GEOLOGY AND GEOCHEMISTRY OF THE OCEANIC ARC-CONTINENT BOUNDARY IN THE WESTERN IDAHO BATHOLITH NEAR MCCALL

Thesis by

Cathryn Clement Allen Manduca

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ABSTRACT

A major lithospheric boundary is preserved within the western Idaho Batholith. The juxtaposition of two suites of supracrustal rocks, exposed as sheets within intrusive rocks, is the expression of this boundary at the level of exposure. The western suite of mafic layered gneisses are inferred to be metamorphosed oceanic arc rocks; the eastern suite of biotite schist, quartzite and calc-silicate gneiss are inferred to be metamorphosed continental sedimentary rocks. Three broadly Cretaceous, plutonic and meta-plutonic complexes record the presence of the boundary at greater depth. The Hazard Creek complex, west of the supracrustal boundary, is comprised of epidote-bearing intrusives. The Little Goose Creek complex is comprised primarily of the porphyritic orthogneiss that intruded the supracrustal boundary. The Payette River complex, east of the supracrustal boundary, is comprised of large bodies of tonalite and granite.

Each complex has a distinct geochemical character. The Hazard Creek complex is dominantly a tonalite-trondhjemite suite characterized by R_i less than .7045, $\delta^{18}O$ less than 8.4, high Sr, Na_2O and Al_2O_3 concentrations and low MgO, Rb and K_2O concentrations. Porphyritic orthogneiss in the Little Goose Creek complex has a remarkable range in Ri and $\delta^{18}O$ (.7042-.7097,8.0-10.7). The porphyritic orthogneiss is interpreted as dominated by two components: one similar in composition to the Hazard Creek complex and a second, modeled as Precambrian sedimentary material, with high R_i, $\delta^{18}O$ and K_2O concentrations and lower Sr concentration. The Payette River complex has generally high Ri (.7076-.7100) and variable $\delta^{18}O$ (7.2-10.4). The geochemical changes indicate that the supracrustal boundary is the surface expression of a steeply-dipping structure which juxtaposes oceanic-arc lithosphere against continental lithosphere. An abrupt geochemical discontinuity preserved within the porphyritic orthogneiss, near the change in supracrustal rocks, may reflect an

abrupt discontinuity at depth or may be due to the juxtapositon of portions of a stratified pluton. The juxtaposition of lithospheric blocks must have occured prior to intrusion of the porphyritic orthogneiss approximately 111 Ma, and most probably, occured before 118 Ma, prior to the beginning of plutonism. No structural evidence for the initial formation of the boundary is recognized; it is proposed to form by transform faulting or by rifting followed by convergence.

Episodic or continuous deformation along the boundary began prior to 118 Ma and produced four sets of structures. The oldest structures are foliation and isoclinal folding of crystalloblastic gneisses which may have formed during rapid burial of oceanic-arc rocks west of the boundary. Compressive deformation, forming north-south striking steeply-dipping foliations and steeply-plunging lineations in the eastern portion of the Hazard Creek complex, was broadly coeval with its emplacement. Igneous foliation and lineation with similar orientation formed during emplacement of the Payette River complex around 90 Ma. The youngest penetrative deformation formed similarly oriented, mylonitic fabrics in a 10 km wide zone centered on the boundary. All but the oldest structures are inferred to have formed by flattening and vertical flow in response to east-west compression. Deformation is interpreted to represent the response of a preexisting lithospheric boundary to compressive stresses related to subduction of material to the west.

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CHAPTER 1: INTRODUCTION

1.1 OVERVIEW

1.1.1 Statement of Problem

This study examines a portion of the western border of the Idaho Batholith (Figure 1.1) where it intrudes the regional boundary observed between metamorphosed, cratonally derived, sedimentary rocks and metamorphosed oceanic arc rock assemblages. The Idaho Batholith underlies most of central Idaho and is one of a string of great Cretaceous batholiths in the western Cordillera believed to have been produced along the western margin of the North American Plate during an extended interval of plate convergence. The batholith is divided into two lobes, the smaller, northern Bitterroot lobe and the southern Atlanta lobe, by an area of abundant wall rocks often referred to as the Salmon River Arch (Armstrong, 1975). The interior of both lobes is comprised of large masses of two-mica granite and biotite granodioite (Lewis et al., 1987; Hyndman, 1983; Toth, 1987). This core is surrounded by a somewhat older margin comprised primarily of tonalite.

The location of the boundary between late Proterozoic and Paleozoic (?) miogeoclinal sedimentary rocks and the Blue Mountain arc assemblage can now be identified by a systematic change in the lithology of pendants within the western tonalitic margin of the Idaho Batholith in a distance of 5 km. This contact differs from the boundary between miogeoclinal and eugeoclinal rocks observed along much of the length of the Cordillera in that the Paleozoic miogeosynclinal section is either missing entirely or only partially preserved. The change from shallow water cratonal sedimentary sequences to oceanic-arc material is abrupt. Neither transitional sedimentary

Figure 1.1 Regional geologic map of the Idaho Batholith and its wall rocks in Idaho, eastern Oregon, southeastern Washington and western Montana showing study area. Compiled from Bond, 1978; Ross et. al., 1955; Lewis et. al.,1987; Walker, 1977; Lund, 1984.

Cretaceous Idaho Batholith

Blue Mountain Arc

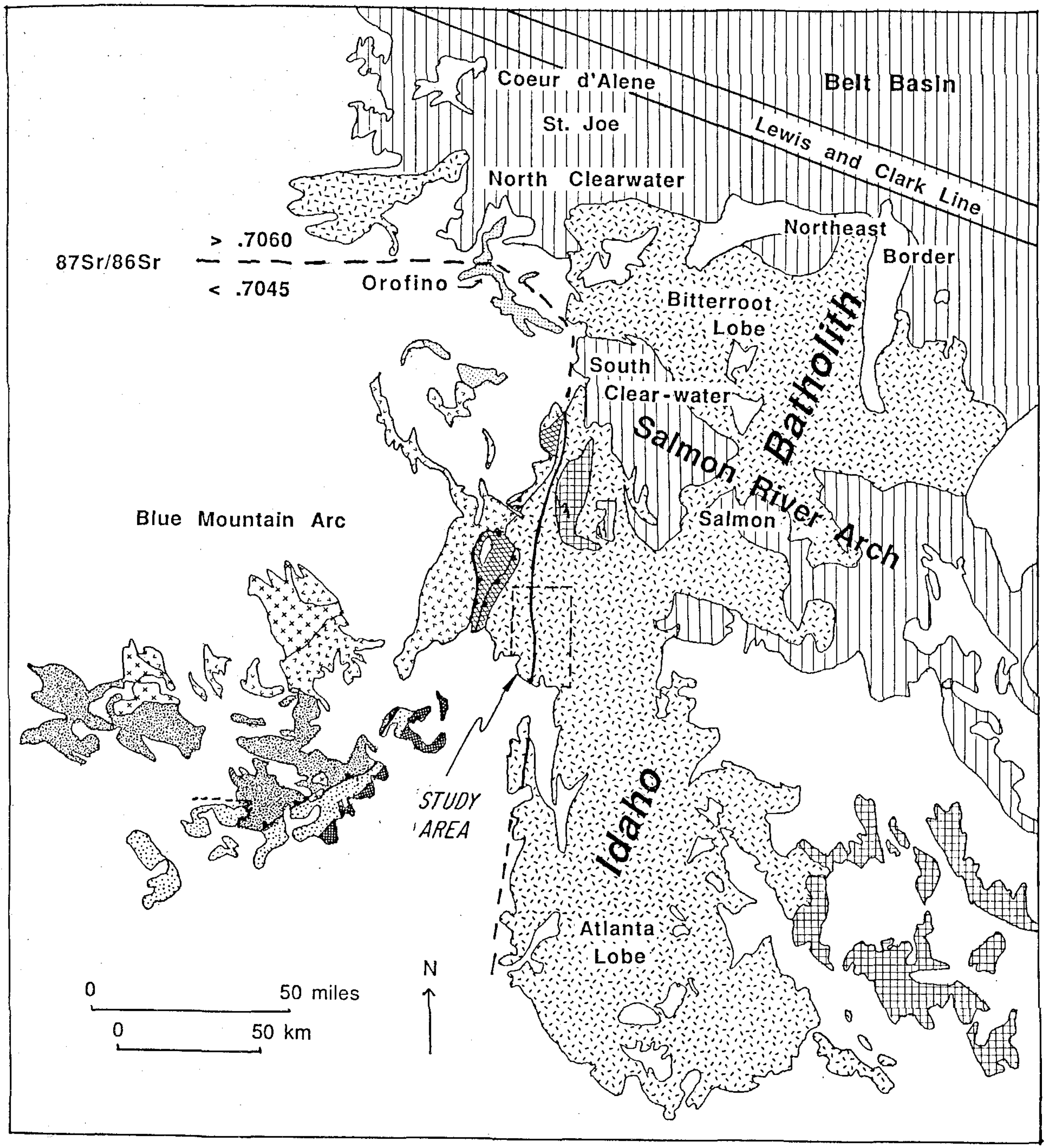
- Huntington Arc
- Wallowa/Seven Devils Arc
- Central Melange Terrane
- Mesozoic Clastic Assemblage
- Riggins Group
- Jurassic Plutons

Post Batholithic Rocks

Cratonally Derived Rocks

- Precambrian sediments and gneisses including the Belt/Purcell and Salmon supergroups and their metamorphic equivalents and possible basement gneisses.
- Paleozoic and Precambrian Z (?) sediments and their metamorphic equivalents.
- Rocks of continental or oceanic arc affinity and plutons of unknown age

Idaho Batholith and Wall Rocks



assemblages nor a large subduction complex has been recognized (Lund, 1984). This paleogeographic anomaly suggests that the boundary has been tectonically modified.

The abrupt change in the lithology of pendants in the Cretaceous batholith coincides regionally with a dramatic change in the Sr and O isotopic systematics of the intrusive rocks (Armstrong et al., 1977; Fleck and Criss, 1985). Intrusives with oceanic-arc wall rocks have $^{87}Sr/^{86}Sr$ (R) less than .7045; intrusives with continental wall rocks have R greater than .707 and often closer to .710. A change in R of this magnitude across such a narrow zone is unknown elsewhere in the Cordillera. A similar though less dramatic shift is observed in the $\delta^{18}O$ of the intrusives.

These marked discontinuites have led to the suggestion that the continent was truncated in this area by rifting or strike-slip faulting (Davis et al., 1978; Lund, 1984; Hamilton, 1978). The juxtaposition of accreted arc material against the continent has been interpreted to have occurred either during or after truncation. A zone of Cretaceous deformation localized along the arc-continent boundary has been inferred to have developed during accretion.

Many fundamental properties of this boundary have not been well characterized. The boundary has been carefully mapped only locally and its geologic history is still not fully analyzed. While many interpretations have been offered, no conclusive evidence for the mechanisms or timing of initial juxtaposition of arc material against continent has been put forward. The deformational character and history of the boundary and its relationship to plutonism are still being explored. In many places, intrusion and deformation of the plutons is believed to postdate deformation of wall rocks. The role of various deformational phases in the history of the arc-continent boundary is still the subject of much debate.

The fundamental relationship between the boundary in supracrustal rocks and variations in the geochemistry of the intrusive units of the batholith is also not well understood. The change in pendant lithology is known to coincide regionally with the change in strontium and oxygen isotopic systematics in the intrusives, perhaps implying a steep lithospheric boundary. However, in detail, it is known that changes in intrusive rock chemistry do not coincide exactly with changes in wall rock lithology. The distribution of plutonic units with respect to changes in isotopic systematics is not known. Rare earth element patterns are known to change from west to east across the area; however, the changes in many other important chemical parameters and their relationship to the changes in isotopic systematics have not been investigated in detail. The lack of these data has hindered the understanding of the tectonic and petrogenetic history of the area.

1.1.2 Organization of Study

The study area, located in west-central Idaho north of McCall (Figure 1.1), was selected to investigate this major boundary. It lies along the western side of the Idaho Batholith extending from the tonalitic margin into the granitic interior. The study area is in the southern Atlanta lobe.

Field and petrographic observations, Sr and O isotopic data, and major and trace element geochemical data are combined to address two major questions. 1) What is the nature and tectonic history of the boundary between arc and continent? 2) What can properties of the plutonic rocks tell us about the sources and generation of batholithic magmas and their role in the evolution of continental crust? The major goals of the study were to characterize the distribution, geochemistry and deformation of the plutonic rocks and to study the relationship between the change in wall rocks and the geochemistry of the intrusives. The results appear to place several new and important constraints on the nature of the boundary between arc and continent, the time and conditions of its formation and its history since formation.

The work is presented in five chapters. The remainder of this chapter summarizes the previous work in the study area, describes the regional geologic setting and discusses previous models put forth for the formation of the arc-continent boundary. Chapter Two describes the field relationships and petrography of the map units defined in the study area. This section addresses several key questions. 1) What is the relative age and distribution of plutonic rocks around the boundary between arc-related and continentally derived material? 2) Are rocks west of this boundary part of the Cretaceous batholith or are they older intrusives associated with the arc terrane? 3) Did some or all of the plutonic rocks intrude the arc-continent boundary? Evidence for a long intrusive and deformational history is presented. The arc-continent boundary was intruded in late Early Cretaceous time.

Chapter Three describes observations on structural features and models for the deformational history of the area. It addresses several key questions. 1) What structural evidence exists for formation of the arc-continent boundary? 2) Has it been substantially modified by younger deformation? 3) What is the relationship between deformation and plutonism? 4) Is there evidence of strike-slip motion during the Cretaceous deformation or at any other time? Penetrative deformation that produced north-south striking, steeply-dipping foliations, and down-dip lineations and fold axes along the arc-continent boundary is described. No evidence for strike-slip deformation was observed. Deformation and plutonism occurred simultaneously for much of the history of the area; however, deformation along the boundary outlasted plutonism. The localization of deformation in a narrow zone along the arc-continent boundary is inferred to suggest that the boundary formed prior to this deformation.

Chapter Four describes the geochemistry of the plutonic rocks, and addresses three key questions. 1) How do the Sr and O isotopes vary in detail and how does this variation correlate with changes in plutonic units and wall rocks? 2) What other geochemical parameters correlate with these changes? 3) What constraints can be placed on the origins of these geochemical changes and what do they tell us about the nature and distribution of the source regions of the plutonic rocks? Geochemical constraints confirm that the arc-continent boundary predates the emplacement of late Early Cretaceous plutonic rocks and suggests that this boundary is steep and extends to lower crustal or lithospheric depths.

The study concludes in Chapter Five with an interpretation of the geologic history of the area, a discussion of the relationship between this history and that observed elsewhere along the oceanic arc-continent boundary, and the suggestion of a preferred tectonic model for the development and deformation of the boundary.

1.2 REGIONAL SETTING

In order to evaluate the history of the arc-continent boundary it is necessary to understand the histories of the rocks on either side. In the following section these histories are described. The analysis begins with the oceanic assemblages of the Permian to Jurassic Blue Mountain Arc. These rocks crop out to the west of the boundary and extend into eastern Oregon (Figure 1.1). It is generally accepted that they have experienced two major compressional events, a long lasting event in the Permian to Triassic, and a brief event in the Jurassic. These assemblages may have formed as a fringing arc to North America, or may represent far travelled exotic fragments. Next, the rocks of the continent that crop out east of the boundary are described. These are primarily shallow water sedimentary sequences of Precambrian to Paleozoic age that occur within and to the east of the Idaho Batholith. The distribution and age of the underlying crystalline basement material is poorly constrained. The continental rocks have a longer history, which includes local evidence for deformation in Precambrian, Cambrian, Ordovician and Late Devonian to Early Mississipian times. Finally, evidence is put forth for a major Cretaceous deformation that affected both the eastern Blue Mountain Arc and the continental rocks .

1.2.1 Oceanic Rocks: The Blue Mountain Arc

The Idaho Batholith is bordered to the west by Permian to Jurassic rocks of oceanic and oceanic-arc affinity. These are exposed in inselburgs within the Tertiary Columbia River Basalts in the Blue Mountains and are known collectively as the Blue Mountain Arc (Pessagno and Blome, 1986) or the Blue Mountain terrane (Vallier and Brooks, 1986; Figure 1.1). The Blue Mountain Arc has been subdivided into four lithotectonic assemblages: a central melange terrane, a Mesozoic clastic assemblage, and two arc assemblages, the Huntington arc and the Wallowa/Seven Devils arc (Vallier et al., 1977; Brooks and Vallier, 1978; Dickinson and Thayer, 1978; Brooks, 1979).

The Wallowa/Seven Devils and Huntington arcs are large volcanic and volcaniclastic piles. The Wallowa/Seven Devils arc formed during Permian and Triassic time (Vallier, 1977). As arc activity waned, limestone (Martin Bridge Formation) and shale (Hurwal or Lucile Formation) were deposited on top of the arc (Hamilton, 1963a; Brooks and Vallier, 1978). Volcanism in the Huntington arc did not begin until Triassic time and continued through the time of deposition of Martin Bridge and Hurwal Formations (Brooks and Vallier, 1978).

The central melange terrane is formed primarily of chert argillite melange, which was deposited during Permian and Triassic time (Dickinson and Thayer, 1978; Brooks and Vallier, 1978; Morris, 1986). Subduction related melange formation, which incorporated blocks of limestone, volcanic and ultramafic rock, occurred throughout this interval (Dickinson and Thayer, 1978; Ave Lallemant et al., 1980; Morris and Wardlaw, 1986) and continued into early Jurassic time (Dickinson and Thayer, 1978). The strata are interpreted as forming in fore-arc (Dickinson, 1979; Mullen, 1985) or intra-arc (Ave Lallemant et al., 1980) settings. Triassic blueschists (K-Ar, Hotz et al., 1977) in the westernmost portion of the central melange terrane further suggest deformation was related to subduction. While it is generally held that the central melange terrane and Huntington arc formed adjacent to one another (Dickinson and Thayer, 1978), the Wallowa/Seven Devils arc need not have been adjacent to the other assemblages at this time.

During Jurassic time flysch containing a large component of arc-derived detritus was deposited forming the mesozoic clastic assemblage (Dickinson and Thayer, 1978; Brooks and Vallier, 1978). Similar age flysch is locally preserved overlying the Wallowa/Seven Devils arc (Brooks and Vallier, 1978; Pessagno and Blome, 1986; Vallier et al., 1987). Flysch deposition ceased abruptly in the Middle Callovian (Imlay, 1986) as the basin was closed in a short lived deformational event in the late Jurassic (Brooks and Vallier, 1978). This deformation affected all members of the Blue Mountain Arc. Deformation was strongest in the east; however, south and southeast vergent folds and thrusts developed throughout the arc (Ave Lallemant et al., 1980). This deformation marks the final amalgamation of the arc. Undeformed calc-alkaline plutons intruded all four terranes and the boundaries between them in the late Jurassic to middle Cretaceous (Walker, 1986; Brooks and Vallier 1978). Paleomagnetic data suggest that the Blue Mountain Arc has experienced 65^{o} of clockwise rotation since the emplacement of the Jurassic plutons (Hillhouse et al., 1982; Wilson and Cox, 1980). Fifteen degrees of this rotation occurred after the extrusion of the Columbia River Basalts (summarized in Wilson and Cox, 1980).

Two lines of evidence suggest that the Blue Mountain Arc (possibly excluding the Wallowa/Seven Devils arc) formed as a fringing arc to North America. The Grindstone terrane in the westernmost central melange terrane contains upper Paleozoic strata and fossils which are very similar to those found to the south in the McCloud arc (Miller, 1988). The McCloud arc contains both faunal and stratigraphic ties to North America (Miller, 1988). Second, the pattern of deformation in both the late Triassic-early Jurassic and the late Jurassic is similar to that seen in the McCloud arc in the northern Sierra (Oldow et al., 1984). Paleomagnetic data allow this possibility (Hillhouse et al., 1982). No detritus of continental origin is recognized in the rocks of the Blue Mountain Arc.

The Wallowa/Seven Devils arc was suggested by Jones et al. (1977) to be part of the Wrangellia terrane, a sequence of Triassic lavas overlain by Norian limestone which crops out in British Columbia and Alaska. This idea has been both promoted and rebutted based on the chemistry of the lavas (Sarewitz, 1983; Mortimer, 1986; Vallier, 1986) and the similarity of the stratigraphic sections (Jones et al., 1977; Stanley, 1986; Pessagno and Blome, 1986).

1.2.1.a Riggins Group

The easternmost assemblage of rocks with oceanic arc affinity is a somewhat enigmatic assemblage known as the Riggins Group (Hamilton, 1963a, 1969; Lund, 1984; Lund and Snee, 1988; Myers, 1982; Onasch, 1987). These rocks always appear between the lower grade assemblages of the Wallowa/Seven Devils arc and the continental rocks (Figure 1.1). They are thrust over the Wallowa/Seven Devils arc in western Idaho and are intruded by batholithic rocks to the east. The Riggins Group is comprised of volcanic and volcaniclastic rocks and associated sedimentary rocks which have been metamorphosed to greenschist and amphibolite facies. The lower Riggins Group (Fiddle Creek Schist and Lightning Creek Schist), comprised primarily of tuffs and flows, is separated from the upper Riggins Group (Squaw Creek Schist) by a zone of intermittent shearing which contains pods of amphibolite inferred to be metamorphosed mafic and ultramafic rocks (Hamilton, 1963a). The upper Riggins Group is comprised of metamorphosed flysch and does not contain abundant volcanic material (Hamilton, 1963a). The metamorphism and deformation of the Riggins group and its emplacement over the Wallowa/Seven Devils arc occurred during the Cretaceous (Lund and Snee, 1988). Evidence for this deformation is preserved only in a narrow zone in western Idaho (Hamilton, 1963a, 1969; Onasch, 1987; Lund, 1984; Myers, 1982).

The Riggins group has been correlated with other terranes in the Blue Mountain Arc in a variety of ways. Brooks and Vallier (1978) correlated the lower Riggins Group with the central melange terrane and the Squaw Creek Schist with the Mesozoic clastic terrane. Lund (1984) correlated the entire Riggins group with the Seven Devils arc; however, Squaw Creek Schist is apparently absent from her study area. Sm-Nd systematics suggest that the Pollock Mountain Amphibolite is comprised of metamorphosed Triassic arc rocks (Aliberti et al., 1987) which are contemporaneous with the upper Wallowa/Seven Devils arc or lower Huntington arc. All workers agree that the Riggins group and amphibolite facies rocks are of oceanic affinity and that they may not be correlative with any Blue Mountain strata.

1.2.2 Continental Rocks

The cratonally derived material that occurs within the western Idaho Batholith has been metamorphosed to amphibolite facies. The origin and age of this material is the subject of some controversy and the rocks have been correlated with all of the major cratonal sequences which surround the batholith to the north and east. In the following section these cratonal rocks and their geologic history are described. The description is important for two reasons. First, in order to evaluate what deformation may be related to the formation or deformation of the arc-continent boundary, it is necessary to know which deformation may be related to earlier tectonics. Second, the geochemical nature of the crustal section intruded by the batholith is deduced from the information contained in the wall rocks. In this context, it is important to know what basement material may be present and the range of possiblities for overlying strata.

The nature of the basement material is poorly constrained. The only known crystalline basement material occurs to the southeast of the Batholith where Archean gneisses occur in the cores of the Pioneer and Albion ranges (Armstrong and Hills, 1967; Dover, 1983). The boundary between Archean and Proterozoic basement is exposed in block uplifts of southern Montana and is striking toward southern Idaho. The Idaho Batholith thus might be expected to have intruded Proterozoic crust. While no exposures of this crust have been documented in Idaho, several occurrences of paragneiss, schist and orthogneiss have been proposed to be Proterozoic crystalline basement. These occurrences are in the North Clearwater, South Clearwater and Salmon areas.

Cratonally derived sedimentary rocks can be divided into three groups. To the north and northeast the batholith intrudes Proterozoic sedimentary rocks of the Belt-Purcell supergroup. Within the Salmon River Arch, and continuing to the southeast, are variably deformed and metamorphosed sedimentary rocks of the Salmon supergroup (Evans, 1981). Finally, to the east and southeast, the batholith intrudes Paleozoic miogeoclinal sedimentary strata.

The Proterozoic Belt-Purcell supergroup is exposed in a large area roughly extending from Spokane, Washington to Helena, Montana and north to the latitude of Banff in the Canadian Rockies. Approximately 70,000 feet of fine-grained monotonous sediment are preserved, most of which have experienced only low-grade metamorphism (Harrison, 1972). In Canada these strata are thought to have been deposited in a miogeocline, while in the United States they are interpreted as deposits formed in a marine embayment on continental crust (Harrison et al., 1974). The lithologies are primarily quartz-rich argillites and siltites. Carbonate-rich material is abundant in the middle Belt carbonate unit. Belt-Purcell deposition is generally held to have occurred between 1450 and 850 Ma (Harrison, 1972; Obradovich and Peterman, 1968).

The Salmon supergroup (Evans, 1981), in south-central Idaho, is found in a series of thrust slices which have been transported eastward (Ruppel, 1975). The Salmon supergroup consists of more than 30,000 feet of quartz-rich sedimentary rocks

(Ruppel, 1975). Much of this material may be correlative with the Belt-Purcell supergroup; some may be older basement material.

Paleozoic strata crop out to the southeast of the Idaho Batholith in a series of thrust slices (summarized by Skipp, 1981). Cambrian through Silurian strata are dominated by carbonates and quartzites with less predominate shales and argillites. Devonian through Permian strata are dominatly argillaceous. Sedimentation is not recorded from late Permian to latest Jurassic time with the exception of two occurrences of lower Triassic beds (Skipp et al., 1979). Latest Jurassic to early Cretaceous shales, silts and sands (Scholten, 1973) are overlain by synorogenic gravels and conglomerates deposited from Albian to early Eocene time (Ryder and Scholten, 1973; Lowell and Klepper, 1953).

1.2.2.a Deformation of the Cratonal Rocks

The Precambrian and Paleozoic sections record evidence of several tectonic disturbances. Precambrian rocks of the Belt and Salmon supergroup are metamorphosed to greenschist and amphibolite facies in the South Clearwater-Salmon area, to the north in the North Clearwater-St. Joe-Couer d'Alene area, and to the northeast in the Northeast Border area (Figure 1.1). Some of the metamorphism and related deformation is Precambrian in age. The augen gneisses (1370 Ma) emplaced into the lower Salmon supergroup postdate the formation of folds and greenschist facies metamorphism in the South Clearwater-Salmon area (Evans, 1981). In the Coeur d'Alene area, Precambrian mineralization (Zartman and Stacey, 1971) postdates early faulting and folding along the Lewis and Clark line (Hobbs et al., 1968).

During Cambrian to Ordovician time several syenitic bodies emplaced into the Precambrian and lower Paleozoic section at high structural levels (Evans and Zartman, 1981; Skipp, 1981; Scholten and Ramspott, 1968). Syndepositional faulting also occurred during deposition of Ordovician sediments (Garmezy and Scholten, 1981).

Other lower Paleozoic sections suggest deposition in fault controlled basins as well (Skipp, 1981).

During the latest Devonian to early Mississippian Antler orogeny, Devonian strata in south central Idaho may have been folded and exposed. Juxtaposition of Devonian deep water and continental shelf strata probably occurred at this time (Skipp and Hall, 1980). However, Antler age thrusts have not been documented (Dover, 1980b). Thick sections of Mississippian flysch present in south-central Idaho are interpreted to have been deposited east of the Antler highland (Dover, 1980b; Skipp and Hall, 1980). Flysch deposited west of the highland may also be preserved in eastern Idaho (Skipp, 1981).

The major deformation of the Paleozoic and Precambrian sediments east of the batholith occurred during Cretaceous to Paleocene time. East-directed thrusts of the Sevier orogeny stacked the rocks in their present configuration. Thrusting which began in the west and migrated to the east (Ruppel et al., 1981; Skipp and Hait, 1977), began in middle Cretaceous time prior to the emplacement of local batholithic intrusives and continued into the Eocene ceasing prior to the Challis episode (Hyndman et al., 1975; Ruppel et al., 1981; Ryder and Scholten, 1973; Ruppel, 1978).

The Couer d'Alene-St. Joe-Clearwater area (Figure 1.1) experienced a protracted dynamothermal metamorphism which probably also occurred during Cretaceous time, beginning prior to the emplacement of the Idaho Batholith. Several workers have suggested that much of the deformation occurred during the Precambrian age (Reid et al., 1981; Hietanen, 1984; Hobbs et al., 1965; Reid et al., 1973); however, two lines of evidence suggest that some of the deformation and higher grade metamorphism is Mesozoic. 1) The deformation is localized around the northern border of the batholith with metamorphic grade generally increasing toward the batholith (Hietanen, 1962). 2) Estimates of pressure and temperature conditions for emplacement of the intrusives and for metamorphism of the gneisses are comparable

(Rice, 1987). Field relationships clearly indicate that this deformation was complicated, perhaps long lasting, and began prior to local intrusion of batholithic rocks.

The dynamothermal event is characterized by multiple generations of deformation and metamorphism. Recumbent isoclinal folds with generally northwest trending axes and vergences to both north and south are refolded into open folds which are both northwest and northeast trending (Reid et al., 1981; Hietanen, 1984). Strike-slip activity along the western portion of the Lewis and Clark line is associated with some of these folds (Reid et al., 1981; Hobbs et al., 1968). Emplacement of Cretaceous to early Tertiary plutons related to the Idaho Batholith clearly postdates formation of recumbent folds (Reid et al., 1981) and some faults (Hobbs et al., 1968). In the St. Joe area northeast trending structures that postdate recumbent folds are associated with the emplacement of Cretaceous intrusives (Reid et al., 1981).

Sillimanite-facies deformation and metamorphism in the Northeast Border area (Figure 1.1) are also believed to be Mesozoic in age (Chase, 1973). As in the Clearwater-St. Joe-Couer d'Alene area, deformation and the peak of metamorphism clearly predate batholith emplacement. The absence of a static metamorphic aureole around the batholithic rocks is cited as evidence that peak metamorphic conditions only slightly preceded batholith emplacement (Chase, 1973).

1.2.3 Summary

Continentally derived sedimentary materials were deposited intermittently from Proterozoic to Permian time. There is good evidence of some Precambrian deformation in the South Clearwater-Salmon area where folding and greenschist metamorphism preceded intrusion of augen gneisses. Sedimentation in fault bounded basins and the intrusion of Cambrian and Ordovician plutons suggest another pulse of tectonic activity during the Cambrian and Ordovician. Thrusting during the Mississippian to Devonian

Antler orogeny may have affected sediments in south-central Idaho; thick sequences of flysch in south-central Idaho were deposited as a result of the deformation. There is no known record of the Antler orogeny elsewhere in Idaho. Beginning in the Cretaceous, Sevier thrusting stacked Precambrian and Paleozoic sediments. Much of the amphibolite-facies metamorphism and deformation of Precambrian strata near the batholith appears to have occurred during this time. Intrusion of the Idaho Batholith occurred during these deformational events.

1.2.4 Idaho Batholith

The Late Cretaceous Idaho Batholith underlies most of central Idaho. It is divided into two lobes, the smaller northern Bitterroot lobe and the southern Atlanta lobe, by the metamorphic rocks in the South Clearwater-Salmon area (Figure 1.1), often referred to as the Salmon River Arch (Armstrong, 1975). The interior of the batholith consists primarily of large masses of two-mica granite and biotite granodiorite (Lewis et al., 1987; Hyndman, 1983; Toth, 1987). Although dating these rocks has proved to be difficult, they are believed to range in age from approximately 85 to 70 Ma in the Atlanta Lobe (summarized by Lewis et al., 1987). Ages as young as 46 Ma are reported for intrusives in the Bitterroot lobe (Chase et al., 1983). An older tonalitic phase is present extensively on the western side of the batholith and also occurs locally on its eastern margin (Lewis et al., 1987; Toth, 1987). Contacts between the batholith core and the older tonalites are both gradational and intrusive. The tonalite phase appears to have been emplaced between 85 and 95 Ma (summarized by Lewis et al., 1987). Older intrusives are present along the western margin of the batholith and are described in detail as part of this study.

1.2.5 The Arc-Continent Boundary

This study explores a portion of the boundary between the rocks of oceanic-arc and continental affinity within the Idaho Batholith. Several other studies of this boundary have been completed elsewhere (Figure 1.2). These studies document Cretaceous deformation along the boundary and bring out some features of the boundary that are consistent along its length.

Myers (1982) studied the boundary in the Harpster area (Figure 1.2). He describes three blocks of pre-Miocene rocks separated by steeply east-dipping reverse fault systems (Figure 1.3). The western block contains rocks of the Seven Devils Group intruded by Jurassic gabbro-tonalite (Toth and Stacey, 1985, per. comm.). The central block contains rocks correlated with the Squaw Creek Schist of the Riggins Group intruded by the trondhjemitic Blacktail pluton of Cretaceous (?) age. The eastern block contains shallow water, cratonally derived, sedimentary material tentatively correlated with the Orofino series (Anderson, 1930; Hietanen, 1962) intruded by tonalite, trondhjemite and adamellite wedges of the Idaho Batholith. The fault systems separating the blocks are believed to have been active in Cretaceous time before, during and after batholith emplacement and were reactivated in Tertiary time. Deformation is described as "progressive comminution and small scale penetrative displacement along interlensing slippage surfaces with closely associated shear folding, reverse faulting and metamorphism." Myers reports that metamorphic grade increases across the area from lower greenschist facies to upper amphibolite facies. Metamorphic gradients are described within blocks, and isograds are cut by faults. The deformation and metamorphism is attributed to the emplacement of the Idaho Batholith.

A structural study of the southern portion of the Harpster area was undertaken by Hoover (1986). She recognized two deformational events in arc rocks west of the boundary, accompanied by a single thermal metamorphic event. The first deformational

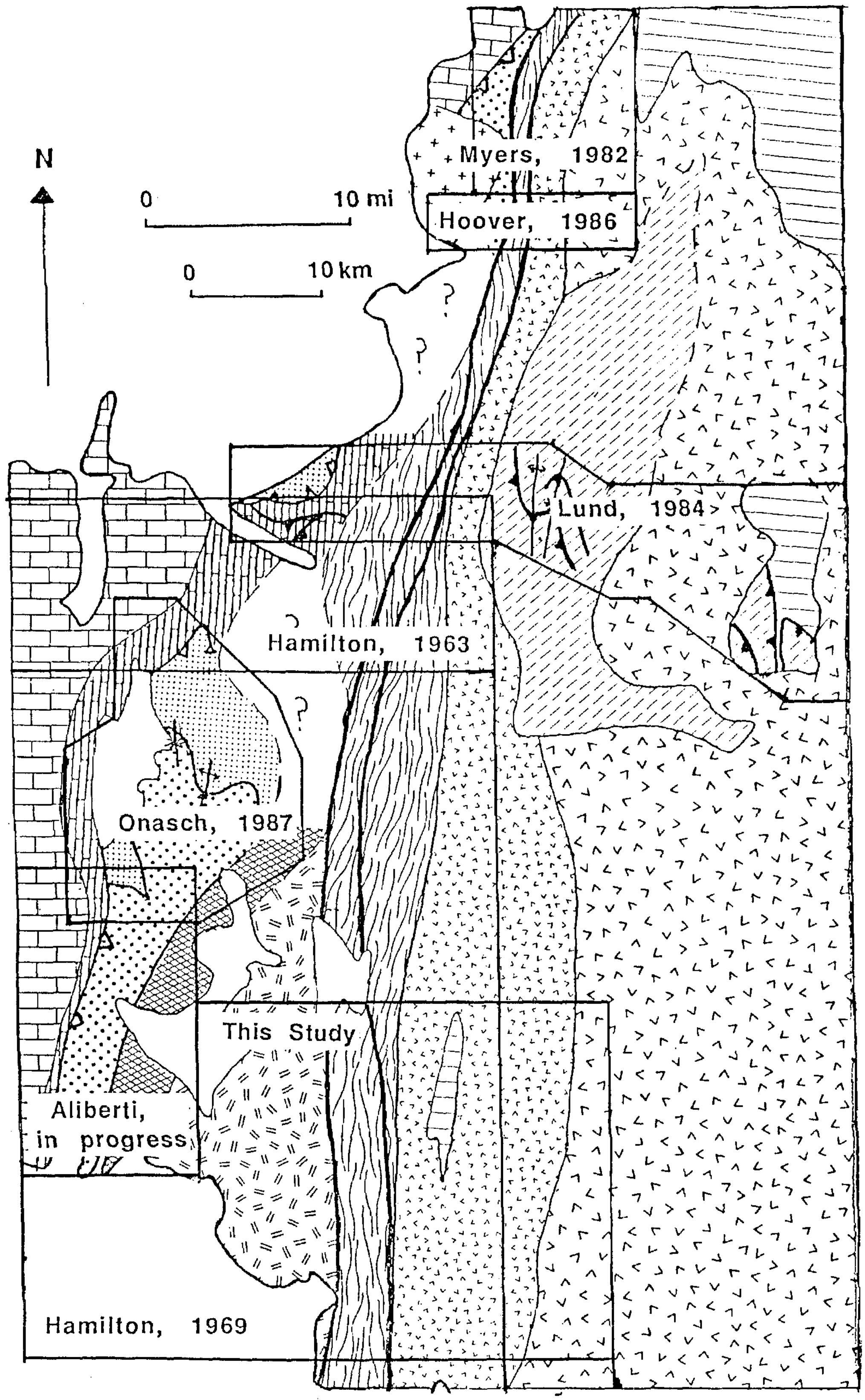


Figure 1.2 Location map for studies of the arc-continent boundary.

Figure 1.3 Geologic map of arc-continent boundary compiled from Bond, 1978, sources shown in Figure 1.2 and current work.

Rocks of Oceanic Affinity

Wallowa Terane

- Seven Devils Group
- Martin Bridge and Lucile Formations

Riggins Group

- Upper: Squaw Creek Schist
- Lower: Fiddle Creek and Lightning Creek Schists

- Pollock Mountain Amphibolite

- Areas of metamorphic and plutonic rocks of uncertain affinity

Rocks of Continental Affinity

- Metamorphosed Pz? sedimentary rocks
- Metamorphosed PC? sedimentary rocks

Intrusive Rocks

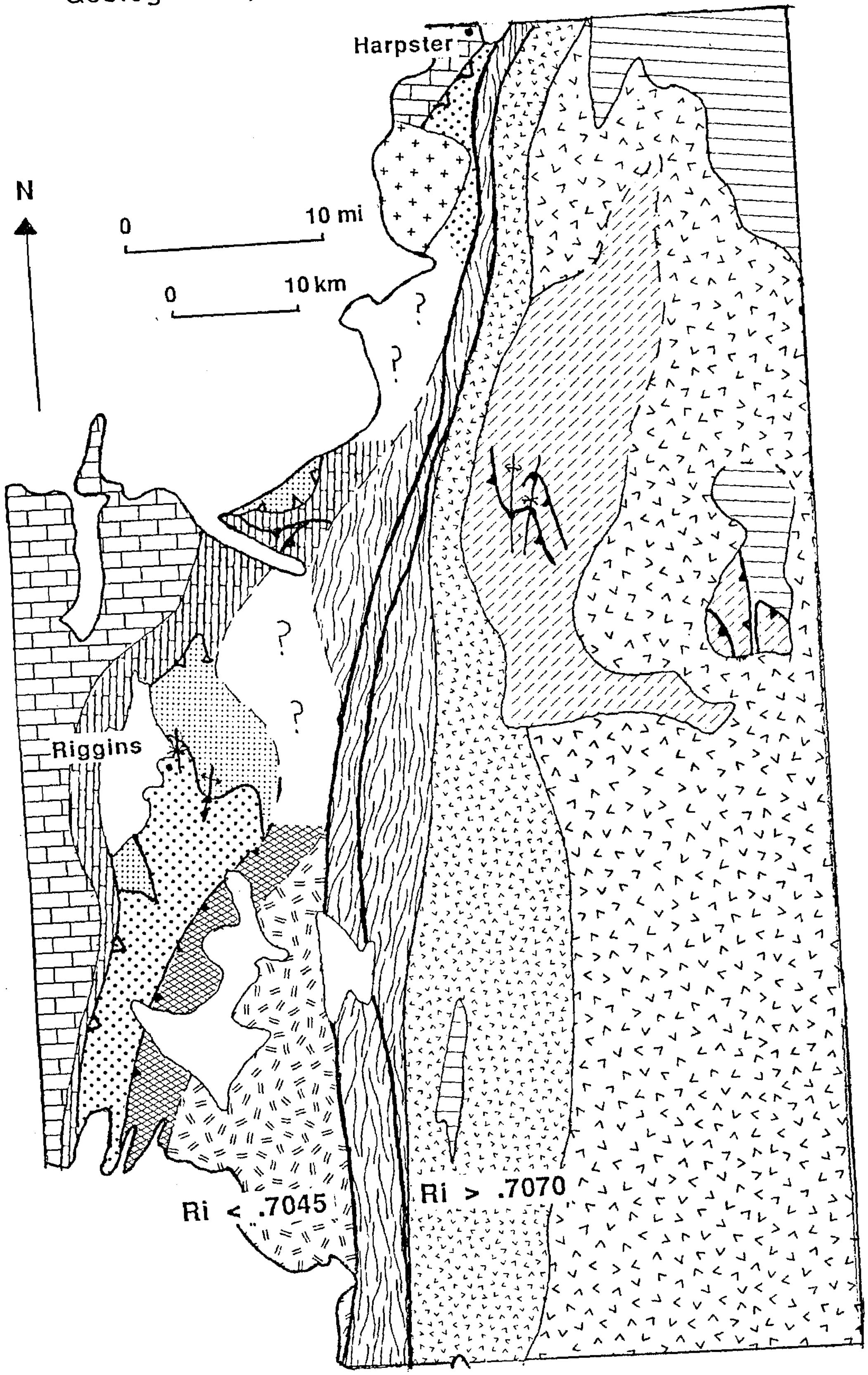
- Hazard Creek Complex
- Mylonitized tonalite and granodiorite
- Biotite hornblende tonalite
- Undiferentiated intrusives of the Idaho Batholith
- Blacktail Stock

- Tertiary Basalt and Quaternary Cover

- Rapid River Thrust Fault
- Other thrust faults
- Ri isopleth

- Anticline-trace of axial plane
- Syncline-trace of axial plane

Geologic Map of Arc-Continent Boundary



McCall •

event was characterized by bulk flattening. Pegmatites and aplites emplaced during or prior to this event were boudinaged or shortened in the plane of the schistosity. La Salata (1982) estimated up to 500% shortening. The second deformational event is characterized by northwest verging folds and thrust faults. The deformation is better expressed in rocks farther from the arc-continent boundary. Metamorphism continued after deformation had ceased. In continental rocks three deformational events are recognized. During the first event a schistosity developed, which was folded into an open north-northeast trending synform during the second deformation. These deformations occurred during thermal metamorphism. Correlation of lithologies with tectonostratigraphic units of Lund (1984) requires a third set of fault structures that separate these units. Movement on these faults may have postdated metamorphism. In contrast to Myers (1982), Hoover infers that the only change in metamorphic grade occurs in arc rocks where kyanite zone gneisses to the west are juxtaposed against sillimanite zone gneisses to the east across a thrust fault formed during the second period of deformation. All rocks west of this thrust in her study area were interpreted to be in the sillimanite zone of the amphibolite facies. Hoover subdivided the plutonic rocks near the suture into a porphyritic orthogneiss with a strong mylonitic fabric which contains the boundary between arc and continental pendants and a tonalite/granodiorite with a hypidiomorphic texture which lies to the east of the arc-continent boundary (Figure 1.3). Arc and continental pendants are interleaved within the porphyritic orthogneiss. Intrusion of these units is inferred to postdate metamorphism, as they were not metamorphosed. A narrow zone of mylonitic deformation developed along the arc-continent boundary after emplacement of these plutons. Lund (1984) studied the arc-continent boundary in the Buffalo Hump-Gospel Peak area (Figure 1.2). She describes four types of pre-Cretaceous rocks (Figure 1.3). The westernmost rocks are metamorphosed Seven Devils Group volcanics and overlying Martin Bridge and Lucile Formations. These are tectonically overlain by rocks of the lower portion of the Riggins Group. Eastern assemblages are metamorphosed shallow-water, cratonally derived, sedimentary rocks, which Lund divides into several lithotectonic packages. The easternmost packages are correlated with Precambrian Belt age strata. The more westerly packages are correlated with Paleozoic strata. All pre-Cretaceous rocks were metamorphosed prior to tectonic stacking. Metamorphic grade increases toward the arc-continent boundary from greenschist to upper amphibolite facies in the arc rocks and from lower to upper amphibolite facies in the continental rocks. In the arc rocks two dynamothermal events are recognized with another overprinting static metamorphic event. Metamorphic minerals yield $^{40}Ar/^{39}Ar$ ages of 118 to 101 Ma (Lund and Snee, 1988). Thrust faults that stack the pre-Cretaceous rocks cut metamorphic isograds. In continental rocks, vergent folds and thrusts developed during stacking. Folds and thrusts are more recumbent and more strongly asymmetric away from the arc-continent boundary. Following stacking, a second set of west to north west plunging cross folds developed. The arc rocks several generations of west directed thrusts and folds developed during stacking. This deformation is interpreted to have occurred during the formation of the arc-continent boundary. Following deformation, tonalitic rocks were emplaced on both sides of the boundary. This is the only location where post-deformational plutons are reported west of the arc-continent boundary. Both plutons and their wall rocks were deformed together in a 5 km wide zone along the arc-continent boundary, forming a steep schistosity parallel to the boundary and steep south-east plunging mineral lineation. $^{40}Ar/^{39}Ar$ age spectra suggest that some of the tonalites were emplaced around 93 Ma and deformation occurred about 88 Ma (Lund and Snee, 1988). This deformation is similar in style and location to the mylonitic deformation described by Hoover (1986). Lund attributes this deformation to uplift.

