake. The zone separates unmetamorphosed rocks on the west side of Harrison Lake from metamorphosed rocks on the east side, and is thought to have been active from the Middle Cretaceous to
4.19 Photomicrograph (in plane-polarized light) of post-metamorphic shear bands wrapping biotite and garnet porphyroblasts at site 180-33 in the western structural domain. Distributed shear on structures such as these may have contributed to displacement on the Breakenridge fault.
Late Tertiary. Although inferred to lie within the lake along most of its length, it was mapped within the chlorite schist unit of this study, where it was interpreted to juxtapose Brokenback Hill Formation against Slollicum Schist (Figure 1.3).

No evidence for a discrete fault was observed in this study. Traverses across the proposed fault show no break in lithology within the chlorite schist unit, which is assumed here to be Slollicum Schist. Furthermore, isograds indicate that rocks interpreted to be Brokenback Hill Formation, an unmetamorphosed metavolcanic rock unit on the west side of Harrison Lake, were actually metamorphosed to biotite and garnet grades. In addition, NW-SE trending lineations, which define the Harrison Lake Shear Zone to the south, were not observed in this area.

Eastern boundary fault for the Breakenridge Orthogneiss

The eastern contact of the Breakenridge orthogneiss was mapped previously as an unnamed west-vergent thrust which juxtaposes Slollicum Schist over the Breakenridge antiform (Monger, 1989; Journeay and Friedman, 1993; Figure 1.3). South of the orthogneiss, the fault was mapped along Big Silver Creek and truncated by the Harrison Lake Shear Zone in Harrison Lake. Mapping of the eastern orthogneiss contact in this study shows that the Breakenridge orthogneiss is intrusive into Slollicum Schist. A distinct pelitic marker unit is continuous across the trace of the proposed fault in Big Silver Creek, indicating that no fault exists there. A possible location for such a fault might lie along the contact between metavolcanic and pelitic schists on the eastern side of the area. This contact appears to be aligned with the eastern edge of the Jurassic sill to the north, suggesting that it might represent a structural break.
Central Coast Belt Detachment

The Central Coast Belt Detachment is cited by Journeay and Friedman (1993) as an east-dipping, mylonitic thrust fault bounding the east side of the Breakenridge antiform in the Coast Belt Thrust System (Figures 1.3 and 1.4). In this model, the Breakenridge fault and the Central Coast Belt Detachment isolate the antiform as an allochthonous wedge (Figure 1.4). It is described as an interlocking network of ductile shear zones in which down-dip stretching lineations record an east-side-up shear sense. Although the fault is not exposed in the area of this study, its presence immediately to the east is significant in determining if the Breakenridge antiform is allochthonous.

The detachment cuts the 90-84 Ma Scuzzy batholith, and is therefore a post-metamorphic fault. It is therefore of appropriate age to have formed coeval with $D_2$ of this study. The Breakenridge fault, Breakenridge antiform, and Central Coast Belt Detachment are of similar age, and maintain a geometry consistent with their mutual development as a fault-fold structure in a west-vergent orogenic belt, as proposed by Journeay and Friedman (1993).

Summary

Two major deformation events are recorded in the study area. During the first deformation event, $S_1$, $L_1$, and $F_1$ developed. Lineation orientations define a western domain of down-dip lineations, and an eastern domain of strike-parallel lineations. Metamorphism was probably of intermediate grade, because biotite and garnet are the only metamorphic index minerals syn-kinematic with $D_1$. $D_1$ fabrics are overprinted by high-
pressure, post-tectonic mineral assemblages accompanied by regional static recrystallization textures.

The second deformation event, $D_2$, involves development of the Breakenridge antiform. $S_1$ and $L_1$ are folded in an incipient sheath fold pattern in the antiform, and in rare small scale folds associated with the larger structure. Localized areas developed secondary fabrics associated with $D_2$ folding. Low-grade constituent minerals in these fabrics indicate that $D_2$ occurred at intermediate to shallow levels in the crust, suggesting that the event is related to uplift.

Development of the Breakenridge fault represents the last major deformation event in the area. Telescoping of metamorphic gradients, based on a large jump in metamorphic pressure, and the truncation of isograds, suggests that the fault is a distributed shear zone. The Breakenridge fault is equal to or younger than the Breakenridge antiform in age, and may have formed in unison with the Breakenridge antiform and the Central Coast Belt Detachment as a fault-fold structure during $D_2$. 
V. GEOCHRONOLOGY

Introduction

Geochronology was conducted by isotope analysis of U-Pb in zircon and $^{40}\text{Ar}/^{39}\text{Ar}$ in hornblende and mica. Analysis by the U-Pb method was conducted on a sample of Breakenridge orthogneiss for comparison with previously determined U-Pb pluton crystallization ages in the area. Argon-argon geochronology was conducted to determine cooling ages for high-grade rocks, and thus establish a minimum age bracket for peak metamorphism. The U-Pb geochronology was carried out by W.C. McClelland and the Ar-Ar geochronology by A. Calvert, both at the University of California at Santa Barbara.

U-Pb Geochronology

Three previous U-Pb zircon ages have been determined in the study area (Figure 5.1). Gabites (1985) reported an age of $105 \pm 5$ Ma in a felsic phase of Breakenridge orthogneiss along the southeast margin of the pluton. Parrish and Monger (1992) reported an age of $96 \pm 1$ Ma in a mafic phase of Breakenridge orthogneiss in the core of the pluton. They also reported an age of $102 \pm 1$ Ma on crystal lithic tuff of the Slollicum Schist west of the Big Silver Creek delta along the shore of Harrison Lake.

An additional U-Pb zircon date was determined for the Breakenridge orthogneiss in this study. The sample was collected from a felsic phase adjacent to the intrusive contact at the southern margin of the pluton (Figure 5.1). A concordia plot for this sample is shown in Figure 5.2, and the data are listed in Appendix C. It records an age of $103.8 \pm 0.5$ Ma, based on the concordance of fractions a and b. This result is similar to the age of $105 \pm 5$ reported by Gabites (1985), but it has a smaller error. When compared with previous ages
Figure 5.1 Geochronology site localities in the field area. The 96 Ma age of Parrish and Monger (1992) on Breakenridge orthogneiss was sampled outside the northern edge of the field area.
Figure 5.2 U-Pb concordia plot of sample 180GN.
mentioned above, this result suggests that igneous emplacement and crystallization of the
Breakenridge orthogneiss occurred between 104 - 96 Ma. The limited amount of data
suggests that felsic phases are older than mafic phases of the plutonic complex.

Argon Geochronology

Argon geochronology was conducted on biotite, muscovite, and hornblende
collected from three separate samples from the nose area of the Breakenridge antiform
(Figure 5.1). All three samples were collected in close proximity to each other to represent
a shared cooling history between sites. Sample HBL-180 was collected from plagioclase-
hornblende schist for analysis on hornblende. Sample OGN-180, from the Breakenridge
orthogneiss, was collected for analysis on biotite. Sample M-180, from a layer of garnet-
muscovite schist, was collected for analysis on muscovite.

Mineral separation was conducted by the author using facilities at Western
Washington University. Following grinding in a disk-mill, samples were sieved into three
grain-size fractions. The fraction containing the largest, non-composite grains of the target
mineral was chosen for separation (250-180 micron fraction for hornblende and biotite;
180-150 micron fraction for muscovite). This fraction was first cleaned in water using
ultrasound. The sample was then run through a Frantz magnetic separator. Heavy liquids
were used to complete the separation process.

Results from the step-heating $^{40}$Ar/$^{39}$Ar method are shown as age spectra and
inverse isochron plots for each sample (Figures 5.3, 5.4, and 5.5). K/Ca plots, which
indicate the compositional homogeneity of the sample, are shown for reference. Data tables
are listed in Appendix C. Age spectra plots portray apparent age vs. cumulative $^{39}$Ar
released during the step-heating process. Two ages are cited. The total fusion age (TFA)
represents an average age that would be obtained if all of the gas were to be released at once
by melting the sample. The weighted mean plateau age (WMPA) takes into account only those temperature steps in the step-heating process that fall within ±1 sigma error of each other. A rigorous plateau age is obtained only if the steps within error are contiguous and correspond to 50% or more of the total 39Ar released (Fleck et al., 1977; Dalrymple and Lanphere, 1969).

The inverse isochron plot correlates 36Ar/40Ar and 39Ar/40Ar ratios for gas released during each heating step, where the reference isotope is 40Ar. The Y- and X-axis intercepts correspond to the isotopic composition of trapped and radiogenic argon, respectively. This plot is valuable in that it portrays the isotopic composition of trapped argon, and therefore allows for the identification of contamination from external sources. Atmospheric argon has a 40Ar/36Ar composition of 295.5 (Steiger and Jäger, 1977). Good correlation of heating step compositions along a line intercepting the isotopic composition of atmospheric argon suggests an absence of contamination other than atmospheric argon (McDougall and Harrison, 1988). The isochron age is determined from the X-axis intercept value.