The nearest extensive exposures of arc assemblage rocks to the study area are in the Riggins area (Figure 1.3). These rocks were first studied by Hamilton (1963a,b, 1969; Figure 1.2). He divided pre-Cretaceous rocks into two packages separated by a major north-south striking, west verging structure: the Rapid River Thrust. The higher grade Riggins Group occurs above the Rapid River Thrust with the rocks of the Wallowa/Seven Devils arc below (Figure 1.3). Metamorphic grade increases from west to east in each package, with the thrust cutting both the metamorphic fabric and offsetting isograds. Shearing accompanied and locally outlasted metamorphism. A compound metamorphic fabric developed in some places in association with complex larger structures. Trondhjemite dikes are both cut by and cut retrograde shear zones related to this thrusting episode, indicating that thrusting and intrusion happened in a relatively short period of time. Hamilton notes that despite the locally complex deformation, there is no evidence that suggests that all deformation and metamorphism did not take place in the first half of the Cretaceous, and that the present mineral assemblages probably formed during a single period of metamorphism. In the Seven Devils Group, a Jurassic metamorphic fabric and plutonic episode was also recognized.

Onasch (1987) undertook a more detailed study of the structure in the Riggins area. His work brings out the complexity of the deformation. He recognized five phases of mesoscopic folding in the Riggins Group and three to five phases in rocks of the Wallowa/Seven Devils arc. First generation folds are isoclinal in both packages. Second generation folds are overturned to the west in the Riggins Group and to the east in the Seven Devils Group. Later folds are open or upright. Intrusion of batholithic rocks and metamorphism occurred throughout the deformational history. The relative timing of events varies.

Aliberti has studied the Rapid River Thrust System in more detail (Aliberti and Wernicke, 1986a,b). The thrust system roots in a north-northeast striking, steeply dipping zone characterized by bulk flattening. It shallows to the north, placing higher grade Riggins Group over lower grade Wallowa/Seven Devils arc. A second major fault, the Pollock Mountain Fault, places garnet-bearing banded amphibolites of the Pollock

Mountain Amphibolite over lower grade schists of the Riggins Group (Aliberti et al., 1987; Figure 1.3).

The metamorphic petrology of these rocks is being studied by Selverstone (Selverstone et al., 1987) who reports that garnets in the Riggins Group record isothermal decompression to final equilibration at 500 to 550°C and 6 kbars. Garnets in the Pollock Mountain Amphibolite indicate burial and heating to 600-650° C and 8-9 kbars. All garnets studied formed after the development of the foliation. Rolled garnets from similar rocks were dated using Nd-Sm systematics as 144 Ma (Aliberti et al., 1988). This age suggests that some deformation may have occurred in earliest Cretaceous or Jurassic time.

These studies indicate several attributes of the arc-continent boundary that are observed along its length is west-central Idaho. Some portion of the Riggins Group always lies between the Wallowa/Seven Devils arc and the continental rocks; the lithologic members present vary along strike. Riggins Group rocks were multiply deformed and metamorphosed to greenschist or amphibolite facies in Cretaceous time. This deformation also affected the eastern margin of the Wallowa/Seven Devils arc rocks, overprinting an older Jurassic deformation. West-vergent folds and thrusts developed west of the arc-continent boundary at a variety of times during the Cretaceous deformation. Most of these structures juxtapose rocks of differing metamorphic grade. Intrusion of trondhjemitic material began prior to and continued after thrusting. East of the arc-continent boundary east directed thrusts formed in metamorphosed cratonal material. These structures developed prior to batholith intrusion and after metamorphism. Lund believes they formed at the same time as the west-vergent structures west of the arc-continent boundary. Tonalitic plutons intruded rocks on both sides of the arc-continent boundary during and after deformation. A narrow zone of mylonitic deformation developed along the boundary after emplacement of the plutons obscuring older relationships along the boundary.

1.2.6 Tectonic Models Proposed for the Formation of the Arc-Continent Boundary

Several models have been put forth to explain the abrupt transition from oceanic arc to cratonal material in the wall rocks of the batholith. This work attempts to test and place further constraints on these models. Davis et al. (1978) suggested that the continental crust was truncated by a spreading center active during the Late Jurassic through Early Cretaceous (Figure 1.4). The rift is postulated to have been terminated to the north by a transform fault which extended from the bend in Sr isopleths, near Orofino, Idaho, through the Methow Trough (Figure 1.4). This model was proposed to explain the doubling of Upper Paleozoic to Jurassic rocks in north-central Washington and southern British Columbia. The late Jurassic Ingalls ophiolite was inferred to form during rifting. Input of crystalline material into the Methow trough during mid-Cretaceous time was interpreted as marking the juxtaposition of the southern section against the northern section. The location of the rift was chosen to explain the absence of the Paleozoic-Jurassic sequence in western Idaho. The Blue Mountain Arc (excluding the Wallowa/Seven Devils arc) was suggested to have formed in a preexisting embayment in the continental margin. The Wallowa/Seven Devils arc was inferred to be emplaced against the continental margin in Early Cretaceous time, occupying the hole formed by rifting. Emplacement is constrained to be post Jurassic by the presence of the Late Jurassic marine strata of the Coon Hollow formation on the eastern side of the arc. In western Idaho, the absence of a subduction complex and rift related intrusives are the chief difficulties with this model.

A second model, put forth by Lund (1984), was based primarily upon her studies of the metamorphic and plutonic rocks in western Idaho. She concluded that the lack of interfingering relationships between rocks of continental and arc origin, and of sediments of mixed provenance, precluded the presence of a back-arc basin between the

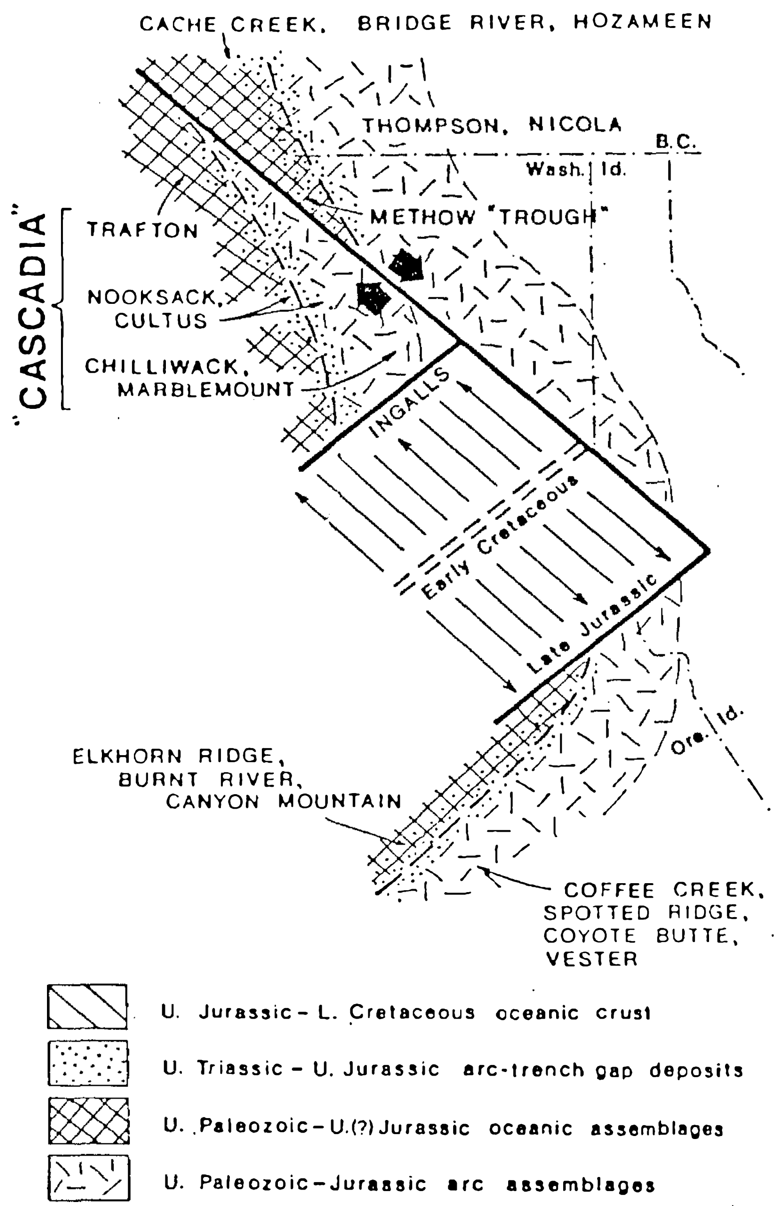


Figure 1.4 Proposed rifting of continental margin from Davis et al., 1978.

continent and arc. She also found no material representing melange or accretionary prism preserved between the arc and continent. Citing this lack of intermediate material between arc and continent, the abrupt change in isotopic systematics, and the lack of blueschist metamorphism, Lund suggested that the boundary did not represent a closed ocean basin.

From her structural study, Lund concluded that the rocks on either side of the boundary had no common history until the late stages of deformation and metamorphism in the middle Cretaceous (Lund, 1984; Lund and Snee, 1988). This deformation involved movement up and away from the boundary, forming steep structures near the boundary that shallow both to the east and west. These structures are interpreted as flower structures developed along a slightly convergent transcurrent fault. Lund (1984) suggests that the Wallowa/Seven Devils arc was accreted to the continent by subduction in central Oregon, as proposed by Ave Lallemant et al. (1980), and then carried northward along a transcurrent fault boundary (Figure 1.5). The Idaho Batholith was emplaced during the final stages of this movement in response to simultaneous subduction west of the arc. Later clockwise rotation gives the Seven Devils their current east-west orientation.

A third model for the development of the crustal boundary, put forth by Wernicke (1984), tried to explain the doubling of the Paleozoic to Jurassic section in central Washington by transcurrent faulting. Wernicke noted the indentation into the craton by the Seven Devils arc and the left lateral drag fold geometry in the Blue Mountains Arc. He suggested that the Cretaceous Cordillera at this latitude has experienced east-west shortening accommodated by the removal of blocks of crust to the north and south along strike-slip faults. He proposes that this shortening was a response to the impingement of Wrangellia as represented by the Wallowa/Seven Devils arc.

The other models proposed for the arc-continent boundary in western Idaho involve subduction. Hamilton (1978) proposed that the Riggins groups represented an

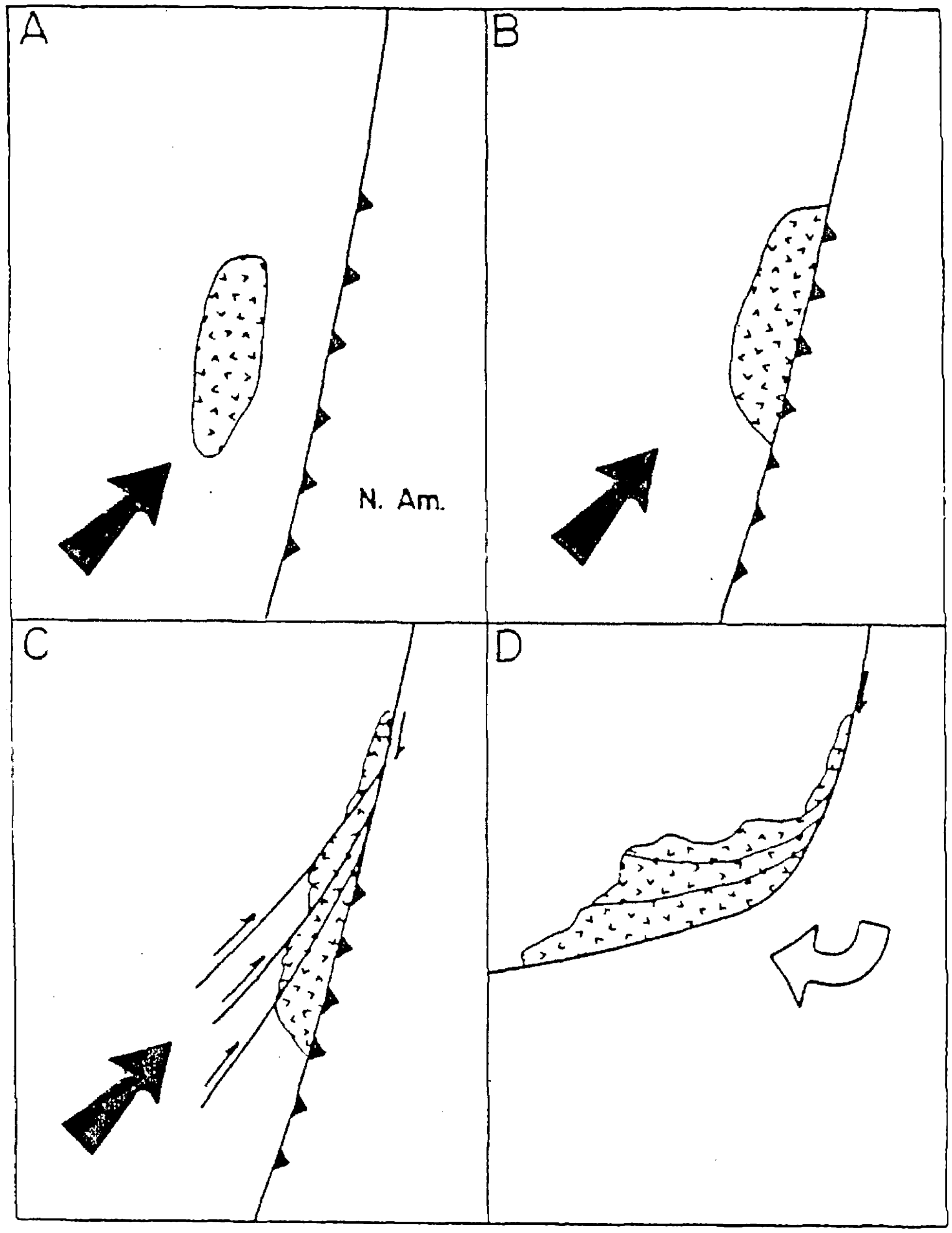


Figure 1.5 Proposed model for the accretion and subsequent transcurrent movement of the Wallowa/Seven Devils arc from Lund, 1984.

overturned arc-melange sequence which was accreted by late Triassic time. He interpreted this as the first of several arcs which have been accreted to the continent at this latitude. He placed accretion in the Triassic because he suspected that plutons intruding the deformed rocks were Triassic. These plutons are now known to be Cretaceous, relaxing the time constraint. Talbot and Hyndman (1975) proposed that the Seven Devils and Riggins group represent a subduction complex. Ave Lallemant et al. (1985) suggested that the Cretaceous deformation in Idaho was due to collision and clockwise rotation of the Blue Mountain Arc. Details of the mechanism of collision were not given.

Two endmembers emerge from these models: accretion by subduction or juxtaposition by transform faulting. One key to distinguishing between these models lies in the Riggins Group and its interpretation. Those favoring juxtaposition by transform faulting cite the absence of material of mixed provenance and the absence of a subduction complex. Those favoring subduction interpret parts of the Riggins Group as a subduction complex. It is notable that Hamilton, who favors subduction, worked in the most complete section of Riggins Group while Lund, who favors transform faulting, worked in a section where only the lower Riggins Group is preserved. Both workers agree, however, that there is no evidence of continental detritus within any member of the Riggins Group.

1.2.7 Post Cretaceous Geology

The study area lies along the eastern edge of exposures of Cenozoic Columbia River Basalt. These basalts, which are present locally in the study area, have been offset by Cenozoic faults which produced the major valleys in the area. The distribution of Columbia River Basalt flows within the Salmon River Canyon suggests that this faulting began prior to and continued during Columbia River Basalt eruption, 17.5 to 6 Ma (Hooper, per. comm., 1987). Pleistocene glaciation affected the area at elevations above 5000 feet. There is no record of faulting that postdates glaciation (Ken Pierce, pers. comm.).

1.3 PREVIOUS WORK IN THE STUDY AREA

Early maps that included parts of the study area are by Lindgren (1900) and Livingston and Laney (1920). Larsen and Schmidt (1958) made a petrographic traverse of the area recognizing tonalite, porphyritic granodiorite and quartz monzonite as major rock types. Schmidt's (1964) study of the western border zone of the batholith extends into the southern portion of the study area. He described the border zone as a 35 mile wide smooth gradation from migmatite and schist to directionless granite, with an internal zone of shearing and recrystallization.

Hamilton (1963a, 1969) mapped the western portion of this study area at a scale of 1:125,000. He divided the plutonic rocks into quartz monzonite/granodiorite, gneissic quartz diorite, quartz monzonite/quartz diorite migmatite, and trondhjemite. Hamilton first recognized the trondhjemites that are common in the western portion of the study area, and reported on their chemistry and occurrence (1963b). He interpreted the metamorphism and deformation in these rocks to represent ongoing deformation during emplacement, which was synchronous with thrusting in the metamorphic rocks to the west. He attributed this deformation to be generally related to the emplacement of the batholith. Hamilton also reports some chemical data for intrusives into the Riggins Group (1963a).

Taubeneck (1971) produced an excellent study of the gneissic border zone of the batholith in the area south of Cascade. He recognized the four belts of plutonic rocks described in this study and described them as extending south to the Snake River Plain. He disputed Schmidt's idea of a smooth gradation between migmatite and directionless granite noting reversals in rock composition and intrusive contacts between the massive core of the batholith and its gneissic border. He also recognized that epidote-bearing plutons of trondhjemitic affinity occur all along the westernmost margin of the batholith. In a separate report, Taubeneck (1973) notes the occurrence of continentally derived sedimentary material in pendants near McCall, and concludes that the continental margin must have extended this far west and that it is juxtaposed abruptly against material of oceanic origin.

Zen and Hammarstrom (1984) and Zen (1985) described the epidote that occurs in intrusives in the western portion of the area and discussed its petrologic and tectonic significance. They concluded on the basis of textural arguments that the epidote is a primary magmatic phase of the intrusives. Experimental data from Naney (1983), Liou (1973) and Holdaway (1972) were cited as constraining formation of the epidote-bearing assemblages to pressures greater than 6 kbars and most likely greater than 8 kbars (25 km). The emplacement of these plutons into metamorphosed supracrustal rocks led Zen to the conclusion that collisional tectonics produced crustal thickening followed by rapid uplift. Data on hornblende composition in these intrusives was collected by Hammarstrom and Zen (1986) and is similar to that from other epidote bearing intrusives in the Cordillera. This data was used to develop an empirical igneous geobarometer, based on the aluminum in hornblende, which was tentatively confirmed by Hollister et al. (1987).

Isotopic studies that included the study area have been conducted by Armstrong et al. (1977) and Criss and Fleck (1987). Armstrong et al. (1977) analyzed several samples for Rb, Sr and $^{87}Sr/^{86}Sr$. These samples indicated that the abrupt change in isotopic systematics passed through this area. Criss and Fleck (1987) confirmed this conclusion. Oxygen isotopic analyses (Criss, 1981; Criss and Fleck, 1987) also indicated a change in geochemistry of the plutonic rocks across the area. These data are discussed in more detail in Chapter 4. Two late Cretaceous K-Ar apparent ages were determined on biotite from intrusives in the area (McDowell and Kulp, 1969).

CHAPTER 2: DISTRIBUTION AND DESCRIPTION OF LITHOLOGIES IN THE WESTERN IDAHO BATHOLITH NEAR MCCALL

2.1 INTRODUCTION

This chapter describes the rocks present in the study area, located in west-central Idaho north of McCall. It extends generally from 44° 55' to 45° 15' north and from 115° 52.5' to 116° 22.5' west. Along the margins of this area some portions were not examined due to poor outcrop or access. The mapping was done by Dr. Mel Kuntz of the U.S. Geological Survey and myself, with assistance from Jim LaFortune, during the summers of 1982 to 1985. A compiled geologic map of the study area is shown on Plate 1. The map area includes the Indian Mountain, Hazard Lake, Black Tip, Granite Lake, Brundage Mountain 7.5 minute quadrangles and portions of the Bally Mountain, Meadows, McCall, Box Lake and Victor Peak 7.5 minute quadrangles.

The classification of the rocks into units, and their descriptions, are the integrated result of field and petrographic studies. The thin-section data are tabulated in Appendix 1 with sample localities shown on Plate 3. The modal mineralogy was estimated from thin sections using a spot chart; all samples except those with names containing "z" were stained for alkali feldspar. Comparison of estimated modes, with norms for those samples that were analyzed chemically, suggest that quartz was often overestimated and plagioclase underestimated by ten to twenty percent, particularly in samples from the Little Goose Creek complex where plagioclase is untwinned. Alkali feldspar appears to be overestimated in samples from the Trail Creek Granite. Most alkali feldspar in the Little Goose Creek complex is in large megacrysts, thus modes estimated from thin sections tend to underestimate the alkali feldspar present in the unit. Accessory phases are occasionally overestimated: an estimate of 2% or 3% was used to indicate a very abundant phase and may not truly reflect the volume percent.

The rock names assigned are based both on thin-section analysis and on field estimates of the mineralogy. Field estimates were relied on more heavily in describing units in the Hazard Creek complex due to the extreme variability of these rocks. The classification used is that of Streckeisen et al. (1973, Figure 2.1). In addition, tonalites of low color index containing biotite, or biotite and muscovite, are classified as trondhjemites. Chemical analyses corroborate that these rocks are trondhjemites as defined by Barker (1979). This definintion and the implication of the presence of trondhjemites are discussed more thoroughly in Chapter 4. The distinction between gabbro and diorite was made primarily on the basis of hand sample appearance: the name diorite was assigned to rocks with a salt and pepper appearance, gabbro to rocks that were dark. Unit names are capitalized; rock names are lower case (e.g., the unit Epidote-Biotite Tonalite is comprised of the lithology epidote-biotite tonalite). The term orthogneiss is appended to unit names where the majority of rocks in the unit have mylonitic or crystalloblastic textures.

The field area studied is underlain primarily by variably deformed and metamorphosed Cretaceous intrusive rocks of the Idaho Batholith. These intrusives naturally organize into three north-south elongate complexes. These belts from west to east are named the Hazard Creek complex, the Little Goose Creek complex, and the Payette River complex. The regional boundary between arc-related and cratonally-derived supracrustal material is preserved in pendants within these intrusives. Wall rocks within the Hazard Creek complex, and most of the Little Goose Creek complex, are amphibolite, schist and gneiss inferred to be metamorphosed volcanic or volcaniclastic rocks and associated sedimentary rocks, formed in an oceanic-arc. Screens and inclusions within the Payette River complex and locally within the eastern Little Goose Creek complex are sillimanite-bearing schist, biotite schist, quartzite, quartzofeldspathic gneiss, and calc-silicate gneiss interpreted to be metamorphosed cratonally-derived shallow water sedimentary material. The boundary between wall

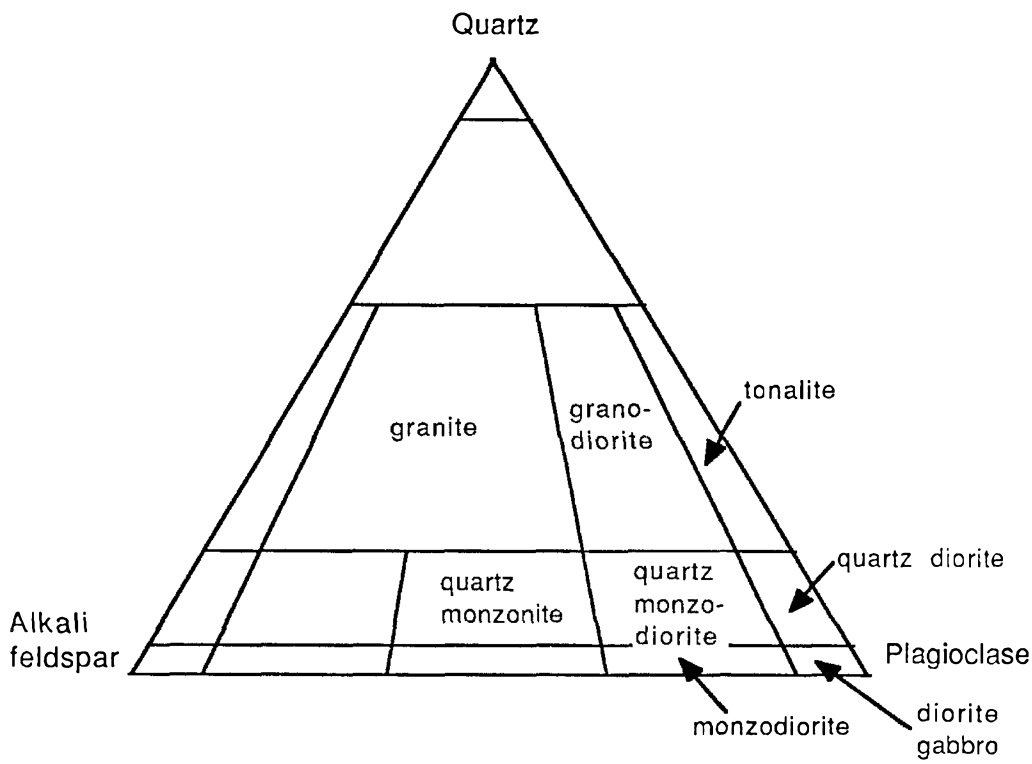


Figure 2.1 Rock names used in text. Classification of Streckeisen et al., (1973).

rocks of oceanic-arc and continental affinity is located locally within the Little Goose Creek complex but generally along its eastern margin. The change in wall rock can be demonstrated to occur within two kilometers.

The oldest intrusive material in the study area is contained in the Hazard Creek complex. This complex consists of variably deformed and recrystallized orthogneisses ranging in composition from diorite to granite. The bulk of the material is tonalitic and usually contains magmatic epidote. The complex is inferred to have formed as a series of intrusives emplaced during ongoing deformation.

The Little Goose Creek complex is a narrow band of strongly deformed intrusives that crops out directly to the east of the Hazard Creek complex. The most abundant rock in this unit is a distinctive microcline megacrystic orthogneiss called porphyritic orthogneiss. It is interleaved on centimeter to 100 meter scales with other orthogneisses that range in composition from pyroxenite and gabbro to granodiorite, and are inferred to be both younger and older than the porphyritic orthogneiss. Most original intrusive contacts within the complex were transposed by synplutonic deformation or by ductile deformation after emplacement. Ductile deformation also transposed the contacts between the Little Goose Creek complex and the Hazard Creek complex to the west, and the Payette River complex to the east, obliterating any intrusive relationships.

The Payette River complex lies directly to the east of the Little Goose Creek complex. This unit is composed primarily of Payette River Tonalite and Trail Creek Granite. Payette River Tonalite crops out in a north-south striking band three to ten kilometers wide. It is similar in composition and texture to tonalites mapped along the length of the west side of the Idaho Batholith. The Payette River Tonalite has a strong primary foliation and locally has a hornblende lineation. Mylonitization of its western margin is the only penetrative subsolidus deformation that has occurred within the unit.

The tonalite is intruded by granite sills that are broadly concordant with the foliation. These are interpreted to be granite differentiates emplaced shortly after the tonalites.

The easternmost intrusive in the study area is the Trail Creek Granite, which is part of the granitic core of the Idaho Batholith. To the west this unit intrudes the Payette River Tonalite; it extends out of the study area to the east. The western contact consists of a zone of abundant granite dikes in tonalite that give way to granite containing blocks of tonalite. The Trail Creek Granite is comprised of one or more massive undeformed plutons that are foliated only locally.

In the following sections, the supracrustal rocks present in the area are described. Then each complex is introduced more thoroughly and the map units of intrusive origin are described in detail.

2.2 SUPRACRUSTAL ROCKS

Metamorphosed supracrustal rocks occur as inclusions, sheets, and pendants in intrusive rocks throughout the study area. These are described in the following sections. Metamorphosed oceanic-arc rocks occur in the Hazard Creek complex and the Little Goose Creek complex. Mafic Layered Gneiss is comprised primarily of these rocks and they form a significant component of Tonalite with Gneiss Pods, Dark Gneiss Assemblage and Light Gneiss Assemblage. Metamorphosed oceanic-arc rocks in the Little Goose Creek complex are particularly common in the west; they occur as sheets interlayered on meter scales with porphyritic orthogneiss. Metamorphosed continentally-derived sedimentary rocks occur in the Payette River complex and rarely in the Little Goose Creek complex. All supracrustal rocks contain amphibolite facies mineral assemblages.

2.2.1 Oceanic Supracrustal Rocks

2.2.1.a Hazard Creek Complex

Mafic Layered Gneiss

Mafic Layered Gneiss is comprised of medium to fine-grained gneisses that are compositionally layered on scales from a centimeter to several tens of meters. All have crystalloblastic fabric and show no evidence of post-metamorphic mylonitization. Locally the unit contains sheets and dikes of younger intrusive rocks. The layers can be subdivided based on composition.

The most common lithology is a gneiss of dioritic composition comprised of plagioclase, hornblende, biotite, quartz ± clinozoisite/epidote, garnet, diopside (Figure 2.2). This gneiss is relatively homogeneous and accounts for 50% of the unit. The color index is between 30 and 40; hornblende is more abundant than biotite comprising 100 to 60% of the mafic minerals (Figure 2.2). Two suites of hornblende, biotite, and epidote are recognized. Suite one has blue-green hornblende, light to dark brown biotite, and highly birefringent epidote. Suite two has olive-green hornblende, foxy biotite, and clinozoisite or epidote with low birefringence. Suite two was observed only along the Little Salmon River. In both suites plagioclase is generally unzoned, anhedral andesine. Suite one contains some small or medium-sized grains with relict oscillatory or normal zoning. Garnet where present is pink in hand sample and clear in thin-section. In one sample, foliation is warped around the garnets while biotite within the garnet is aligned in the foliation plane, suggesting the garnets formed at the same time as the foliation without deformation. Some garnets show two stages of growth: homogeneous cores are surrounded by a rim containing abundant inclusions.

Coarse and fine-grained amphibolite layers with color index between 30 and 70 are common. The mineralogy of the amphibolites suggests that their composition is

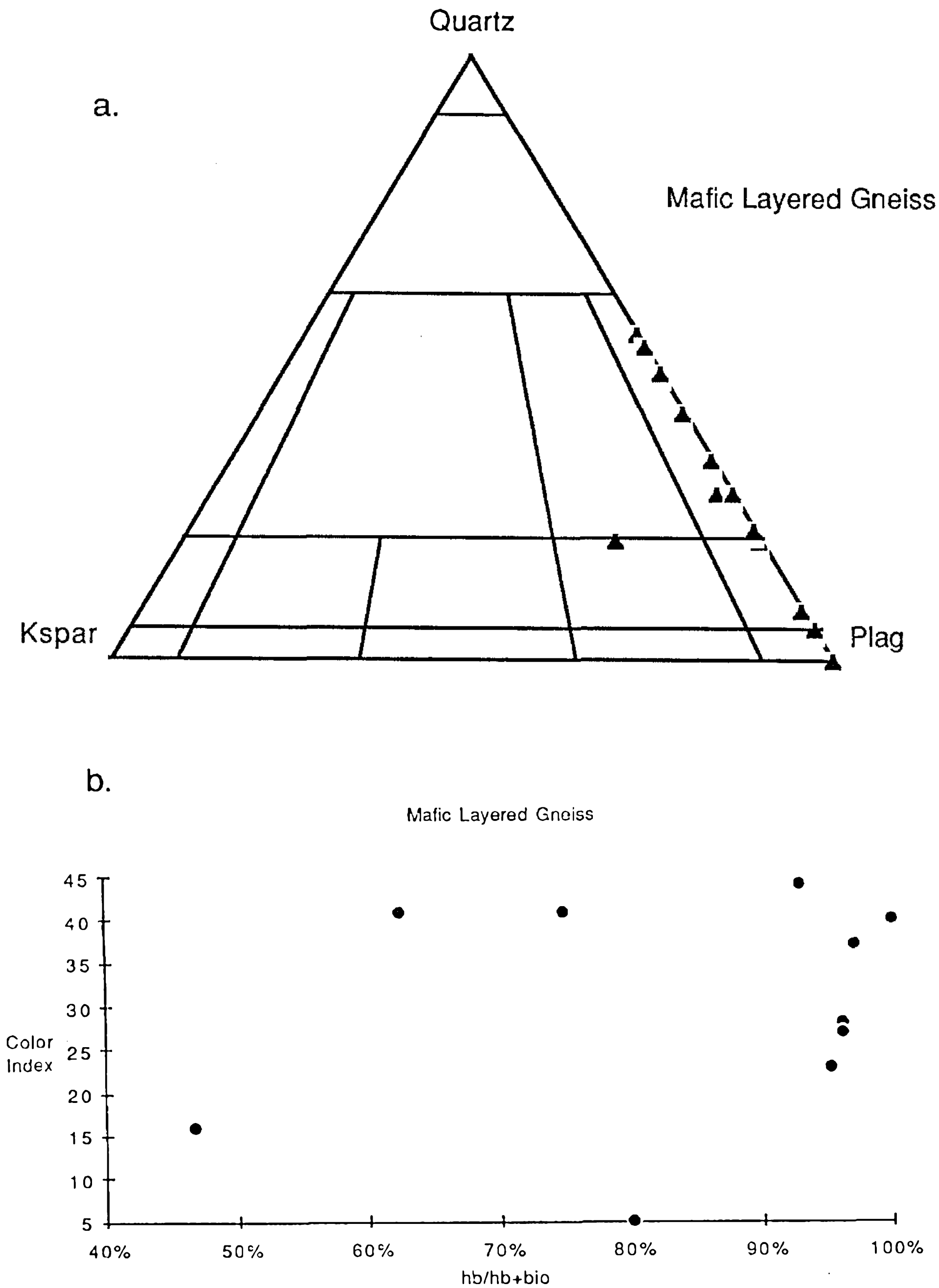


Figure 2.2 Estimated modal composition of Mafic Layered Gneiss: a) classification after Streckeisen, 1973; b) color index versus hornblende/hornblende+biotite.

similar to that of basaltic andesite and basalt. Hornblende is generally blue-green in color similar to hornblende in the first suite of andesitic gneiss.

Leucocratic gneiss layers have a color index of seven with foxy-red biotite as the mafic mineral. Clinozoisite, muscovite and garnet are present in small amounts. Plagioclase, which forms 65% of the rocks, has two forms: small well twinned grains, and larger two to three millimeter, poorly twinned, zoned grains believed to be porphyroclasts from an intrusive protolith. These layers appear to be metamorphosed tonalite and trondjhemite sills similar in composition to younger intrusives in the area.

The least abundant layers in the gneisses are quartzite and calc-silicate gneiss. The calc-silicate gneisses are dark rocks with abundant hornblende comprised of hornblende, epidote, plagioclase, diopside, quartz and sphene ± garnet. Hornblende is altering to cummingtonite locally.

Inclusions of Mafic Layered Gneiss are common in Epidote-Biotite Tonalite, Light Gneiss Assemblage and Dark Gneiss Assemblage. One sample of altered ultramafic rocks was also collected from the Dark Gneiss Assemblage.

Tonalite with Gneiss Pods

In the central portion of the Hazard Creek complex metamorphosed supracrustal rocks occur as sheets and pods within epidote-biotite-hornblende tonalite. The metamorphosed supracrustal rocks are amphibolite, marble, calc-silicate gneiss and quartzofeldspathic gneiss with fine to medium-grained crystalloblastic fabrics. These rock types occur both together and in monolithologic pods. The calc-silicate gneisses vary widely in composition containing the mineral assemblage garnet, diopside, calcite, quartz ± scapolite, epidote and sphene. Amphibolites contain hornblende, plagioclase, opaque, epidote ± anthopholite, scapolite, quartz, rutile and sphene. Hornblende comprises 60 to 90% of the rock; no pyroxene was observed. Quartzofeldspathic gneiss is generally tonalitic in composition with a color index of 15 to 50. The screen at Frog

Lake contains abundant amphibolite with lesser calc-silicate gneiss. The screen at Jack's Creek contains quartzofeldspathic gneiss and calc-silicate gneiss. This screen is lighter in apearance and contains more hornblende and less biotite.

The supracrustal pods occur within hornblende-biotite tonalite gneiss. Where inclusions are least abundant, this tonalite is very similar in appearance to the Jack's Creek Pluton. More commonly, it has a variable composition and gneissic fabric. Color index varies from 10 to 45. Hornblende-rich layers are common at Frog Lake; biotite-rich layers are common at Jack's Creek.

2.2.1.b Little Goose Creek Complex

Meter to ten meter scale sheets of fine-grained, mafic gneiss are seen throughout the Little Goose Creek complex except in the north-eastern portion of the unit. These gneisses are comprised of hornblende, andesine ± biotite, diopside, alkali feldspar, quartz and opaque, in proportions suggesting they have andesitic to basaltic compositions. The rocks have a fine-grained, strained crystalloblastic texture. They are often associated with minor amounts of calc-silicate gneiss.

These gneisses of andesitic composition are distinguished from tonalitic layers in the Little Goose Creek complex by the presence of a crystalloblastic fabric and/or the presence of diopside or associated calc-silicate gneisses. Some of the tonalitic layers in the complex may be either mylonitized supracrustal rock or intrusive.

2.2.1.c Origin and Correlation

The gneisses described above are inferred to be metamorphosed volcanic, volcaniclastic and sedimentary rocks of an oceanic island arc. The mineralogy of the mafic gneisses and amphibolites suggests that they have compositions that are basaltic to andesitic. Two chemical analyses corroborate this (Table 4.5). The fine scale layering and presence of calc-silicate, marble and quartzite layers suggest that these rocks are of supracrustal origin. The presence of small oscillatory zoned plagioclase crystals in the Mafic Layered Gneiss suggests they had or were derived from a fine-grained igneous protolith.

While the supracrustal rocks in the Mafic Layered Gneiss, Tonalite with Gneiss Pods and Little Goose Creek complex are generally similar, there are some differences. Leucocratic gneiss layers inferred to be sheets of metamorphosed trondhjemite are seen only in the Mafic Layered Gneiss. Clinozoisite is present in the Mafic Layered Gneiss; scapolite is present in Tonalite with Gneiss Pods; hornblende and diopside are more abundant and epidote and scapolite generally absent in the Little Goose Creek complex. These mineralogical differences may reflect differences in composition or metamorphic conditions.

It is impossible to say with certainty that the gneisses are correlative as they show much outcrop-scale variation and occur in small fragments. Gneisses similar to those in the Mafic Layered Gneiss crop out to the west and north of the field area in the Pollock Mountain Amphibolite (Figure 1.3; Aliberti et al., 1988). The similar appearance and close proximity of these gneisses suggest they may be correlative. They contain similar mineral assemblages, including the presence of cummingtonite and two stage garnets (Selverstone et al., 1987). Zoned garnets in these rocks have compositions that imply formation during burial to 8-9 kbars at temperatures of 600-650° C (Selverstone et al., 1987). Alteration of hornblende to cummingtonite is consistent with metamorphism during burial.

Chemical analyses from the layered mafic gneisses are quite similar to analyses from the Riggins Group (Hamilton, 1963a). Na_2O contents and FeO/MgO ratios are more similar to Riggins Group than to Seven Devils Group (Vallier and Batiza, 1978).

2.2.2 Continentally-Derived Supracrustal Rocks

Metasedimentary rocks with crystalloblastic fabrics occur as small screens and pods throughout the Payette River complex and in two locations in the central Little Goose Creek complex. They are comprised of biotite schist, sillimanite-bearing schist, quartzofeldspathic gneiss, calc-silicate gneisses and quartzite, generally interlayered on a half meter scale. Layers as thick as 20 meters have been observed locally . Biotite schists are the most abundant metasedimentary material in the screens. These contain biotite, quartz, plagioclase ± sillimanite, muscovite, alkali feldspar, hornblende, garnet, epidote and opaque. In sillimanite-bearing schists plagioclase is less abundant and hornblende is absent. Biotite and sillimanite-bearing schists contain alkali feldspar, muscovite or both. Calc-silicate gneisses and calcareous quartzites contain varying proportions of some of the following minerals: diopside, quartz, scapolite, calcite, tremolite, garnet, sphene, wollastonite, and hornblende.

Within the Little Goose Creek complex these rocks occur as small deformed pods in the porphyritic orthogneiss. Contacts between metasedimentary rocks and porphyritic orthogneiss are tectonic. Garnet is common in the porphyritic orthogneiss near the pods. Sillimanite-bearing granite crops out near Duck Lake, suggesting that the porphyritic orthogneiss intruded the metasediments.

Within the Payette River complex metasedimentary rocks occur as pods and sheets, meters to hundreds of meters in length. They are intimately mixed with intrusive material ranging in composition from granite to tonalite. This material contains abundant garnet when associated with biotite and sillimanite schist, and hornblende when associated with calc-silicate gneisses, suggesting chemical interaction. The abundance of granitic melt within inclusions in tonalite suggests that the sediments began to melt. These melts may have mixed with tonalitic material forming the range of compositions seen in intrusives associated with the inclusions. The pods and sheets of metasedimentary rocks and the intimately associated intrusive material are mapped together as Metasedimentary Rocks and Hybrid Intrusives where they are of appropriate size. Based on airphoto interpretations made by Mel Kuntz at the U.S. Geological Survey, areas of iron stained rock have been delineated. This does not definitively separate metasedimentary from intrusive material but it does give the flavor of the outcrop pattern. Many other small screens of metasedimentary rock, too small to show on the map, are present throughout the Payette River complex.

2.2.2.a Origin and Correlation

The mineralogy of these schists and gneisses suggests that they are metamorphosed argillaceous and limey sediments with subordinate quartzite. The abundance of quartz suggests that they were cratonally derived. They are generally similar in composition to cratonally-derived metasedimentary rocks in the areas studied by Lund (1984), Hoover (1986) and Myers (1982). These rocks have been variously correlated with Precambrian Belt and Paleozoic miogeoclinal sedimentary rocks. The small size and high metamorphic grade of the sheets and pods area make such a correlation based on lithology impossible.

2.3 HAZARD CREEK COMPLEX

The Hazard Creek complex is characterized by the presence of abundant epidote-bearing, variably deformed, intrusive rocks. The lithologies and outcrop scale structure and deformational style change rapidly on scales from several meters to a kilometer. Rocks types comprise four main categories: layered gneiss, orthogneiss, tonalite and trondhjemite, and alkali feldspar-bearing intrusives. These rocks crop out in elongate north-south trending bands one or more kilometers in width. Some bands are comprised of a single lithologic type; however, many bands contain several rock types that form coarsely layered gneisses and/or injection breccias. Rock types found in units containing only one lithology are also commonly members of more complex units.

Orthogneisses with crystalloblastic fabric occur throughout the unit. The most completely recrystallized gneisses contain few remnants of the original igneous texture other than an occasional porphyroclast of zoned plagioclase. These older gneisses occur as large mappable bodies, as inclusions in younger intrusives, and as layers within the layered gneisses. Most crystalloblastic orthogneiss is tonalitic or quartz dioritic with a relatively high color index; however, lighter tonalitic and trondhjemitic gneiss is also present.

Younger, less deformed intrusives are epidote-biotite-hornblende tonalite, epidote-biotite tonalite and trondhjemite, granodiorite and granite. Epidote-biotite-hornblende tonalite is the most common rock type in the complex. A large body of slightly deformed epidote-biotite-hornblende tonalite, the Jack's Creek pluton, crops out in the central Hazard Creek complex. Epidote-biotite-hornblende tonalite is also abundant in many of the more complex units. More than one generation of epidote-biotite-hornblende tonalite is indicated by numerous cross-cutting and inclusion relationships. Tonalites on the west side of the Hazard Creek complex tend to have more biotite than those on the east. Alkali feldspar is generally absent but occurs in small amounts in the east.

Epidote-biotite tonalite and trondhjemite have a lower color index than the hornblende-bearing intrusives and may contain muscovite and/or garnet. A large, slightly deformed body of this rock type comprises much of the western Hazard Creek complex and may include one or more plutons. This unit contains abundant inclusions of epidote-biotite-hornblende tonalite. Locally, the two lithologies may be gradational; however, in these areas hornblende is usually subordinant to biotite in the darker rock. Epidote-biotite tonalite is also present as inclusions in granodiorite in the eastern

Hazard Creek complex. Crystalloblastic gneisses of similar composition are present in the mafic layered gneiss and in the gneiss complexes.

Granodiorites and granites are subordinate in abundance to tonalite, forming less than 5% of the complex. The alkali feldspar-bearing rocks commonly contain garnet. Epidote is much less common than in the tonalites and quartz diorites. The large bodies of alkali feldspar-bearing intrusives are the youngest rocks in the local sequence. Throughout the Hazard Creek complex, the youngest dikes and sills are alkali feldspar-bearing. These alkali feldspar-bearing rocks may reflect a distinctly younger period of plutonism.

Very few intrusive rocks in the complex have pristine igneous textures. Most samples combine features typical of hypidiomorphic, crystalloblastic and mylonitic textures. In the least deformed rocks, quartz and plagioclase grain boundaries have migrated, forming smooth, sutured boundaries (Figure 2.3a). This texture is referred to as a modified igneous texture. More deformed samples have small polygonal grains of quartz and feldspar along the boundaries of larger grains. These grains grow in size and increase in abundance until a crystalloblastic fabric is present. Samples where small polygonal grains are present are referred to as having an extensively modified igneous texture (Figure 2.3b). When polygonal grains dominate the texture, it is referred to as crystalloblastic (Figure 2.3c). This sequence demonstrates progressive recrystallization. These fabrics are inferred to form by grain boundary migration at temperatures where annealing outpaced strain. The presence of large quartz and feldspar grains that generally resemble igneous grains suggests strains accompanying recrystallization may have been small.