Closure temperatures for minerals used in analysis are as follows: hornblende closure at 500°C (Harrison, 1981), muscovite closure at 350°C (Robbins, 1972; Purdy and Jäger, 1976; Jäger, 1979), and biotite closure at 325°C (Harrison et al., 1985).

Results

**HBL-180**

Analysis of hornblende defines a WMPA of 87.64 ± 0.20 Ma when 53% of the gas is used (Figure 5.3a). This result defines a true plateau, and is within ±2 sigma error of the isochron age of 89.04 ± 0.65 Ma (Figure 5.3b). When 100% of the gas is used, the WMPA is 87.77 ± 0.17 Ma. Although this does not represent a true plateau, the age is still
Figure 5.3 Argon geochronology charts for sample HBL-180 (hornblende). Shaded steps were used to obtain a true plateau. A. Age spectra. B. Inverse isochron plot. The age cited was obtained using shaded steps of the inverse isochron plot above. The isochron line itself represents 100% $^{39}$Ar in order to show correlation of the isotopic composition of contaminant argon with the composition of atmospheric argon. C. K/Ca plot.
a. HBL-180 (hornblende)
TFA = 87.64 ± 0.22 Ma
WMPA = 87.64 ± 0.20 Ma
(±1 sigma shown w/o error in J)

b. HBL-180 (hornblende)
MSWD = 0.88 (< 2.41)
Age = 89.04 ± 0.65 Ma
(±1 sigma shown)

c. HBL-180 (hornblende)
within error of the isochron age of $87.76 \pm 0.19$ Ma. The WMPA of $87.64 \pm 0.20$ Ma is interpreted to represent the age at which hornblende cooled below $500^\circ$C.

**M-180**

Analysis of muscovite defines a WMPA of $81.91 \pm 0.22$ Ma when 86% of the gas is used (Figure 5.4a). This result defines a true plateau, and is within ± 2 sigma error of the isochron age of $82.26 \pm 0.28$ Ma (Figure 5.4b). The use of 100% of the gas defines a WMPA of $82.07 \pm 0.20$ Ma. This result is within error of the isochron age of $82.41 \pm 0.26$ Ma. The WMPA of $81.91 \pm 0.22$ Ma is interpreted to represent the age at which muscovite cooled below $350^\circ$C.

**OGN-180**

Analysis of biotite defines a WMPA of $82.10 \pm 0.15$ Ma when 46% of the gas is used (Figure 5.5a). This result does not define a true plateau because less than 50% of the gas is used. Steps $950^\circ$C - $1110^\circ$C define the closest approximation of a plateau age. They are contiguous, and are within ± 1 sigma error of each other. The WMPA determined with these steps is within ± 2 sigma error of the isochron age of $80.78 \pm 1.04$ Ma (Figure 5.5b). The use of 100% of the gas defines a WMPA of $82.15 \pm 0.12$ Ma. This result is within error of the isochron age of $82.24 \pm 0.15$ Ma. The WMPA of $82.10 \pm 0.15$ Ma is regarded as the age at which biotite cooled below $325^\circ$C.

**Geological Implications**

Metamorphism of the Breakenridge orthogneiss and adjacent rocks in the area of the Ar-Ar samples occurred at temperatures of at least $600^\circ$C (Figure 3.4). Igneous intrusion of the Breakenridge plutonic suite probably occurred at temperatures greater than that.
Figure 5.4 Argon geochronology charts for sample M-180 (muscovite). A, B, and C are the same as in Figure 5.3.
a. M180 (muscovite)
TFA = 81.90 ± 0.23 Ma
WMPA = 81.91 ± 0.22 Ma
(±1 sigma shown w/o error in J)

b. M180 (muscovite)
MSWD = 0.88 (< 1.85)
Age = 82.26 ± 0.28 Ma
(±1 sigma shown)

c. M180 (muscovite)

Cumulative $^{39}$Ar
Figure 5.5 Argon geochronology charts for sample OGN-180 (biotite). A, B, and C are the same as in Figure 5.3.
OGN180 (biotite)

TFA = 82.17 ± 0.13 Ma
WMPA = 82.10 ± 0.15 Ma
(±1 sigma shown w/o error in J)

Cumulative $^{39}$Ar

OGN180 (biotite)

MSWD = 0.23 (< 2.63)
Age = 80.78 ± 1.04 Ma
(±1 sigma shown)

$^{39}$Ar/$^{40}$Ar

OGN180 (biotite)

990°C

$^{39}$Ar

Cumulative $^{39}$Ar
Therefore, the Ar-Ar cooling ages place a minimum age bracket on high-grade metamorphism in the Breakenridge gneiss dome. The hornblende age of 87.64 ± 0.20 Ma indicates that metamorphic temperatures had dropped below 500°C by this time. Muscovite and biotite cooling ages are indistinguishable when error is taken into account, and jointly define an age of 82 Ma that corresponds to cooling below 325-350°C.