Some textures are transitional between those described above and mylonitic texture. These samples contain abundant polygonal grains but also contain strongly sheared quartz and undulatory plagioclase. Grain boundaries are usually jagged (Figure 2.3d). This texture may have formed where shear stresses were higher, where

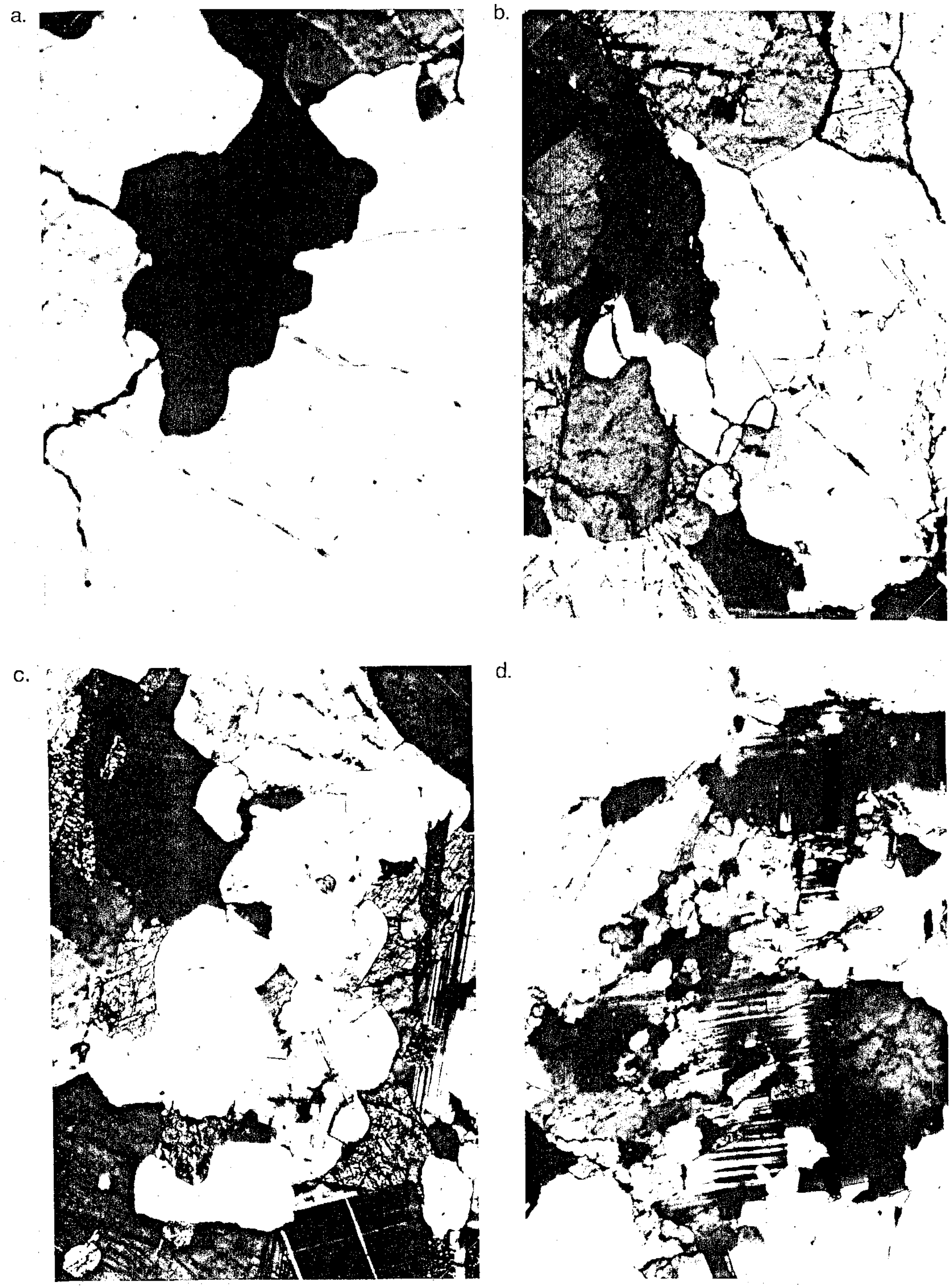


Figure 2.3 Textures observed in Hazard Creek Complex: a) modified igneous fabric: smooth, sutured boundaries with little evidence of unannealed strain; b) extensively modified igneous fabric: abundant small, polygonal grains of quartz and feldspar with little evidence of unannealed strain; c) crystalloblastic fabric; d) strained modified igneous fabric: strongly sheared quartz and feldspar with rough grain boundaries. Field of view 3.26 mm.

annealing was slower (possibly due to lower temperature or fluid pressure), or may reflect multiple episodes of deformation.

Mylonitic fabrics vary in intensity. The least deformed rocks have undulatory quartz in an otherwise annealed or hypidiomorphic fabric. In moderately to strongly deformed samples, quartz occurs as fine-grained mortar or as interlocking medium-sized grains. In many samples plagioclase has undulatory extinction and deformed twins. Locally micas have been deformed. More rarely, plagioclase is rounded and encased in a mortar of fine-grained material. Depending on the strength of the deformation, the igneous or crystalloblastic habit of mafic minerals and less commonly of plagioclase may be preserved. These textures formed in rocks where strain outpaced annealing.

The presence of fabrics that are gradational between igneous, mylonitic and crystalloblastic endmembers suggests that they formed in response to locally variable amounts and rates of strain. Some fabrics may represent more than one episode of strain; however, no cross-cutting fabrics were observed. Modified and extensively modified igneous textures are more common in the western portion of the complex, while mylonitic textures are more common in the eastern portion of the complex. Examples of each texture can be found throughout the complex. Samples with modified igneous texture were observed to occur within several meters of samples with extensively modified igneous or mylonitic textures. The distribution of fabrics is not well understood but does not appear to be strictly related to sample age or composition.

Despite the wide range in compositions and textures of the intrusive rocks within the Hazard Creek complex all but the alkali feldspar-bearing intrusives share many characteristics. Mafic minerals in rocks of all compositions have similar optical characteristics and habit, occurring together in intergrown clumps. Epidote is present in most rocks and occurs as large grains in these clumps. Plagioclase is generally well twinned and unzoned or normally zoned in rocks of high color index. In low color index rocks, oscillatory zoning becomes common.

The Hazard Creek complex is distinguished from the adjacent Little Goose Creek complex by the presence of abundant-epidote bearing orthogneisses. Alkali feldspar-bearing rocks are subordinant and the distinctive porphyritic orthogneiss of the Little Goose Creek complex is absent. The deformational style is more variable and the strong pervasive transposition seen in the Little Goose Creek complex is generally absent. Mylonitization similar to that in the Little Goose Creek complex affected rocks in the eastern Hazard Creek complex; intensity of deformation decreases to the west. This deformation is less pervasive even in the eastern-most Hazard Creek complex than in the Little Goose Creek complex. The contact between the two complexes is observed only near the headwaters of Hazard Creek. Lithologies of both complexes are tectonically interlayered on meter scales over several hundred meters. A similar relationship may be present in Little Goose Creek Canyon; however, poor outcrop makes it difficult to decipher. Elsewhere, the contact is unexposed and contained in major north-south striking stream valleys. North of Hazard Creek, basalt outcrops on one side of the stream, suggesting the contact in these locations is a post-Miocene fault. The linearity of the streams corroborates this interpretation.

Each of the ten units of intrusive or meta-intrusive rocks mapped within the Hazard Creek complex is described below. Three units, Jack's Creek Pluton, Epidote Biotite Tonalite and Round Valley Granite, are bodies of relatively undeformed intrusive with relatively few gneissic inclusions. Eastern Tonalite Gneiss and Quartz Diorite Orthogneiss are relatively homogeneous bodies of crysalloblastic orthogneiss. The remainder of the units consist of mixtures of different intrusives and gneisses. The units were distinguished both on the basis of their components and of the outcrop scale fabric and structure. Garnet Granodiorite Mylonite contains abundant granodiorite sills that form greater than 50% of the crop out. Light and Dark Gneiss consists of interlayered diorite to tonalite with color index greater than 30 and tonalite with color index less than 15. Epidote-Biotite Tonalite Gneiss has a low color index, strong gneissic fabric and contains abundant thin layers of biotite schist. Light Gneiss Assemblage and Dark Gneiss Assemblage contain multiple generations of intrusive rocks with abundant cross-cutting relationships. Dark Gneiss Assemblage is dominated by epidote-biotite-hornblende-tonalite. Epidote-biotite-tonalite is more abundant in the Light Gneiss Assemblage.

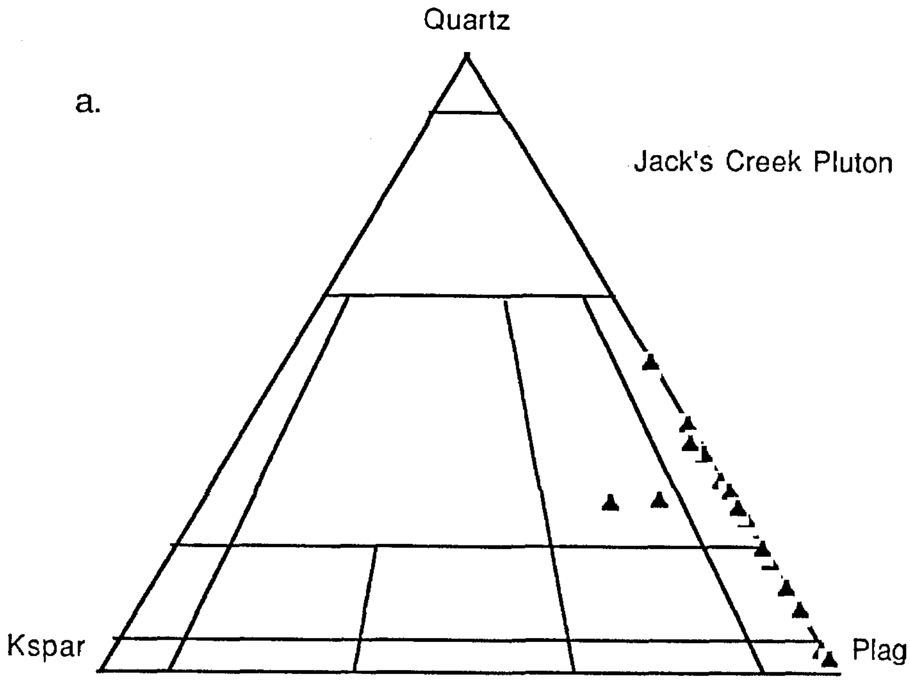
2.3.1 Jack's Creek Pluton

The Jack's Creek Pluton is one of the cleanest, least deformed intrusive bodies in the complex. It is comprised primarily of hornblende-biotite tonalite that usually contains a few percent of epidote (Figure 2.4). Alkali feldspar is generally rare but some granodiorites are present. Color index varies from 5 to 25. The hornblende:biotite varies widely, but is generally near 1. Plagioclase (An 25-40) is subhedral and is commonly normally zoned. Sphene, allanite, opaques, apatite and zircon are common accessory minerals. Sphene and opaque minerals are particularly common and can comprise several percent of the rock.

Epidote occurs in most samples; its abundance can vary from zero to five percent within several meters. Where epidote is in contact with biotite or hornblende, it is euhedral; a syplectic intergrowth with quartz occurs where epidote contacts plagioclase or quartz. Locally epidote appears to be growing at the expense of hornblende or, less commonly, of biotite. Epidote grains cross cut hornblende and biotite grains. The epidote is clear in plain light, has high birefringence and is biaxial negative with a large 2V. Brown allanite cores to epidote grains are not uncommon. Textural evidence clearly indicates that epidote grew late in the paragenetic sequence of the rock. This epidote is similar in texture to that described by Zen and Hammarstrom (1984) as primary magmatic epidote.

Outcrops of the Jack's Creek Pluton appear undeformed except near the margins of the pluton. However, in thin-section all samples show evidence of some subsolidus

a.



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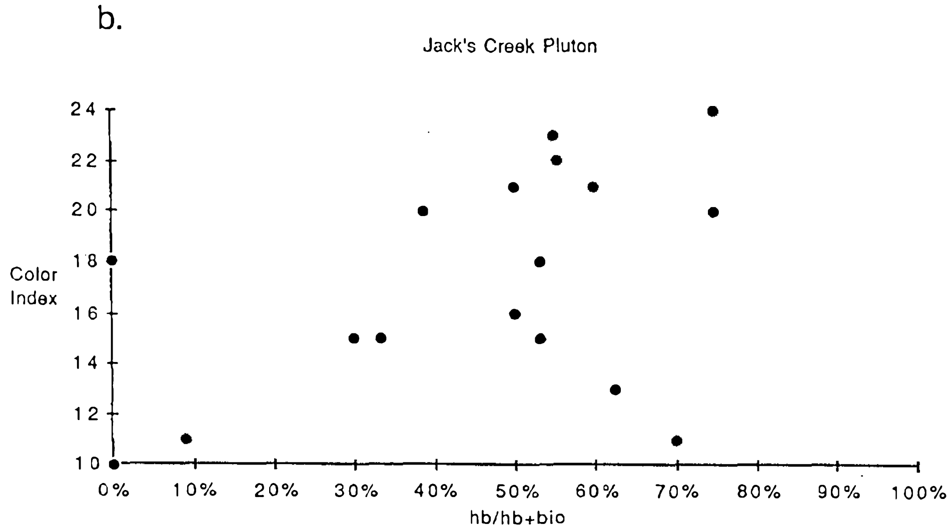


Figure 2.4 Estimated modal composition of Jack's Creek Pluton: a) classification after Streckeisen, 1973; b) color index versus hornblende/hornblende+biotite.

deformation. The intensity of deformation varies over several meters. Quartz is commonly undulatory or recrystallized into foliated subgrains. Plagioclase is less often deformed but can be rounded, undulatory or contain subgrains. Mafic minerals occur both as recrystallized aggregates or deformed grains and as clusters of subhedral or euhedral grains with intergrown hornblende and biotite. The later occurrence is inferred to be an igneous texture as it resembles the occurrence of these minerals in undeformed igneous rocks. Epidote and euhedral sphene also occur in these clusters, which are often aligned to form a foliation. Large grains of both hornblende and biotite within the clusters generally parallel the foliation. This fabric is inferred to be a relict igneous fabric because the mafic minerals in the clusters preserve their igneous habit and are aligned with the foliation. In the most deformed samples, fine-grained mafic minerals, mafic and feldspar augen, and elongate quartz grains define a mylonitic foliation. In some samples the clusters are preserved as augen shapped aggregates with mylonitically deformed finer grained mafic minerals as tails. Where both deformed and undeformed minerals are present the two foliations are parallel.

Pegmatites are common throughout the unit. The youngest pegmatites contain large salmon-colored alkali feldspar crystals. These pegmatites are generally undeformed or weakly deformed and both parallel and crosscut the foliation. They are inferred to be derived from the Garnet Granodiorite Mylonite because of their high potassium content. Epidote-bearing pegmatites are rare.

In the west central portion of the pluton the unit develops a gneissic fabric and slabby parting. This portion has been mapped as a subunit separate from the more massive, typical rock and is inferred to represent a deformed pluton margin that was rich in inclusions. Abundant gneissic layers of mafic-rich material that parallel the foliation are inferred to have originated as mafic inclusions and wall rocks. The slabby parting is due in part to a strong mylonitic fabric developed locally and to the high abundance of biotite; tonalite with more typical biotite concentrations does not exhibit parting even when strongly deformed.

The Jack's Creek Pluton intrudes the Quartz Diorite Gneiss along its southwestern margin. This border is marked by a zone of more mafic tonalitic gneiss that is inferred to have formed by injection and partial assimilation of the Quartz Diorite Gneiss by the tonalite. This zone is characterized by abundant ductile folding of foliation, and layering in the tonalitic gneiss and of quartz diorite gneiss inclusions. Pods of gabbroic material are common in this zone. These are composed primarily of hornblende and plagioclase with minor biotite and a variable amount of quartz. Locally, clinopyroxene cores the hornblende. The gabbros are coarse-grained and locally contain rhythmic layering (Figure 2.5). They are interpreted to have originated as cumulates which may be related to the quartz diorite orthogneiss; similar rocks are observed within that body. The pods generally have a selvage of lower color index, hornblende-rich material suggesting interaction with the Jack's Creek Pluton or the Quartz Diorite Orthogneiss. The tonalitic gneiss is similar to that found in the Tonalite with Gneiss Pods, which is in gradational contact with the Jack's Creek Pluton.

In the southeast the Jack's Creek Pluton is in contact with Garnet Granodiorite Mylonite. This contact is poorly exposed and somewhat enigmatic. It is inferred that the Garnet Granodiorite Mylonite intruded the Jack's Creek Pluton because the tonalite is crosscut by granitic pegmatites. The eastern contact of the Jack's Creek Pluton with the Eastern Tonalite Gneiss is sharp. It is locally a fault and may be one everywhere. The gneiss is distinguished by its higher color index and its strong foliation and lineation.

2.3.2 Epidote-Biotite Tonalite

Epidote-Biotite Tonalite is a relatively clean, slightly deformed intrusive lithology. It ranges in composition from tonalite to granite (Figure 2.6); however, rocks containing alkali feldspar are volumetrically minor. Color index is near 10%

Figure 2.5 Rhythmic layering in hornblende, plagioclase cumulate rocks near Granite Twin Lakes in the Hazard Creek Complex.

with biotite as the dominant mafic mineral. Small amounts of either muscovite or hornblende are also present. Epidote commonly forms 2 to 5% of the rock. This epidote is the same highly birefringent, subhedral epidote found within the Jack's Creek Pluton. It shows the same symplectic intergrowth with quartz and similar relationships to the other minerals. Allanite cores to epidote grains are common. Plagioclase often shows oscillatory or normal zoning. Apatite, sphene and zircon are common accessory minerals.

Most samples from this unit have modifed igneous or extensively modified igneous textures. The primary hypidiomorphic texture was medium to coarse-grained. Quartz always has undulatory extinction and is locally strongly deformed. Deformation of plagioclase or mafic minerals was only recognized in two samples from the eastern portion of the unit. Mafic minerals, occuring as subhedral to euhedral grains that are often intergrown, are aligned in a well developed foliation. In rocks with low color index the foliation is difficult to recognize in the field. This foliation is interpreted to be an igneous foliation because the grains occur in a habit typical of igneous rocks.

Leucocratitic garnet-epidote-muscovite-biotite trondhjemite dikes inferred to be late stage differentiates of the tonalite are common in the western portion of the unit. Pegmatites and aplites are abundant throughout the unit. They are particularly abundant in the western portion of the unit where they are accompanied by granite dikes. These are believed to be related to the Round Valley Granite.

This unit probably contains more than one pluton. The eastern side and northern extension of the unit contain few inclusions and probably comprise a single elongate pluton. The southwestern portion of the area has large areas of clean epidote-biotite tonalite but also contains many areas with abundant stope blocks of older tonalites and gneisses. One or more plutons may be present in this area. Along the Little Salmon River, an area of epidote-biotite tonalite with abundant inclusions is mapped as a separate unit and described below as Light Gneiss Assemblage.

Epidote Biotite Tonalite is in gradational contact with Light Gneiss Assemblage and Dark Gneiss Assemblage. It is in intrusive contact with Mafic Layered Gneiss and Granite Diorite Gneiss, Light and Dark Gneiss, and Tonalite with Gneiss Pods.

2.3.3 Round Valley Granite

Round Valley Granite is leucocratic muscovite granite with lesser granodiorite (Figure 2.6b). Muscovite forms three to seven percent of the rock occurring in books with interleaved biotite and an opaque mineral (ilmenite?). Epidote and garnet are present in small amounts. Alkali feldspar is microcline; plagioclase is anhedral oligoclase that has euhedral oscillatory zoning in some of the larger grains. Apatite and zircon are accessory phases. This unit is very similar in composition to the main granodiorite lithology in the Garnet Granodiorite Mylonite and to granodiorites in the Epidote-Biotite Tonalite (Figure 2.6b).

Samples have modified igneous or extensively modifed igneous textures. The original hypidiomorphic texture was medium to coarse-grained. In all samples quartz has undulatory extinction; two samples have strained plagioclase as well. Secondary mica is present in several samples.

The Round Valley Granite is the youngest intrusive in the western Hazard Creek complex. It has cross cut and incorporated large blocks of Epidote-Biotite Tonalite. Dikes of similar composition and texture crosscut Epidote-Biotite Tonalite, Mafic Layered Gneiss, Dark Gneiss Assemblage, and Light Gneiss Assemblage, and occur as far east as Bally Mountain. It is not clear if the Round Valley Granite is a differentiate of the Epidote-Biotite Tonalite or a distinctly younger intrusive. Granite in Epidote-Biotite Tonalite bears a strong resemblance to Round Valley Granite; both units are relatively leucocratic. Texturally, both units have similar modified igneous fabrics, are relatively coarse-grained and both contain oscillatory zoned plagioclase.

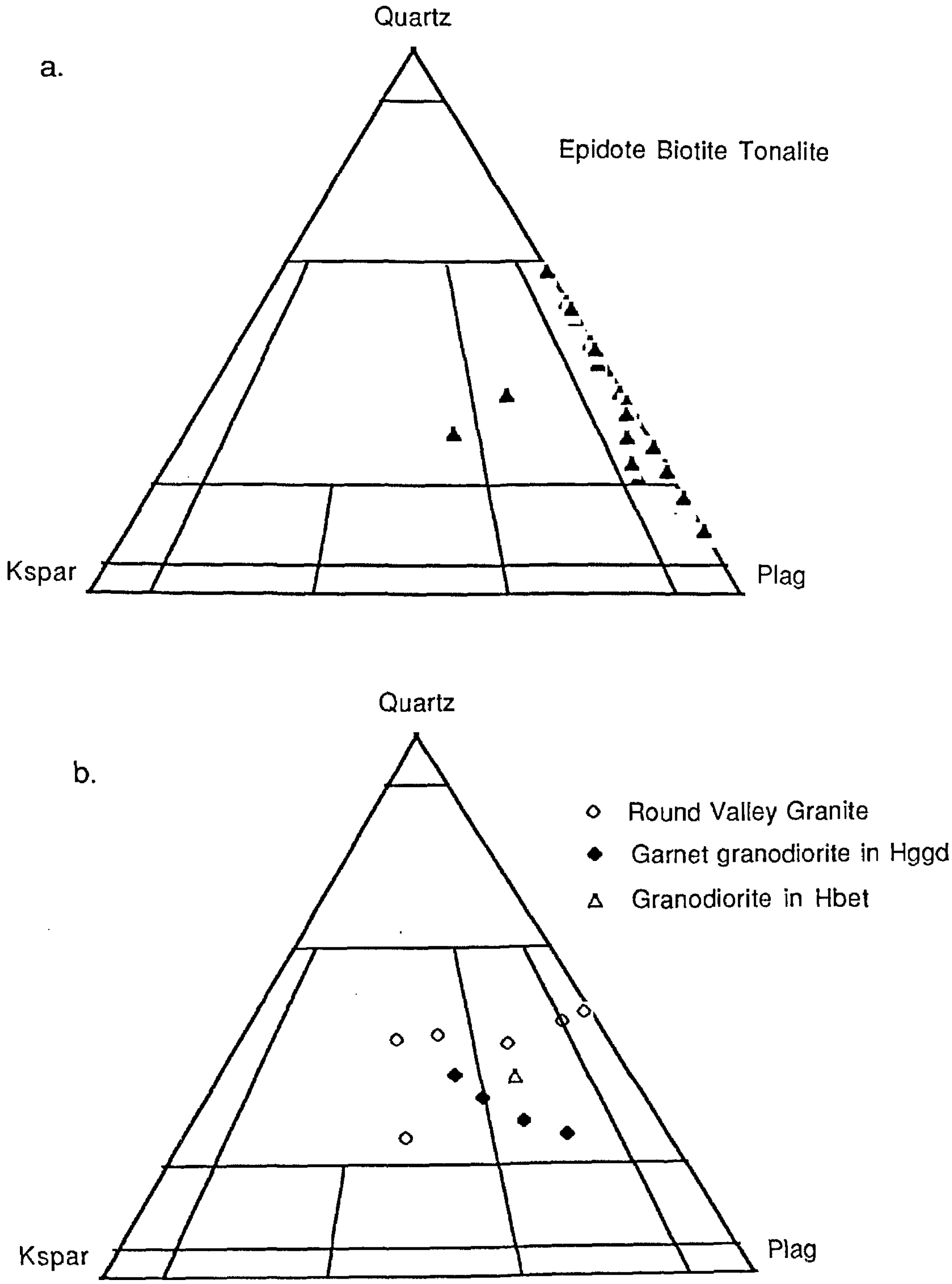


Figure 2.6 a) Classification based on estimated modes of Epidote Biotite Tonalite after Streckeisen, 1973; b) classification based on estimated modes of Round Valley Granite (after Streckeisen, 1973) and comparison with other potassium feldspar bearing intrusives of the Hazard Creek complex.

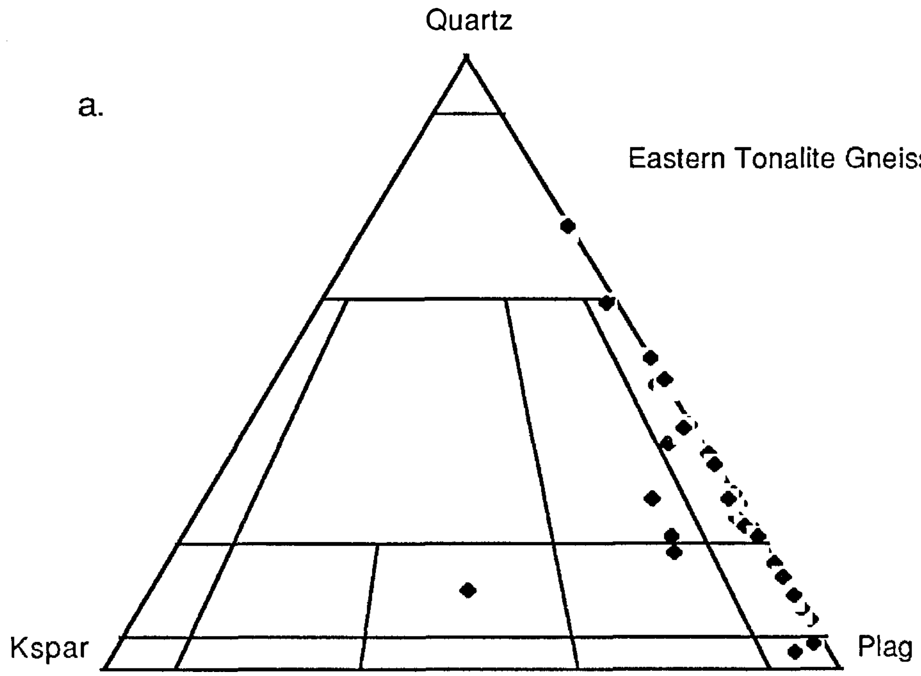
2.3.4 Eastern Tonalite Gneiss

Eastern Tonalite Gneiss is a mylonitic orthogneiss that ranges in composition from epidote-biotite granodiorite to epidote-hornblende diorite (Figure 2.7); tonalitic gneiss with a relatively high color index (20-35) is the dominant lithology. Hornblende:biotite varies considerably but is generally one or more in the dominant tonalitic gneiss (Figure 2.7b). Epidote is always present, varying in abundance from a trace to 10 percent. This epidote is optically and texturally identical to that in the intrusives described above. Allanite, sphene, apatite and zircon are common accessory phases. Opaque phases are uncommon. Leucocratic, alkali feldspar-bearing sills, generally one centimeter in width, are common. These sills are semiconcordant with the foliation and are often folded. Foliation is less commonly folded. Locally these sills crosscut the foliation indicating that the deformation in part predated sill intrusion.

Most samples from this unit have a medium-grained mylonitic fabric with deformed quartz, feldspar and mafic minerals. However, less deformed samples containing relict zoned plagioclase, are scattered throughout the unit, attesting to its igneous origin. Several samples have partially annealed crystalloblastic fabrics. It is not clear if the crystalloblastic fabric formed first and is partially deformed or if the mylonitic fabric has been partially annealed. In all cases mylonitic and crystalloblastic fabrics are parallel.

The tonalite gneiss is similar in composition to the mafic member of the Light and Dark Gneiss and to crystalloblastic gneisses in the Light Gneiss Assemblage and Dark Gneiss Assemblage. It is distinguished from the Western Orthogneiss of the Little Goose Creek complex to the east by the presence of epidote and generally lower alkali feldspar content (Figure 2.8).

a.



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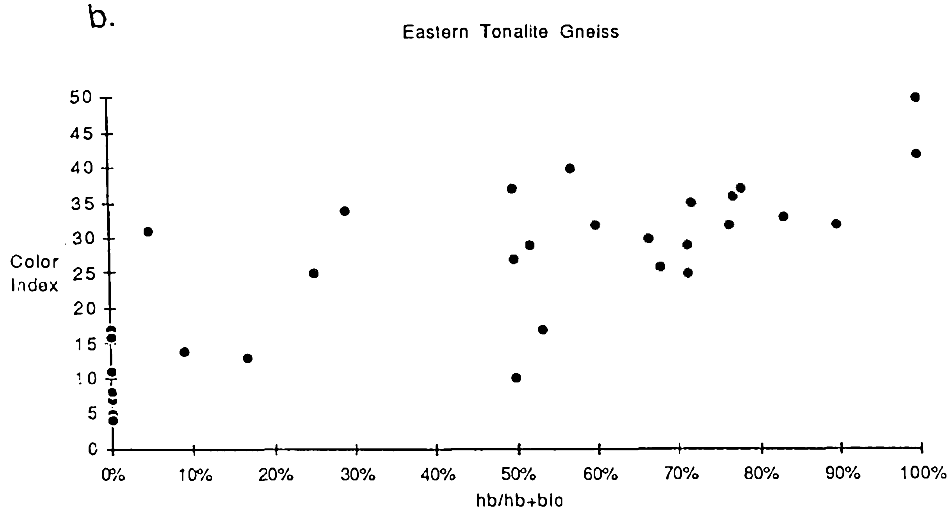

Figure 2.7 Estimated modal composition of Eastern Tonalite Gneiss: a) classification after Streckeisen, 1973; b) color index versus hornblende/hornblende+biotite.

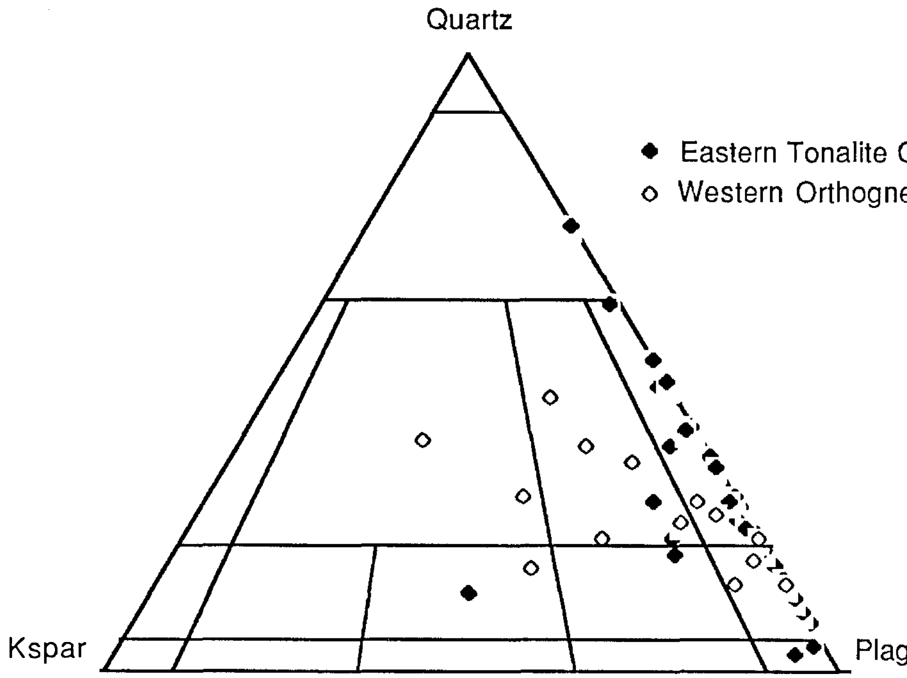

Figure 2.8 Comparison of estimated modal compositions of the Eastern Tonalite Gneiss of the Hazard Creek complex and the adjacent Tonalite Orthogneiss of the Little Goose Creek complex showing higher potassium feldspar content of the latter.

2.3.5 Quartz Diorite Gneiss

The Quartz Diorite Gneiss is primarily composed of epidote-biotite hornblende quartz diorite and tonalite with a very uniform composition and appearance (Figure 2.9). This rock can be recognized by its high color index, abundant epidote, and by the distinctive strong lineation formed by the alignment of hornblende and biotite within the foliation plane. The color index varies between 30 and 45; hornblende is more abundant than biotite (Figure 2.9b). Epidote forms up to seven percent of the rock occurring as subhedral to anhedral grains associated with mafic minerals. It is optically identical to the epidote in the Jack's Creek Pluton and Epidote-Biotite Tonalite. Symplectic intergrowths with quartz and inclusions of plagioclase are present but are much less common than in Jack's Creek Pluton or Epidote-Biotite Tonalite. Plagioclase is anhedral and often polygonal; zoning is rare. Quartz comprises less than 20 percent of the rock occurring as small equant grains scattered throughout the rock and as inclusions in other minerals. Sphene, allanite, zircon, apatite and an opaque are the accessory phases. Sphene and apatite are particularly abundant. Allanite occurs in the cores of epidote grains.

Most samples have a crystalloblastic texture in which strongly foliated mafic minerals are interlayered with plagioclase. Plagioclase is anhedral and often polygonal. Mylonitic deformation superimposed on this fabric is common. All samples have undulatory quartz and in samples with greater than 10% quartz it is deformed into interlocking subgrains. In several samples plagioclase and mafic minerals are also deformed.

Layers with lower color index and biotite as the dominant mafic are present locally. One to ten meter hornblendite and gabbro layers and pods similar to those in the

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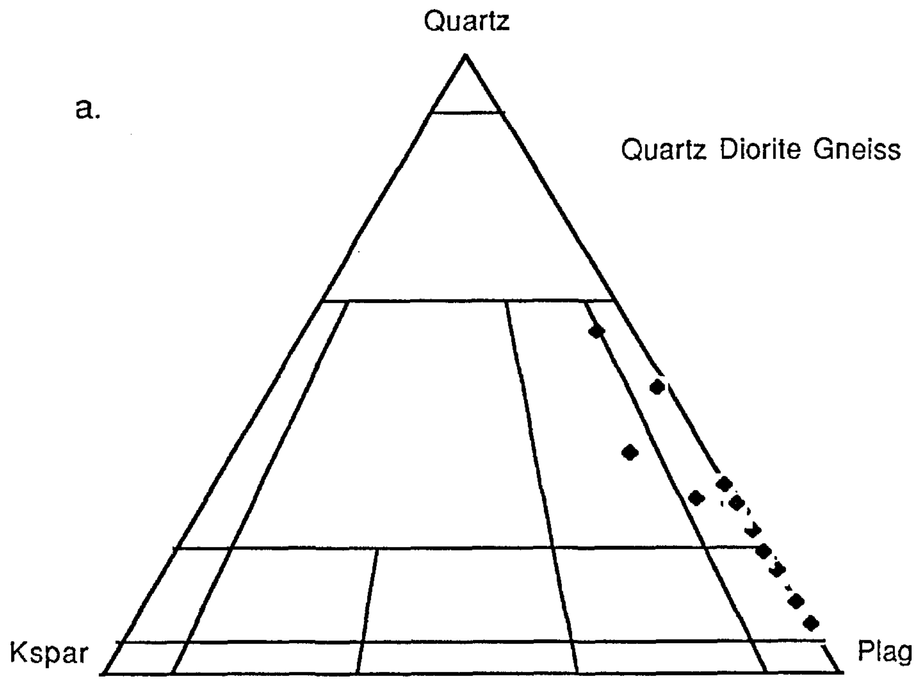

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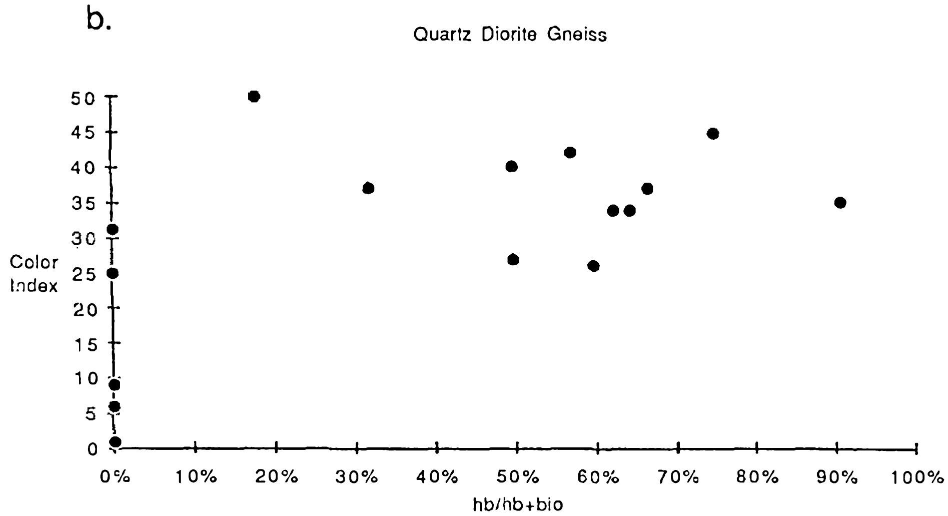

Figure 2.9 Estimated modal composition of Quartz Diorite Gneiss: a) classification after Streckeisen, 1973; b) color index versus hornblende/hornblende+biotite.

border zone of the Jack's Creek Pluton are also present and are particularly common just south of Granite Twin Lakes.

In the northern outcroppings of the unit, near Buck Lake, compositional and textural variations are more common with abundant interlayers of epidote hornblende tonalite gneiss and epidote-biotite tonalite gneiss. The rocks in this area were included in the Quartz Diorite Gneiss unit because they contain a significant component of typical Quartz Diorite Gneiss and because most other material has a high color index and epidote content.

Leucocratic veins several millimeters wide are common throughout the unit. They are roughly concordant but frequently cut across the foliation. Spacing between veins is commonly 20 centimeters. Large discordant pegmatitic veins one to several meters in width are also common. The youngest set of granitic pegmatites are related to the Garnet Granodiorite Mylonite to the east.

Quartz Diorite Gneiss is intruded by Garnet Granodiorite Mylonite to the south and east, and by Jack's Creek Pluton to the north.

2.3.6 Light and Dark Gneiss

The Light and Dark Gneiss is composed of light and dark orthogneisses interlayered on scales from one to several tens of meters. In several locations the light gneiss was observed to intrude the dark gneiss. Locally, the two are deformed and folded together.

The light gneiss usually accounts for 60 to 80 percent of the outcrop. It varies in composition from granite to tonalite (Figure 2.10). Color index is less than 25 with biotite commonly more abundant than hornblende (Figure 2.10b). Epidote forms zero to five percent of the rock. The texture varies from extensively modified igneous to mylonitic and from fine to coarse-grained. No samples preserving good hypidiomorphic textures were obtained. The light gneiss generally resembles the Epidote-Biotite

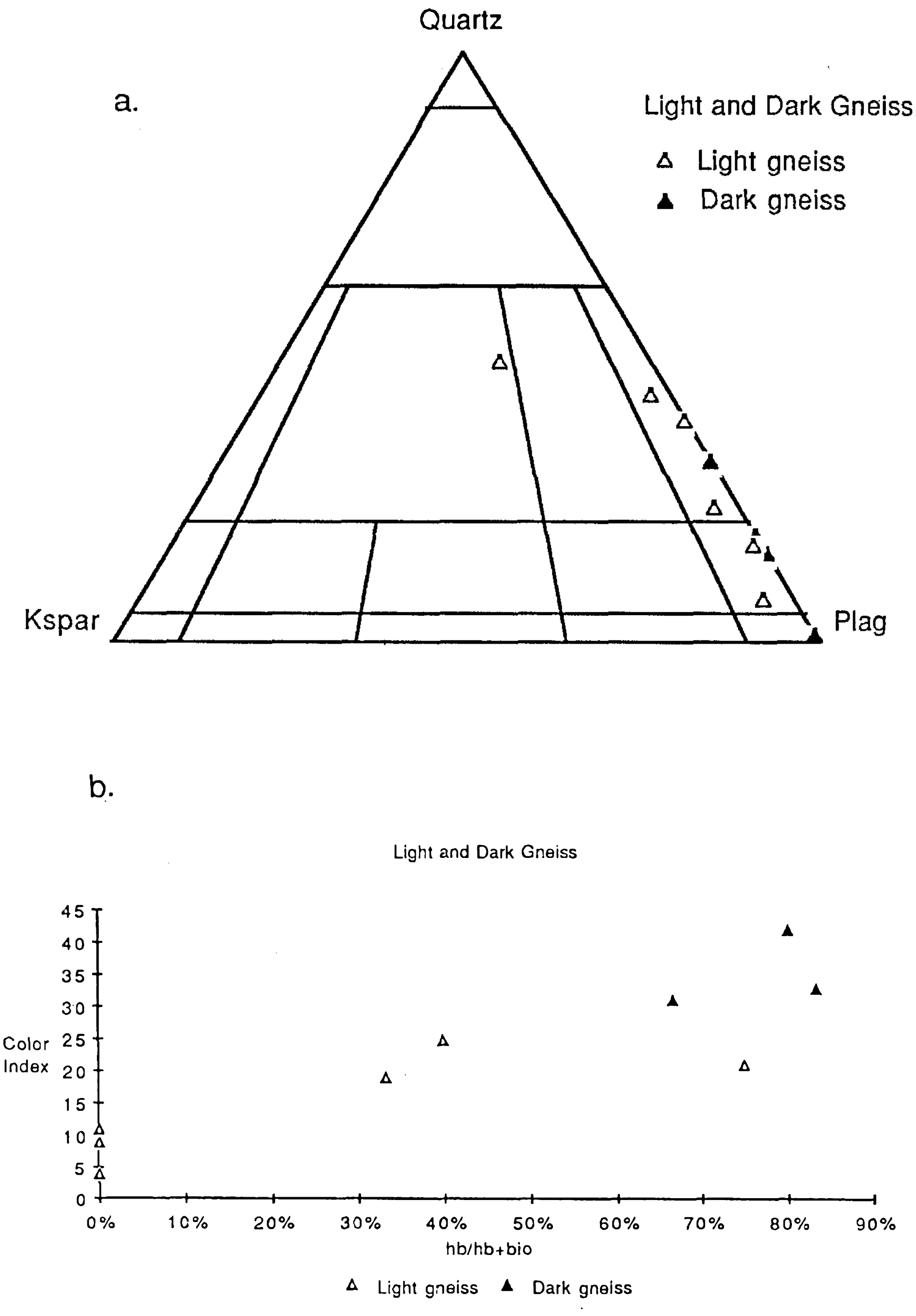

Figure 2.10 Estimated modal composition of Light and Dark Gneiss: a) classification after Streckeisen, 1973; b) color index versus hornblende/hornblende+biotite.

Tonalite mapped to the south and west; however, it contains less muscovite, lacks oscillatory zoning in the plagioclase, and is generally more deformed than the Epidote-Biotite Tonalite.

Outcrops of the light gneiss often appear undeformed; however, in some places they have a strong gneissic fabric similar to that observed in the Epidote-Biotite Tonalite Gneiss unit west of Buck Lake (described below). In these areas thick layers of dark gneiss are generally sparse, and thin biotite-rich layers more common. Thin biotite-rich layers are often concentrated in meter scale intervals.

The light gneiss may have formed from Epidote-Biotite Tonalite, an older intrusive of similar composition, or from several intrusives. Samples showing more pronounced gneissic layering may be older or may only be more deformed.

The dark orthogneiss varies in composition from tonalite to diorite (Figure 2.10a). Layers and pods of hornblendite and fine-grained biotite-hornblende tonalite gneiss are commonly associated with the dark orthogneiss. The orthogneiss has a color index greater than 30 with hornblende:biotite greater than one (Figure 2.10b). Epidote commonly forms up to seven percent of the rock. Textures vary from extensively modified igneous to crystalloblastic. Undulatory quartz and plagioclase were both observed; however, no well developed mylonitic fabrics were observed even in samples where mylonitic light gneiss is present nearby. This may be a reflection of the relatively low quartz content in the dark gneiss samples. The dark gneiss is similar to the Quartz Diorite Gneiss in composition; however, the distinctive fabric seen in the Quartz Diorite Gneiss was not observed anywhere in the dark gneiss. Quartz Diorite Gneiss crops out along strike both to the north and south of the Light and Dark Gneiss. This relationship and the general similarity in composition between the dark gneiss and Quartz Diorite Gneiss suggest that they may be equivalent.

Both light and dark gneisses are cut by fine-grained biotite tonalite dikes. These dikes have sharp non-planar boundaries and are between 10 and 50 centimeters wide.

Pegmatite dikes are the youngest rocks in the unit. These have straight boundaries and vary in width from 10 centimeters to a meter.

Near Disappointment Lake the Light and Dark Gneiss is blocky rather than interlayered. This area has been distinguished as a subunit on the map. The light gneiss lacks a strong foliation and intrudes dark gneiss as dikes with a wide range of orientations. Cross cutting pegmatites and fine-grained biotite dikes also lack a preferred orientation. The dark gneiss is massive when biotite is absent, where biotite is present it is well foliated though the orientation varies widely. Shallow dips are abundant in this area. Both biotite tonalite dikes and older inclusions of amphibolite and biotite tonalite gneiss are more abundant here than elsewhere in the unit.

2.3.7 Garnet Granodiorite Mylonite

Garnet Granodiorite Mylonite is composed of 50 to 80% garnet granodiorite sills, one to several tens of meters wide, which intrude quartz diorite gneiss and epidote-biotite tonalite. The unit name is assigned where garnet granodiorite forms more that half of the outcrop. Quartz Diorite Gneiss and Epidote-Biotite Tonalite are common as inclusions throughout the unit. Amphibolite and hornblendite pods and layers are also present particulary near the north end of the unit.