Hornblende and mica cooling ages can be used to estimate an uplift rate as follows. If a vertical geothermal gradient of 50°C/km is assumed, closure temperatures of 500°C and 350°C correspond to depths of approximately 10 km and 7 km, respectively. Therefore, an uplift of 3 km occurred in approximately 5.6 million years, corresponding to an uplift rate of approximately 0.5 mm/yr. Although 50°C/km is an appropriate geothermal gradient for an active arc, the true gradient cannot be known in such a magmatically and tectonically active region, and may have ranged from 20-100°C/km.
VI. RELATIVE TIMING OF METAMORPHISM, DEFORMATION, AND UPLIFT

Introduction

The sequence of geologic events in the study area is established by simple cross-cutting relationships of lithologic units, the geometry of metamorphic index minerals as related to deformational fabrics, and geochronology. A timeline of events is shown in Figure 6.1.

Onset of Regional Deformation

The earliest deformation event in the high grade rocks east of Harrison Lake is associated with thrust stacking of the Slollicum, Cogburn, and Settler Schist units to the south and east of the Breakenridge orthogneiss (Monger, 1986). Associated with this event is a penetrative, east-dipping foliation, accompanied by a down-dip lineation, in all three units (M. Davis pers. comm. 1996). A lower age bracket for the thrust-stacking is defined by U-Pb zircon ages in the Slollicum Schist south of the study area. Bennett (1989) reported a U-Pb age of 146 Ma from a metavolcanic unit of the Slollicum Schist. The onset of deformation is therefore assumed to have occurred after 146 Ma. Thrust sheets and associated fabrics are cut by the strain-aureole of the 96 Ma Spuzzum pluton, providing a minimum age bracket for thrusting in this area (Brown and Walker, 1993; E.H. Brown pers. comm. 1996).
Figure 6.1 Chronological diagram showing temporal constraints for geologic events discussed in the text. White rectangles represent error bars for ages based on U-Pb zircon or Ar/Ar isotope analysis (black rectangles).
D$_1$ Deformation

The timing of D$_1$ in rocks west of the Breakenridge fault depends on the age of down-dip fabrics relative to initial terrane-stacking thrusts reported to the south. The igneous crystallization age determined by Parrish and Monger (1992) in metavolcanic chlorite schist west of the Big Silver Creek delta (Figure 5.1) requires that deformation occurred after 102 Ma in these rocks. The metamorphic index minerals biotite and garnet are syn-kinematic to the foliation in some areas, indicating that deformation and metamorphism occurred simultaneously. Kinematic indicators show an east-side-up sense of shear. The exact age of these fabrics is not known, however. Fabrics in this area exhibit the same geometry and kinematics as those associated with thrust-stacking in the Slollicum Schist further to the south, suggesting that they may have developed during the same deformational event. This interpretation requires that thrust-related deformation was active after 102 Ma.

The timing of D$_1$ east of the Breakenridge fault is constrained by the intrusion of the Breakenridge orthogneiss. Igneous emplacement occurred between 104 - 96 Ma, as determined from U-Pb zircon dates (Parrish and Monger, 1992; this study). Igneous textures are recrystallized and deformed by D$_1$ fabrics. D$_1$ foliation overprints injection zones, requiring concomitant deformation of both intrusive rocks and country-rock. Strike-parallel lineations formed during D$_1$. The metamorphic index mineral biotite is syn-kinematic to the foliation in the orthogneiss, requiring coeval biotite-grade metamorphism and deformation.

The timing of arc-parallel fabric development relative to arc-normal thrusting is not well-understood. Orogen-normal deformation had ceased by 96 Ma in the Slollicum Schist to the south, based on pinning of terrane-stacking thrusts by the Spuzzum pluton. There is
no evidence to suggest that orogen-normal deformation ceased after 96 Ma within the chlorite schist unit of the study area, however. It is assumed that strike-parallel motion occurred after 96 Ma, since strike-parallel fabrics deform the Breakenridge orthogneiss. Therefore, 96 Ma may represent a transition from orogen-normal to orogen-parallel kinematics. It is also possible that orogen-parallel and orogen-normal kinematics occurred simultaneously by kinematic partitioning in the crust. Partitioning is possible, since the strike-parallel fabrics occur in deeper rocks than thrust fabrics.

Post-Kinematic High-Pressure Metamorphism

High-pressure metamorphism post-dates D₁ east of the Breakenridge fault. High-pressure metamorphic index minerals include garnet and hornblende, both of which are post-kinematic to D₁ fabrics. Garnets are euhedral and cut across foliation. In addition, garnets are in places distinctly zoned with Ca-rich rims, indicating a jump in metamorphic pressures. Hornblende occurs as euhedral prisms oriented randomly in the D₁ foliation. In many cases, hornblende prisms cut across foliation planes. These observations suggest that high-pressure metamorphism occurred statically following the D₁ event. The transition from metamorphism associated with D₁ to high-pressure metamorphism may represent two separate events; or alternatively the whole process could record a single progressive metamorphism, during which peak conditions were reached after deformation ceased. The second interpretation is supported by rare occurrences of aligned hornblende prisms, suggesting that deformation was at least partly contemporaneous with high-pressure metamorphism in localized areas.

The presence of high-pressure metamorphism post-kinematic to D₁ and pre-kinematic to D₂ requires that loading occurred after, and possibly during, strike-parallel deformation.
D₂ Deformation and Uplift

The D₂ deformation event, involving development of the Breakenridge antiform and associated structures, deforms earlier D₁ structures, and therefore post-dates D₁ (see Structural Geology chapter). The highest pressure rocks are exposed in the core of the Breakenridge antiform; isobars are roughly concentric about the core of the dome, decreasing in value on all sides (E.H. Brown, unpublished). Therefore, it appears that the isobaric surfaces are folded, and consequently the Breakenridge antiform post-dates the high-pressure metamorphism. The temporal relationship of D₂ to high-pressure metamorphism is poorly constrained on the outcrop scale, however, due to the post-kinematic nature of high-pressure metamorphic minerals. Deformation after peak metamorphism is difficult to recognize, since there are no high-pressure metamorphic fabrics to deform. Folding of foliation defined by hornblende was not observed in the field.

The timing of uplift is defined by Argon geochronology conducted as part of this study. Hornblende cooled below 500°C by 87 Ma, suggesting a decline in metamorphic temperatures from their >600°C peak by this time. Cooling below 300-350°C occurred by 82 Ma, as determined from cooling ages of muscovite and biotite. If uplift is the result of D₂, then these uplift ages approximate the timing of that event.

The timing of motion along the Breakenridge fault is not well constrained. The fault appears to truncate the metamorphic gradient, suggesting that the structure assisted in uplift of the high-grade rocks during D₂. The position of the fault on the western side of the Breakenridge antiform may suggest that these two structures formed together as a fault-fold structure.
VII. DISCUSSION

The sequence of orogenic events in the vicinity of the Breakenridge orthogneiss occurred during a relatively narrow time window. Plutonism, arc volcanism, deformation, burial, high-pressure metamorphism, and initiation of uplift occurred between 104 - 87 Ma, a time-span of 17 m.y. Age constraints require temporal overlap between some events. In addition, the timing of several events remains poorly constrained. These factors make interpretation of certain aspects of the structural and metamorphic history of the area problematic.

The spatial and temporal relationship between metamorphic events on either side of the Breakenridge fault remains undetermined, because the relative ages of syn- and post-kinematic metamorphic index minerals between the two areas are not known. The occurrence of post-kinematic garnet, moderate pressures, and biotite-garnet metamorphic isograds in the chlorite schist unit suggests that the metamorphic gradient on the west side of the area represents the outer fringes of the high-pressure metamorphic event documented on the east side of the Breakenridge fault. The presence of syn-kinematic garnets in the northern part of the chlorite schist unit may indicate that metamorphism ended with deformation in lower grade regions. Post-kinematic garnets in the higher grade areas would indicate that metamorphism continued after deformation ceased. It should be noted that west of the fault, the timing of syn-kinematic garnet growth relative to peak metamorphic conditions is not known. Thermobarometry and analysis of chemical zoning in syn-kinematic garnets would be a significant contribution to solving this problem.

The cause of high-pressure metamorphism is not known. Loading of the crust must have occurred after strike-parallel motion ceased, since static high-pressure metamorphic index minerals overprint strike-parallel D1 fabrics. Joumeay and Friedman (1993) suggest that two stages of shortening occurred in the region. In their model,
volcanic arc sequences were structurally imbricated along low angle thrust faults during an early stage of shortening. These early stage structures were in turn folded and cut by out of sequence high-angle faults during late stage contraction. Although early low-angle thrusts were not observed within the study area, the presence of such structures outside the study area would allow for thrust loading as a cause for high pressure metamorphism. Late stage high-angle faulting and folding, expressed as the Breakenridge fault, Central Coast Belt detachment, and Breakenridge antiform, would be responsible for emplacement of the high-pressure rocks at shallower structural levels.