The garnet granodiorite protolith is leucocratic and varies in composition from tonalite to granite (Figure 2.11). Biotite forms up to seven percent of the rock. Muscovite, epidote and garnet occur in small amounts. Alkali feldspar augen up to 7 mm in length suggest the intrusive contained coarse-grained alkali feldspar. Plagioclase also occurs as augen; normal zoning is locally preserved. Allanite, apatite, and zircon are common accessory minerals. Mineralogical and textural evidence suggests this gneiss was a medium or coarse-grained intrusive.

Both the garnet granodiorite and the included epidote-biotite tonalite are strongly deformed into a mylonite gneiss. In all samples quartz occurs as fine-grained

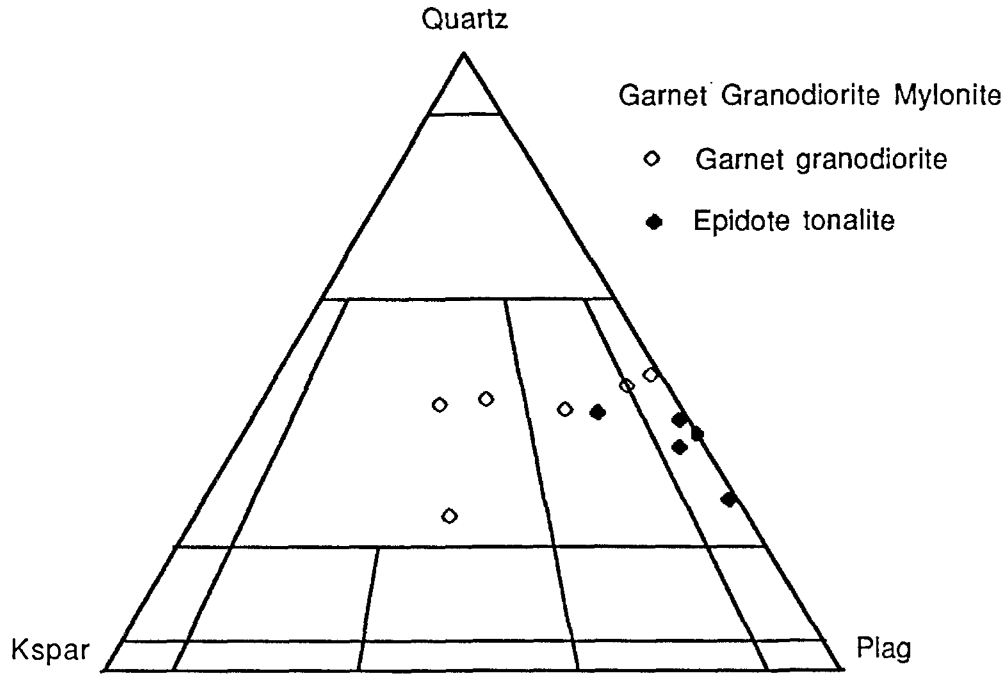

Figure 2.11 Classification based on estimated modes of garnet granodiorite and epidote tonalite in Garnet Granodiorite Mylonite Unit (after Streckeisen, 1973).

interlocking strained grains. Plagioclase is rounded and deforming into subgrains. Alkali feldspar occurs both as fine-grained elongate grains with undulatory extinction and as rounded augen. Myrmekite is abundant. Micas are bent, shredded, or recrystallized in trails of fine-grained material. Much folding occurred near the north end of the unit.

A ten meter layer of hornblendite crops out in the northern part of the unit extending for at least a kilometer south from Granite Twin Lakes. This body is composed of coarse-grained hornblende and plagioclase (C.I. 75-90) with minor biotite, opaque, allanite, apatite, sphene and zircon. The rock is inferred to have been a gabbro or cumulate similar to those preserved in the quartz diorite orthogneiss and in the contact zone with Jack's Creek Pluton.

2.3.8 Epidote-Biotite Tonalite Gneiss

Epidote-Biotite Tonalite Gneiss (Figure 2.12) has a low color index and strong gneissic fabric. Biotite is the primary mafic mineral. Hornblende or muscovite may be present. Epidote and garnet are generally present. Sphene and apatite occur as accessory phases. No zoning was observed in the plagioclase.

Texture is characterized by fine-grained material surrounding larger grains. The fine-grained material is weakly strained, has interlocking grain boundaries and shows no foliation. Biotite forms a strong foliation, occurring as medium-grained books that transect plagioclase grains (Figure 2.13). Formation of the biotite foliation appears to have postdated crystallization of plagioclase suggesting that the igneous protolith was deformed under conditions appropriate for the development of a crystalloblastic fabric. The abundance of fine grained material surrounding these larger grains suggests a mylonitic deformation overprinted the earlier fabrics. In most samples this mylonitic fabric has been partially annealed.

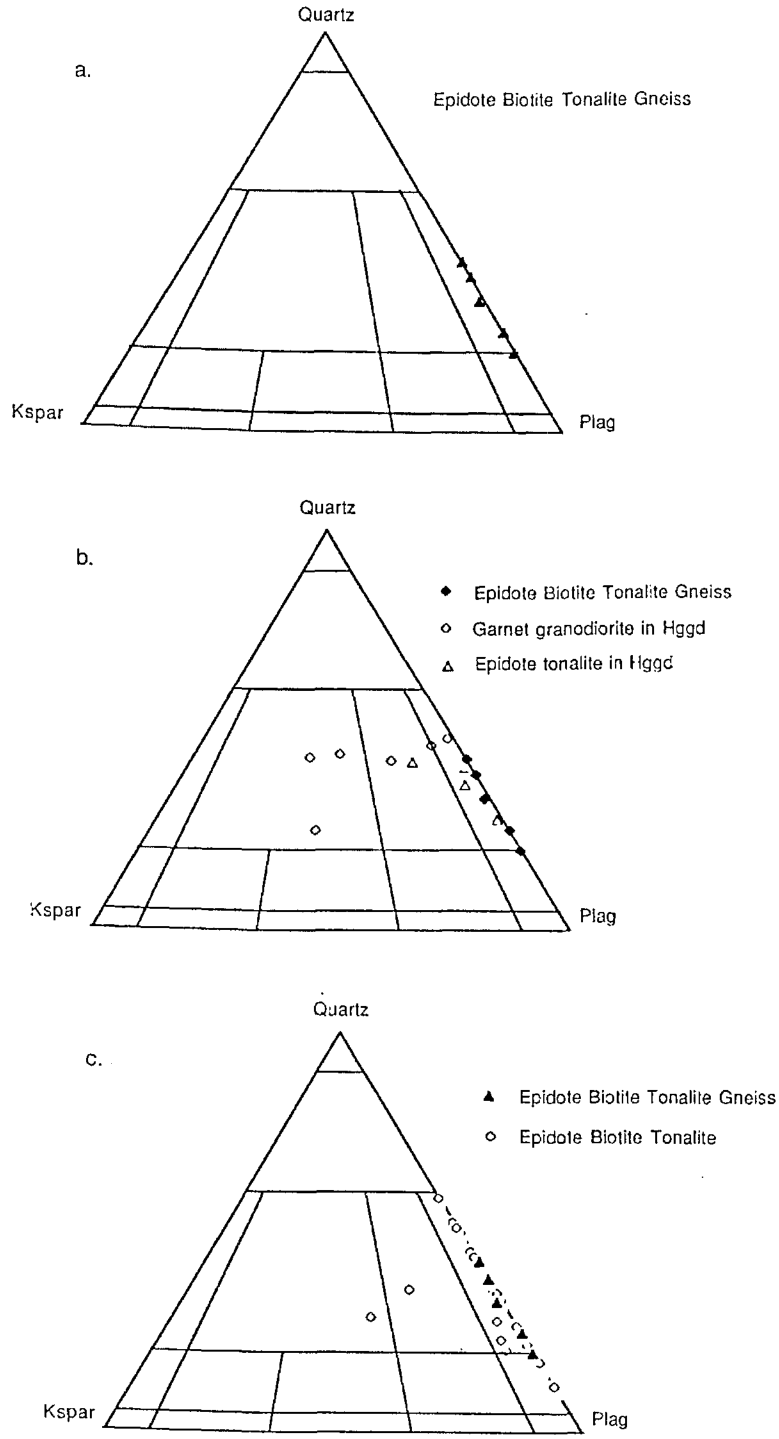

Figure 2.12 a) Classification of Epidote-Biotite Tonalite Gneiss (after Streckeisen, 1973); b) comparison with Garnet Granodiorite Mylonite which contains similar gabbroic pods but contains more potassium feldspar; c) comparison with Epidote-Biotite Tonalite which has similar composition but lacks strong deformed fabric.

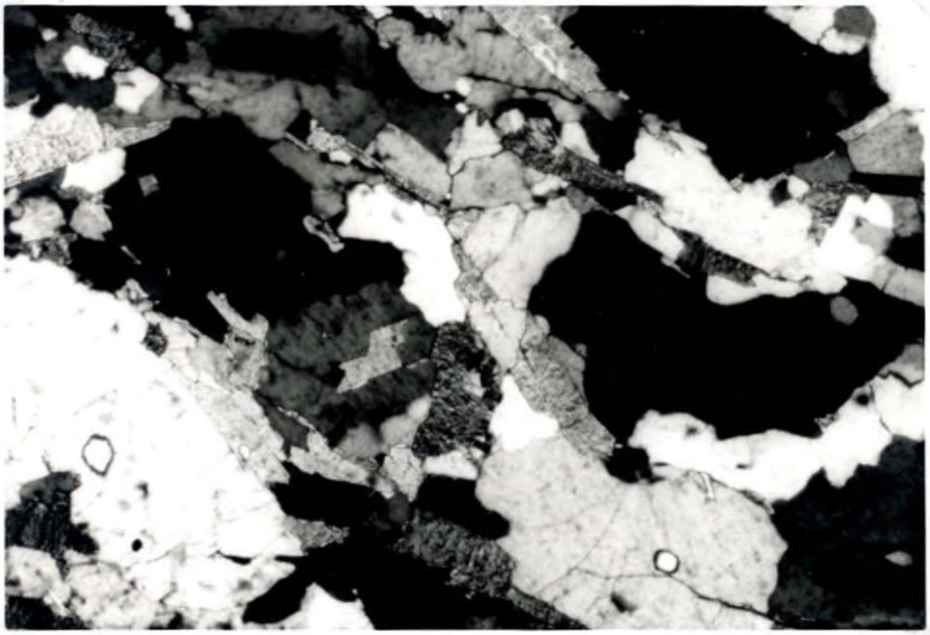

Figure 2.13 Photomicrograph of texture of Epidote-Biotite Tonalite Gneiss showing strong biotite foliation cross cutting plagioclase grains. Field of view 3.26 mm.

Biotite-rich layers and layers of epidote-hornblende tonalite gneiss are common, particularly near Buck Lake. Locally, the gneiss develops a pinstripe layering. Near Black Lake the gneiss is more variable where one to ten centimeter layers are common. Granite pegmatites that both parallel and crosscut the foliation are common near Black Lake and in the eastern portion of the crops out near Buck Lake. Hornblendite and gabbro pods are also abundant near Buck Lake. Gabbros contain both augite and hypersthene. No olivine was observed and one sample contained quartz. The samples are medium to coarse-grained with crystalloblastic fabrics. The similarity in composition to gabbroic pods in the Quartz Diorite Gneiss suggests these occurrences were also of igneous origin.

This unit is inferred to be a deformed intrusive body that contained abundant inclusions, screens and dikes. Some fine scale layering may be due to metamorphic differentiation. The presence of gabbroic pods in a light gneiss is reminiscent of the Garnet Granodiorite Mylonite. However, the composition of the Epidote-Biotite Tonalite Gneiss is more tonalitic (Figure 2.12b) and lacks the abundant intrusive relationships seen in Garnet Granodiorite Mylonite. It is similar in composition to Epidote-Biotite Tonalite (Figure 2.12c); however, the extensively modified fabrics suggest that this gneiss is older. It is probably most closely related to low color index crystalloblastic gneisses in the gneiss complexes to the west. Epidote-Biotite Tonalite Gneiss is interlayered with Eastern Tonalite Gneiss. Other contact relations are not exposed.

2.3.9 Dark Gneiss Assemblage

The main exposure of the Dark Gneiss Assemblage is along Hard Creek in the western Hazard Creek complex, where the unit contains most of the rock types seen in the Hazard Creek complex preserved both as interlayered gneisses and as intrusion breccias. The most abundant lithology in the unit is epidote-biotite-hornbende tonalite (Figure 2.14) similar in composition to the Jack's Creek Pluton. However, it is very

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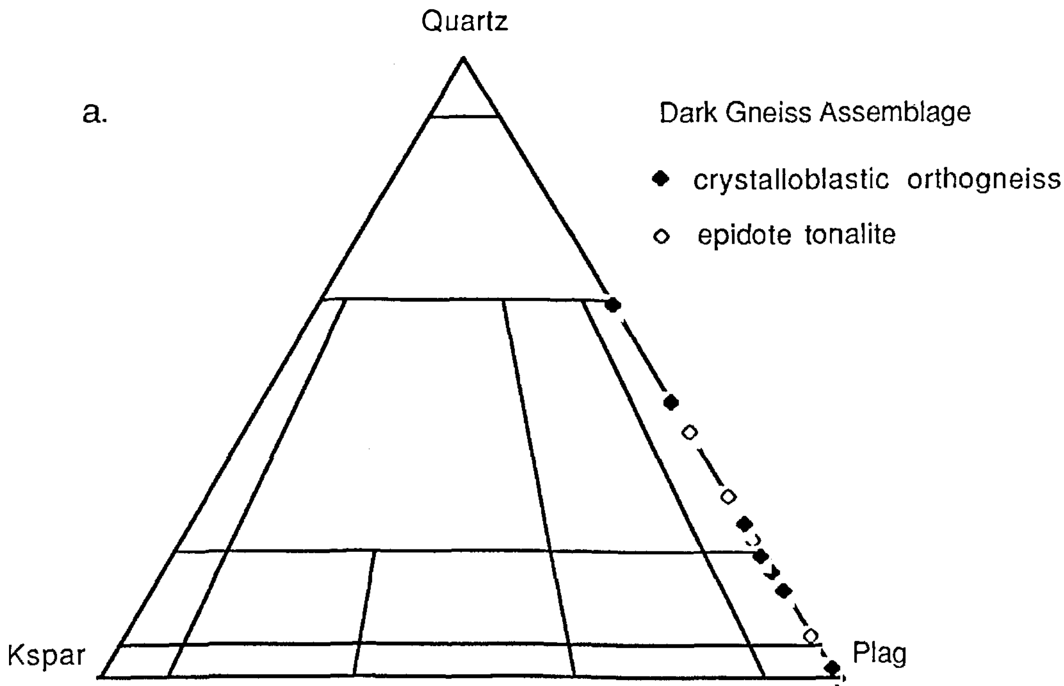

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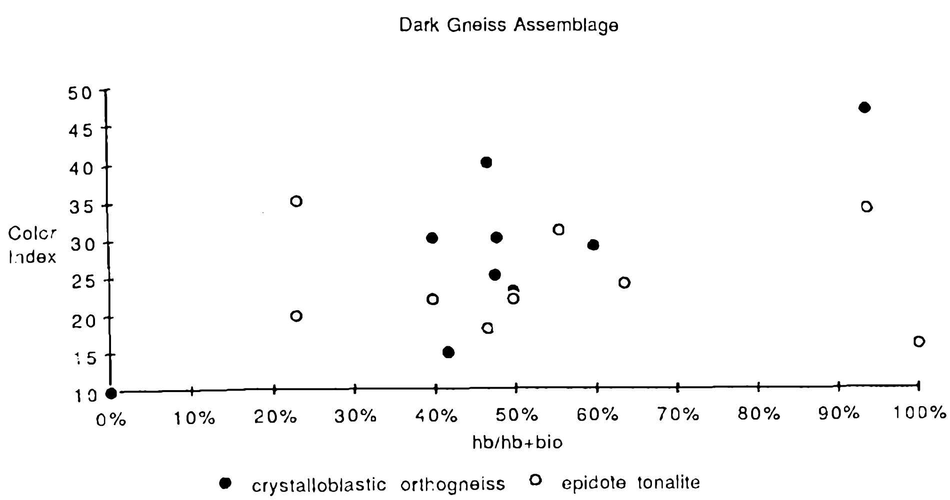

Figure 2.14 Estimated modal composition of Dark Gneiss Assemblage: a) classification after Streckeisen, 1973; b) color index versus hornblende/hornblende+biotite.

inhomogeneous, with abruptly varying color index and grain size; inclusions are abundant. Crystalloblastic orthogneiss formed from quartz diorite, diorite, and tonalite of both high and low color index is common. Some material similar in appearance to the Quartz Diorite Gneiss is present, especially on the eastern side of the unit. The oldest rocks in the complex are Mafic Layered Gneiss. Hornblendite and amphibolite are preserved as pods within other lithologies. Fine-grained tonalite is common both as inclusions and as cross-cutting dikes. The youngest dikes are trondhjemite, pegmatite and aplite similar to those that cross-cut the epidote-biotite tonalite.

Textures within the unit range from modified igneous to polygonal crystalloblastic. Fabrics preserve varying amounts of strain: some samples have only weakly undulatory quartz, others from the same vicinity have strongly strained quartz and feldspar. While some variations in texture appear to be due to the incorporation of older deformed rocks in younger intrusives, textures also appear to vary spatially over short distances. Samples from a single outcrop of epidote-biotite-hornblende tonalite have weakly to strongly modified igneous textures. Samples from a small area have annealed and mylonitic fabrics.

Rock types are both interlayered and agmatitic. It is common to see older interlayered gneisses, cross-cut by younger intrusives and dikes. Interlayering relationships may be due to intrusion along the foliation or to transposition. Several generations of rocks with similar composition are present in many places. It is not clear if the same generation of material is interlayered in some places and cross-cutting in others or whether two different generations are represented.

Rocks in three other areas are included in this unit. Two exposures south of Hard Creek contain abundant layered gneisses of varying composition. Layered gneisses are intruded by variably deformed epidote-biotite-hornblende tonalite, epidote-biotite tonalite and garnet-biotite granodiorite. These occur in areas of poor outcrop where field relationships are difficult to ascertain.

The third exposure is near Corral Lake where agmatite is formed from quartz diorite gneiss, biotite-hornblende epidote tonalite, pegmatites and minor schist. The oldest material present is strongly foliated quartz diorite gneiss, containing garnet and epidote, with a color index of 30 to 40. Small amounts of fine-grained biotite gneiss are included in the quartz diorite gneiss. This gneiss is broken into blocks varying in size from one to ten meters, by injections of epidote-hornblende-biotite tonalite. The dikes are several centimeters to a few meters in width. Large dikes often send off smaller apophysis into the blocks. Both rocks are locally deformed and folded; however, the overall impression is of a breccia, not a gneiss.

Two sets of pegmatites crosscut the breccia. The first set are tonalitic and are locally foliated and/or folded. The second, less abundant set of pegmatites contain large salmon-colored alkali feldspar crystals. These dikes follow the foliation but appear to be undeformed.

Textures vary from crystalloblastic to modified igneous. All samples have strained quartz and plagioclase; some have mylonitic textures. Grain boundaries in all samples are rough and irregular. These textures contrast to those in the gneiss complex along Hard Creek where grains usually have smooth irregular boundaries.

The edges of the screen are abrupt with few inclusions extending into the tonalite to the east or west. The tonalite in the screen may be related to the surrounding tonalite. It lacks the abundance of hornblende seen in the surrounding Jack's Creek Pluton and has a wispier texture due to the size and distribution of mafic minerals. These effects may be due to interaction with material in the screen. However, to the south where small pods of the screen are present, tonalite shows little chemical interaction, suggesting two generations of tonalite may be present.

The Corral Lake Screen is a small body containing relatively few rock types. It is believed to represent a small, well exposed example of the same relationships preserved in the main exposures of Dark Gneiss Assemblage along Hard Creek. The screen appears to represent a partially deformed older intrusion breccia which has been preserved as a large inclusion in younger tonalite.

2.3.10 Light Gneiss Assemblage

The Light Gneiss Assemblage is similar to the Dark Gneiss Assemblage, containing a wide variety of rock types preserved both as interlayered gneisses and as intrusion breccia . The most abundant lithology is epidote-biotite tonalite. Other lithologies are biotite-hornblende tonalite, epidote-biotite tonalite, biotite muscovite trondhjemite dikes, quartz diorite gneiss, biotite tonalite gneiss and layered gneiss. Hornblendite is uncommon. Epidote tonalite and trondhjemite dikes have modified or extensively modified igneous textures. Orthogneisses and layered gneisses have crystalloblastic fabrics. No mylonitic textures were observed in this unit.

Proportions of components vary considerably. Diking relationships and angular inclusions are abundant, giving the unit a blocky character. In some cases angular blocks are oriented parallel to the foliation. Blocks of older gneiss commonly contain more than one rock type. In this case the contacts within the block usually parallel the foliation.

The contact with Epidote-Biotite Tonalite is arbitrary and placed where angular inclusions become very abundant. There are similar inclusions in patches throughout the western exposure of Epidote-Biotite Tonalite. The contact with Dark Gneiss Assemblage is also arbitrary and represents a general change from a gneiss that is predominantly epidote-biotite-hornblende tonalite, to one that is predominantly epidote-biotite tonalite.

Samples of epidote tonalite and trondhjemite studied by Zen and Hammarstrom (1984) and Hammarstrom and Zen (1986), are from this unit which they refer to as the pluton near Round Valley.

2.3.11 Summary

The time and space relationships in the Hazard Creek complex are summarized in Figure 2.15. The oldest rocks are gneisses inferred to be metamorphosed oceanic-arc material which form the Mafic Layered Gneiss unit and pods within the Tonalite with Gneiss Pods. These gneisses are also common in the Light Gneiss and Dark Gneiss Assemblages. They were intruded by quartz diorite and tonalite that have been metamorphosed to crystalloblastic gneiss. In the western part of the complex these crystalloblastic gneisses are present as sheets and blocks in the Dark Gneiss Assemblage. In the central and eastern part of the complex they form mappable units: Epidote-Biotite Tonalite Gneiss, Quartz Diorite Gneiss and Eastern Tonalite Gneiss, as well as parts of the Light and Dark Gneiss. The relative age of these gneisses is unknown. Cross-cutting and inclusion relationships in the Dark Gneiss Assemblage show that more than one generation of crystalloblastic orthogneisses is present. Crystalloblastic orthogneisses are intruded by less deformed tonalite which form the Epidote-Biotite Tonalite in the west and the Jack's Creek Pluton in the east. These units are large, relatively inclusion-free bodies. Similar less deformed tonalite is abundant in Light Gneiss and Dark Gneiss Assemblages, Light and Dark Gneiss, and Tonalite with Gneiss Pods. The youngest rocks in the complex are alkali feldspar bearing: Round Valley Granite in the west and Garnet Granodiorite Mylonite in the east. Young granite and pegmatite dikes throughout the complex are inferred to be derived from these intrusions.

2.3.12 Magmatic Epidote

One of the most striking characteristics of the Hazard Creek complex is the abundance of epidote in all of the intrusive units except the youngest granites and granodiorites. The epidote occurs as large grains with some euhedral boundaries. Several samples from the Hazard Creek complex have been studied by Zen and Hammerstrom (1984) as part of their study of primary igneous epidote. They conclude,

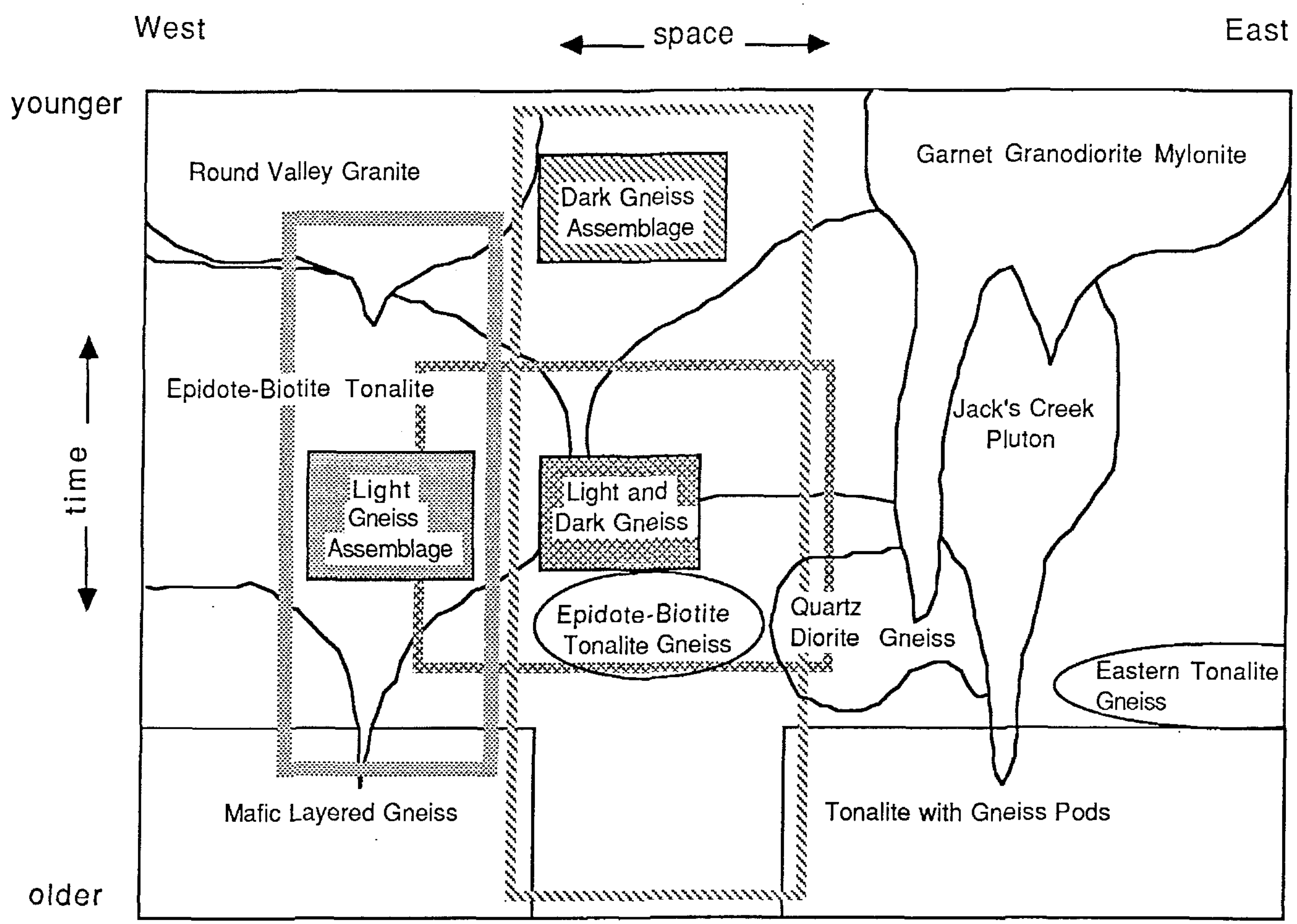

Figure 2.15 Time and space relationships in Hazard Creek complex.

on the basis of textural arguments, that the epidote is a primary magmatic phase of the intrusives. Observations from this study show that epidote is clearly late in the paragenetic sequence. Optically similar epidote is present in samples having modified igneous texture, extensively modified igneous texture and crystalloblastic texture. Zen and Hammarstrom (1984) concluded that experimental data (Naney, 1983; Liou, 1973; Holdaway, 1972) constrain formation of the epidote-bearing assemblages to pressures greater than 6 kbar and most likely greater than 8 kbar (25 km).

2.4 LITTLE GOOSE CREEK COMPLEX

The Little Goose Creek complex is comprised of interlayered intrusive rocks, layered gneisses and metasedimentary rocks that have been deformed together into mylonitic gneisses. The most abundant and distinctive of these gneisses is a microcline megacrystic granodiorite gneiss. All rocks within the Little Goose Creek complex have been strongly mylonitized. Intrusive contacts have been transposed and earlier structures and fabrics obliterated in most cases.

The Little Goose Creek complex is distinguished from the Payette River complex by the presence of the microcline megacrystic granodiorite gneiss. In most places the Payette River complex lacks the strong mylonitic deformation found throughout the Little Goose Creek complex. The contact between the two complexes is exposed only in the area near Brundage Mountain. Here the two complexes are tectonically interlayered and interfingered over a few hundred meters. Mylonitic deformation continues several hundreds of meters into the Payette River Tonalite. In most places undeformed Payette River Tonalite is juxtaposed abruptly against strongly deformed Little Goose Creek complex. The contact is often in a major stream valley and marked by several meters of no outcrop. Locally basalt occurs along the contact suggesting that the contact is a post-Miocene fault.

The Coarse-Grained Tonalite Orthogneiss in the Little Goose Creek complex is mineralogically and chemically (see analyses in Chaper 4) similar to the Payette River Tonalite. The mafic minerals where undeformed also have the same textures in both units suggesting that the Coarse-Grained Tonalite Orthogneiss represents a sheet of Payette River Tonalite emplaced in the Little Goose Creek complex prior to the mylonitic deformation.

2.4.1 Porphyritic Orthogneiss

The most abundant and most distinctive rock type in the complex is a microcline megacrystic granodiorote gneiss (Figure 2.16) called porphyritic orthogneiss. Prominent microcline megacrysts are commonly three to five centimeters in length and two to three centimeters wide. Near Hazard Lake they are larger: five to ten centimeters in length and five centimeters in width. The phenocrysts contain crystallographically oriented, euhedral inclusions of biotite, plagioclase, and quartz that are up to two millimeters in length (Figure 2.17). The presence of these undeformed inclusions within the strongly deformed gneiss suggests that the megacrysts formed as phenocrysts in the igneous protolith. Alkali feldspar also occurs as fine-grained recrystallized material and as deformed elongate grains up to two millimeters in length. Smaller megacrysts of plagioclase, up to one centimeter in length, occasionally show normal zoning with antiperthitic exsolution, also suggesting an igneous origin. Most plagioclase is untwinned with undulatory extinction. The gneiss has a color index between 5 and 15 with biotite as the most common mafic mineral (Figure 2.16b); garnet, hornblende and muscovite are present locally. Apatite, sphene, allanite and zircon are accessory phases; sphene and allanite are more common in rocks containing hornblende.

The porphyritic orthogneiss is interlayered with other orthogneisses on centimeter to hundred meter scales. These sheets, which account for approximately 30% of the unit in the north increasing to 70% in the south, vary in composition from

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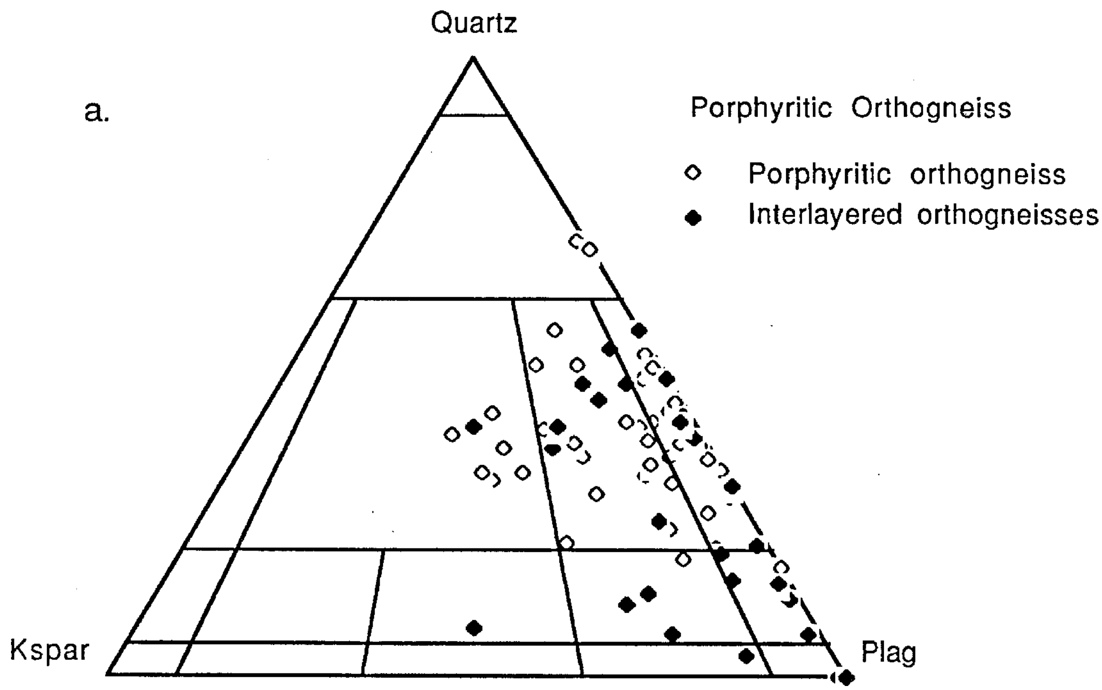

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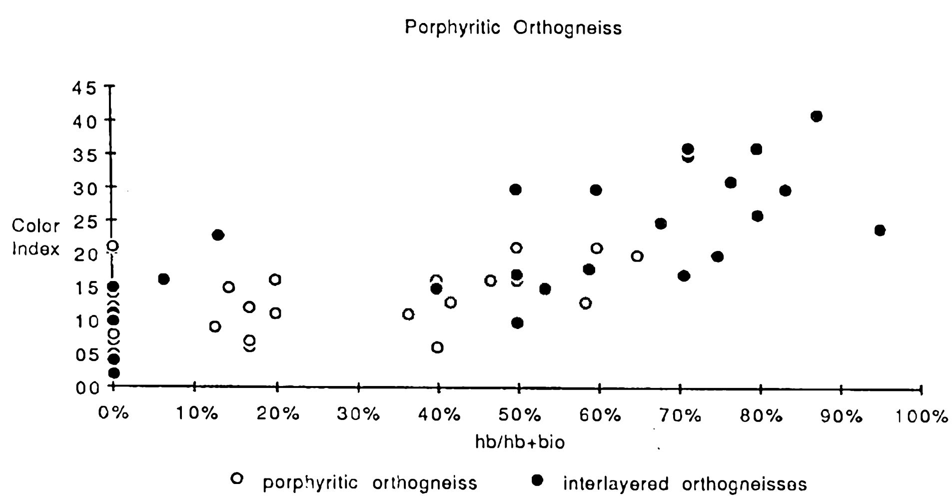


Figure 2.16 Estimated modal composition of porphyritic orthogneiss and interlayered orthogneisses in Porphyritic Orthogneiss Unit: a) classification after Streckeisen, 1973; b) color index versus hornblende/hornblende+biotite.

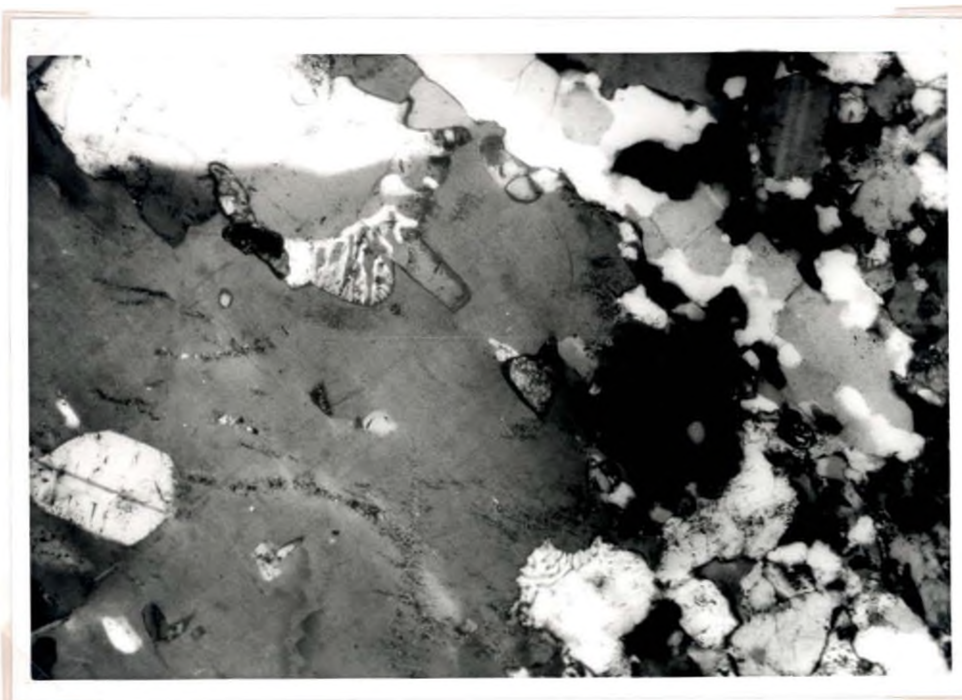


Figure 2.17 Photomicrograph of phenocryst in porphyritic orthogneiss showing euhedral oriented plagioclase and biotite inclusions and lobes of myrmekite. Field of view 3.26 mm.

gabbro to granite (Figure 2.16). In the north, the most abundant layers are tonalite gneisses similar to the tonalitic units described below (Coarse-Grained Tonalite Orthogneiss, Western Orthogneiss, Spotted Orthogneiss). These commonly occur in hundred meter wide cyclical exposure sequences with individual layers centimeters to tens of meters wide. Where they contain hornblende, it is also commonly found in the porphyritic orthogneiss suggesting chemical interaction. In the south, thicker layers of granitic to tonalitic orthogneiss are more abundant. Some of these rock types are similar in composition to the porphyritic orthogneiss but lack the megacrysts. Tonalitic and granodioritic augen gneisses are also abundant, containing plagioclase augen one to two centimeters in length. Deformation has destroyed the primary intrusive relationships between the different orthogneisses.

Meter scale pods of mafic rocks are often present within the tonalitic orthogneisses and more rarely within the porphyritic orthogneiss. Most pods are coarse-grained hornblende gabbro. Locally clinopyroxene cores the hornblende. Biotite is present in small quantities and becomes more abundant near the edges of the pods. The pods are always sheathed in tonalite which is inferred to be a reaction zone. Mafic minerals in the pods are strongly foliated; felsic minerals are strained and deformed. Fine-grained football sized mafic inclusions within the tonalitic orthogneisses have the same mineralogy but a higher color index than the host tonalite. They are inferred to be mafic inclusions emplaced with the tonalites and attest further to the intrusive igneous nature of the protolith to the orthogneiss.

Many sheets also contain cross-cutting fine-grained tonalitic or leucocratic dikes. In a few cases, these continue out into the surrounding porphyritic orthogneiss, but in most cases the screen is separated from microcline megacrystic orthogneiss by a shear surface.

All rocks in the unit have a mylonitic fabric and are strongly foliated and lineated. Larger plagioclase and alkali feldspar crystals are progressively deformed into augen. In most samples some plagioclase has recrystallized into small polygonal grains. Abundant lobes of myrmekite have formed on the edges of alkali feldspar megacrysts. It is common to see fragments of myrmekite in the fine-grained sheath surrounding megacrysts. Myrmekite is not observed where plagioclase is included in alkali feldspar. The myrmekite lobes are similar to those described by Simpson (1985) as developing during mylonitization. Quartz occurs as medium to fine-grained anhedra with sutured grain boundaries and undulatory extinction. In a few samples coarse-grained quartz with strongly undulatory extinction is preserved. Biotite generally occurs in small grains that have recrystallized. Some shredded or bent fragments of biotite and hornblende are present. In samples with higher color index, larger grains and aggregates of biotite and hornblende are preserved. These aggregates are generally eye-shaped with tails of finer grained mafic material. A few samples preserve relatively coarse-grained fabrics with mafic minerals in their igneous habit. When both coarse and fine-grained biotite are present they are optically similar and no ilmenite is observed to exsolve, suggesting that biotite composition remained constant during deformation.

The textures form a continuum between very fine-grained gneisses with little annealing, and largely recrystallized gneisses. The most abundant texture is medium to fine-grained with some annealed plagioclase and biotite. The formation of abundant myrmekite, the absence of fracturing from all minerals, and the presence of annealed grains in most samples suggests that deformation took place under middle amphibolite facies conditions (Simpson, 1985).

There is no apparent geographical pattern to the intensity of deformation, which varies locally throughout the complex. Fabrics are stronger in rocks with more quartz; however, the intensity of deformation also varies between rocks of equal quartz content. Gabbroic and quartz dioritic gneisses commonly contain local zones of intense deformation. Very fine-grained textures with little annealing are more abundant in the southern portion of the complex. The variations in fabric are inferred to reflect variability in local conditions: amount and rate of strain, temperature and water content.

2.4.2 Coarse-Grained Tonalite Orthogneiss

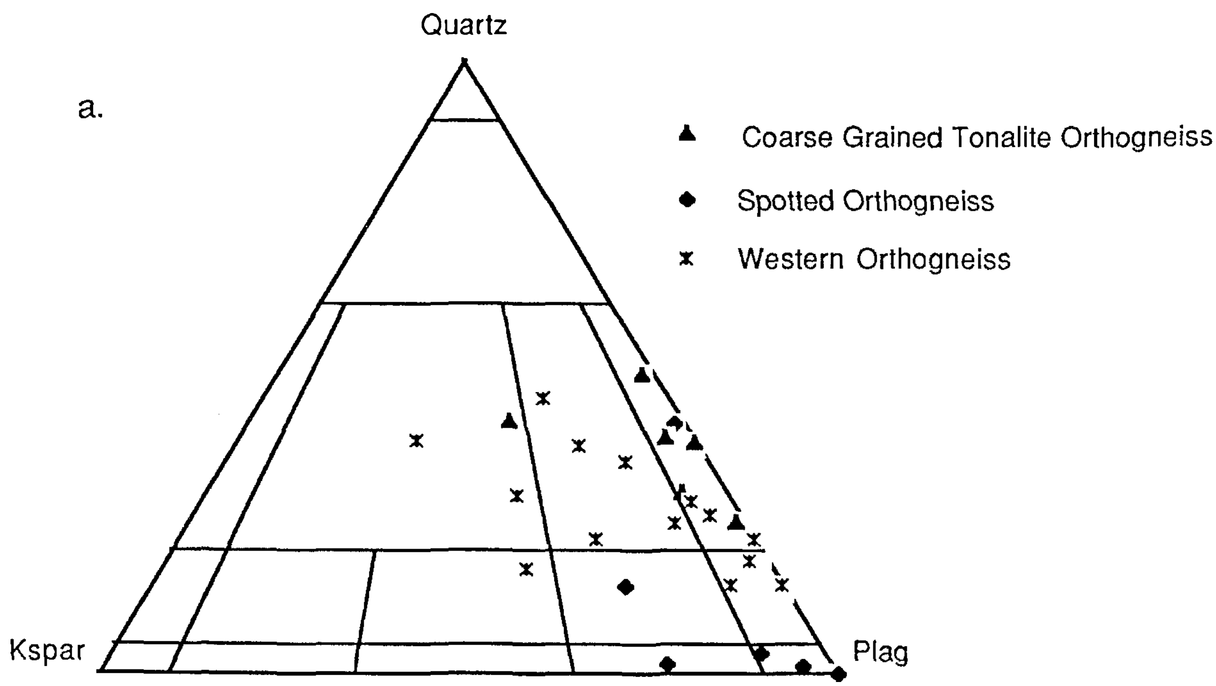
Coarse-Grained Tonalite Orthogneiss typically has a larger grain size than other tonalitic rocks in the complex. In particular, the mafic minerals occur as large grains and aggregates of grains several centimeters in length. Hornblende and biotite are equally abundant (C.I. = 20, Figure 2.18). Plagioclase is andesine which is not zoned or more rarely normally zoned. Orthoclase is present in one sample. Sphene, allanite, opaque, apatite and zircon are the accessory phases. Mafic inclusions are sparsely distributed throughout the unit. Pegmatite dikes are abundant. This tonalite is tentatively correlated with the Payette River Tonalite, as they are comparable in composition and have similar habit of mafic minerals.

The unit has been mylonitized, causing a general decrease in grain size. Quartz is deformed into elongate subgrains with sutured boundaries. Plagioclase occurs both as two to four centimeter augen and as fine-grained annealed material. Fine-grained annealed mafic minerals are present as sheathes around larger grains and aggregates. Contacts between the Coarse-Grained Tonalite Orthogneiss and surrounding Porphyritic Orthogneiss are tectonically transposed; the two units are interlayered over 20 to 100 meters.

2.4.3 Spotted Orthogneiss

Spotted Orthogneiss is distinguished by large, one to two centimeter aggregates of mafic minerals that give the gneiss a spotted appearance. The gneiss varies in composition from diorite to monzodiorite and granodiorite (Figure 2.18). Color index varies from 25 to 40 with both hornblende and biotite present in the clots (Figure 2.18b). It is generally quartz poor. Sphene and apatite are abundant.

a.



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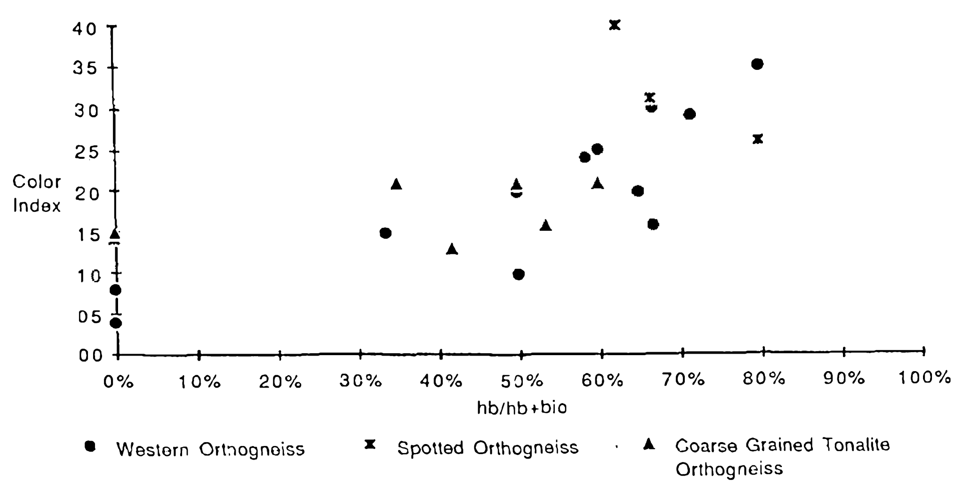


Figure 2.18 Estimated modal composition of tonalitic units in Little Goose Creek complex: a) classification after Streckeisen, 1973; b) color index versus hornblende/hornblende+biotite.