Alternatively, high-pressure metamorphism may have been caused by magma-loading. In this model, batholiths intruding at intermediate to shallow crustal levels balloon and load the underlying country rock (Brown and Walker, 1993). The remnants of plutons of appropriate age for magma-loading (post 96 Ma - pre 87 Ma) may be exposed as the Urquhart and Mt. Mason plutons (Figure 1.2).

The spatial and temporal relationship of orogen-normal and orogen-parallel kinematics depends on the relative timing of fabric development between the western and eastern lineation domains. Assuming that down-dip fabrics are related to terrane-stacking thrusts requires that thrusting in the western lineation domain occurred after 102 Ma, and therefore developed at the same time the Breakenridge orthogneiss was intruding (104-96 Ma).

It is also possible that thrusting continued after 96 Ma, during which the terrane-stacking thrusts to the south were inactive. This interpretation would allow for down-dip fabric development in the western lineation domain after 96 Ma as part of a continuous thrusting event that was coeval with strike-parallel fabrics in the eastern lineation domain. It would also require that strain partitioning in the crust was active during this time. This model is consistent with the lack of overprint of one lineation domain on the other. The two domains were metamorphosed at different structural levels, based on the steep pressure
gradient between them, making kinematic partitioning in the crust a viable model. A continuous orogen-normal contractional event would allow for the crustal loading necessary for high-pressure metamorphism after strike-parallel fabric development. It also allows for the late-stage crustal shortening required for D$_2$, during which the Breakenridge antiform and the Breakenridge fault were active.

Two tectonic models, both of which incorporate strain partitioning, are suggested to explain the orogenic events that have affected the Breakenridge orthogneiss. Transpression may have resulted in development of an orogen-scale positive flower structure in which orogen-parallel and orogen-normal kinematics are partitioned (Figure 7.1a). In this model, the strike-slip component of transpression is accommodated by ductile flattening and strike-slip flow along vertical shear zones in deep rocks such as the Breakenridge orthogneiss. The compressional component is accommodated by upward flow along orogen-normal thrusts and associated fabrics, such as those in the western lineation domain of this study. This model does not explain the geometry of fabric development in the Breakenridge orthogneiss, however. Because foliation formed sub-parallel to the flat-lying sills of orthogneiss, development of vertical strike-slip shear zones in the orthogneiss is problematic. A second model, in which a detachment partitions strain between shallow orogen-normal thrusts and deep, flat-lying orogen-parallel shear zones, better explains the fabric geometries in the Breakenridge orthogneiss. Final juxtaposition of the two fabric domains occurs along the Breakenridge fault during or after D$_2$. Both of these models are consistent with structural data elsewhere in the Cascades of Washington, where both orogen-parallel and orogen-normal structures formed in different provinces of the orogen at approximately the same time (Brandon et al., 1988; Brown, 1987; Brown and Talbot, 1989; McGroder, 1989).

Alternatively, orogen-normal and orogen-parallel kinematics may have occurred separately at different times within the field area. In this model, D$_1$ thrusting ceases, and is
Figure 7.1 Transpressional magmatic arc models, where strain is partitioned in the crust. 
A. An orogen-scale positive flower structure. B. A sub-horizontal detachment model.
immediately followed by strike-parallel movement at around 96 Ma. Strike-parallel movement ceases prior to $D_2$, which represents a return to convergent tectonics.

Plate motion reconstructions of the Pacific Basin (Kelley, 1993) indicate that the Farallon plate maintained southeast-directed oblique convergence relative to the North American plate between 118-95 Ma, at the latitude of the present day Olympic Peninsula. At 95 Ma, the obliquity decreased during a change in relative plate motion, but convergence remained southeast-directed until 83 Ma. Therefore, sinistral oblique plate convergence was active during the deformation events recorded in the area of study. Transpression, resulting from oblique plate convergence, may have been expressed by the coeval development of contractional and strike-slip deformational structures in the orogenic belt, and is therefore supportive of the strain partitioning models presented above. The decrease in obliquity at 95 Ma may be responsible for the late stage shortening that formed the Breakenridge antiform.
VIII. CONCLUSIONS

The conclusions of this study are as follows:

(1) The Breakenridge orthogneiss intruded metavolcanic rocks of the Jurassic-Cretaceous Sollucum Schist between 104-96 Ma. Intrusive contacts along the margins of the orthogneiss indicate that it was not emplaced along faults as mapped previously. Intrusive emplacement occurred as a sheeted-sill complex, based on the apparent composite sill structure of the pluton.