The Spotted Orthogneiss has a deformed medium-grained fabric with abundant annealed grains. Aggregates of one to three millimeter mafic and polygonal plagioclase grains appear as eye-shaped inclusions in the mylonitic fabric. These may have formed in an earlier metamorphism prior to mylonitization. Locally, screens of Spotted Orthogneiss have an older foliation that is partially transposed into the plane of the mylonitic foliation, further suggesting an earlier history for the unit. Porphyritic orthogneiss and associated pegmatites locally intrude the Spotted Orthogneiss.

2.4.4 Western Orthogneiss

Western Orthogneiss is a medium-grained, mylonitic gneiss ranging in composition from quartz diorite to granite; field observations indicate tonalitic gneiss is the most abundant. The color index is between 20 and 30, with hornblende more abundant than biotite (Figure 2.18). Sphene is very abundant and allanite and apatite are also common.

This gneiss is interlayered on ten centimeter to ten meter scales with less abundant, lower color index orthogneisses ranging in composition from granite to tonalite. Both biotite and hornblende-biotite gneisses are present. Some layers resemble the porphyritic orthogneiss but lack phenocrysts. Sparse one centimeter to one meter layers of fine-grained gneiss are also observed. These gneisses have mylonitized crystalloblastic fabrics and are mineralogically identical to the tonalite. The abundance of sphene in the layer sampled suggests it was a cognate inclusion rather than a xenolith of paragneiss.

All gneisses within this unit have a mylonitic fabric similar to that seen in the porphyritic orthogneiss described above. Foliation and lineation are well developed. Quartz has developed foliated subgrains and locally has recrystallized to annealed grains. Plagioclase forms augen and small annealed polygonal grains. Alkali feldspar has recrystallized as fine-grained material, and myrmekite lobes are abundant. Large mafic minerals have been deformed into augen and small recrystallized grains have formed as sheathes. Biotite is often bent or shredded.

Leucocratic, granitic intrusives, one centimeter to one meter wide, are abundant in the Western Orthogneiss. Orientations range from concordant to highly discordant and many sheets appear to have been transposed. The presence of alkali feldspar in the intrusives suggests that they may be related to the Porphyritic Orthogneiss, indicating that the Western Orthogneiss is the older of the two units.

Along its eastern edge the Western Orthogneiss is interlayered over tens to hundreds of meters with Porphyritic Orthogneiss. The layers are most commonly one to 20 meters wide, but where the zone of interlayering is widest, layers up to 300 meters wide were observed. To the west the Western Orthogneiss is in abrupt contact with the Eastern Tonalite Gneiss of the Hazard Creek complex.

2.4.5 Mafic Orthogneiss

Four pods of mafic intrusives crop out in the eastern Little Goose Creek complex. These pods are comprised of hornblende gabbros, hornblende-biotite diorite, quartz diorite and tonalite. Pods vary in character: the pod near Brundage Mountain is very heterogeneous, while 90% of the pod near Slab Butte is homogeneous diorite. Gabbros form 10 to 30% of the pods. They always contain hornblende that often mantles one or two pyroxenes. Color index varies from 25 to 95. Hypersthene is often encased and intergrown with clinopyroxene. An opaque mineral is often associated with the hornblende. Plagioclase has sharp twins. Gabbros have coarse-grained cumulate or crystalloblastic textures that are locally strained.

Hornblende biotite diorite, quartz diorite, and tonalite in the pods have color indices from 10 to 40. Biotite is more abundant in these rocks than in the gabbros, and pyroxene is generally absent. The diorite has a distinctive pale hornblende and red biotite, and contains abundant opaque minerals. The mafic pods are inferred to be cumulates and mafic intrusives. The abundance of hornblende and presence of quartz in many samples suggests they are intrusives related to calc-alkalic arc magmatism.

Medium to coarse-grained igneous and crystalloblastic fabrics are often mylonitically deformed. Deformation within the pods has been taken up heterogeneously. It is often concentrated in thin zones between areas that appear undeformed. Contacts with surrounding tonalite or porphyritic orthogneiss are tectonic, and rock types are interlayered over tens of meters.

2.4.6 Summary

The time and space relationships in the Little Goose Creek complex are summarized in Figure 2.19. The oldest rocks in the complex are mafic gneisses and metasedimentary rocks. Metasedimentary rocks occur in the north-central portion of the complex. Mafic gneisses occur throughout the complex except in the area east of the metasedimentary rocks. Mafic gneisses are inferred to be metamorphosed oceanic-arc rocks; metasedimentary rocks are inferred to be cratonally-derived. This change in lithology marks the regional boundary between oceanic-arc and continental supracrustal rocks. The supracrustal rocks are present as sheets in the porphyritic orthogneiss. Tonalite, quartz diorite, and diorite that form Western Orthogneiss and Spotted Orthogneiss are the oldest intrusives in the complex. Mafic Orthogneiss may also be of this generation. These lithologies are also present as sheets in Porphyritic Orthogneiss, the most distinctive and abundant lithology in the complex. The Coarse-Grained Tonalite Orthogneiss is inferred to be the youngest intrusive in the complex because it is mineralogically, chemically, and texturally similar to the Payette River Tonalite.

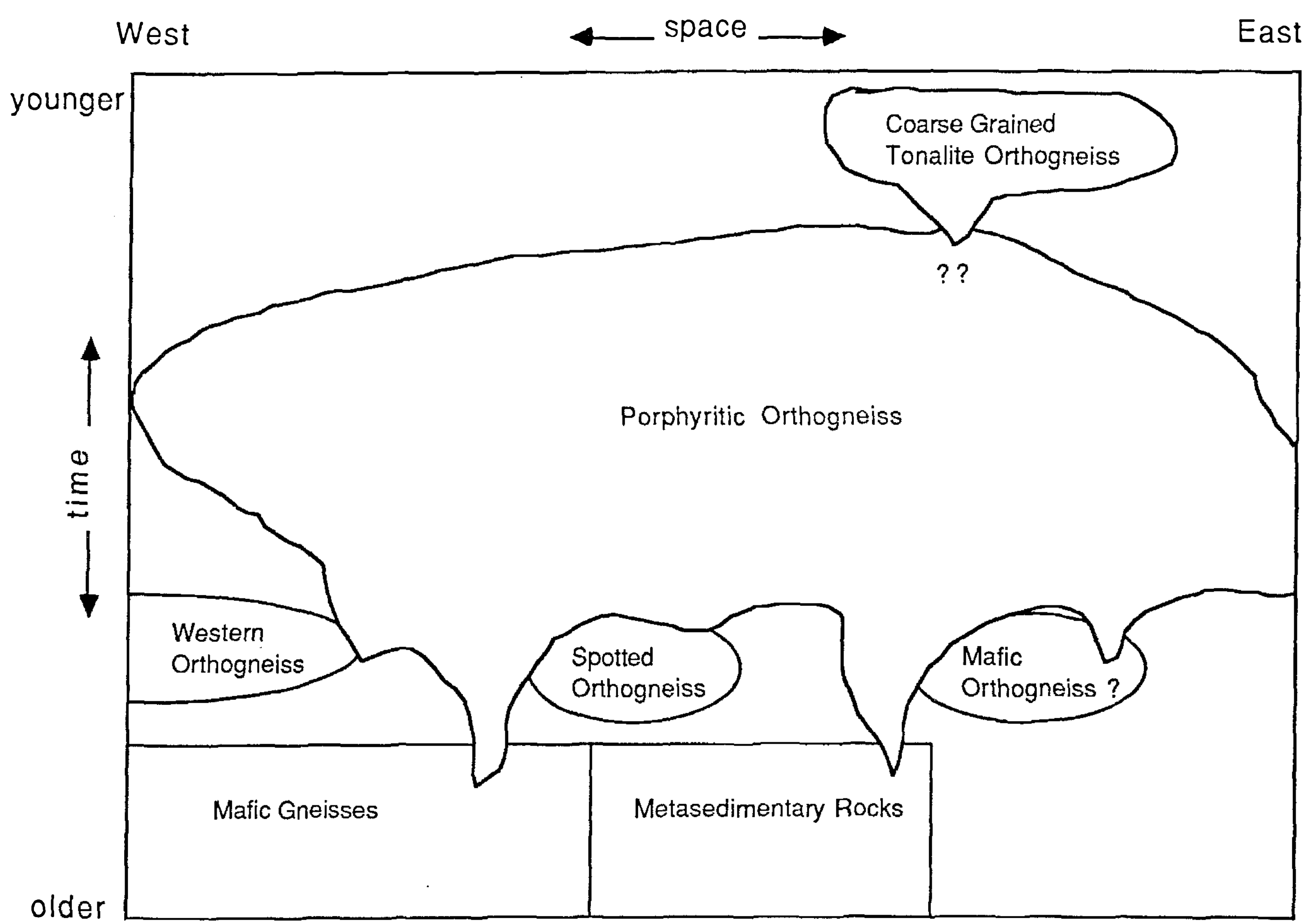


Figure 2.19 Time/space relationships in Little Goose Creek complex.

2.5 PAYETTE RIVER COMPLEX

The Payette River complex consists of large undeformed intrusive bodies. The most voluminous of these are the Payette River Tonalite and the Trail Creek Granite. Payette River Tonalite consists of tonalite and granodiorite with a strong primary foliation. It is concordantly intruded by a number of small granitic bodies that are grouped together as the Payette River Granite. The Payette River Tonalite is part of the coarse-grained biotite-hornblende tonalite which occurs within the western margin of the Idaho Batholith along its length. The Payette River Tonalite has been passively intruded along its eastern border by a large intrusion of granite, here called Trail Creek Granite. Trail Creek Granite extends out of the map area on the east and is part of the central two-mica granite of the Idaho Batholith.

2.5.1 Payette River Tonalite

Payette River Tonalite is composed primarily of medium to coarse-grained biotite-hornblende tonalite and hornblende-biotite granodiorite with a color index of 15 to 30 (Figure 2.20). Variations in hornblende:biotite are very common both on outcrop and larger scales. Hornblende:biotite is generally near one (Figure 2.20b). Plagioclase is andesine that is commonly normally zoned. Oscillatory zoning is rare. Orthoclase occurs as subhedral grains and as interstitial material between plagioclase laths. The granodiorite generally contains less hornblende than the tonalite. Both tonalite and granodiorite contain abundant sphene, allanite, apatite and zircon. Sphene may often be recognized in hand specimen.

Epidote is present locally in the Payette River Tonalite. It occurs as small polygonal grains with the mafic minerals and as larger anhedral grains rimming allanite. Clinozoisite and epidote are both present. Neither large grains that cross-cut

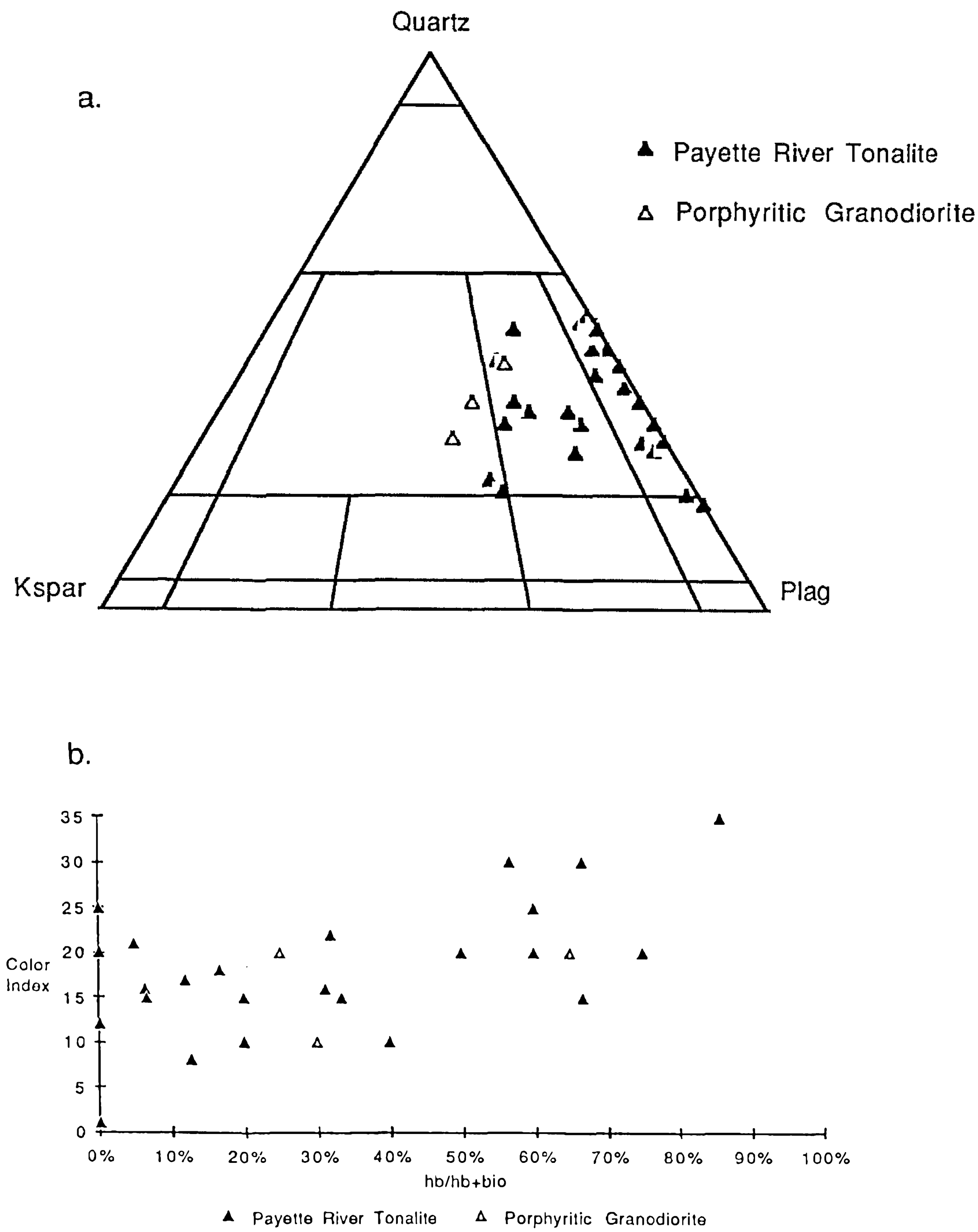


Figure 2.20 Estimated modal composition of Payette River Tonalite and Porphyritic Granodiorite of Payette River Complex: a) classification after Streckeisen, 1973; b) color index versus hornblende/hornblende+biotite.

mafic minerals, nor symplectic intergrowth common in the Hazard Creek complex were observed.

Tonalite forms approximately 70% of the unit, granodiorite 30%. Locally, the granodiorite is seen to intrude the tonalite; however, most contacts are gradational. The granodiorite is believed to be a phase of the tonalite, which is locally late crystallizing.

Schlieren and football shaped inclusions of fine-grained amphibolite, hornblende-biotite tonalite gneiss, and porphyritic biotite granodiorite are very common throughout both tonalite and granodiorite (Figure 2.21). Patches of amphibolite and of schlieren-rich material can be up to several tens of meters in diameter. Pegmatite and aplite dikes up to a meter in width are abundant, as are small pods (one to 100 meters) of Payette River Granite and metasedimentary rock.

Most rocks within the unit have a strong steeply dipping foliation formed by the alignment of biotite and hornblende. Many also have a strong down-dip lineation formed by the alignment of the long axes of hornblende. Most samples have an undeformed, hypidiomorphic texture, which indicates that the fabric is a primary igneous feature. Inclusions and schlieren are aligned in the foliation with long axis parallel to the lineation. Some rocks show evidence of a weak subsolidus deformation in which quartz has been strained or recrystallized. This texture is only observed west of the Payette River.

2.5.2 Porphyritic Granodiorite

Porphyritic Granodiorite is composed of biotite-hornblende granodiorite and granite with a color index between 10 and 20 (Figure 2.20). Biotite is generally more abundant than hornblende. The rock contains abundant orthoclase phenocrysts up to three centimeters in length. The phenocrysts have abundant inclusions of plagioclase, biotite and quartz. Apatite, allanite and zircon are common accessory phases. No sphene was observed. Mafic inclusions are distributed sparsely through the unit.

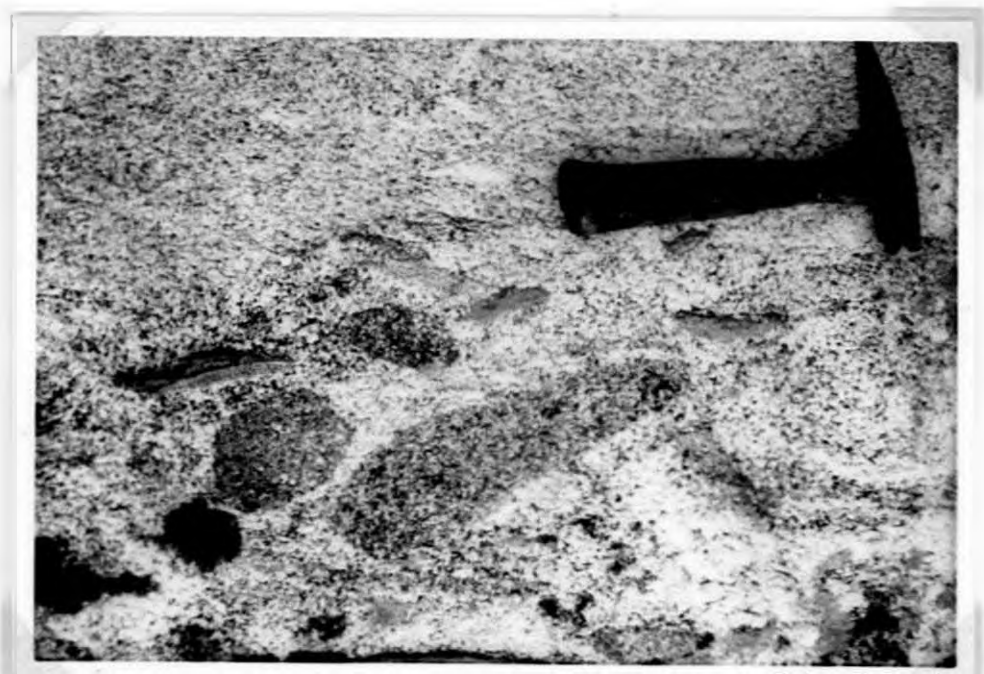


Figure 2.21 Schlerien and football shaped mafic inclusions in Payette River Tonalite.

The rock has a medium to coarse-grained strained igneous fabric. Quartz has undulatory extinction and is locally deformed into elongate grains. Large feldspar grains are rimmed by small recrystallized grains. Myrmekite lobes are abundant on the edges of orthoclase phenocrysts. The deformed fabric is stronger than, but similar to, fabrics elsewhere in the western Payette River Tonalite. Contacts with the Payette River Tonalite were not exposed. This unit is most likely a porphyritic phase of the Payette River Tonalite as it is similar in appearance. It is unlikely that it is related to the porphyritic orthogneiss as the color index is higher and the phenocrysts are orthoclase.

2.5.3 Payette River Granite

Mappable sills and pods, as well as small pods, dikes and sills of biotite granite intrude the Payette River Tonalite across its width (Figure 2.22). Color index varies from 2 to 15 but is generally low; muscovite or hornblende are rare. Plagioclase occurs in tablets that are normally zoned or unzoned. Alkali feldspar occurs in both untwinned and microcline twinned grains; phenocrysts are uncommon. Allanite, apatite and zircon are common accessory minerals.

These granites have a medium to coarse-grained hypidomorphic fabric with a primary foliation formed by the alignment of biotite. In addition, many samples show evidence of varying degrees of mylonitization. The intensity of this deformation varies substantially within individual granite bodies. It is equally common in granites from all parts of the complex and may be related to emplacement of the granites. Both primary and mylonitic foliations parallel the foliation in the tonalite.

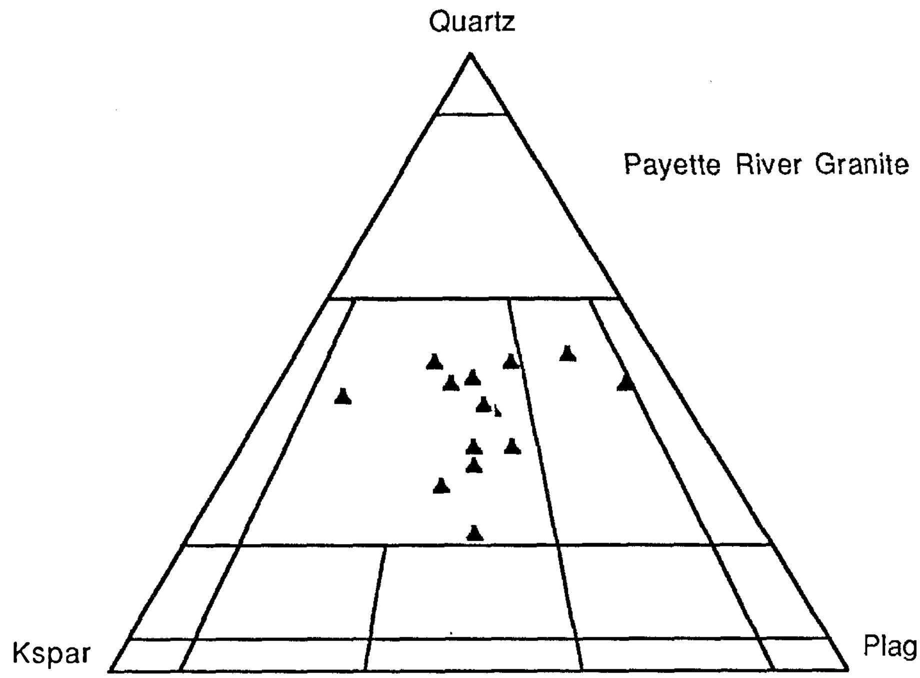


Figure 2.22 Classification based on estimated modes of Payette River Granite (after Streckeisen,1973).

These granites are thought to be related to the Payette River Tonalite. They share the same fabric elements, suggesting that they were emplaced at a similar time. They occasionally contain hornblende like the tonalite and, as in the tonalite, allanite is a common accessory mineral. Allanite is absent from the Trail Creek Granite. Contacts with tonalite are commonly migmatitic or interfingering. These bodies are often associated with abundant metasediments.

2.5.4 Trail Creek Granite

The Trail Creek Granite is composed of medium to coarse-grained granite and granodiorite with one or two micas (Figure 2.23). Red-brown biotite is the dominant mafic mineral forming one to 15 percent of the rock. Muscovite forms zero to five percent of the rocks. Microcline occurs locally as phenocrysts up to three centimeters in length. Plagioclase occurs as subhedral tablets with well developed oscillatory zoning. Apatite and zircon are common accessory minerals. Zircon inclusions in biotite are abundant and have prominent pleochroic haloes.

The unit is massive or weakly foliated. Foliation is better developed in the porphyritic rocks. Blocks of tonalite up to several meters in diameter are included in the granite. These are particularly common when approaching the western contact with the Payette River Tonalite.

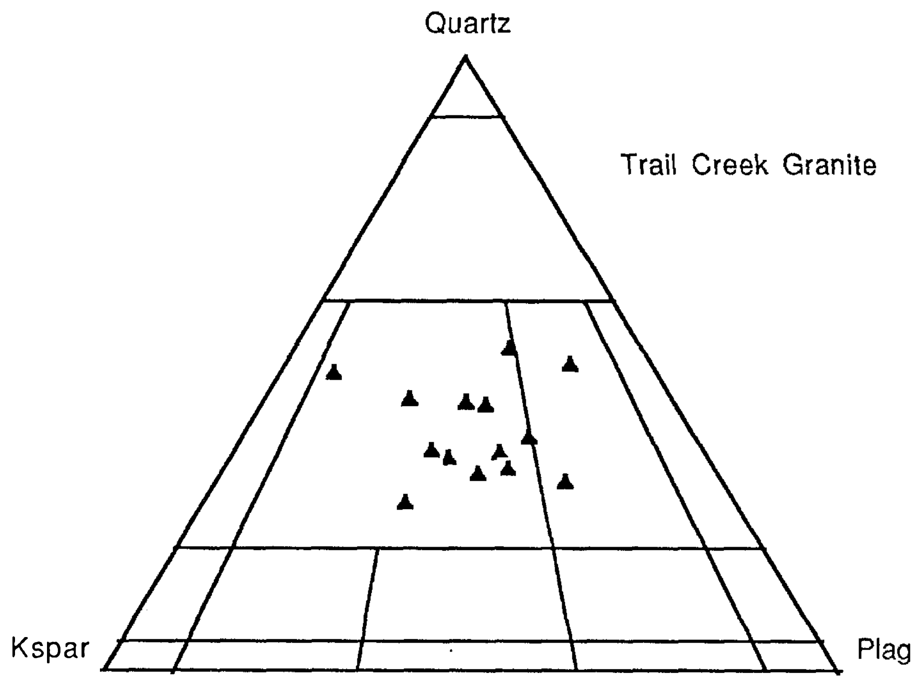


Figure 2.23 Classification based on estimated modes of Trail Creek Granite (after Streckeisen,1973).

2.5.5 Trail Creek Granite Border Zone

Trail Creek Granite has intruded the Payette River Tonalite, creating a zone of mixed rocks. Blocks of tonalite in granite give way progressively to dikes and pods of granite in tonalite. The border zone unit covers the area of abundant dikes and pods of granite in tonalite. Dikes are biotite granite with a color index less than 10. Plagioclase is oscillatorily zoned as in the Trail Creek Granite. Dikes have sharp contacts and are up to three meters wide. They show no preferred orientation, in contrast to the concordant sills of Payette River Granite.

2.5.6 Summary

The space and time relationships in this complex are summarized in Figure 2.24. The oldest rocks in the complex are metasedimentary rocks inferred to be cratonally-derived sedimentary rocks. These occur as inclusions in all of the intrusive units. The oldest intrusive is the Payette River Tonalite. Porphyritic Granodiorite in the central portion of the complex is inferred to be a phase of the Payette River Tonalite. Sills of Payette River Granite, which parallel the primary foliation in the Payette River Tonalite, are inferred to be a younger differentiate of that body. Trail Creek Granite intrudes and stopes the Payette River Tonalite along its eastern margin. Trail Creek Granite, which extends out of the study area to the east, is part of the central granitic core of the batholith.

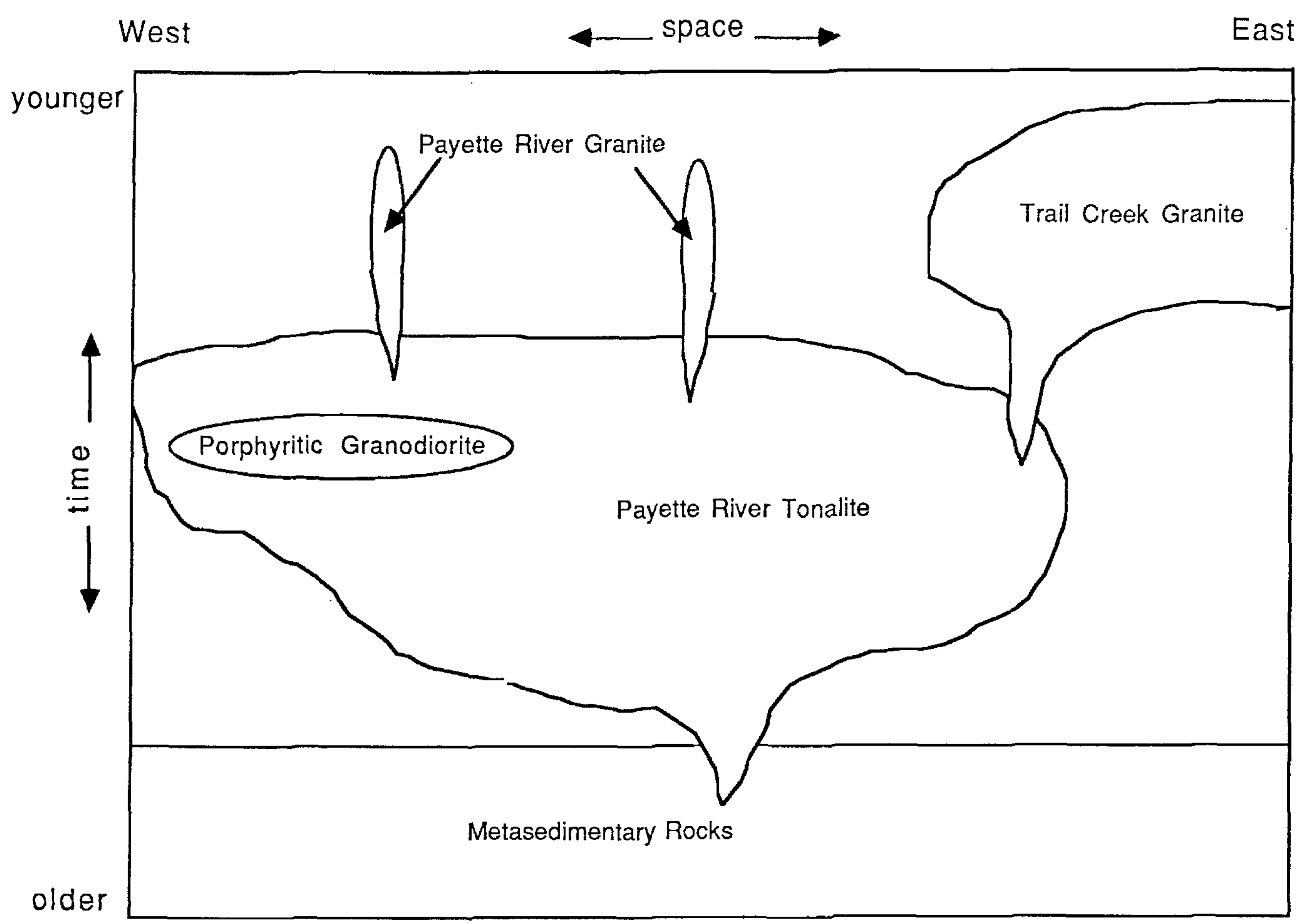


Figure 2.24 Time/space relationships in Payette River complex.

2.6 RADIOMETRIC AGE DATA

The absence of intrusive relationships between units of different complexes necessitates the reliance on radiometric age data to establish their relative age. This data also allows some constraint to be placed on the absolute ages of intrusive episodes. There may be significant age variations within individual units; this possibility cannot be evaluated with the data at hand.

Three types of radiometric data are available: U-Pb isotopes in zircons, $^{40}Ar/^{39}Ar$ data from hornblende, and conventional K-Ar data from hornblende and biotite. U-Pb data is from preliminary work by Dr. Leon T. Silver and myself, done in his laboratories at the California Institute of Technology. Multiple fractions from three samples have been analyzed. All samples show internal discordance. The ages reported are the lower intercept of a linear regression with concordia. These ages are preliminarily interpreted to be the emplacement age of the plutonic rocks because zircon is not easily affected by metamorphism. However, further work is necessary to ascertain the effect of metamorphism and to better characterize the lower intercepts.

$^{40}Ar/^{39}Ar$ data was obtained by Dr. Lawrence Snee at the U. S. Geological Survey. This data gives information on the last time hornblende became closed to the diffusion of argon. Hornblende is commonly accepted to close in the temperature interval 580-480^{o}C (Harrison, 1981). $^{40}Ar/^{39}Ar$ ages are younger than the emplacement age of a plutonic rock. If the rock cooled quickly, they may be only slightly younger. Amphibolite facies metamorphism can reset or partially reset this age.

K-Ar ages for three samples are reported by Armstrong et al. (1977). These ages also record the time a mineral closed to argon diffusion and are younger than the emplacement age. They provide minimum age constraints.

2.6.1 Hazard Creek Complex

The sample from the Jack's Creek Pluton was analyzed to obtain an U-Pb age. As summarized above (Figure 2.15) field relationships show that this pluton is one of the youngest epidote-bearing intrusive rocks in the complex. It is cross-cut only by dikes related to the Garnet Granodiorite Mylonite. The age of this unit therefore provides a maximum constraint for the emplacement of the Garnet Granodiorite Mylonite protolith and a minimum constraint for older epidote-bearing intrusive rocks and orthogneisses. The ages of rocks in the western Hazard Creek complex are not constrained as it is not possible to uniquely correlate rocks across the complex. The preliminary U-Pb age for the Jack's Creek Pluton is 118±5 Ma. Hornblende from the same sample has an $^{40}Ar/^{39}Ar$ age of 95.1±.3 Ma. This age, significantly younger than the U-Pb age, may reflect slow cooling or reheating of the intrusive.

2.6.2 Little Goose Creek Complex

A sample of porphyritic orthogneiss from the western side of the Little Goose Creek complex was analyzed to obtain an U-Pb age. The porphyritic orthogneiss intrudes Spotted Orthogneiss; granitic dikes inferred to be derived from the porphyritic orthogneiss intrude Western Orthogneiss. The age of the porphyritic orthogneiss provides a minimum age constraint for the emplacement age of these units. The preliminary U-Pb age is 111±5 Ma. Hornblende from the same sample has a total gas $^{40}Ar/^{39}Ar$ age of 87 Ma. The spectrum does not contain a good plateau (Snee, per. comm., 1987) suggesting that a younger thermal event has affected the sample. Biotite from a biotite-hornblende quartz diorite gniess in the Little Goose Creek complex has a K-Ar age of 77 Ma (Armstrong et al., 1977), which may also reflect a younger thermal event.

2.6.3 Payette River Complex

A sample from tonalite in the eastern part of the Payette River Tonalite was analyzed to obtain a U-Pb age. Granodiorite within the Payette River Tonalite locally intrudes the tonalite indicating it is younger in part. Payette River Granite intrudes Payette River Tonalite. The similarity of primary fabrics in the Payette River Tonalite, Porphyritic Granodiorite and Payette River Granite suggest that these units were emplaced during the same interval. In contrast, Trail Creek Granite cross-cuts the Payette River Tonalite fabric and has a distinct suite of accessory phases suggesting it belongs to a separate intrusive interval. The age of the Payette River Tonalite provides a maximum constraint for the ages of both Payette River and Trail Creek Granites.

The Coarse-Grained Orthogneiss of the Little Goose Creek complex is correlated with the Payette River Tonalite on the basis of its mineralogy, relict igneous texture and chemical composition. The age of the Payette River Tonalite constrains this age as well.

The Payette River Tonalite sample yields a poorly constrained U-Pb age of 90 Ma. Marked inheritance complicates interpretation of this age. Hornblende from this sample has an $^{40}Ar/^{39}Ar$ of 81± 4 Ma (Snee, per. comm., 1987). Biotite from a sphene-hornblende-biotite tonalite within the Payette River Tonalite has a K-Ar age of 65±2 Ma (Armstrong et al., 1977). Other tonalitic rocks with similar mineralogy and texture present along the western side of the Idaho Batholith have K-Ar and $^{40}Ar/^{39}Ar$ ages between 75 and 95 Ma (Criss and Fleck, 1987; Lund and Snee, 1988; Lewis et al., 1987). A U-Pb age of 89±5 Ma is reported for one of these rocks (Lund and Snee, 1988). These data suggest that the Payette River Tonalite and other similar tonalites along the western side of the Idaho Batholith were emplaced in the early part of the Late Cretaceous.

The Trail Creek Granite is part of the two-mica granite core of the Idaho Batholith. K-Ar and $^{40}Ar/^{39}Ar$ analyses from similar two-mica granites elsewhere in the Atlanta lobe of the batholith give ages between 69 and 77 Ma (Lewis et al., 1987; Lund and Snee, 1988).

2.6.4 Summary of Age Constraints

The age data are summarized in Figure 2.25. Emplacement of the Hazard Creek complex began prior to 118 Ma, the age of the sample from Jack's Creek Pluton. Garnet Granodiorite is younger. Some tonalite and quartz diorite in the Little Goose Creek complex may have been emplaced during this time. The Porphyritic Orthogneiss was emplaced around 111 Ma. The Payette River Tonalite, Payette River Granite, and Porphyritic Granodiorite units in the Payette River complex, and the Coarse-Grained Tonalite Orthogneiss unit in the Little Goose Creek complex are interpreted to have been emplaced in the early Late Cretaceous. The Trail Creek Granite was emplaced after 90 Ma and probably prior to 69 to 77 Ma.

The data show that most or all of the Hazard Creek complex is older than the porphyritic orthogneiss in the Little Goose Creek complex. The Payette River complex is significantly younger than the other complexes.

2.7 SUMMARY OF THE INTRUSIVE HISTORY

The oldest rocks in the study area are metamorphosed supercrustal rocks. In the western portion of the area, these are mafic layered gneisses, generally of andesitic composition. Some or all of these rocks are probably correlative with the Pollock Mountain Amphibolite (Aliberti et al., 1987) to the west. Rock composition and isotopic systematics suggest that the Pollock Mountain Amphibolite protolith was Triassic arc volcanics and related material (Aliberti et al., 1987). They may be correlative with the upper part of the Wallowa/Seven Devils arc or the Huntington arc.

These rocks give way abruptly to a second suite of supracrustal rocks in the western Little Goose Creek complex and the Payette River complex. This suite is

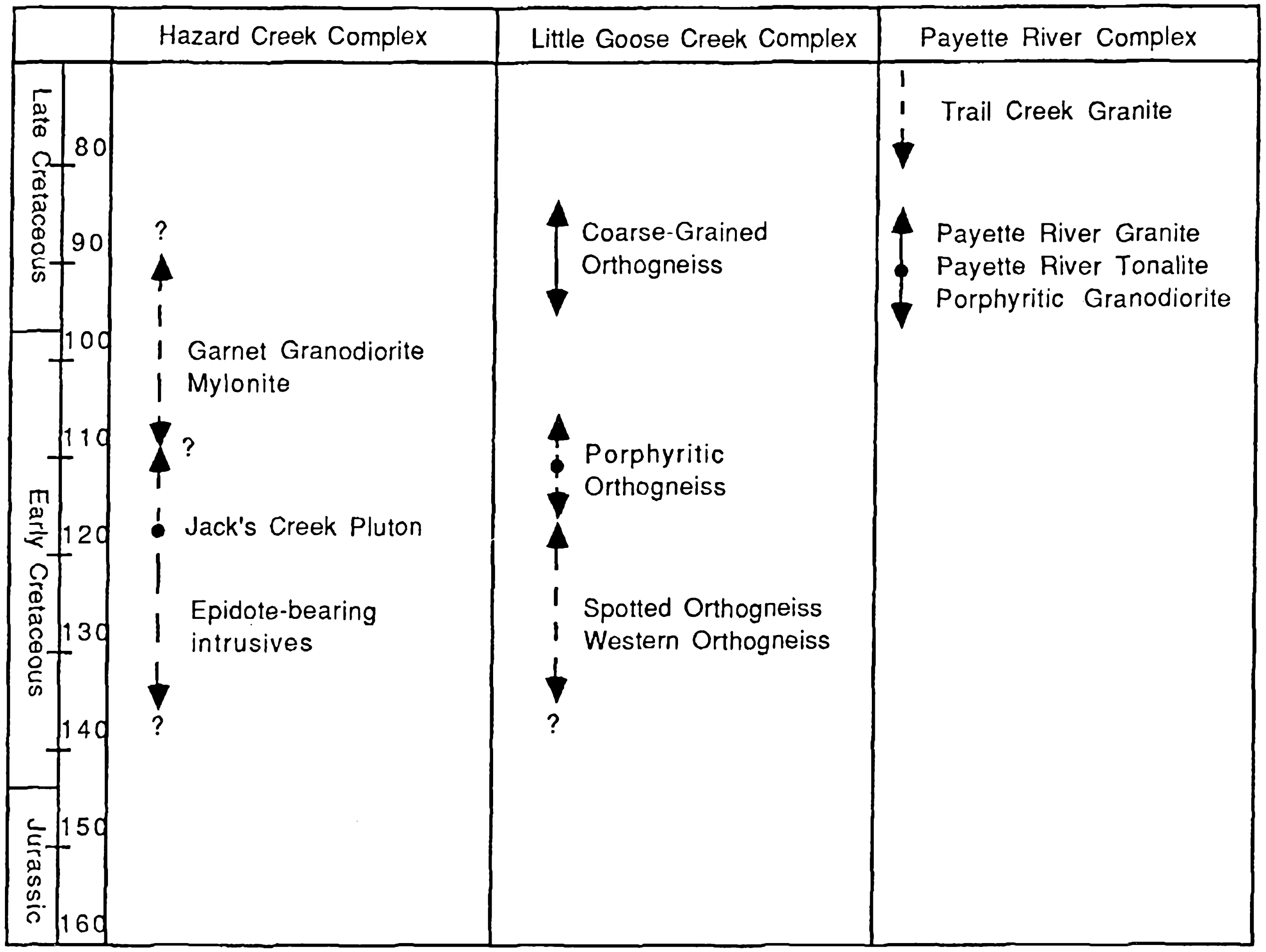


Figure 2.25 *Summary of age constraints.* Black dots indicate preliminary U-Pb ages discussed in text.

comprised of biotite schist, silimanite schist, quartzite and calc-silicate gneiss. These rocks are inferred to be metamorphosed cratonally-derived sedimentary material, which was deposited in a shallow water environment. These rocks may be correlative with Precambrian Belt sediments or Paleozoic miogeoclinal sediments. The boundary between oceanic and continentally material is constrained to within two kilometers. No interlayering of material or intermediate material was observed. The supracrustal rocks have been metamorphosed to amphibolite facies.

Intrusive material was emplaced into both the oceanic and continental supracrustal rocks. The Hazard Creek complex, west of the boundary between oceanic and continental supracrustal rocks, is formed from a series of intrusives emplaced during ongoing deformation. Intrusives are primarily quartz diorites, tonalites and trondhjemites most of which contain epidote. The Jack's Creek Pluton, one of the youngest intrusives in the complex, has a U-Pb age of 118 Ma. The presence of epidote suggests emplacement at depths greater than 24 km (Zen and Hammarstrom, 1984). These depths are similar to those recorded in the garnets of the Pollock Mountain Amphibolite (Selverstone et al., 1987).

The Little Goose Creek complex was emplaced across the boundary between oceanic and continental supracrustal rocks. Porphyritic orthogneiss, emplaced around 111 Ma, contains screens of both supracrustal suites. Tonalitic sheets are inferred to be both older and younger than Porphyritic Orthogneiss. Coarse-Grained Tonalite Orthogneiss is correlated with the Payette River Tonalite to the east. All units within this complex have been strongly mylonitized.

The Payette River Tonalite was emplaced east of the boundary between oceanic and continental supracrustal rocks in the early Late Cretaceous. This unit is comprised of tonalite and granodiorite containing screens of metamorphosed continental supracrustal rock. Amphibolite facies metamorphism and partial melting of sedimentary material occurred during emplacement of the Payette River Tonalite. Payette River Granite was emplaced into the tonalite as sills, which parallel the igneous foliation. These bodies are inferred to be closely related differentiates of the tonalite.

Trail Creek Granite stopes and dikes the Payette River Tonalite. This unit is composed of biotite and two-mica granite, which is part of the central core of the Idaho Batholith. It extends out of the map area to the east.

CHAPTER 3: DEFORMATIONAL HISTORY OF THE WESTERN IDAHO BATHOLITH NEAR MCCALL

3.1 INTRODUCTION

The most striking regional characteristic of the study area is the pervasive north-south structural grain (Figure 3.1, Plate 1). This is reflected in the elongate map pattern of lithologic units and in the fabric within units: most rocks have a generally north-south striking, steeply-dipping foliation and down-dip lineation. This north-south grain parallels the boundary between oceanic-arc and continental supracrustal rocks. It weakens and is lost away from the boundary. The geometry of this fundamental crustal boundary influenced the orientation of younger intrusive bodies and localized several deformations.

In contrast to many areas where multiple deformational events can be recognized by fabric orientation, all structural fabrics in this area share a common orientation. Evidence for distinguishable periods of deformation is based on the age of affected plutons. Four periods of fabric formation are recognized as described below; however, some unrecognized deformational events may have occurred, or deformation may have been continuous over the intervals of time discussed. Some of the deformation cannot be constrained in time by field relations, and thus cannot be assigned uniquely to any of these deformations. Not all of the events are recognizable in each part of the study area and thus the discussion is organized spatially.

The evidence for the oldest recognized deformations is preserved in the Hazard Creek complex. Two early episodes of deformation can be distinguished: a folding associated with amphibolite facies metamorphism, and a subsequent deformation broadly coeval with the emplacement of the epidote-bearing intrusives. The next structures formed in the Payette River complex during emplacement of the Payette River Tonalite.

Figure 3.1 Simplified geologic map of study area.

Simplified Map of Western Idaho Batholith

Hazard Creek Complex

- Ht — Epidote Tonalites: Hhbet, Hbet
- Hg — Older Gneisses: Hqdg, Hbetg, Hetg
- Hgc — Gneiss Complexes: Hldg, Hdg, Hlg, Hpg
- Hk — K rich Suite: Hggd, Hbmgr
- Hlmg — Mafic Layered Gneiss

Little Goose Creek Complex

- Gtg — Tonalites: Ghbtg, Gsg, Gmg
- Gpg — Porphyritic Orthogneiss
- Gt — Coarse Grained Tonalite Orthogneiss

Payette River Complex

- Pt — Tonalite: Pt, Ppgd
- Pgr — Payette River Granite
- Ptgr — Trail Creek Granite
- Pbz — Trail Creek Border Zone

ms — Metasedimentary Rocks and Associated Hybrid Igneous Rocks

Columbia River Basalt and Quaternary Cover

30 Strike and dip of foliation

10 Bearing and plunge of lineations and fold axes

Mylonitic Fabric


Hazard Creek Complex
Little Goose Creek Complex
Payette River Complex
Ht
Hg
Hgc
Hlmg
Hk
Glg
Gl
Gpg
ms
Pgr
Ptgr
Pbz
Pt
N
0
2 mi
0
2 km

Schematic Cross Section Showing Structural Features

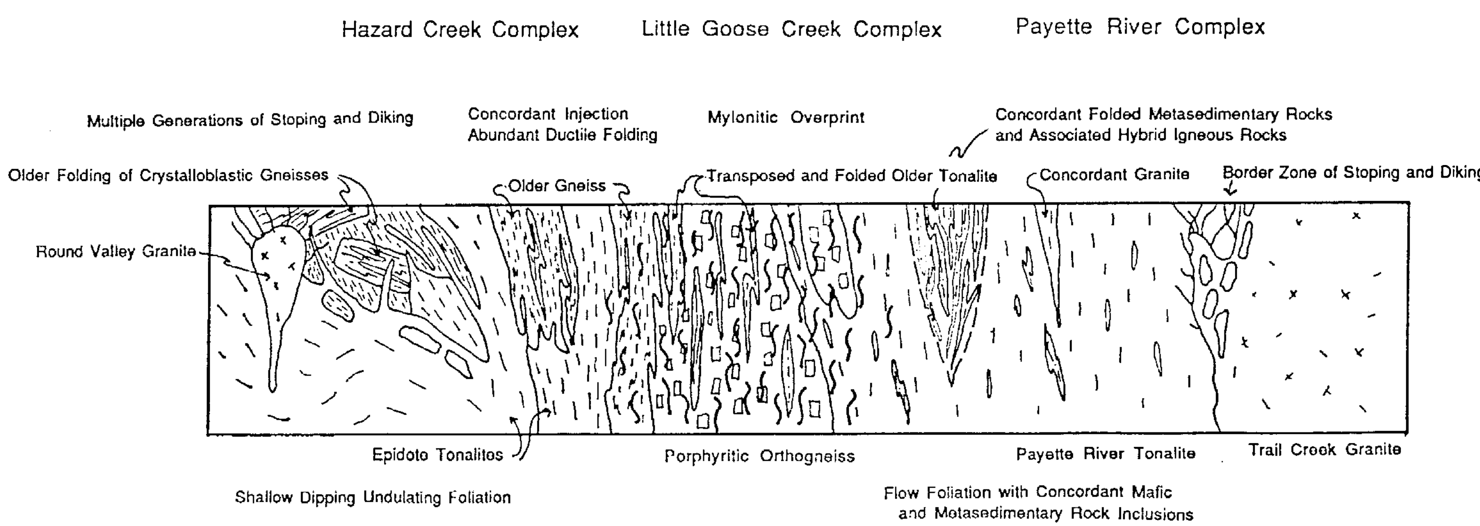


Figure 3.2 Schematic cross section across study area.

The youngest deformation is a major mylonitic event centered on the Little Goose Creek complex. Structural relationships are summarized schematically in Figure 3.2.

3.2 DEFORMATION IN THE HAZARD CREEK COMPLEX

The oldest deformational fabrics in the area are preserved in the Hazard Creek complex. Three deformational events can be distinguished in these rocks. The oldest structures are preserved in the crystalloblastic gneisses in the western Hazard Creek complex. A second deformation is associated with the emplacement of the epidote-bearing intrusives over a protracted period of time. In the eastern Hazard Creek complex, a post-intrusive event deforms all the rocks including the youngest pegmatites. This deformation appears to increase in strength to the east and is inferred to be related to the mylonitization centered on the Little Goose Creek complex.

3.2.1 Older Deformation of Crystalloblastic Gneisses

Crystalloblastic gneisses, which include both mafic layered gneiss and orthogneiss, preserve evidence of the oldest deformation seen in the area. These gneisses, which comprise the Mafic Layered Gneiss and the older portions of the Light Gneiss and Dark Gneiss Assemblages, have been metamorphosed to upper amphibolite facies (andesine, hornblende, biotite, quartz, clinozoisite± diopside). Compositional layering in paragneiss parallels the mineral foliation. The presence of intrafolial isoclinal fold noses suggests that the compositional banding is bedding that has been transposed on fold limbs. The metamorphism and deformation of both orthogneiss and mafic layered gneiss likely occurred during one event after the intrusion of the orthogneiss protolith, but may reflect ongoing deformation during the emplacement of these oldest intrusives.

These gneisses are the only folded units recognized in the western Hazard Creek complex, demonstrating that folding predates intrusion. Foliation in these gneisses dips moderately in all directions (Figure 3.3). Lineations and fold axes are generally parallel plunging moderately to the east or west (Figure 3.3).

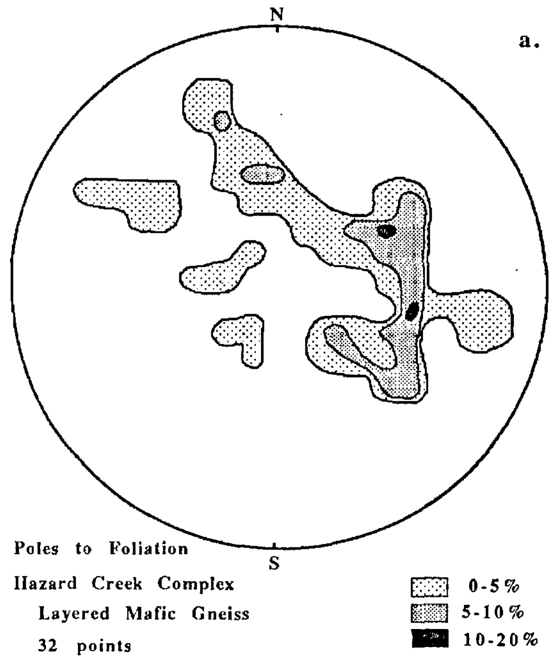
3.2.2 Deformation Coeval with Emplacement of Epidote-Bearing Intrusions

Most of the deformation in the Hazard Creek complex was broadly coeval with the emplacement of the epidote-bearing intrusives. This deformation was very heterogeneous: the style of intrusion and deformation and the fabric orientation change markedly across the complex. In the eastern part of the complex, foliations are steeply-dipping, lithologic units are elongate, and ductile folds are abundant. In the western part, foliations dip shallowly in all directions, lithologic units are irregularly shaped and broad in plan view, and folds are present only in the oldest gneisses. The intrusives are thought to be broadly coeval because they are chemically and mineralogically similar.

The gneisses and intrusive rocks of the eastern Hazard Creek complex all have steeply dipping north to north-northwest striking foliation. Dips shallow from east to west across the eastern part of the complex from 80 to 90^{o} in the Eastern Tonalite Gneiss to 50 to 70^{o} in the central portion of the complex. Planar features include gneissic layering, schistosity, igneous foliation, concordant intrusive margins and mylonitic foliation.

Within the Eastern Tonalite Gneiss a strong gneissic foliation, which developed prior to the emplacement of cross-cutting granitic pegmatites, is attributed to this deformation. Subsequently, the gneiss and pegmatites were mylonitized and folded. Transposition of the gneissic fabric has not been observed, nor is there any correlation between the intensity of mylonitization and the orientation of the gneissic fabric. This suggests that gneissic foliation was steeply dipping prior to mylonitization.

Several generations of concordant to semi-concordant intrusive sheets, centimeters to kilometers wide, are present in the gneisses of the eastern Hazard Creek



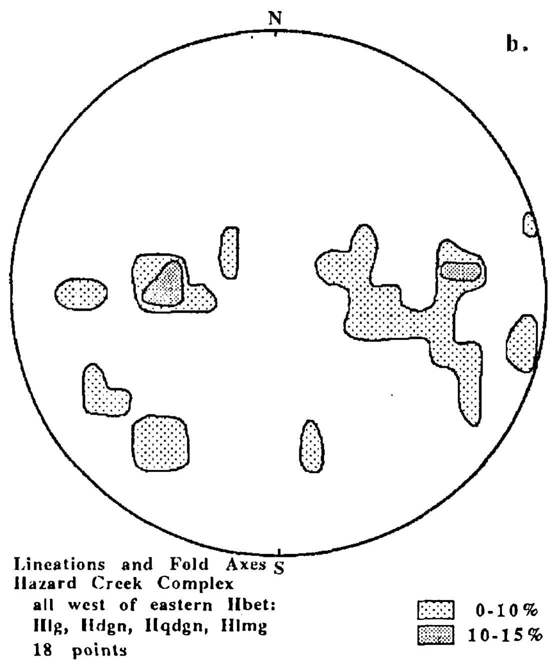


Figure 3.3 Contoured lower hemisphere equal area stereonet plots of poles to foliation, mineral lineations and fold axes in crystalloblastic gneisses of western Hazard Creek complex.

complex. These sheets are variably deformed; however, the preservation of abundant igneous relationships suggests that they have not been transposed. Dikes that cross-cut foliation often send off concordant apophyses. Fabrics are often better developed in older plutons, further suggesting that deformation was broadly coeval with intrusion.

The Jack's Creek Pluton and the eastern portion of the Epidote-Biotite Tonalite form larger concordant sheets (Plates 1 and 2). They have a strong foliation (Figure 3.4 a,c) defined by the orientation of mafic minerals, flattened mafic inclusions, and screens of gneissic material. Locally, lineations formed by the alignment of hornblende plunge 50 to 70^{o} to the east to north-east (Figure 3.5 a,c). In both bodies dikes of pegmatite and aplite are generally subparallel to the foliation direction.

The fabric within these bodies is inferred to be largely a primary igneous feature. The texture of the rocks indicates they have experienced some subsolidus deformation: quartz and feldspar are often somewhat strained and have developed undulatory extinction, serrated grain boundaries or subgrains. Mafic minerals generally do not appear to be deformed. Locally in the Jack's Creek Pluton, however, mylonitic textures have locally overprinted the igneous fabric completely. In general, the subsolidus deformation does not appear to have been strong enough to reorient mafic inclusions or screens of gneissic material. There is no correlation between the amount of subsolidus deformation and the orientation of the fabric, and nowhere was an earlier foliation observed to be partially transposed. These features all suggest that the orientation of the foliation is a primary igneous feature, and not due to subsequent subsolidus deformation.

Ductile folding is common in the eastern portion of the complex. Folds are found locally in all lithologies. Folding is heterogeneous: unfolded material passes abruptly into zones of abundant folding with no apparent change in lithology. Folding is particularly common along the margins of larger intrusions and in areas with abundant dark gneissic or amphibolitic layers. Older folded intrusive material is often cross-cut

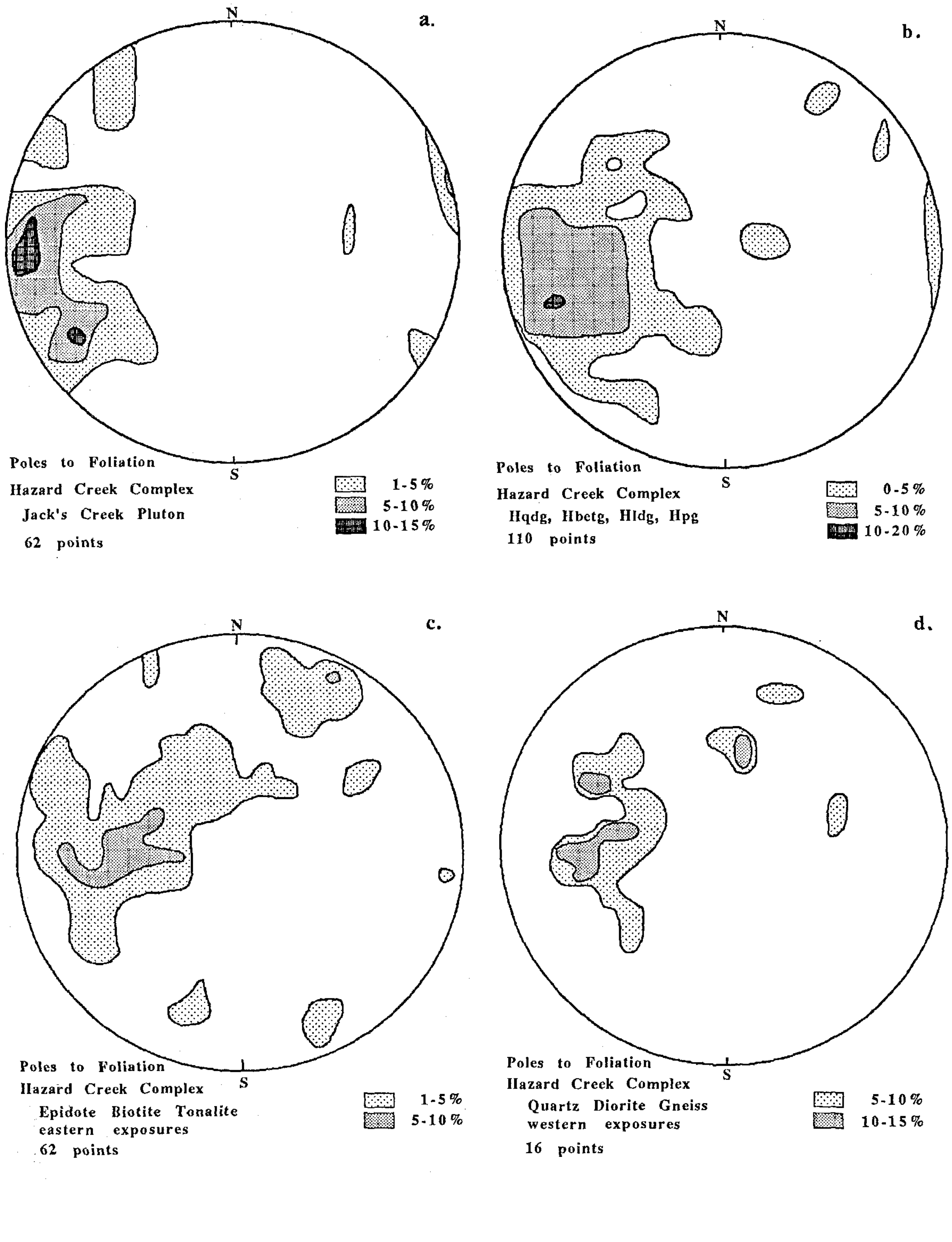


Figure 3.4 Contoured lower hemisphere equal area stereonet plots of poles to foliation in domains of the Hazard Creek complex.

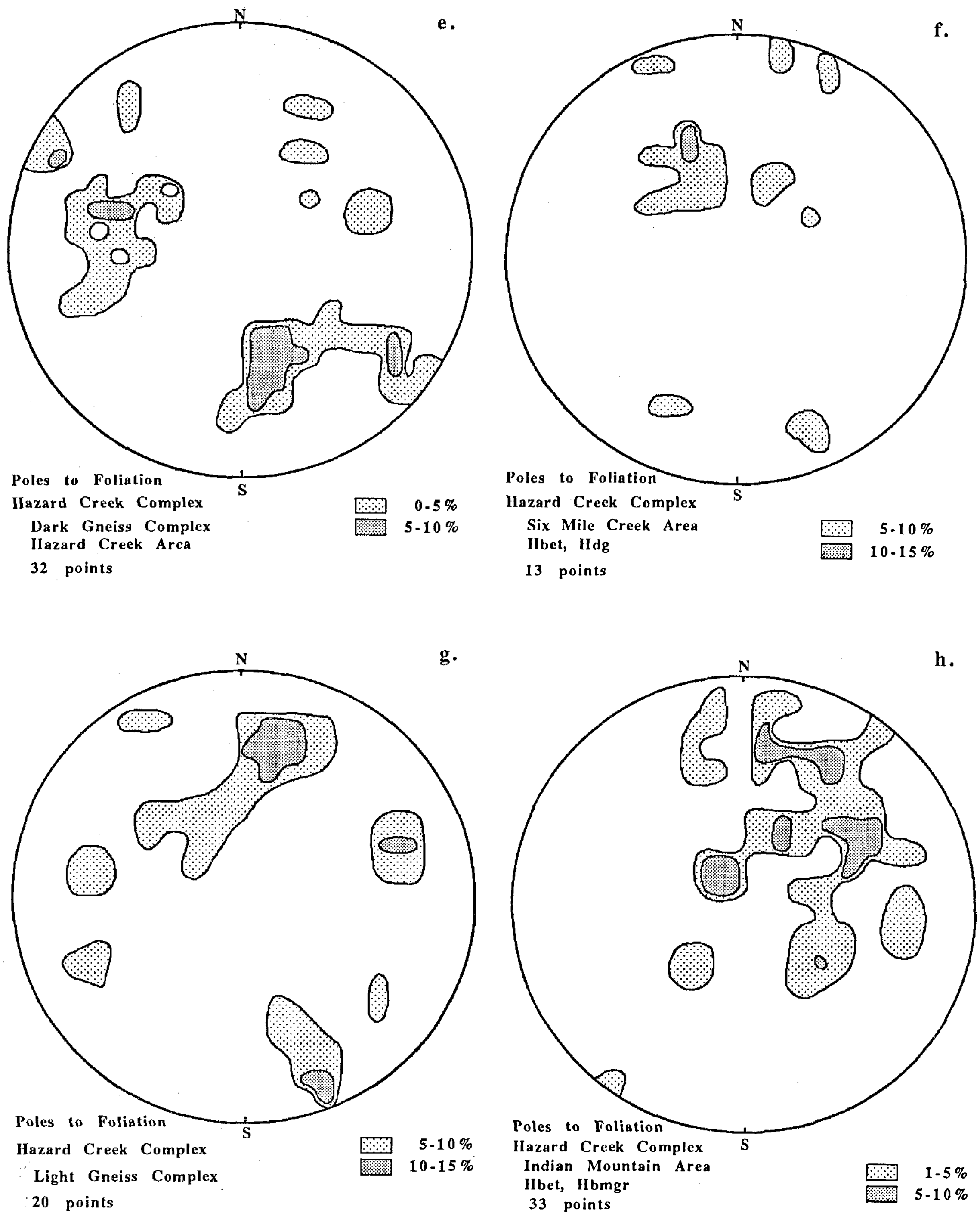


Figure 3.4 continued

Figure 3.5 Contoured lower hemisphere equal area stereonet plots of mineral lineation and fold axes in domains of the eastern Hazard Creek complex.

by younger unfolded material. Folds are tight to isoclinal and generally plunge steeply; north and south shallowly plunging fold axes were also observed. Folds are often asymmetrical but show no consistent vergence. The localization of folding near intrusive margins suggests that some deformation may be related to deviatoric stress generated by the emplacement of the intrusive body or to thermal anomalies causing local increases in ductility.

Several lines of evidence suggest that much of the deformation in the eastern Hazard Creek complex was broadly coeval with plutonism. 1) Intrusive bodies on all scales are semiconcordant with the foliation; large bodies have concordant flow foliations. Where dikes cross-cut the foliation small sills of material often parallel the foliation. 2) There are numerous examples of older more strongly foliated material cut by younger less strongly foliated material. 3) Folding is concentrated near the margins of intrusive bodies or in zones where gneissic material is common. 4) Older folds are occasionally cut by younger sills. These relationships suggest that the eastern Hazard Creek complex formed during a protracted period of intrusion accompanied by ongoing deformation. They indicate that some deformation occurred during the broad span of intrusion.

Deformation within the western Hazard Creek complex differs in style from that in the east. Gneisses have moderately dipping foliations with no preferred orientation (Figure 3.4 e,f,g,h; Plates 1 and 2). Planar features include igneous and metamorphic foliation, gneissic layering and more rarely concordant intrusive margins. Textures range from hypidomorphic to crystalloblastic; mylonitic textures are rare. There are no lineations or ductile folds.

There is much evidence for stoping and injection along the margins of intrusions. Abundant blocks of older intrusives, orthogneiss and paragneiss are found in the younger intrusives. There are numerous examples of rotated more-strongly foliated blocks in less-strongly foliated host material. Locally, older foliation in these blocks is overprinted with a second foliation parallel to that in the host rock. Inclusions of older gneissic or amphibolitic material are usually blocky and injected by several generations of cross-cutting dikes. Locally, where the host material is strongly foliated the inclusions are concordant.

Epidote-Biotite Tonalite in the western portion of the complex has an irregular near-equidimensional outcrop pattern (Plate 1). Randomly oriented, cross-cutting, undeformed trondhjemite dikes, pegmatites and aplites are abundant. Locally, an igneous foliation formed by the alignment of mafic minerals is moderately developed to well-developed. Rare flattened mafic inclusions parallel the mineral foliation. Strikes and dips vary on both outcrop and regional scales (Figure 3.4h). Stikes and dips of foliation in epidote-biotite-hornblende tonalite from a single outcrop in the Dark Gneiss Assemblage show a large variation, suggesting a lack of original planar orientation. There is no lineation. The fabric shows no evidence of shearing; rather, a partially developed crystalloblastic fabric is often observed overprinting the igneous fabric. This fabric is characterized by the migration of quartz and feldspar grain boundaries and by the development of equant quartz grains throughout the rock. Large quartz grains are either strain-free or have weak undulatory extension. Thus, synplutonic deformation in the western Hazard Creek complex involved the development of warped shallow to moderately dipping foliation, accompanied by much stopping and dike injection. Thermal annealing outlasted deformation.

The relationships described above suggest that the epidote-bearing intrusives were emplaced passively in the western portion of the complex. Strongly elongate bodies and folding did not develop. Shallow-to-moderately dipping primary foliation shows no preferred orientation and is warped. The mineralogical similarity between the intrusive rocks of the western and eastern Hazard Creek complex suggests that the intrusives are broadly coeval. Their chemistry, discussed in the next chapter, also suggests that they are related. Thus, during the broad interval of epidote-bearing pluton emplacement, a strong directional grain and associated folding developed in the eastern portion of the complex but is absent from the west. Plutons in the west show abundant evidence of stoping and diking that is absent from the east. Fabrics developed during this time are steeply dipping in the east and moderately to shallowly dipping in the west.

The transition between the eastern and western domains is complex. On the large scale, foliation transitions from steep to shallow gradually (Figure 3.4; Plates 1 and 2). Locally, narrow zones of steep fabric are present within the intermediately dipping packages, suggesting local zones of concentrated deformation. The transition from intrusion accompanied by ductile deformation, to intrusion by stoping, is more obscure. Blocky Light and Dark Gneiss is characterized by abundant randomly oriented dikes and blocky inclusions. This zone is bounded on both sides by ductilely deformed material. Foliations in the ductilely deformed material dip moderate-to-steeply to the east, while foliations in the blocky material dip more shallowly and are more variable in orientation (Plate 2). The blocky fabric may have developed first in this area and been overprinted by the more steeply dipping ductile fabric. The westernmost observed ductile fabrics occur along Hazard Creek, along the eastern border of Quartz Diorite Gneiss. West of this point the gneiss is cross-cut by randomly oriented dikes. The transition from consistent foliation orientation with intermediate easterly dip, to variable foliation orientation with relatively shallow strikes and dips, appears to occur fairly abruptly near this point. A structure separating the two deformational styles has not been observed.

3.2.3 Younger Deformation of the Eastern Hazard Creek Complex

The youngest deformation in the Hazard Creek complex is present in the eastern part of the unit. A strong mylonitic fabric is present in the Eastern Tonalite Gneiss and the Garnet Granodiorite Mylonite. A strong foliation defined by orientation of mafic minerals, feldspar augens, and flattened quartz grains, strikes north-south and dips 80°

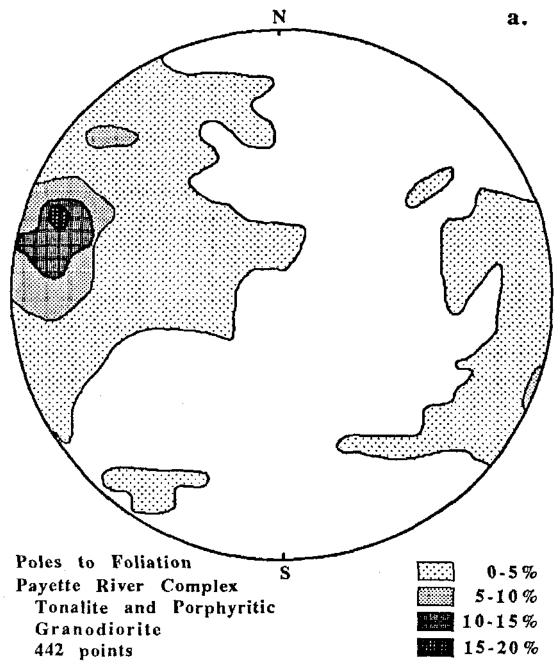
east (Figure 3.10b). Lineation defined by the orientation of hornblende and feldspar augen, and by the streaking out of biotite and quartz plunge down-dip (Figure 3.11b). Garnet Granodiorite Mylonite is the youngest unit in the eastern Hazard Creek complex, indicating that deformation postdated emplacement of the eastern portion of the complex. Some deformation of the Eastern Tonalite Gneiss occurred after the emplacement of potassium feldspar-bearing pegmatite dikes inferred to be related to the Garnet Granodiorite Mylonite. The Eastern Tonalite Gneiss of the Hazard Creek complex and the Little Goose Creek complex are interlayered on scales from several meters to several hundred meters, and all contacts are transposed parallel to the foliation. The mylonitic fabric is similar in style and orientation to that in the Little Goose Creek complex, suggesting that the youngest deformation of the Hazard Creek complex was comtemporaneous with the mylonitization of the Little Goose Creek complex.

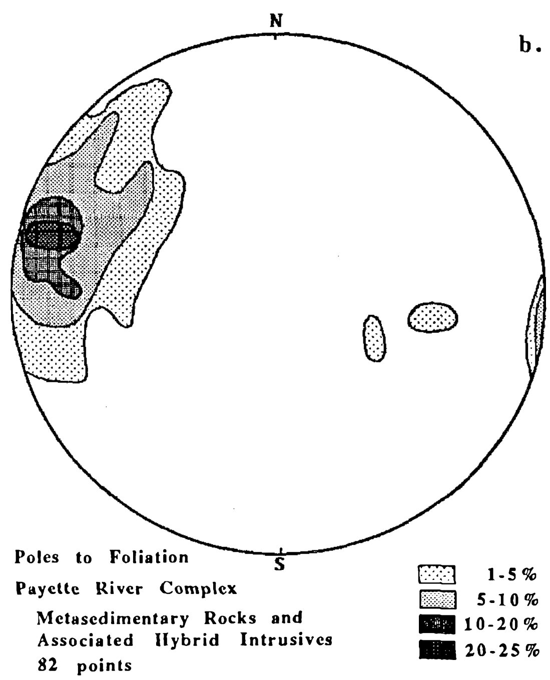
Other structures attributed to this deformation within the Hazard Creek complex occur within Quartz Diorite Gneiss and Jack's Creek Pluton near Granite Twin Lakes where cross-cutting granite pegmatite dikes are sheared along their margins and develop a weak boudinage structure. Locally, they have been folded into similar isoclinal folds. Axial planes of these folds are parallel to older gneissic foliation. Mylonitic fabrics observed in Quartz Diorite Gneiss and Jack's Creek Pluton described above may have formed during this deformation.

3.3 DEFORMATION OF THE PAYETTE RIVER COMPLEX

3.3.1 Deformation Coeval with Plutonism

Structural fabrics in the Payette River complex are associated primarily with the intrusion of plutons. A strong igneous foliation developed during intrusion, which is preserved in the alignment of mafic minerals and of flattened mafic inclusions. Foliation strikes N10E and dips 70 to 80^{o} east (Figure 3.6a). A locally developed lineation,





<u>Figure 3.6</u> Contoured lower hemisphere equal area stereonet plots of poles to foliation: a) flow foliation in Payette River Tonalite and Porphyritic Granodiorite; b) layering and metamorphic foliation in Metasedimentary Rocks and Associated Hybrid Intrusives of Payette River complex.

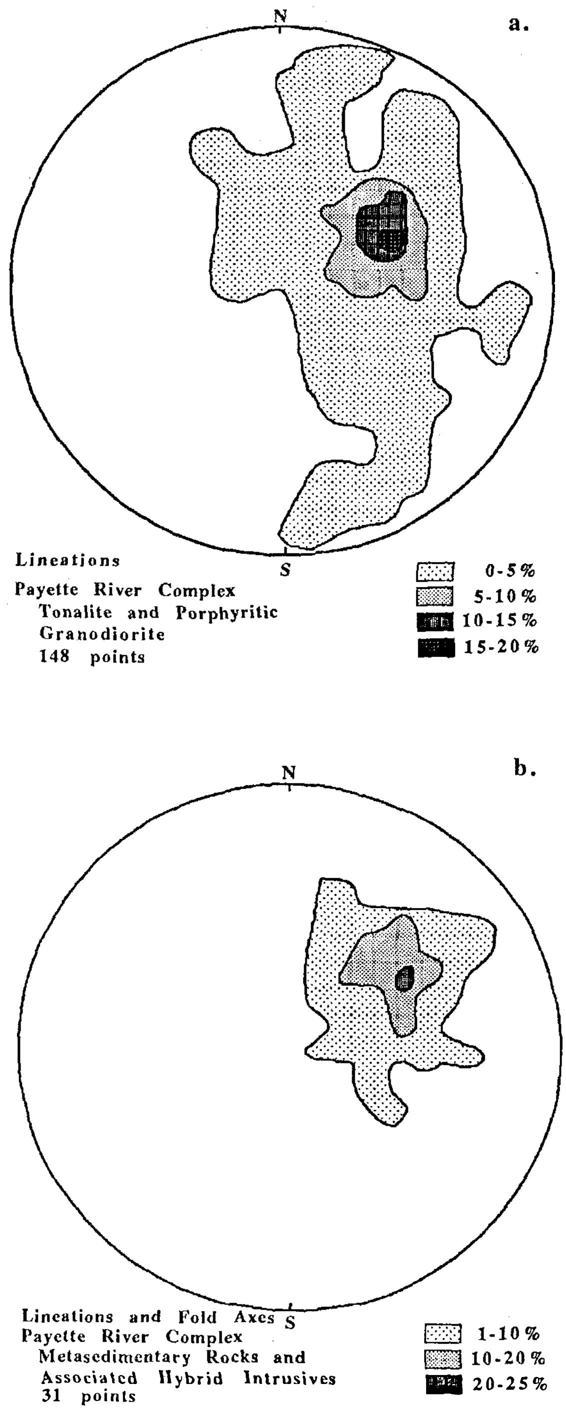


Figure 3.7 Contoured lower hemisphere equal area stereonet plots of: a) flow mineral lineation in Payette River Tonalite and Porphyritic Granodiorite; b) mineral lineation and fold axes in Metasedimentary Rocks and Associated Hybrid Intrusives of Payette River complex.

defined by the alignment of hornblende and the long axes of mafic inclusions, generally plunges 50 to 60° to N55E (Figure 3.7a). Sheets of granite up to two kilometers wide are semi-concordant with the foliation throughout the complex (Plates 1 and 2).

Metasedimentary rocks in screens within the Payette River complex have foliation that parallels the flow foliation in the intrusive rocks (Figure 3.6). Compositional layering, metamorphic foliation and local layers of mobilzed material are deformed into tight-to-isoclinal folds with wavelengths of several centimeters to a meter. Hybrid igneous material inferred to have formed when tonalitic melt assimilated metasedimentary material is also folded. Folds are often chaotic. However, in general, fold axes and metamorphic mineral lineations parallel the lineation in the tonalite (Figure 3.7). The concordance of structures and the presence of folded hybrid material within the metasedimentary screens suggests that partial melting, deformation, and folding occurred during emplacement of the Payette River Tonalite. Older transposed structures may also be present. The strong primary foliation and steeply northeast plunging lineation in the intrusives suggest they were emplaced by flow upward with a southerly component under east-west compression.

3.3.2 Younger Mylonitic Deformation of the Western Payette River Complex

Along the western border of the complex, near Brundage Reservoir, tonalites have been strongly deformed into mylonites with fabrics that are similar in style and parallel to those in the Little Goose Creek complex (Figures 3.10 and 3.11). This deformation is contiguous with that in the Little Goose Creek complex, and deforms the contact between the two units, suggesting it was contemporaneous with that in the Little Goose Creek complex.

Elsewhere in the western portion of the Payette River complex, west of the Payette River, primary igneous fabrics are weakly deformed. Quartz is locally deformed into subgrains and plagioclase is bent; mafic minerals are unaffected. These fabrics may be related to the mylonitic deformation. Rocks in the eastern Payette River complex show no evidence of deformation after emplacement.

3.4 DEFORMATION OF THE LITTLE GOOSE CREEK COMPLEX

3.4.1 Older Deformation Prior to and Coeval with Plutonism

A gneissic foliation in the Spotted Orthogneiss is folded and cross-cut by Porphyritic Orthogneiss (Figure 3.8). A younger mylonitic fabric is superimposed on the older fabric, locally at a high angle.

The strong gneissic layering in the Western Orthogneiss is cross-cut by leucocratic granitic intrusives, inferred to be related to the Porphyritic Orthogneiss. Dikes are commonly isoclinally folded with wavelengths from 10 to 30 centimeters. Fold axes are commonly vertical though some shallow plunging fold axes were also observed. The variable deformation of the dikes suggests that they were intruded during deformation (Figure 3.9).

These relationships show that some deformation of the Little Goose Creek complex occurred prior to, and during, emplacement of the porphyritic granodiorite orthogneiss. Fabrics formed in this early deformation have been overprinted by a younger mylonitic event.

3.4.2 Mylonitic Deformation

The youngest set of deformational structures in the study area is centered on the Little Goose Creek complex. These structures are marked by a penetrative mylonitic deformation that has transposed most lithologic contacts and preexisting structures. Deformational effects varied spatially in strength. Layering is locally completely

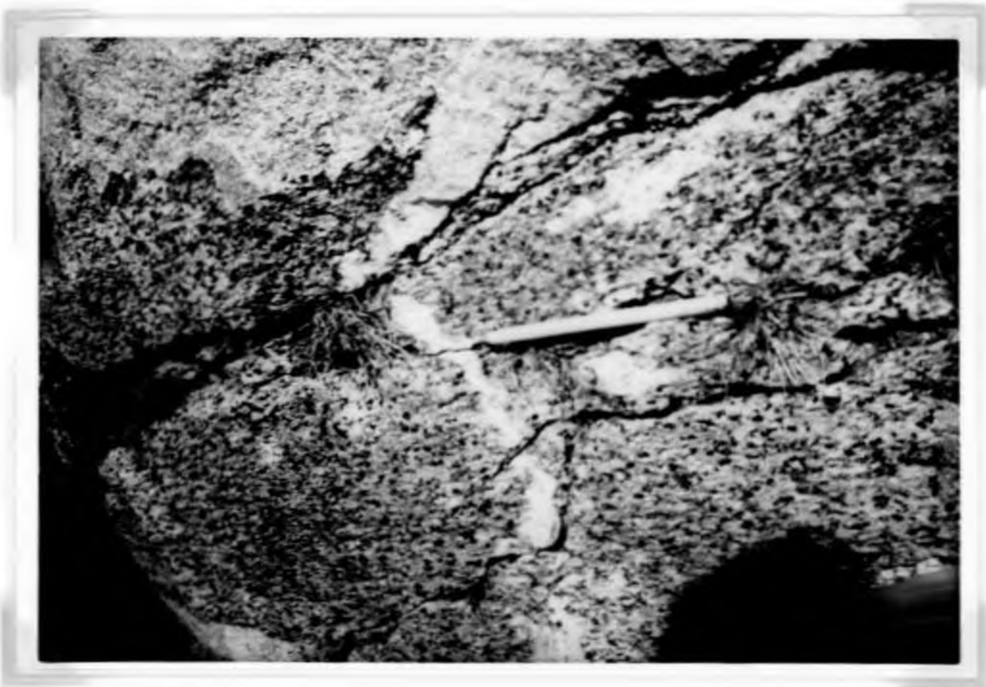


Figure 3.8 Porphyritic orthogneiss intruding Spotted Orthogneiss.

Figure 3.9 Folded and transposed potassium feldspar-bearing dikes emplaced in Tonalite Orthogneiss during ongoing deformation. Rock hammer for scale.

transposed; elsewhere, although a mylonitic fabric is well developed, earlier intrusive relationships are preserved.

A strong foliation due to the alignment of mafic minerals, flattened feldspar augen and quartz grains strikes north-south and dips 80° east (Figure 3.10). A strong down-dip lineation (Figure 3.11) is defined by trails of small biotite crystals, elongate quartz grains and oriented feldspar augen and hornblende crystals. This type of lineation is generally accepted to have formed by simple shear, and parallels the transport direction. Boudins are not abundant; however, where present they are flattened and elongate vertically indicating east-west flattening and vertical elongation.

Folded compositional layering, inclusions, dikes and sills are common. Folds are tight, several centimeters to a meter in width, and commonly asymmetric but with no consistent vergence. Axial planes parallel the foliation; plunges of axes range from shallow to steep, though steep plunges are more common (Figure 3.12). Offsets and folds perpendicular to foliation and parallel to lineation are also present but again lack a consistent sense of vergence. Folded mylonitic foliation is rare and was observed only near the margins of the complex.

The structural data all indicate that the maximum compressive stress was horizontal, oriented east-west, and was accomodated by flattening vertical transport. The inconsistent asymmetry to the folds and small offsets suggests that on a large scale pure shear may have been more important than simple shear; that is, that flattening was more important than vertical shearing. The similarity in metamorphic grade throughout the area also suggests that there may not be large offsets across the zone.

During mylonitization all minerals deformed ductilely, causing a general reduction in grain size. Cores of formerly large grains of alkali feldspar, plagioclase, and hornblende are preserved as rounded augen. Biotite has recrystallized to small ragged books; optical data suggest the biotite has not changed composition. Myrmekite has developed along the boundaries of alkali feldspar grains similar to that described by

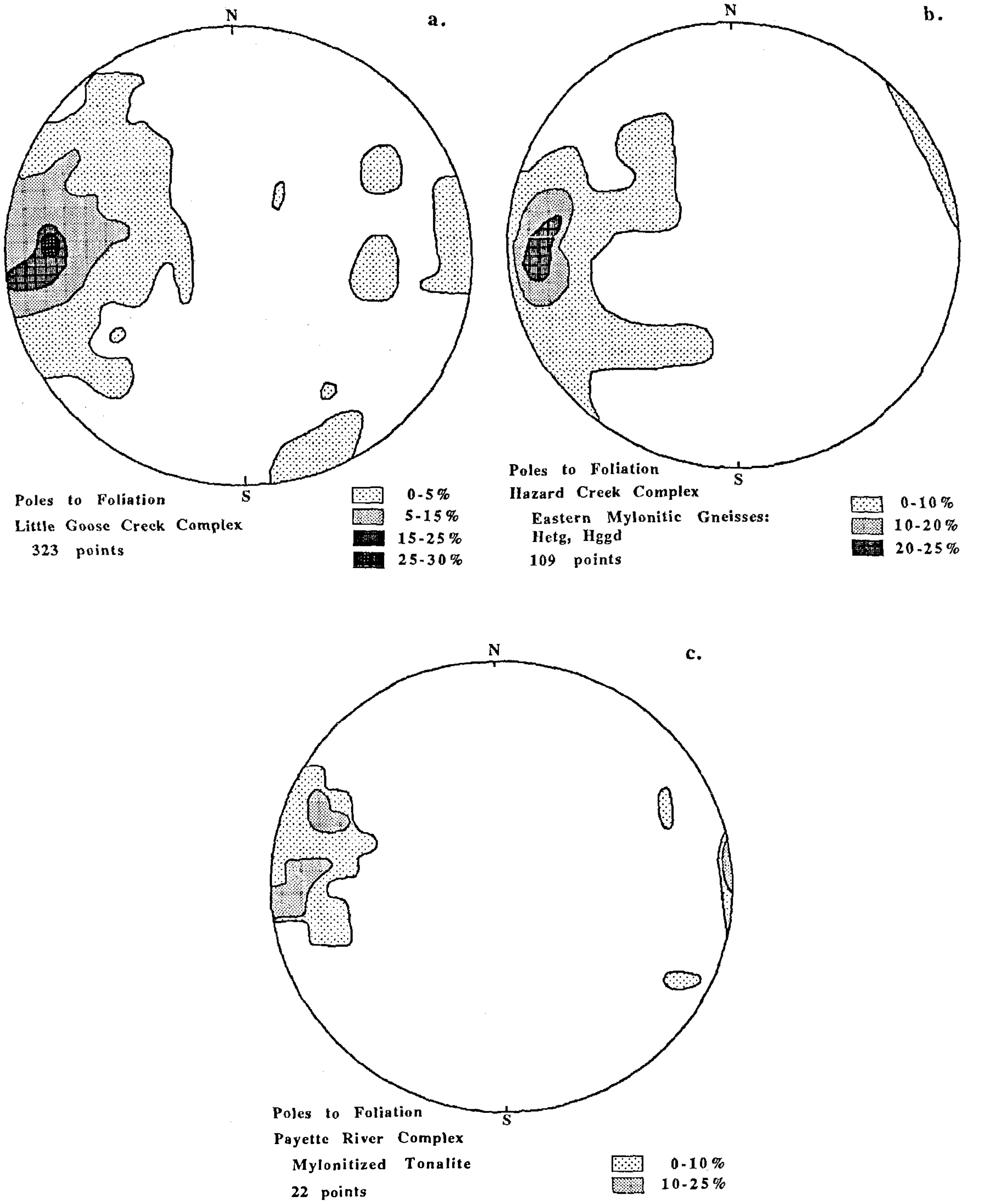


Figure 3.10 Contoured lower hemisphere equal area stereonet plots of poles to foliation in *mylonitic* rocks of a) Little Goose Creek complex, b) eastern Hazard Creek complex and c) western Payette River complex.

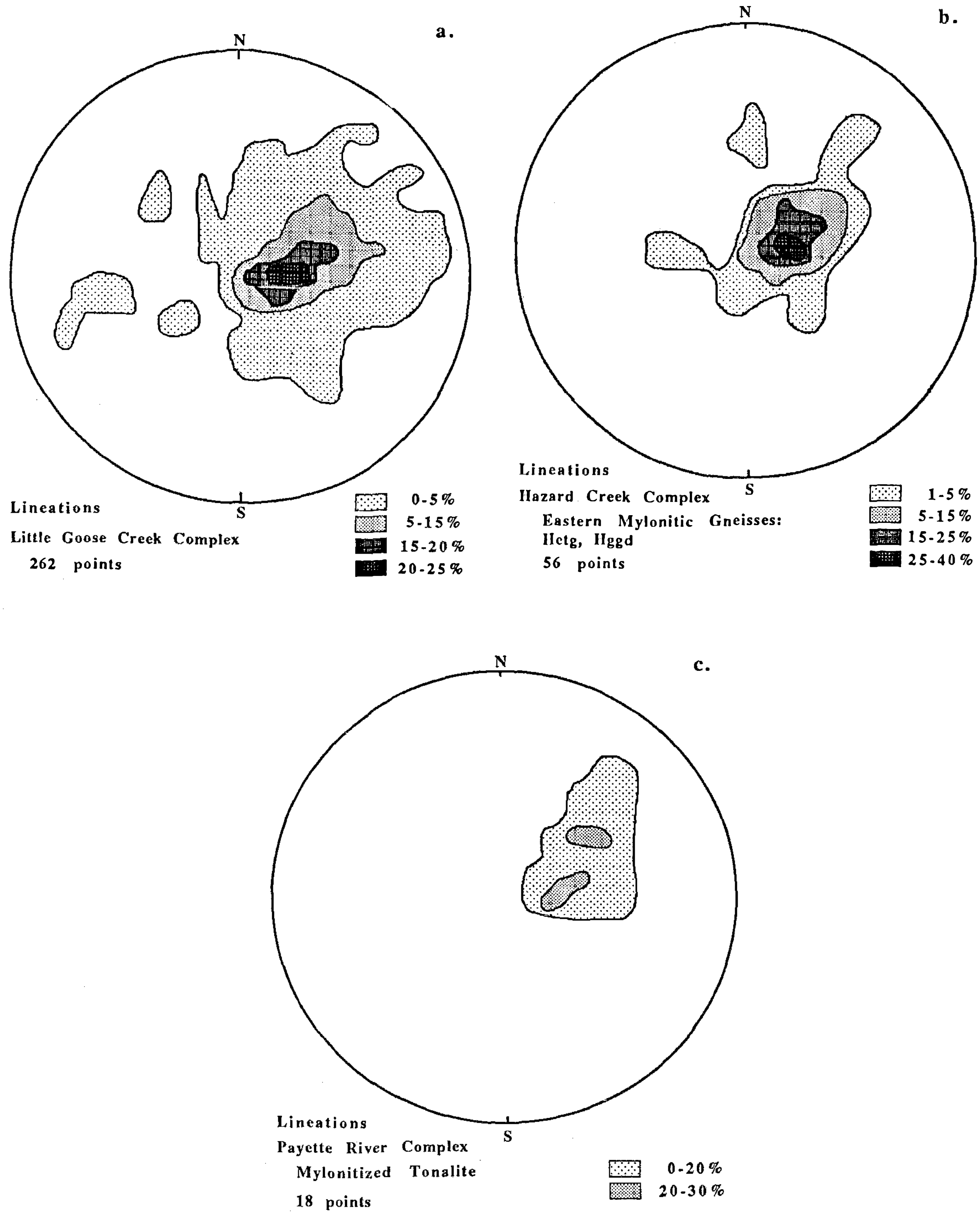


Figure 3.11 Contoured lower hemisphere equal area stereonet plots of stretching lineations in mylonitic rocks of a) Little Goose Creek complex, b) eastern Hazard Creek complex and c) western Payette River complex.