(2) An initial deformational event ($D_1$), accompanied by medium to high-pressure Barrovian metamorphism, occurred after 96 Ma. Metamorphic grade ranges from greenschist facies (biotite zone) to amphibolite facies (kyanite zone). Protolith lithologies are recrystallized and overprinted by solid-state, penetrative foliation and lineation. Post-102 Ma down-dip lineations on the lower grade, western flank of metamorphic gradient are attributed to orogen-normal thrusting. Post-96 Ma strike-parallel lineations in higher grade gneisses suggest orogen-parallel strike-slip deformation. Fabrics in higher-grade rocks are overprinted by post-kinematic high-pressure metamorphic index minerals, suggesting that metamorphism continued after deformation ceased in the highest grade areas.

(3) A second deformation event records folding of the orthogneiss sill-complex and $D_1$ fabrics in the Breakenridge antiform. The doubly-plunging geometry of the antiform, accompanied by small-scale $D_2$ folds, suggest that the structure developed as a regional-scale incipient sheath fold that resulted in upward translation of deep-seated, high-grade metamorphic rocks in its core. The large scale folding of $D_2$ is attributed to continued arc-normal shortening in the orogen.

(4) $^{40}$Ar/$^{39}$Ar isotope geochronology indicates that high grade metamorphism ended by 87 Ma. Uplift and cooling, which may have been partly contemporaneous with $D_2$, occurred between 87-82 Ma.
(5) Although not seen in the field, the presence of the Breakenridge fault is inferred based on an abrupt change in metamorphic pressure, truncation of metamorphic isograds, and the abrupt change in lineation orientation on the western side of the field area. The structure is best described as a post-metamorphic distributed shear zone that may have developed during or after formation of the Breakenridge antiform.

(6) The spatial and temporal relationship of down-dip and strike-parallel lineations may be explained by a transpressional tectonic setting. Orogen-scale positive flower structure and detachment partitioning models are proposed.
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APPENDIX A. MINERAL ASSEMBLAGES IN THIN SECTION

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Qtz = quartz; Plag = plagioclase; Biot = biotite; Musc = muscovite; Chl = chlorite; Epi = Epidote; Cz = Chinozoisite; Hbl = hornblende; Gt = garnet; Marg = margarite; Op = opaque minerals; Sph = sphen; Rut = rutile; Tour = tourmaline; Gr = graphite; Calc = calcite
APPENDIX B. MINERAL COMPOSITIONS USED IN THERMOBAROMETRY (in formula proportions)*

Sample 180-191

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Sample 180-192

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* analyses by E.H. Brown
APPENDIX C. GEOCHRONOLOGY DATA

I. U/Pb Zircon Data and Analytical Methods for Sample 180GN*

<table>
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<tr>
<th>Fraction-size* (μm)</th>
<th>Wt (mg)</th>
<th>Concentrationb (ppm)</th>
<th>Concentrationc (ppm)</th>
<th>Isotopic compositiond</th>
<th>Apparent Ages (Ma)e</th>
<th>Th-corrected Ages (Ma)e</th>
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<tr>
<td></td>
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<td></td>
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<td></td>
<td>206Pb 206Pb 206Pb</td>
<td>206Pb* 207Pb* 207Pb*</td>
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<tr>
<td></td>
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<td></td>
<td></td>
<td></td>
<td>238U 235U</td>
<td>238U 235U</td>
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<tr>
<td>a 63-80</td>
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<td>148</td>
<td>2.4</td>
<td>1753</td>
<td>17.679 7.623</td>
<td>103.7 103.9 ± 0.3 108</td>
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<tr>
<td>b 63-80</td>
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<td>136</td>
<td>2.2</td>
<td>1461</td>
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<td>c 80-100</td>
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<td>130</td>
<td>2.1</td>
<td>1659</td>
<td>17.40 7.770</td>
<td>104.6 105.6 ± 0.3 128</td>
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<tr>
<td>d 100-350</td>
<td>0.2</td>
<td>101</td>
<td>1.7</td>
<td>990</td>
<td>15.582 6.900</td>
<td>106.9 109.4 ± 0.3 165</td>
</tr>
</tbody>
</table>