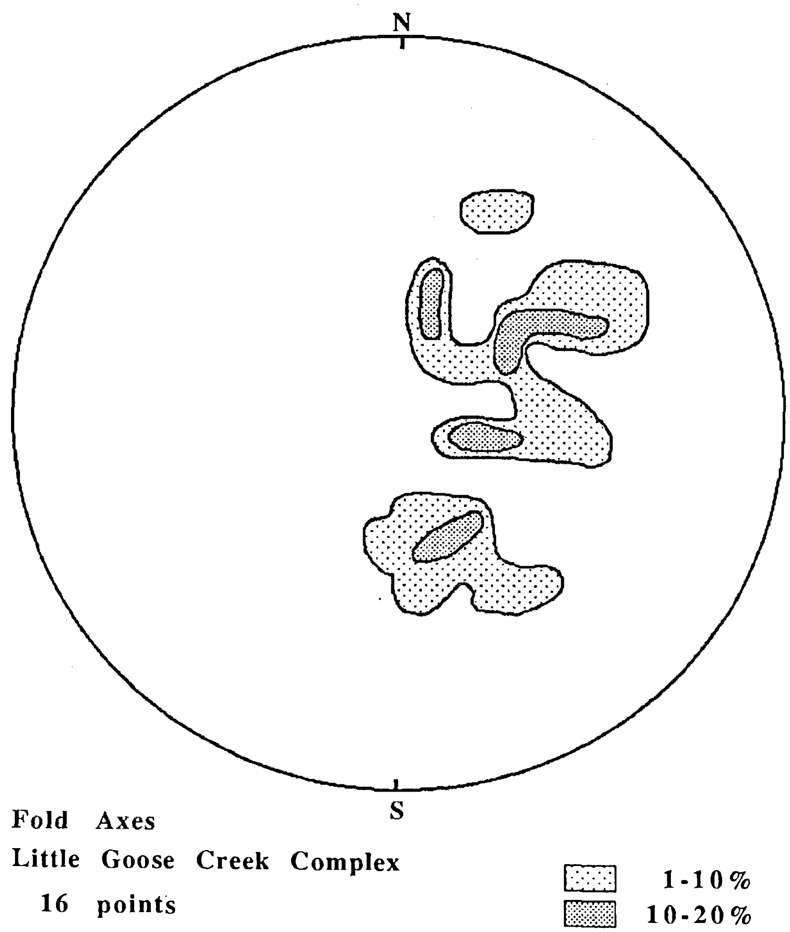


Figure 3.12 Contoured lower hemisphere equal area stereonet plot of fold axes in Little Goose Creek complex.

Simpson (1985). The ductile deformation of all grains and the development of myrmekite suggests this deformation took place at middle amphibolite facies conditions (Simpson, 1985).

Paragneisses in the zone contain the metamorphic mineral assemblages: hornblende, biotite, diopside, quartz ± potassium feldspar, sphene and opaque; and biotite, sillimanite, andesine, garnet and quartz ± potassium feldspar. These are upper amphibolite facies assemblages that may have formed prior to the deformation. However, the minerals are deformed and affected by mylonitization and show no signs of retrogression, suggesting that deformation occurred at upper amphibolite facies conditions. Under these conditions, annealing should proceed rapidly (Simpson, 1985). The preservation of a zoned pattern (see Chapter 4) in the oxygen isotopes suggests the deformation was very dry. The absence of abundant fluid may have discouraged retrogression or allowed unannealed deformation at relatively high temperatures. These data all suggest that mylonitic fabrics formed at middle to upper amphibolite facies conditions.

The mylonitic fabric formed after the emplacement of all units in the Little Goose Creek complex. As described above, similar fabrics are the youngest penetrative structures in the western Payette River complex and the eastern Hazard Creek complex. The boundaries between the three complexes are also deformed.

3.5 CENOZOIC FAULTING

The youngest structures in the area are brittle faults that offset Miocene Columbia River Basalts; they are not known to cut Pleistocene glacial deposits. The faults are most easily recognized where they offset Columbia River Basalt. They are also inferred where intrusive contacts between units are cut out and where undeformed and deformed rocks are juxtaposed. In the eastern portion of the area, these faults generally trend north-south. The most extensive eastern faults form much of the contact between the Hazard Creek and Little Goose Creek complexes and the Little Goose Creek and Payette River complexes. They follow the older contacts between the units, which appear locally along the faults. The fault traces lie in large, prominent north-south trending glacial valleys. Where basalt is present along the faults, throws appear to be between 300 and 500 feet. Similar faults downdrop basalt into the center of Payette Lake. They appear to merge at the north end of the lake and continue north along the river. These faults are not detectable in the bedrock geology suggesting that many more unrecognized faults may be present.

In the western portion of the area, Columbia River Basalt is more common and many more faults are recognized. These faults strike north-south, northeast and northwest. Locally they are closely spaced with less than 500 meters between faults. Fault planes were not observed (Plate 1). However, the faults are not deflected across slope reversals, suggesting steep dips. The sense of movement is not known. Both east and west sides are locally downdropped. Generally, basalt becomes more common to the west suggesting a net cumulative displacement of west side down. This is in keeping with observations northwest of Riggins, which suggest that the Salmon River occupies a graben that was active prior to, during and after basalt deposition (Hooper, per. comm., 1987).

3.6 SUMMARY AND DISCUSSION OF STRUCTURAL GEOLOGY

The distribution of the six sets of structures and fabrics described is summarized in Figure 3.13. The oldest set are metamorphic foliation and intrafolial folds preserved in crystalloblastic gneisses in the western Hazard Creek complex. The second set formed during the interval of emplacement of epidote-bearing intrusives. Fabrics formed in the eastern Hazard Creek complex are north-south striking, steeply dipping and have associated down-dip lineation. Intrusive bodies are elongate and folding is abundant. Fabrics formed in the western Hazard Creek complex are shallow-to-moderately dipping

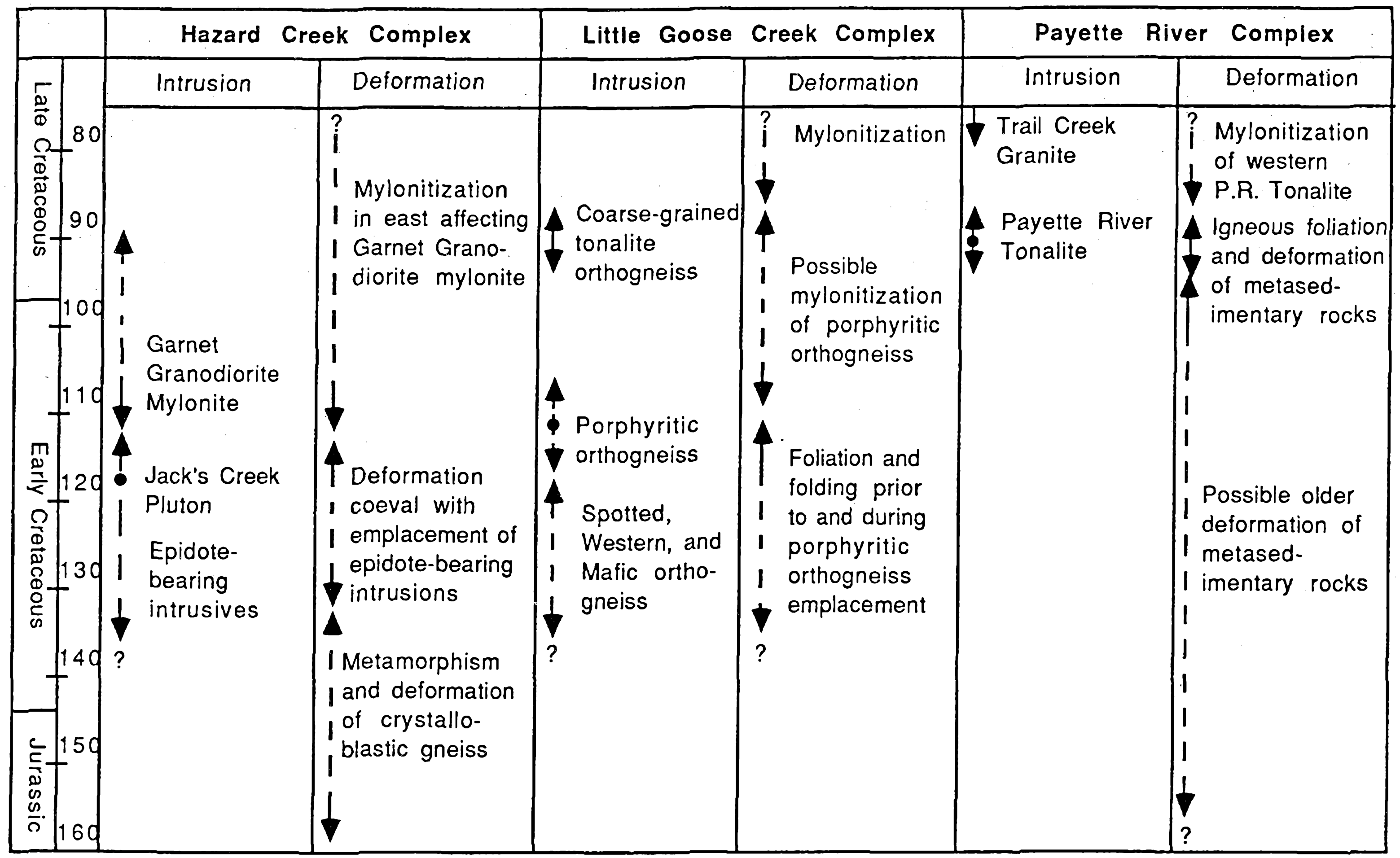


Figure 3.13 *Summary of intrusive and deformational episodes with time constraints.* Black dots indicate age dates discussed in Chapter 2.

and have no preferred orientation. Intrusive bodies are equidimensional with abundant stoping and diking relationships. Lineation and folding are not observed.

The change in intrusive and deformational style across the Hazard Creek complex described above suggests that conditions during intrusion were different on the east side than on the west. Steeply dipping foliation and down-dip lineation indicate localized east-west flattening and upward flow in the east. The presence of abundant tonalitic material throughout the complex indicates that the localization of deformation was not related to the presence or absence of magma. The area of localized compression is currently adjacent to the boundary between oceanic-arc and continental supracrustal material, suggesting that some aspect of this boundary was important in localizing the deformation. Two theories are proposed. The deformation may reflect motion between the different basement blocks along a pre-existing structure lubricated by the magma. The stresses driving this movement may have been regional or related to the emplacement of the magmas. Alternately, the craton may have acted as a physical buttress along which compressional stresses were localized. In either case, the localization of deformation suggests that the boundary was a well-defined structure at the time of emplacement of the epidote-bearing intrusives and that the complex was intruded adjacent to this boundary.

The third set of structures and fabrics are gneissic foliation and folding in the Little Goose Creek complex that formed prior to and during emplacement of porphyritic orthogneiss. The fourth set formed during the emplacment of the Payette River complex. North-south striking, steeply dipping igneous foliation and steeply north-east plunging lineation developed in the tonalite. Transposition and folding of metasedimetary screens occurred and concordant sheets of granite were emplaced.

The fifth set of structures and fabrics are the result of mylonitic deformation. This deformation transposed most intrusive contacts in the Little Goose Creek complex and deformed the boundaries between the three complexes. Fabrics, which formed in all units of the Little Goose Creek complex, in the eastern Hazard Creek complex and the western Payette River Tonalite, are centered on the boundary, suggesting it again localized deformation. Steeply dipping north-south striking foliation and down-dip lineation indicate that east-west compression was accommodated by flattening and vertical motion. The absence of a consistent vergence to folds and dike offsets is consistent with a significant flattening component to the deformation. The similarity in metamorphic grade throughout the area also suggests that large offsets may not have occurred across this zone.

The final set of structures are post-miocene faults that offset basalt. These faults form the present day contacts between complexes in most places.

Our limited knowledge of the age of plutonic rocks in the area places some constraints on the timing of deformation. These relationships are summarized in Figure 3.13. The oldest structures in the Hazard Creek complex deform mafic layered gneiss that is correlated with the Pollock Mountain Amphibolite (see Chapter 2). These structures may have formed at any time after the deposition of the protolith in Triassic time (Aliberti et al., 1988; discussed in Chapter 1). Deformed epidote-bearing trondhjemitic and tonalitic crystalloblastic gneiss is unconstrained in age; however, the gneiss is mineralogically and chemically similar to younger Cretaceous intrusives in the complex, suggesting some of the oldest deformation may be Cretaceous.

Deformation associated with the emplacement of epidote-bearing intrusives is constrained only by the 118 Ma age of the Jack's Creek Pluton. This is one of the youngest plutons in the sequence; deformation and plutonism must have begun earlier, perhaps much earlier. Deformation of the Little Goose Creek complex prior to emplacement of the porphyritic orthogneiss is constrained by its age of 111 Ma. This deformation may be contemporaneous with the deformation associated with emplacement of epidote-bearing intrusives in the Hazard Creek complex.

Structures associated with the emplacement of the Payette River Tonalite are constrained by its early Late Cretaceous age. Older transposed structures that may be present in the metasedimentary rocks could have formed at any prior time.

Mylonitic fabrics and associated structures affect all intrusives in the Little Goose Creek complex, the eastern Hazard Creek complex and the western Payette River Tonalite. The youngest affected rock of known age is the 90 Ma Payette River Tonalite. Fabric development could have begun earlier in older rocks. Hornblende from the porphyritic orthogneiss has a total gas $^{40}Ar/^{39}Ar$ age of 87 Ma (Snee, per. comm., 1987). The spectrum does not contain a good plateau. This data may be interpreted as suggesting that Ar was lost from the hornblende during deformation at approximately 87 Ma. However, the data are preliminary and may be interpreted in other ways. Brittle faulting offsets Miocene basalt, indicating that it is post-Miocene. No glacial debris is deformed.

The similarity in orientation of all deformational fabrics near the boundary between oceanic and continental supracrustal rocks suggests that flattening and vertical flow in response to east-west compression occurred along this boundary for an extended period (Figure 3.13). This deformation may have been continuous or episodic. Structures from several episodes of deformation may be unresolved.

There is no structural evidence of transpressive or strike-slip motion preserved within the area. If the complexes have been juxtaposed after their emplacement, this must have been accomplished by thinning of intervening units or the evidence for transcurrent motion must have been obliterated. Structures in all complexes suggest formation near a discrete zone of extensive deformation which was localized along the boundary between oceanic and continental supracrustal rocks.

CHAPTER 4: GEOCHEMISTRY OF INTRUSIVE ROCKS OF THE WESTERN IDAHO BATHOLITH NEAR MCCALL

4.1 INTRODUCTION

The boundary between oceanic-arc and continental supracrustal material in the study area is part of a regional boundary that continues south from Idaho, passing through the central Sierra Nevada Batholith and the eastern Mojave. Geochemical changes in these plutonic rocks have been related to this boundary (Armstrong et al., 1977; Kistler and Peterman, 1973). One such change is the transition from low to high initial $^{87}Sr/^{86}Sr$ (R_i). In Idaho this transition is unusually sharp, as was first recognized by Armstrong et al. (1977) who noted that R_i changed from less than .7043 to greater than .7055 across a discrete boundary less than ten kilometers wide, and that in general, this boundary coincided with the change from Precambrian rocks to eugeosynclinal strata. They attributed this abrupt change in isotopic systematics to assimilation of crustal materials by magmas generated in the mantle. Several of their analyses indicate that this boundary passes through the study area.

This isotopic discontinuity was studied in more detail by Fleck and Criss (1985, Criss and Fleck, 1987), who looked at both Sr and O variations. Utilizing data from three densely sampled traverses north of the present study area, they ascertained that R_i changed abruptly but continuously from less than .704 to greater than .708 across a five to 15 kilometer-wide zone. They interpreted the isotopic gradient to be due to a mixture of mantle-derived melts with Precambrian wall rocks. They also inferred that the presence of material with intermediate R_i indicated that the Precambrian and eugeosynclinal terranes were juxtaposed prior to intrusion of the plutons in the Late Cretaceous. $\delta^{18}O$ values correlated with R_i: in particular, the +8 contour correlated with the .704 contour that they interpreted to mark the eastern edge of accreted material.

This study integrates field data with geochemical data in order to understand the nature of the relationship between the change in supracrustal rocks and the geochemistry of the intrusive rocks. In this chapter, the Sr and O isotopes and major and trace element chemistry of the intrusives are described and discussed in an effort to describe the deeper levels of the oceanic arc-continent boundary and to understand its effect on the geochemistry of the plutons. Each complex recognized in the field has distinct geochemical characteristics that are described in the following sections.

4.2 DESCRIPTION OF SAMPLING DESIGN AND SOURCES OF ANALYTICAL DATA

44 samples were analyzed for Sr and O isotope ratios at the California Institute of Technology. Sample locations are shown on Plate 3. Analytical techniques and precision are discussed in Appendix B. $\delta^{18}O$ is considered precise to ±0.2 per mil and is referenced to SMOW; $^{87}Sr/^{86}Sr$ is precise to ± .00004-.0001 and concentration data to 1%. Initial ratios were calculated using ages of 120, 110, 90 and 60 Ma for samples from the Hazard Creek complex, Little Goose Creek complex, Payette River Tonalite and Trail Creek Granite respectively. Field observations indicate that many Hazard Creek complex samples and Little Goose Creek tonalite samples are older than the age used in the calculation. Errors introduced into R_i are small, however, relative to variation in the study area: calculations for an age of 150 Ma, an appropriate upper limit, changes R_i by less than .0002 for Little Goose Creek tonalites and .00005 for most Hazard Creek samples.

XRF major and trace element chemical analyses for 36 samples from the study area were obtained from Dr. Mel Kuntz of the U.S. Geological Survey. Duplicate analyses are shown in Appendix B. Most duplicate analyses of the same powder have precision of

±0.5% for major elements. Trace element data have much lower precision, with variations as large as 40%. Rare earth element (REE) and additional trace element data for five samples were obtained by Amy Hoover at Oregon State University using INAA.

Comparison of calculated norms with estimated modes suggests that quartz was often overestimated and plagioclase underestimated by ten to twenty percent, particularly in samples from the Little Goose Creek complex where plagioclase is untwinned. Alkali feldspar appears to be overestimated in samples from the Trail Creek Granite. Most alkali feldspar in the Little Goose Creek complex is in large megacrysts, thus modes estimated from thin sections tend to underestimate the alkali feldspar present in the unit.

Samples from each complex are subdivided into groups of like composition and texture. Samples from the Hazard Creek complex form three groups. The first group, called epidote tonalites, are slightly to moderately deformed tonalites and trondhjemites. These are the most abundant rocks in the complex. 83z1 is from the eastern part of Epidote Biotite Tonalite. 83z3 and 83z9 are from the Jack's Creek Pluton. 85z11 is an epidote-hornblende-biotite tonalite in the Light Gneiss Assemblage. 85z17 is from a two meter trondhjemite dike. The second group, called older gneisses, are crystalloblastic, quartz dioritic to tonalitic orthogneiss with color index 25 to 35. Sample 83z8 is from Quartz Diorite Gneiss. 85z15 is from the Light Gneiss Assemblage. The third group, called K-rich suite, contains only one sample (83z12) of granite from Garnet Granodiorite Mylonite.

Additional chemical data for the Hazard Creek complex were obtained from Zen and Hammarstrom (in prep.) and from Barker et al. (1979). Zen and Hammarstrom present chemical data from the Light Gneiss Assemblage along the Little Salmon River. The location of samples from Barker et al. is not known. This data should be treated with care as samples from dikes within gneisses to the north and west of the study area may be included. Their relationship to the Hazard Creek complex and its plutonic history have not been established.

Samples from the Little Goose Creek complex form four groups. The first group, called Little Goose Creek complex tonalites, contains four samples: 85z18 from the Spotted Orthogneiss, 84z1 from the Western Orthogneiss, and 84z3 and 84z4 which are from one to ten meter screens in the Porphyritic Orthogneiss. These samples range in composition from monzodiorite to quartz-monzodiorite; they are called tonalites because of their high color index.

Samples of porphyritic orthogneiss form two groups: western samples that have R_i less than .706 (83z14, 84z2a, 84z2b, 84z8b, 85z6, 84z15, 84z6a, 86z11, 85z8, 86z53) and eastern samples that have R_i greater than .7065 (85z1, 85z2, 85z12, 86z20, 86z6, 85z13, 86z52). These samples are of the porphyritic orthogneiss lithology within the unit Porphyritic Orthogneiss. The protolith to this orthogneiss is inferred to have been emplaced in a single intrusive episode. The samples were selected to give profiles across the unit. The distinction between eastern and western porphyritic orthogneiss is not obvious on the basis of field or petrographic data, although eastern samples tend to contain more alkali feldspar and often contain garnet. The final group contains one sample (85z9) from Coarse Grained Tonalite Orthogneiss.

Samples from the Payette River complex form three groups: Payette River Tonalite (83z11, 83z2, 85z14, 85z4, 83z6, 86z24, 86z21, 86z18, 86z47), Payette River Granite (84z21) and Trail Creek Granite (83z15, 83z16a, 86z17). Samples 84z21 and 86z17 are from bodies within the Payette River Tonalite. Samples 83z15 and 83z16a are from the main masses of Trail Creek Granite. The chemical analyses suggest that much of the Payette River Tonalite is quartz-dioritic in composition.

4.3 PRESENTATION OF DATA

4.3.1 Sr and O Isotopic Data

The results of Sr and O isotopic analyses are listed in Table 4.1 and plotted on the geologic map in Figures 4.1 and 4.2. The abrupt change in isotopic ratios noted by earlier workers is a combination of variablity within and between major intrusive complexes. The rapid change from low to high ratios is a feature of the Little Goose Creek complex, which also contains the boundary between oceanic-arc and continental supracrustal rocks. The complete range of values are observed within the porphyritic orthogneiss, indicating that the isotopic boundary is a mixing phenomenon and is not dueto the juxtaposition of plutonic rocks of different isotopic character. The Hazard Creek complex to the west has low R_i (generally less than .704) and $\delta^{18}O$ (8). The Payette River complex to the east has generally high R_i and moderate $\delta^{18}O$, which are not geographically organized. More is learned about the cause of the isotopic variations by analyzing each complex individually as a geochemical unit.

4.3.1.a Hazard Creek Complex

Samples from the Hazard Creek complex have the lowest R_i in the area. Epidote tonalite samples range between .7037 and .7039. These values are typical of oceanic-arc magmas, which often have R_i between .703 and .704 (Gill, 1981). They are also similar to values measured in the western portions of the Peninsular Ranges Batholith (Taylor and Silver, 1978; Silver et al., 1975; Silver et al., 1979) and the Sierra Nevada Batholith (Kistler and Peterman, 1973). They are higher than the typical values for normal modern MORB, which have a range of .7022 to .7035 and a mean of .7029 (Hofmann and Hart, 1978).

Samples from the Hazard Creek complex also have the lowest $\delta^{18}O$ in the study area. $\delta^{18}O$ in epidote tonalite samples is between 7.6 and 8.4. These values are higher

Table 4.1a: Sr and O isotopic measurements

HAZARD CREEK COMPLEX

	Rb	Sr	Rm	error	Ri	O18	Age
Epidote Tonalites							
83z1	22	768	.70386	3.6E-05	.70372	8.4	120
83z3	3	1404	.70390	7.5E-05	.70389	7.9	120
83z9	16	963	.70394	2.6E-05	.70386	7.9	120
85z11	25	718	.70396	3.2E-05	.70379	7.6	120
85z17-dike	79	781	.70400	5.8E-05	.70350	10.0	120
Older Gneiss							
83z8	37	630	.70471	5.4E-05	.70442	8.0	120
High K Suite							
83z12	57	315	.70475	2.4E-05	.70386	9.0	120
Hornblendite							
84z25a	3	202	.70400	8.8E-05	.70392	6.4	120

Table 4.1b: Sr and O isotopic measurements

LITTLE GOOSE CREEK COMPLEX

	Rb	Sr	Rm	error	Ri	O18	Age
Tonalites							
84z1	62	797	.70436	3.7E-05	.70401	8.0	110
84z3	25	864	.70482	3.8E-05	.70469	8.5	110
84z4	62	713	.70458	3.9E-05	.70419	8.0	110
85z9	39	662	.70785	3.7E-05	.70758	9.9	110
Western Porphyritic Orthogneiss							
83z14	34	1079	.70486	2.7E-05	.70472	8.8	110
84z2a	51	1003	.70521	3.4E-05	.70498	8.9	110
84z2b	56	856	.70532	2.8E-05	.70502	8.9	110
84z8b	35	908	.70558	3.1E-05	.70540	9.5	110
84z15	37	890	.70527	3.9E-05	.70509	9.5	110
85z6	43	935	.70524	3.9E-05	.70503	9.1	110
85z8	27	1067	.70549	3.6E-05	.70538	9.3	110
86z11	36	992	.70468	3.4E-05	.70451	9.3	110
86z53	6	1760	.70599	2.6E-05	.70597	9.7	110
Eastern Porphyritic Orthogneiss							
85z1	62	820	.70712	3.4E-05	.70678	10.9	110
85z2	122	681	.70946	2.6E-05	.70865	10.4	110
85z12	79	662	.70928	5.6E-05	.70874	10.5	110
85z13	85	518	.71041	3.9E-05	.70967	10.7	110
86z6	43	902	.70788	2.7E-05	.70766	9.6	110
86z20	57	707	.70735	3.6E-05	.70698	10.2	110
86z52	75	708	.70942	3.1E-05	.70894	10.6	110

Table 4.1c: Sr and O isotopic measurements

PAYETTE RIVER COMPLEX

	Rb	Sr	Rm	error	Ri	O18	Age
Payette River Tonalite							
83z2	69	666	.70785	3.7E-05	.70747	9.3	90
83z11	63	675	.70866		.70832	8.8	90
85z4	140	400	.70979	2.2E-05	.70849	10.2	90
85z14	41	644	.70780		.70756	8.5	90
86z18	53	873	.70829	2.5E-05	.70806	9.5	90
86z21	50	751	.70889	2.9E-05	.70865	9.3	90
86z24	56	742	.70933	2.7E-05	.70905	9.2	90
86z32	79	405	.71027		.70954	10.2	90
86z47	88	583	.70904	3.7E-05	.70848	7.2	90
Trail Creek Granite							
83z15	87	514	.70917		.70875	10.4	60
83z16a	145	369	.71058	3.4E-05	.70961	9.6	60
86z17	98	565	.70852	2.5E-05	.70809	10.2	60
Payette River Granite							
84z21	74	538	.71032	2.3E-05	.70981	10.3	90
WALL ROCKS							
Layered Mafic Gneisses							
86z41	15	1116	.70672	2.4E-05	.70667	9.7	90
85z10	23	90	.70621	2.7E-05	.70527	10.2	90
Continentally Derived Schist							
86z31	103	111	.74619	4.3E-05	.74274	12.3	90

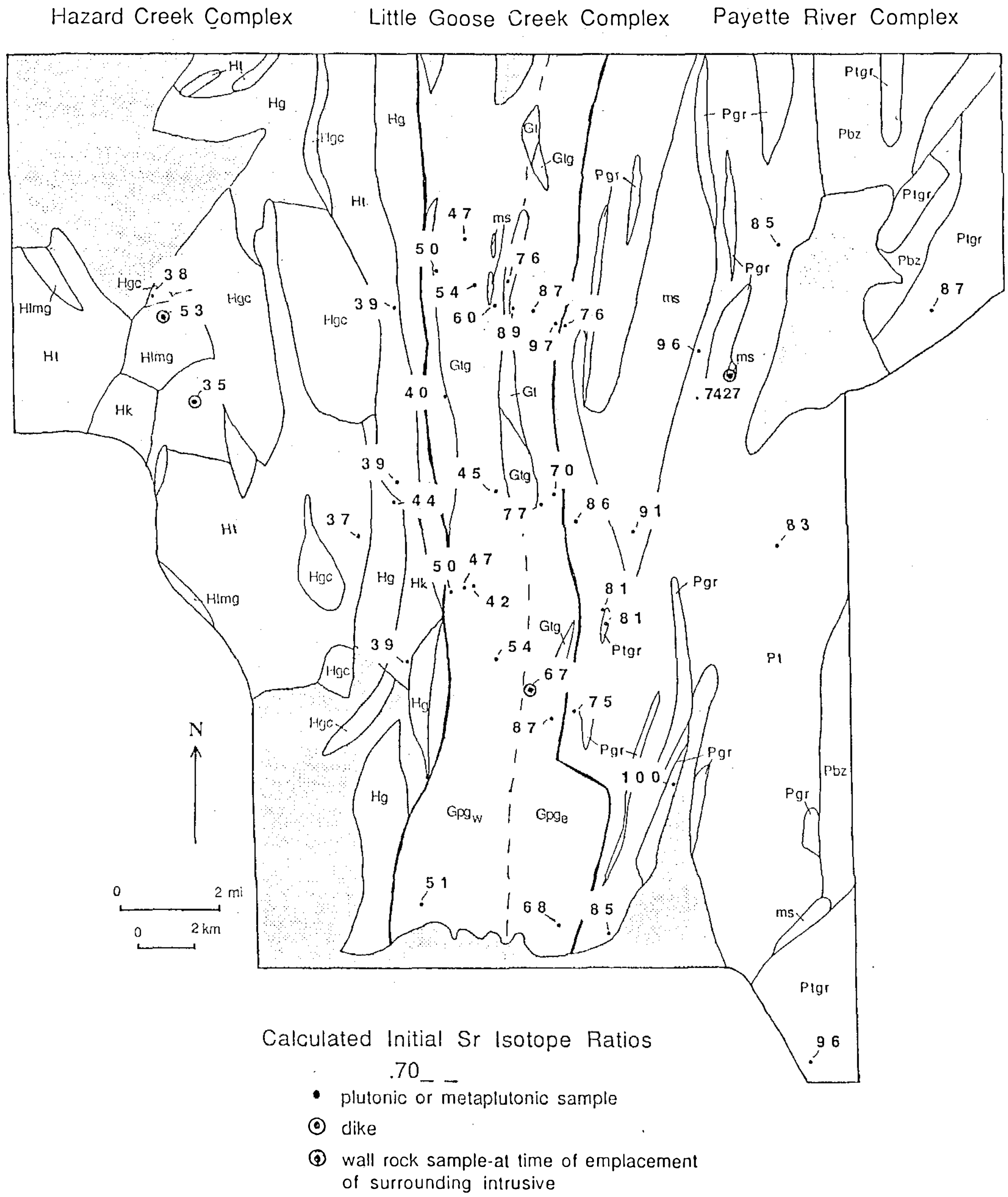


Figure 4.1 a) Geologic map showing calculated R_i. Legend for geologic map is shown in Figure 2.1.

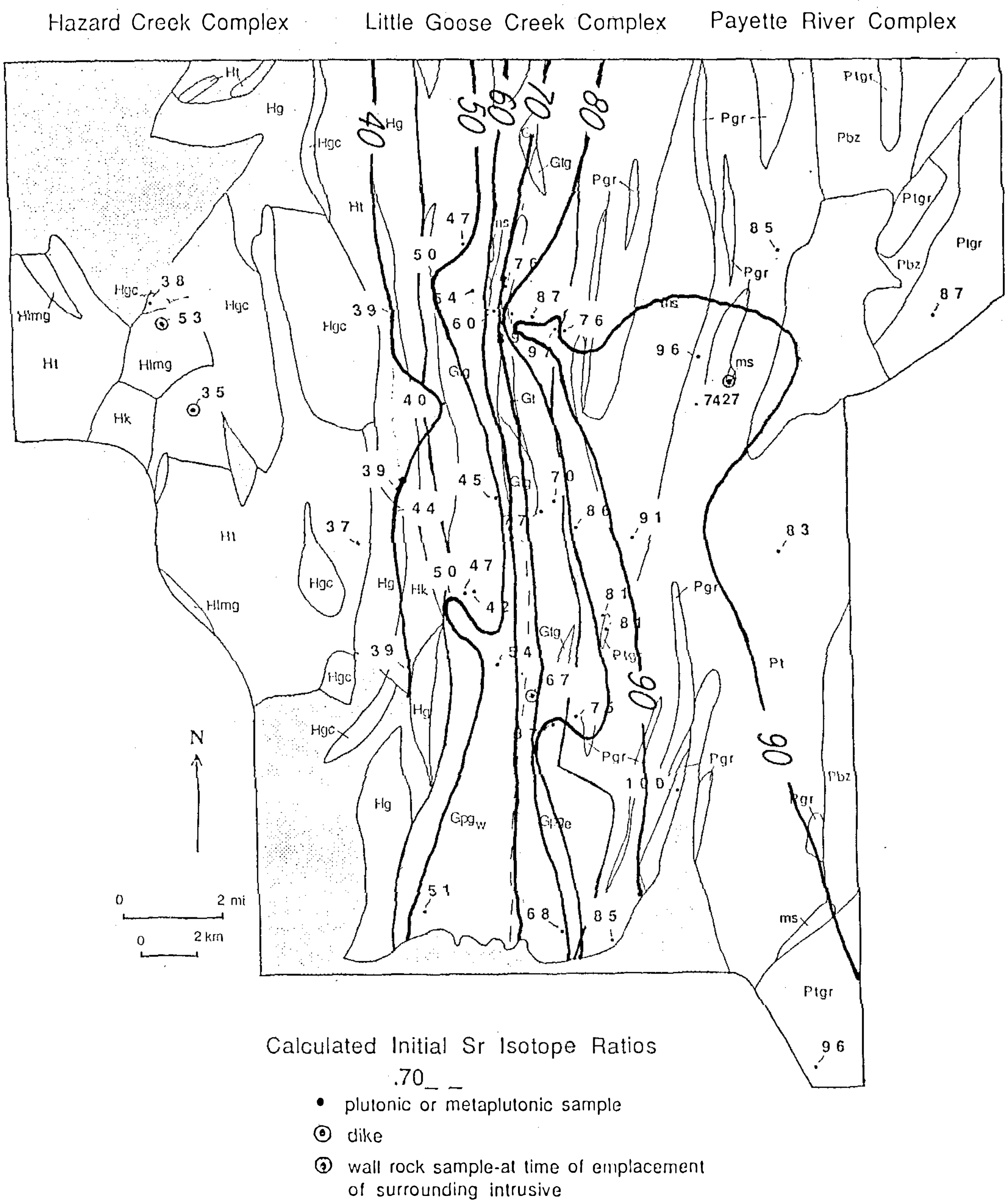


Figure 4.1 b) Contour map of R_i.

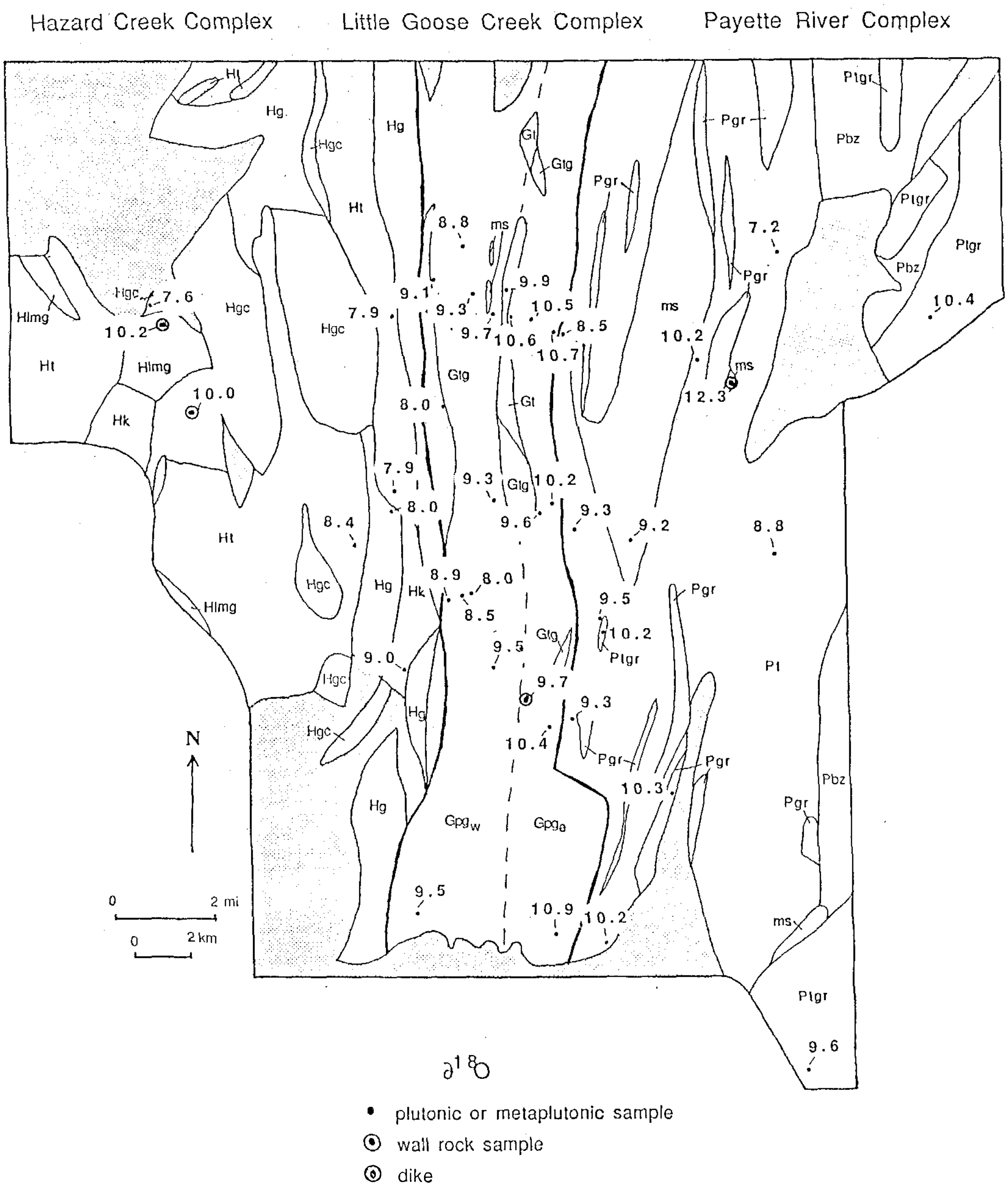


Figure 4.2 a) Geologic map showing δ18O. Legend for geologic map shown in Figure 2.1.

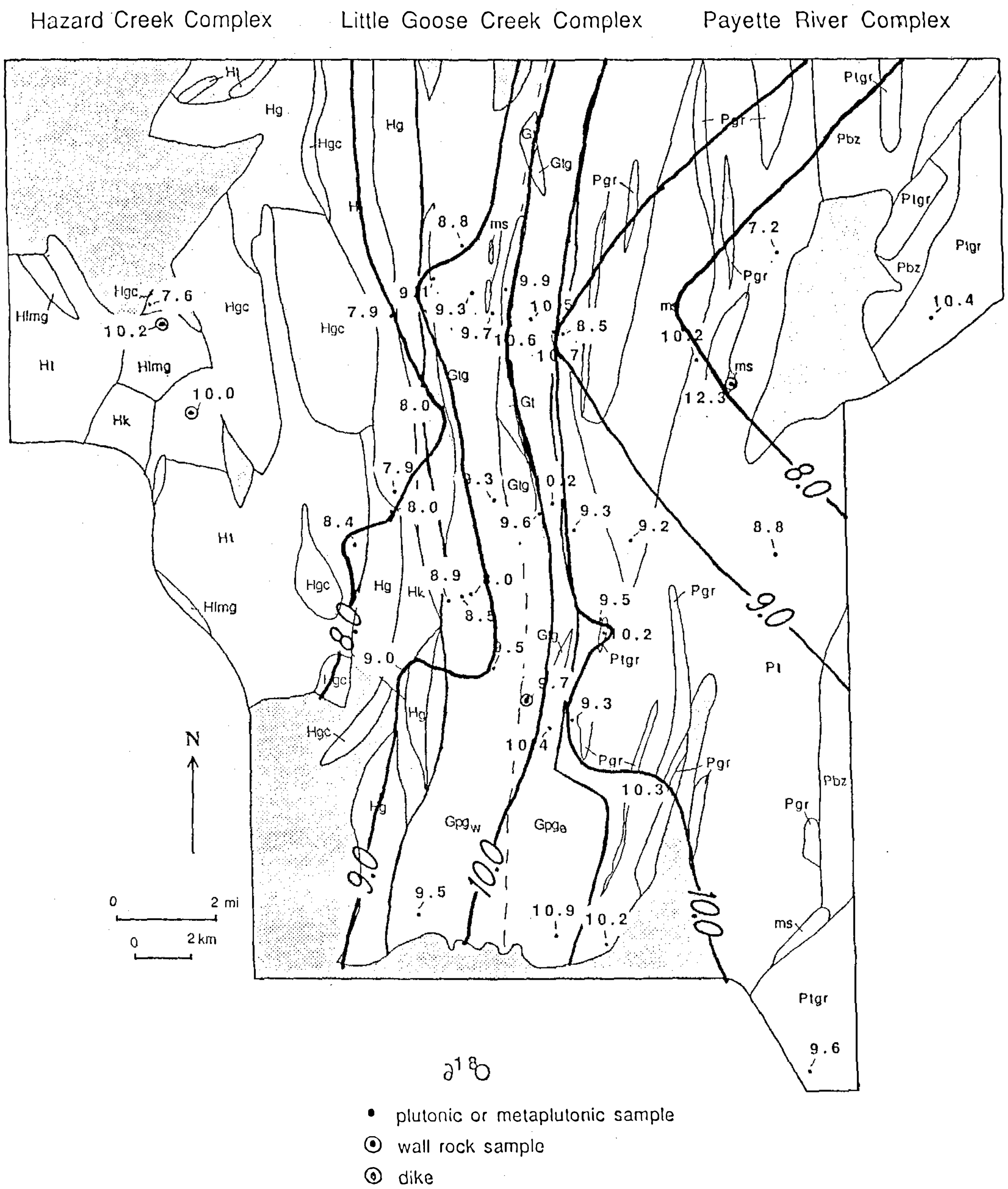


Figure 4.2 b) Contour map of δ18O.

than is typical for oceanic-arcs, which are usually in the range 5.5 to 7.0 (Gill, 1981). $\delta^{18}O$ in the Hazard Creek complex is similar to that in the western Peninsular Ranges Batholith (Taylor and Silver, 1978; Silver et al., 1975; Silver et al., 1979) and the southern and northern sections of the western Sierra Nevada Batholith (Masi et al., 1981). $\delta^{18}O$ is not correlated with R_i (Figure 4.3a). The high value of 10 for dike sample 85z11 may reflect exchange with its wall rocks of mafic layered gneiss that have $\delta^{18}O$ of 10. The high value of 9 obtained from the K-rich suite is not well understood.

Samples from the Hazard Creek complex have a wide range in Sr concentration accompanied by little change in R_i (Figure 4.4a). Epidote tonalites have Sr concentrations between 718 and 1404 ppm. A petrologic model for this suite must produce a high but variable Sr concentration, with little variation in R_i, and account for the elevated $\delta^{18}O$.