*  a, b, etc. designate conventional fractions; All zircon fractions are non-magnetic on Frantz magnetic separator at 1.8 amps, 15° forward slope, and side-slope of 2°.
  b Pb* is radiogenic Pb.
  c Reported ratios corrected for fractionation (0.125 ± 0.038%/AMU) and spike Pb. Ratios used in age calculation were adjusted for 10-20 pg of blank Pb with isotopic composition of 206Pb/204Pb = 18.6, 205Pb/204Pb = 15.5, and 208Pb/204Pb = 38.4, 2 pg of blank U, 0.25 ± 0.049%/AMU fractionation for UO₂, and initial common Pb with isotopic composition approximated from Stacey and Kramers (1975) and assigned uncertainty of 0.1 to initial 206Pb/204Pb.
  d Uncertainties reported as 2 sigma. Decay constants: 238U = 1.5513 E-10, 235U = 9.8485 E-10; 238U/235U = 137.88.
  e (Analysis by Dr. W.C. McClelland at the University of California, Santa Barbara, CA.)
II. $^{40}\text{Ar}/^{39}\text{Ar}$ GEOCHRONOLOGY DATA*

Sample: HBL-180 Hornblende  $J=0.0081809$

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<th>$T(\degree C)$</th>
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<th>$40/39$</th>
<th>$37/39$</th>
<th>$36/39$</th>
<th>$\text{K/Ca}$</th>
<th>$\Sigma 39\text{Ar}$</th>
<th>$40\text{Ar}^*$</th>
<th>Age (Ma)</th>
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<td>97.5 ± 21.3</td>
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<tr>
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<td>4.3139</td>
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<td>0.266</td>
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<td>4.7143</td>
<td>0.0262</td>
<td>0.10</td>
<td>0.005</td>
<td>0.266</td>
<td>94.9 ± 12.8</td>
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<td>6.6826</td>
<td>0.0087</td>
<td>0.073</td>
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<td>0.266</td>
<td>94.9 ± 12.8</td>
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Total fusion age, $\text{TFA}= 87.64 \pm 0.22 \text{ Ma (including J)}$
Weighted mean plateau age, $\text{WMFA} = 87.64 \pm 0.20 \text{ Ma (including J)}$
Inverse isochron age = $89.04 \pm 0.65 \text{ Ma. (MSWD} =0.88)$
$40\text{Ar}/36\text{Ar} = 297.0 \pm 4.2 \text{ (100\% } \Sigma 39 \text{ Ar)}$
Steps used: 1040, 1060, 1080, 1100, 1120, 1140 (53\% $\Sigma 39\text{Ar}$)
$40(\text{mol})$ = moles corrected for blank and reactor-produced $40$.
$\Sigma 39\text{Ar}$ is cumulative, $40\text{Ar}^*$ = rad fraction.
* (Analysis by A. Calvert at the University of California, Santa Barbara, CA.)
Sample: M180 Muscovite J=0.0070653

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<th>T</th>
<th>40(mol)</th>
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<th>37/39</th>
<th>36/39</th>
<th>K/Ca</th>
<th>Σ 39Ar</th>
<th>40Ar*</th>
<th>Age (Ma)</th>
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Total fusion age, TFA= 81.90 ± 0.23 Ma (including J)
Weighted mean plateau age, WMPA= 81.91 ± 0.22 Ma (including J)
Inverse isochron age =82.26 ± 0.28 Ma. (MSWD =0.88)

40Ar/36Ar=241.5 ± 55.0 (100% Σ 39Ar).
Steps used: 550, 650, 750, 785, 815, 850, 875, 900, 925, 950, 990, 1035, 1070 (86% Σ 39Ar)
40(mol) = moles corrected for blank and reactor-produced 40.
Σ 39Ar is cumulative, 40Ar* = rad fraction.
### Sample: OGN180 Biotite J=0.0070488

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<th>T</th>
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<th>40/39</th>
<th>37/39</th>
<th>36/39</th>
<th>K/Ca</th>
<th>Σ 39Ar</th>
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</tbody>
</table>

Total fusion age, TFA= 82.17 ± 0.13 Ma (including J)

Weighted mean plateau age, WMPA= 82.10 ± 0.15 Ma (including J)

Inverse isochron age =80.78 ± 1.04 Ma. (MSWD =0.23)

40Ar/36Ar=507.4 ± 87.5 (100% Σ 39 Ar)

Steps used: 950, 990, 1035, 1070, 1110 (46% Σ 39Ar)

40(mol) = moles corrected for blank and reactor-produced 40.
Σ 39Ar is cumulative, 40Ar* = rad fraction.