4.3.1.b Little Goose Creek Complex

Samples from the Little Goose Creek complex show the greatest range in R_i. Within the porphyritic orthogneiss, which is believed to be a single intrusive complex, R_i has an extraordinary range of .7045 to .7097. R_i is lowest on the west and rises gradually to an area in the east central portion of the unit, where values jump abruptly, rising from values between .7045 and .7060 west of the step, to values between .7077 and .7090 east of the step. All traverses show a large step from relatively low values on the west to high values on the east. Values west and east of the step vary from traverse to traverse and isopleths do not parallel the step in detail. This step is defined as the boundary between the western and eastern porphyritic orthogneiss. It correlates generally but not in detail with the change in supracrustal rocks within the complex (Figure 4.1). In many places, the step occurs across the Coarse Grained Tonalite Orthogneiss, which has R_i (.70758) between that of the porphyritic orthogneiss on either side of it. Little Goose Creek complex tonalites have slightly lower R_i than the

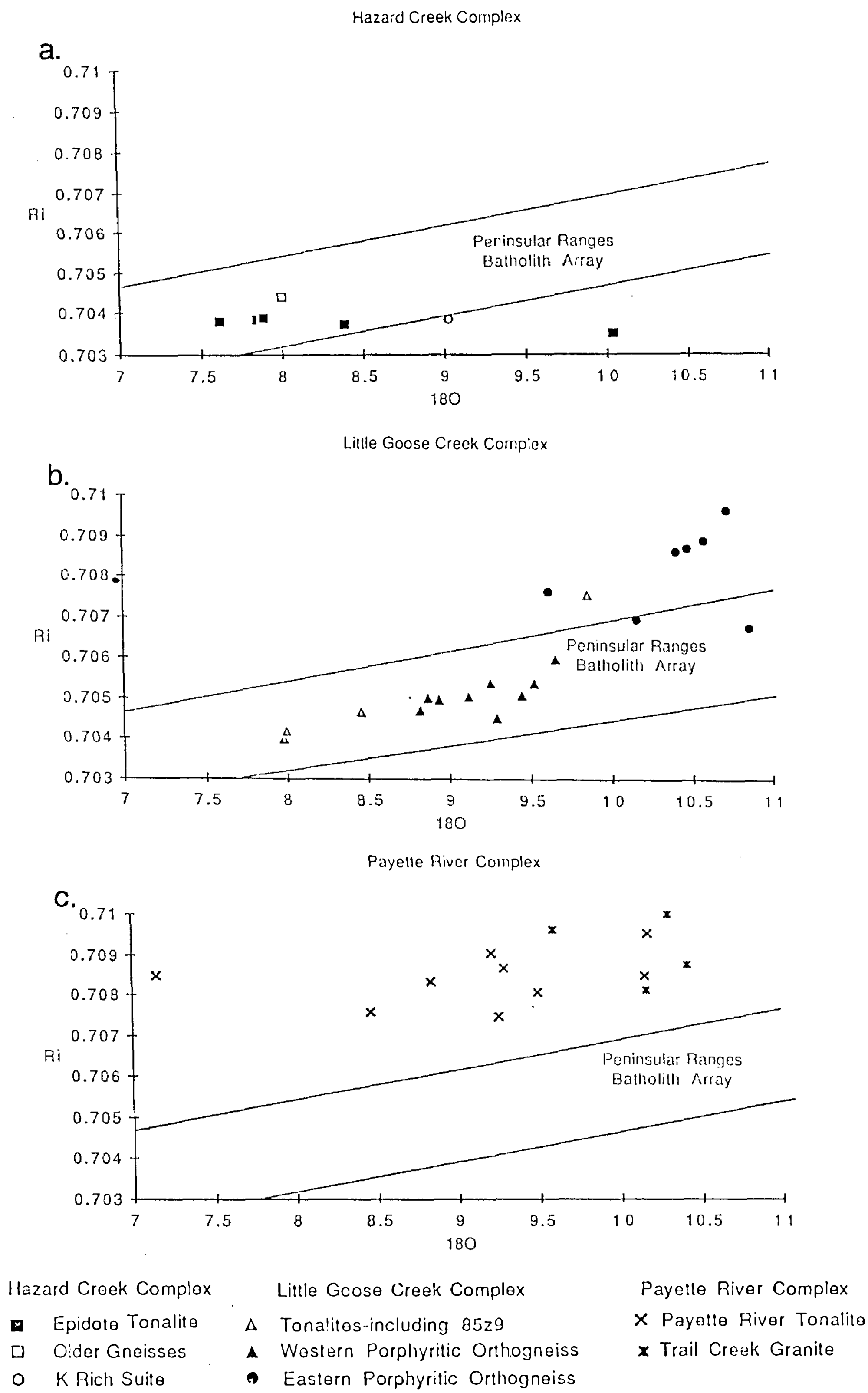
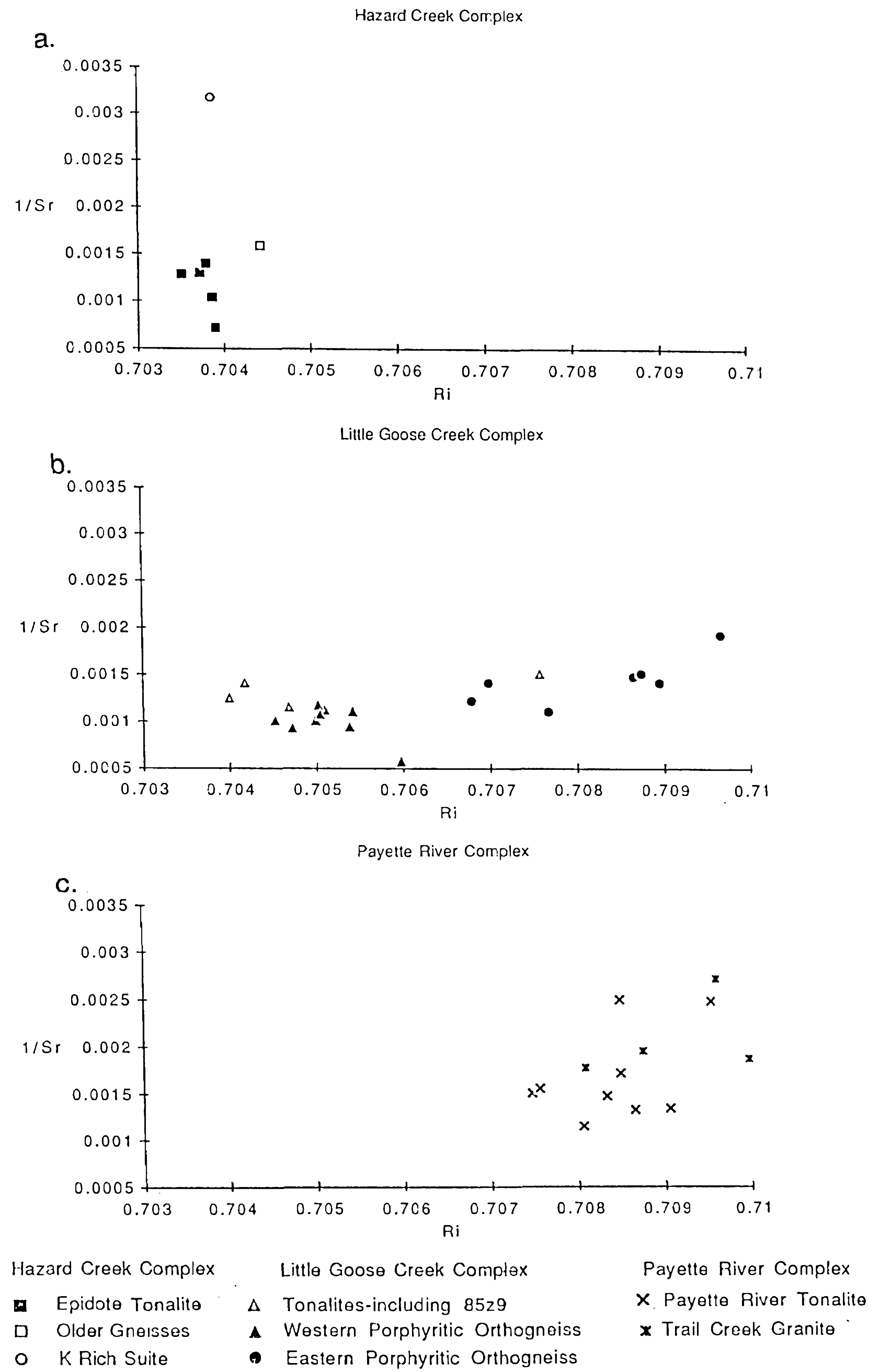


Figure 4.3 R_i versus $\delta^{18}O$ for plutonic and metaplutonic samples: a) Hazard Creek complex b) Little Goose Creek complex c) Payette River complex Peninsular Ranges Batholith Array from Taylor and Silver, 1978.



<u>Figure 4.4</u> 1/Sr versus R_i for plutonic and metaplutonic samples: a) Hazard Creek complex b) Little Goose Creek complex c) Payette River complex.

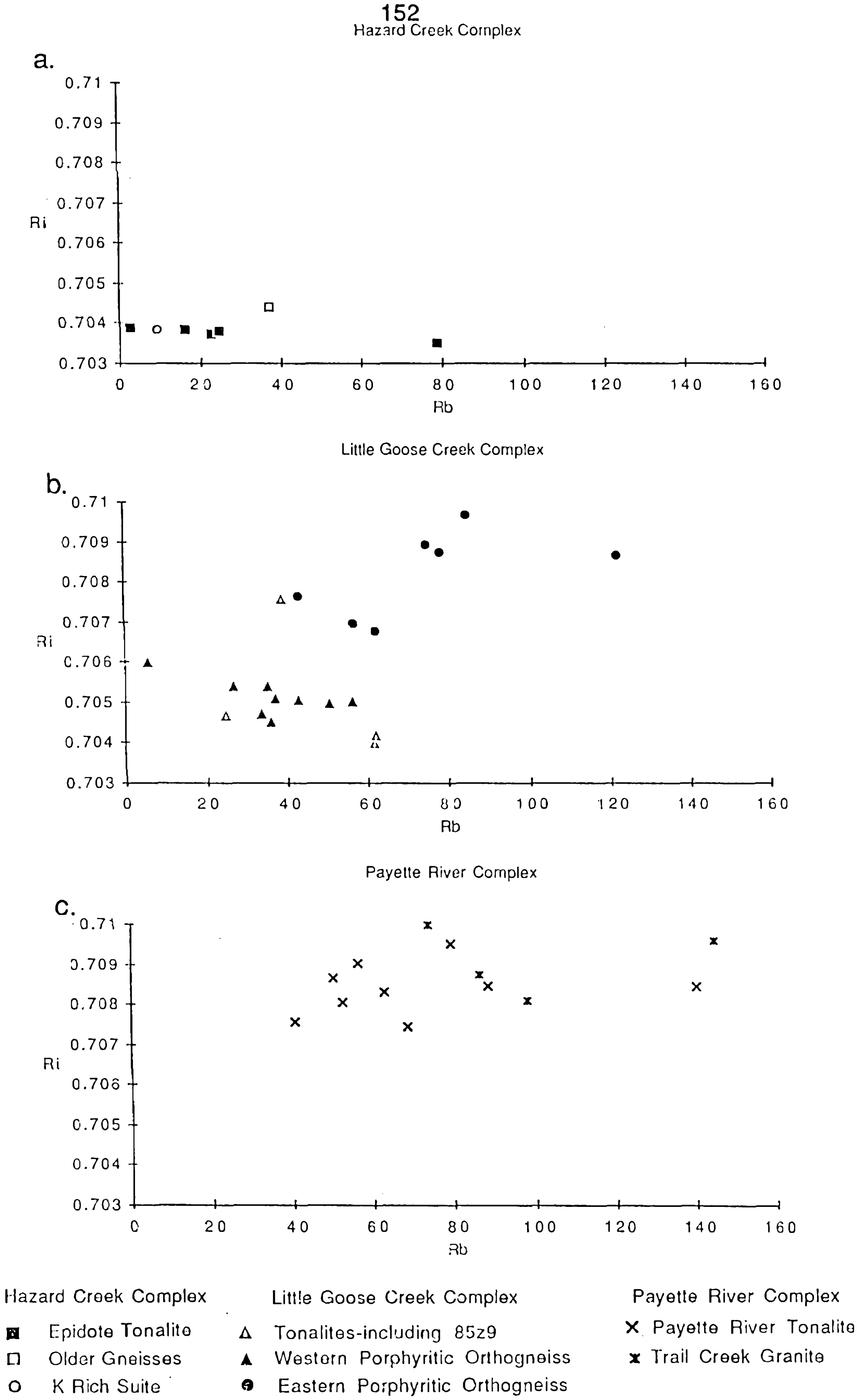


Figure 4.5 R_i versus Rb for plutonic and metaplutonic samples: a) Hazard Creek complex b) Little Goose Creek complex c) Payette River complex.

porphyritic orthogneiss near them. R_i correlates well with Sr concentration (Figure 4.4b).

$\delta^{18}O$ within the Little Goose Creek complex also has a large range (8.0 to 10.8), and correlates well with R_i (Figure 4.3b). Notable exceptions are samples 86z6 and 86z20, which have $\delta^{18}O$ of 10.6 and 10.2, typical of the east side of the complex, but R_i of only .7070 and .7077, lower than values observed to the north and south. The rise in $\delta^{18}O$ is gradational and lacks a large step. Little Goose Creek complex tonalites again have values slightly lower than the porphyritic orthogneiss near them.

R_i and $\delta^{18}O$ in the western porphyritic orthogneiss are similar to the western portion of the eastern Peninsular Ranges Batholith (Taylor and Silver, 1978) and to some samples from the southern Sierra Nevada Batholith (Masi et al., 1981). Several samples have $\delta^{18}O$ between 8.5 and 9.0, which fall into the thinly populated gap observed in the Peninsular Ranges Batholith data. Much of the eastern Little Goose Creek complex has higher R_i for a given $\delta^{18}O$ than observed in the Peninsular Ranges Batholith (Figure 4.3b). No samples of similarly high $\delta^{18}O$ are known from the Sierra. These samples are most similar to samples from the Great Basin (Solomon and Taylor, 1981, Farmer and DePaolo, 1983). R_i and $\delta^{18}O$ in Little Goose Creek complex tonalites are similar to values measured in the eastern portion of the western Peninsular Ranges Batholith (Taylor and Silver, 1978; Silver et al., 1975; Silver et al., 1979) and parts of the western Sierra Nevada Batholith (Masi et al., 1981).

The generally strong correlation between R_i, Sr concentration and $\delta^{18}O$ suggest that mixing of a high R_i, high $\delta^{18}O$, low Sr, high Rb material with a low R_i, low $\delta^{18}O$, high Sr, low Rb material was important in the formation of the porphyritic orthogneiss. The lack of complete correlation between Sr concentration, R_i and $\delta^{18}O$, however, requires a more complicated process than simple mixing of homogeneous endmembers.

4.3.1.c Payette River Complex

Samples from the Payette River Tonalite have moderately high R_i that is not correlated with longitude. Many values are lower than those in the eastern Little Goose Creek complex, creating a local reversal in R_i (Figure 4.1). Sr concentrations show no correlation with R_i (Figure 4.4c). $\delta^{18}O$ is generally lower than in the eastern Little Goose Creek complex (7.1 to 9.5) and is not correlated with R_i. R_i values are higher for a given $\delta^{18}O$ than those seen in the Little Goose Creek complex, forming a discrete field on a plot of R_i vs $\delta^{18}O$ (Figure 4.3c).

R_i and $\delta^{18}O$ in the Payette River Tonalite are similar to those reported for the Cretaceous intrusives in the Mojave, for quartz diorites from the Coast Ranges of California and for some samples from the southern Sierra Nevada Batholith (Masi et al., 1981; Figure 4.6). They define a trend toward lower $\delta^{18}O$ at moderately high R_i similar to that formed by samples from the San Jacinto Block and upper plate of the eastern Peninsular Ranges Batholith (Hill et al., 1986).

Samples of Payette River Granite and Trail Creek Granite have R_i (.7081 to .7100) and $\delta^{18}O$ (10.4 to 9.6) similar to that in the eastern Little Goose Creek complex. Granites from the Idaho Batholith commonly have R_i above .710 (Fleck and Criss, 1985). Although unusual, values as high as .710 are known from the Sierra Nevada Batholith and Peninsular Ranges Batholith; samples with ratios as high as .714 are unknown. Similarly high values are more common in granites that intrude the miogeocline in Nevada and Arizona (Farmer and DePaolo, 1983: Kistler et. al, 1981).

4.3.1.d Discussion of Chemical Mobility During Deformation

Deformation such as that observed in the Little Goose Creek and Hazard Creek complexes may have a strong effect on the isotopic systematics and distribution of elements in rocks (Wickham and Taylor, 1985; Lamphere et. al, 1964; Armstrong and

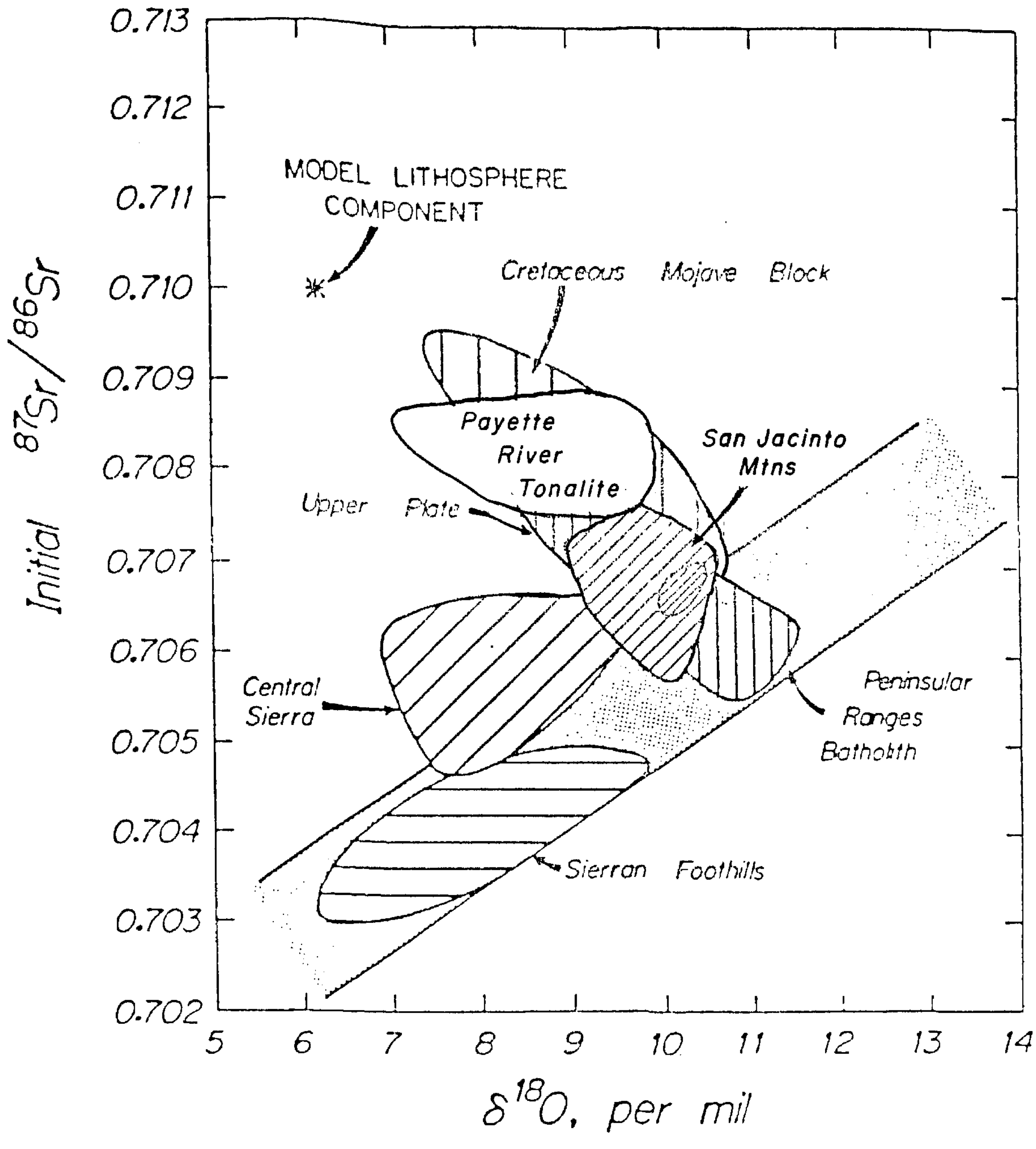


Figure 4.6 Plot of R_i versus $\delta^{18}O$ showing Peninsular Ranges Batholith trend and divergence of samples from the San Jacinto Block and upper plate of the eastern Peninsular Ranges Batholith, the Cretaceous Mojave Block, the Central Sierra and the Payette River Tonalite from this trend. Model lithosphere component proposed by Hill is also shown. Diagram is modified from Hill et al., 1986. Data are from Taylor and Silver, 1978, Hill et al., 1986, Masi et al., 1981 and this study.

Hills, 1967). In order to place constraints on the sources of the igneous rocks it is important to consider if the measured values are the original igneous values.

The preservation of strong zonation in $\delta^{18}O$ within the porphyritic orthogneiss suggests that deformation was not accompanied by fluid fluxing and large scale chemical mobility. Wickham and Taylor (1985) showed that during metamorphism, large scale fluid fluxing will homogenize $\delta^{18}O$. Samples 84z2a, 84z2b, 84z3 and 84z4 were collected in an area less than 100 m wide. Even on this scale, isotopes were not homogenized. The correlation between $\delta^{18}O$, R_i and chemical composition in the Little Goose Creek complex further suggests that the zonation is a primary feature. Small scale mobility of Rb may have occurred during mylonitization; however, the variation in R_i is sufficiently large that it is not obliterated by this effect. The measured $\delta^{18}O$, R_i and Sr concentrations are considered to be close to their original values.

4.3.2 Major and Trace Element Chemistry

4.3.2.a General Characteristics

Major and trace element analyses and calculated norms are listed in Table 4.2. Additional data for the Hazard Creek complex from Zen and Hammarstromm (in prep) and Barker et al. (1979) are listed in Table 4.3. Just as each complex has a distinct set of isotopic systematics, chemical characteristics are also distinct. There are several features, however, which characterize the data set as a whole. Major element data generally form smooth trends on Harker diagrams (Figure 4.7) consistent with formation by magmatic processes. On an AFM diagram (Irvine and Baragar, 1971), data plot in the calc-alkaline field with no evidence of extreme Fe enrichment (Figure 4.8). Most samples also fall within the calc-alkaline field on a FeO*/MgO vs SiO_2 diagram as defined by Miyashiro (1974; Figure 4.9). TiO_2 in all three complexes is low and correlates inversely with SiO_2 as is typical for calc-alkaline suites formed at

Table 4.2a: XRF chemical analyses and CIPW norms

HAZARD CREEK COMPLEX

	Epidote Tonalites					Older Gneisses		High K suite
Sample	83z3	83z9	85z11	83z1	85z17	85z15	83z8	83z12
SiO2	57.0	60.7	61.8	68.9	71.7	55.4	59.3	74.4
Al2O3	21.4	18.4	18.2	16.8	16.6	19.4	17.1	14.1
FeTo3	4.37	5.04	5.08	2.45	0.48	7.54	6.74	0.85
Mgo	1.46	1.85	2.13	0.83	0.17	2.82	2.56	0.18
CaO	5.87	5.98	5.95	3.93	2.99	8.30	6.12	1.29
Na2O	7.00	4.82	4.19	4.68	5.85	4.52	3.74	3.71
K2O	0.62	0.96	1.10	1.13	0.80	0.97	2.23	4.69
TiO2	0.80	0.66	0.52	0.28	0.04	0.78	0.68	0.06
P2O5	0.40	0.30	0.20	0.14	<0.05	0.37	0.28	<0.05
MnO	0.05	0.10	0.08	0.03	0.01	0.18	0.15	0.01
qtz	0.00	13.67	16.84	27.13	27.91	4.41	12.38	31.56
or	3.71	5.75	6.56	6.74	4.80	5.72	13.34	27.93
ab	59.8	41.2	35.7	39.9	50.1	38.1	32.0	31.6
an	25.4	26.0	27.8	18.8	14.8	29.7	23.5	6.2
dio	1.38	1.83	0.65	0.00	0.00	7.54	4.52	0.00
hy	4.00	6.42	7.88	3.40	0.72	7.63	8.15	0.97
il	1.54	1.27	1.00	0.54	0.08	1.48	1.31	0.11
mt	2.74	3.17	3.18	1.54	0.30	4.67	4.24	0.53
ap	0.88	0.66	0.44	0.31	0.09	0.81	0.62	0.09
corr	0.00	0.00	0.00	0.72	0.67	0.00	0.00	0.56
ol	1.22	0.00	0.00	0.00	0.00	0.00	0.00	0.00
ne	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Rb	3	16	25	22	79	(27)*	37	57
Sr	1404	963	718	768	781	(988)	630	315
Nb	<10	<10	<10	<10	<10	13	13	14
Zr	178	160	116	93	69	117	128	66
Y	<2	10	9	7	<2	22	22	7
Cu	17	<2	<2	<2	<2	3	10	<2
Ni	<5	9	12	<5	6	7	6	<5
Zn	96	87	93	61	30	104	90	46

*Rb and Sr concentrations by isotope dilution except for values in parentheses

Table 4.2b: XRF chemical analyses and CIPW norms

LITTLE GOOSE CREEK COMPLEX

	Tonalites					Western Porphyritic Orthogneiss							Eastern Porphyritic Orthogneiss		
Sample	84z1	85z18	84z4	84z3	85z9	83z14	84z2a	85z6	85z8	84z8b	84z15	84z2b	85z1	85z2	85z12
SiO2	54.1	54.5	57.9	58.1	61.0	64.4	66.4	66.9	67.1	68.8	69.0	69.8	67.7	70.6	71.7
Al2O3	18.0	20.5	17.1	17.8	18.3	18.0	17.5	17.1	17.9	16.6	16.4	15.9	15.8	15.0	15.5
FeTo3	8.65	6.12	7.20	6.80	4.88	3.32	2.85	2.48	2.18	2.13	2.19	1.89	2.54	1.90	1.33
MgO	3.20	2.03	2.54	2.21	2.35	1.01	0.65	0.71	0.65	0.55	0.54	0.46	0.71	0.71	0.4
CaO	7.30	6.67	6.39	6.34	5.92	4.54	3.93	3.67	3.85	3.16	3.16	3.24	3.33	2.11	2.41
Na2O	4.01	4.97	3.81	4.09	4.18	4.91	4.40	4.83	5.25	4.10	4.71	3.80	3.91	3.09	4.09
K2O	2.21	3.07	2.72	2.57	1.56	2.08	2.94	2.46	1.68	2.76	2.29	3.66	3.13	4.67	3.28
TiO2	0.93	0.70	0.81	0.75	0.78	0.50	0.37	0.36	0.35	0.28	0.31	0.24	0.37	0.40	0.21
P2O5	0.41	0.51	0.36	0.35	0.25	0.18	0.12	0.13	0.12	0.10	0.09	0.07	0.13	0.06	0.05
MnO	0.19	0.13	0.16	0.17	0.07	0.06	0.07	0.05	0.02	0.01	0.03	0.03	0.01	0.01	0.01
qtz	2.79	0.00	8.82	8.32	14.57	16.71	19.87	20.50	20.60	26.23	25.01	26.10	25.09	28.78	29.23
or	13.20	18.31	16.25	15.33	9.29	12.43	17.52	14.74	10.03	16.58	13.72	21.85	18.96	28.03	19.60
ab	34.2	40.6	32.5	34.9	35.6	41.9	37.5	41.4	44.8	35.2	40.3	32.4	33.8	26.5	34.9
an	24.8	24.7	21.7	22.8	26.7	21.1	19.0	17.7	18.6	15.3	15.4	15.6	16.2	10.3	11.8
dio	7.63	4.61	6.68	5.63	1.13	0.48	0.00	0.00	0.00	0.00	0.00	0.17	0.00	0.00	0.00
hy	9.23	6.22	7.15	6.64	7.62	3.92	3.15	3.03	2.62	2.45	2.44	2.06	3.02	2.49	1.61
mt	5.43	3.83	4.52	4.26	3.05	2.08	1.78	1.56	1.37	1.34	1.38	1.19	1.62	1.20	0.84
il	1.78	1.34	1.55	1.44	1.49	0.96	0.71	0.69	0.67	0.54	0.60	0.46	0.72	0.77	0.40
ap	0.90	1.12	0.79	0.77	0.55	0.40	0.26	0.29	0.26	0.22	0.20	0.15	0.29	0.13	0.11
corr	0.00	0.00	0.00	0.00	0.00	0.00	0.17	0.08	0.69	1.34	0.61	0.00	0.19	1.15	0.94
ol	0.00	9.07	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
ne	0.00	0.94	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Rb	62	(88)*	62	25	39	34	51	43	27	35	37	56	62	122	79
Sr	797	(1510)*	713	864	662	1079	1003	935	1067	908	890	856	820	681	662
Nb	18	12	22	17	<10	15	20	16	<10	<10	<10	10	<10	15	13
Zr	182	125	173	195	144	226	254	195	192	178	189	145	200	331	159
Y	27	13	26	18	10	8	13	10	<2	<2	<2	7	8	14	8
Cu	3	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2	<2
Ni	7	<5	9	12	<5	6	6	<5	<5	<5	<5	<5	<5	<5	<5
Zn	108	90	91	101	109	97	85	80	73	68	78	46	60	42	51

*Rb and Sr concentrations by isotope dilution except for values in parentheses

Table 4.2c: XRF chemical analyses and CIPW norms

PAYETTE RIVER COMPLEX

	Payette River Tonalite					Trail Creek Granite	
Sample	85z14	83z11	83z2	83z6	85z4	83z16a	83z15
SiO2	58.3	59.6	60.3	62.3	68.6	72.5	73.8
Al2O3	17.8	17.5	17.9	18.1	15.3	14.5	14.2
FeTo3	6.72	6.20	5.66	4.80	3.12	1.69	1.23
MgO	3.36	3.18	2.83	2.22	1.38	0.32	0.37
CaO	6.54	6.76	6.40	6.06	3.66	2.10	2.02
Na2O	3.33	3.08	3.33	3.67	3.21	3.26	3.72
K2O	1.64	1.90	1.89	1.39	3.78	4.69	3.33
TiO2	0.85	0.91	0.76	0.60	0.42	0.17	0.14
P2O5	0.24	0.29	0.22	0.22	0.13	<0.05	<0.05
MnO	0.10	0.09	0.08	0.07	0.05	0.02	0.01
qtz	13.32	15.42	15.62	18.90	25.56	30.34	34.25
or	9.81	11.3	11.3	8.27	22.4	27.9	19.9
ab	28.5	26.2	28.3	31.2	27.2	27.7	31.8
an	29.1	28.4	28.4	28.9	16.2	10.3	9.9
dio	1.89	2.89	1.79	0.02	1.03	0.00	0.00
hy	11.07	9.58	9.10	8.06	4.53	1.74	1.58
il	1.63	1.74	1.45	1.15	0.80	0.33	0.27
mt	4.22	3.87	3.54	3.00	1.95	1.06	0.77
ap	0.53	0.64	0.48	0.48	0.28	0.09	0.09
corr	0.00	0.00	0.00	0.00	0.00	0.22	0.80
ol	0.00	0.00	0.00	0.00	0.00	0.00	0.00
ne	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Rb	41	62	69	(51)*	140	145	87
Sr	644	675	666	(753)*	400	369	514
Nb	10	12	12	<10	17	15	13
Zr	167	180	157	142	135	133	112
Y	16	14	15	12	14	14	8
Cu	<2	<2	<2	<2	<2	<2	<2
Ni	6	10	15	<5	<5	<5	<5
Zn	91	96	95	85	61	47	38

* Rb and Sr concentrations by isotope dilution
except for values in parentheses

Table 4.3: Additional chemical analyses of tonalite and trondhjemite from the Hazard Creek Complex and Riggins area

	Little Salmon River (Zen and Hammarstrom, in prep)						Riggins area (Barker, 1979)				
							R1	R2	R3	R4	R5
SiO2	55.6	65.2	67.0	71.7	72.8	73.5	66.1	68.1	69.6	71.2	75.7
Al2O3	19.8	18.6	17.5	16.3	15.7	15.5	18.1	17.7	17.8	17.0	15.5
Fe2O3	2.20	0.52	0.90	0.31	0.29	0.37	1.30	1.10	1.30	0.60	0.48
FeO	3.60	2.80	1.70	0.72	0.28	0.20	1.40	1.50	0.95	0.64	0.13
MgO	2.30	1.10	0.98	0.34	0.14	0.16	0.94	0.95	0.52	0.36	0.27
CaO	7.30	5.50	4.80	2.90	2.40	2.60	4.90	4.30	5.00	3.00	1.20
Na2O	4.80	5.00	4.70	5.60	5.20	5.60	4.70	4.60	4.70	5.50	5.10
K2O	1.00	0.09	0.82	0.66	1.60	0.60	0.96	1.20	0.50	0.95	1.50
TiO2	0.87	0.41	0.33	0.12	0.05	0.06	0.33	0.34	0.23	0.18	0.02
P2O5	0.35	0.17	0.18	0.06	<0.05	<0.05	0.14	0.09	0.01		
MnO	0.09	0.08	0.05	0.03	0.02	0.02	0.06	0.10	0.04	0.02	0.09
Rock Type				dike	dike						peg

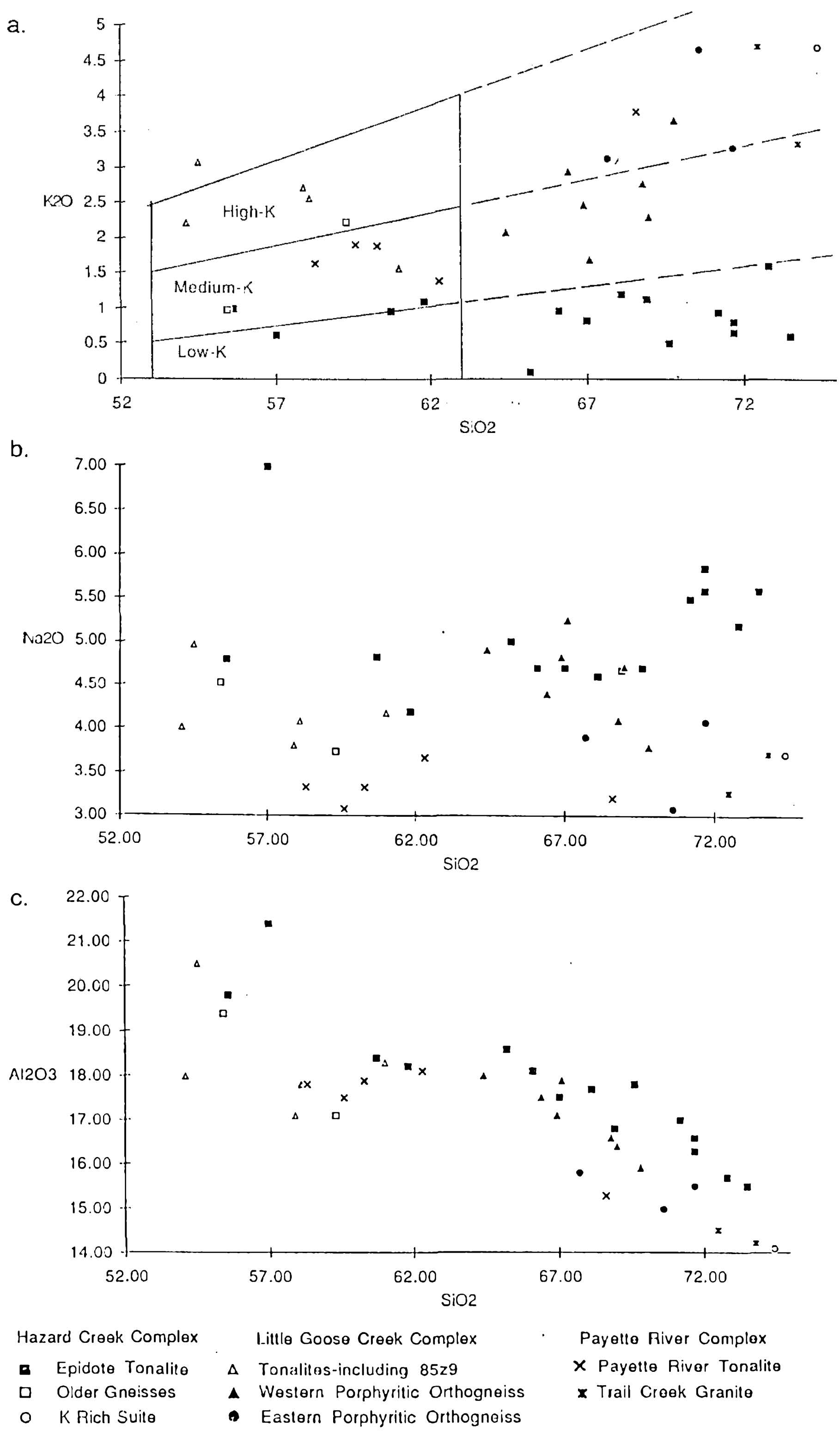


Figure 4.7 Harker variation diagrams for major elements. Data for Hazard Creek complex include values from Zen and Hammarstrom (in prep) and Barker et al. (1979). K_2O classification from Gill (1981).

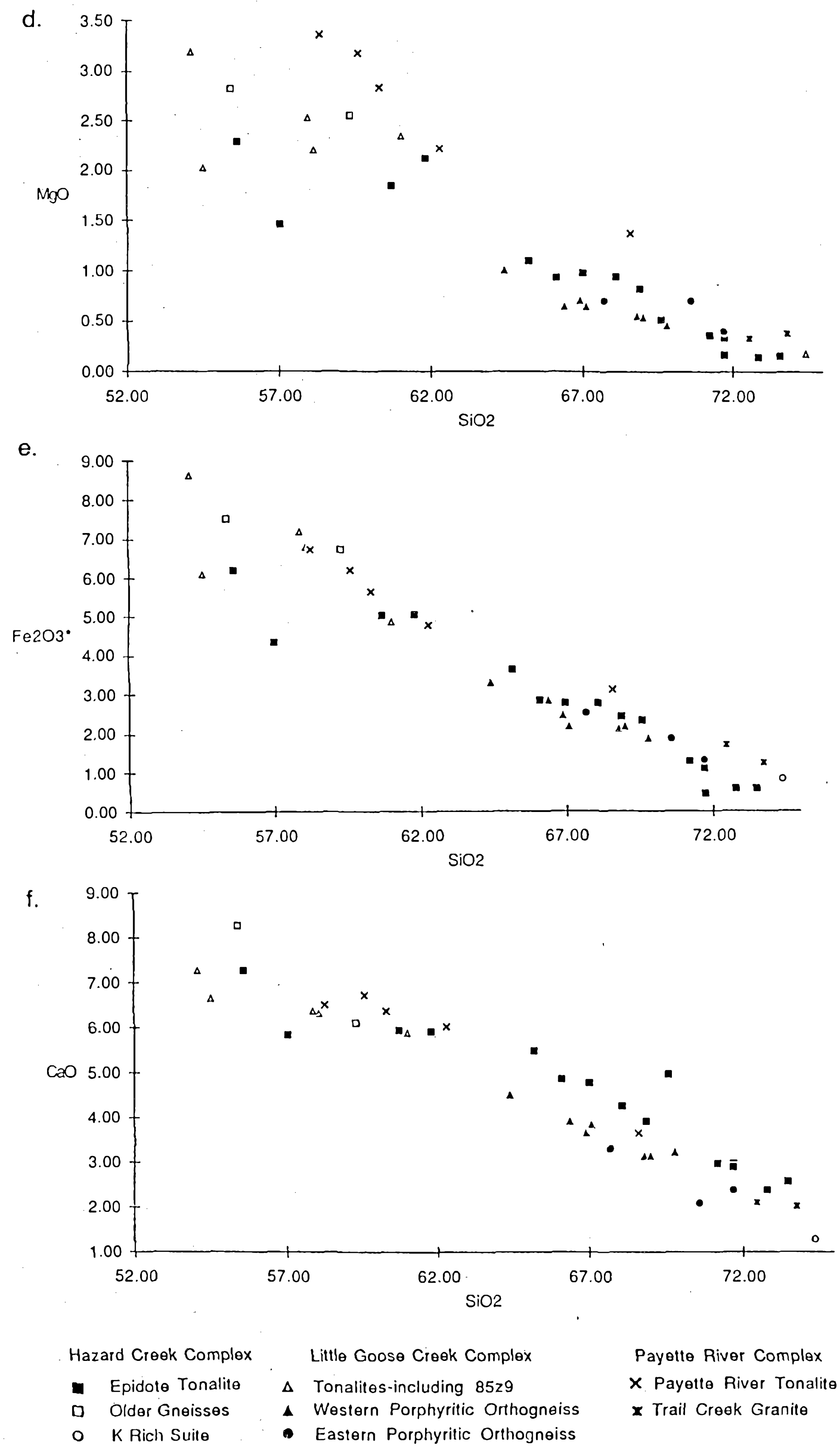
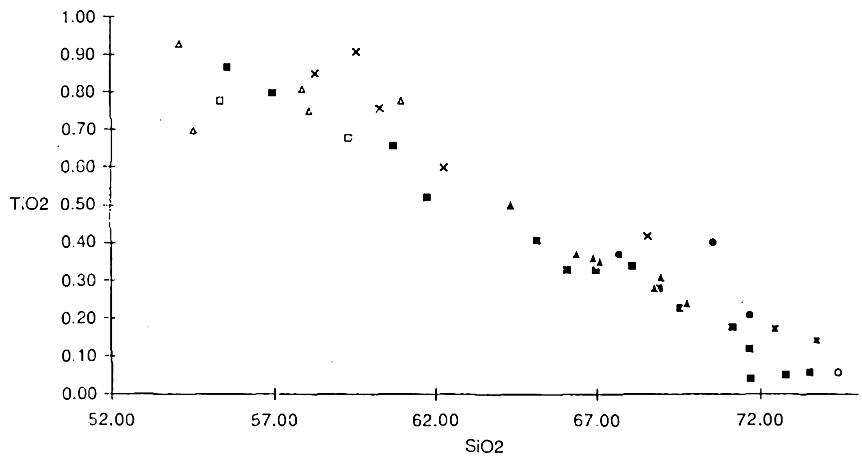


Figure 4.7 continued



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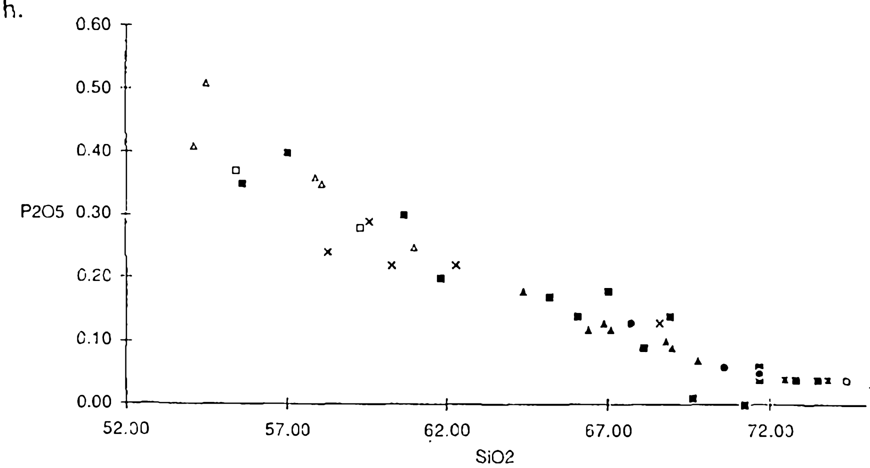

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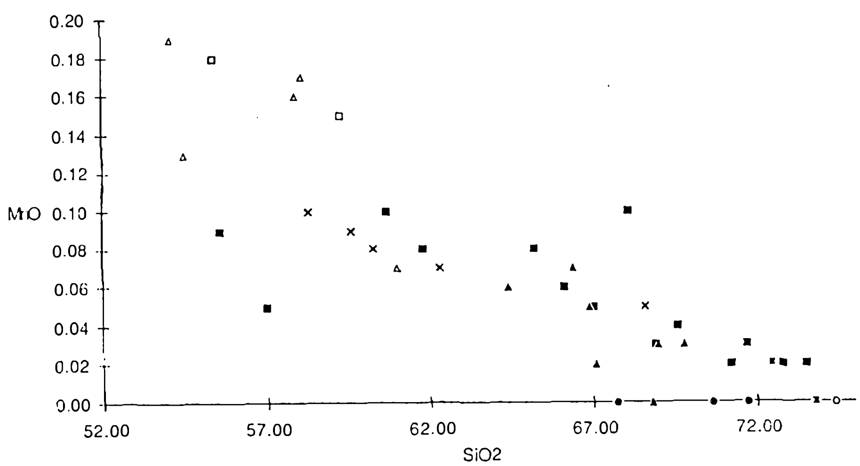
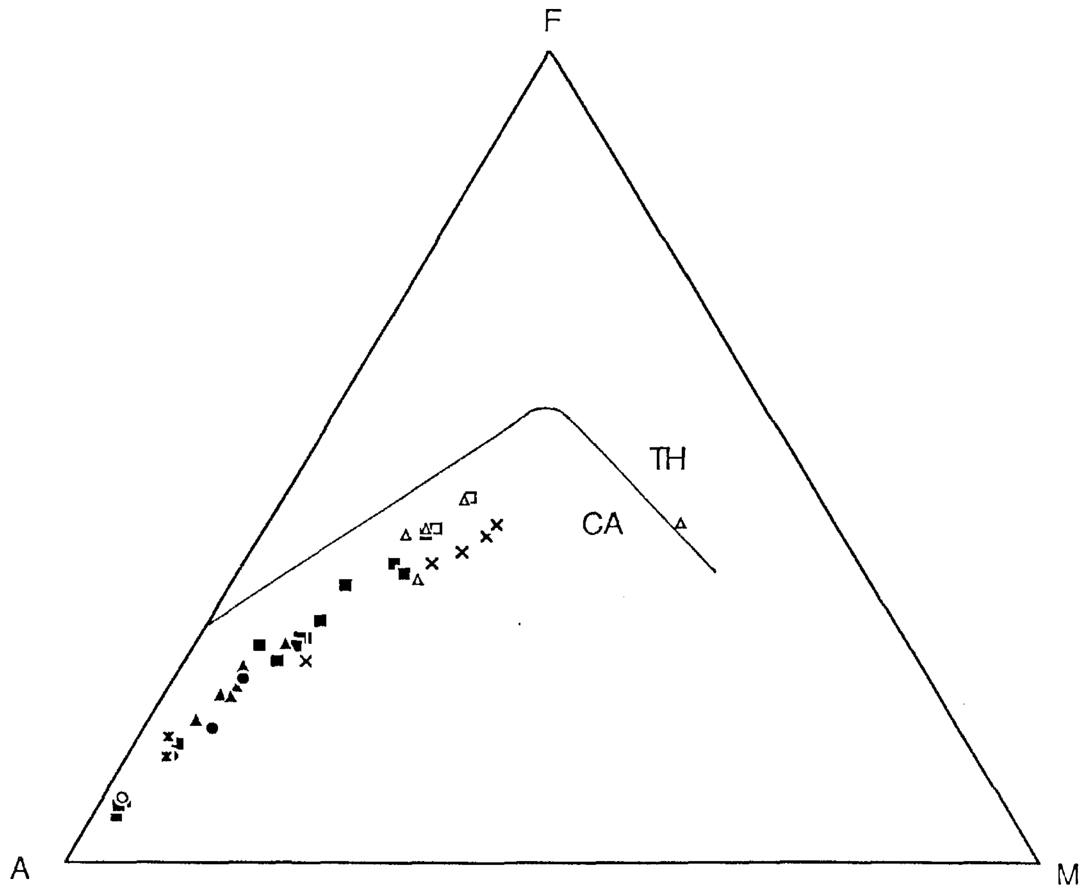


Hazard Creek Complex	Little Goose Creek Complex	Payette River Complex
■ Epidote Tonalite	△ Tonalites-including 85z9	× Payette River Tonalite
□ Older Gneisses	▲ Western Porphyritic Orthogneiss	✖ Trail Creek Granite
○ K Rich Suite	● Eastern Porphyritic Orthogneiss	

Figure 4.7 continued



Hazard Creek Complex	Little Goose Creek Complex	Payette River Complex
■ Epidote Tonalite	△ Tonalites-including 85z9	X Payette River Tonalite
□ Older Gneisses	▲ Western Porphyritic Orthogneiss	✖ Trail Creek Granite
o K Rich Suite	● Eastern Porphyritic Orthogneiss	

Figure 4.8 AFM diagram; solid line separates calcalkaline (CA) and tholeiitic (TH) fields of Irvine and Baragar (1971).

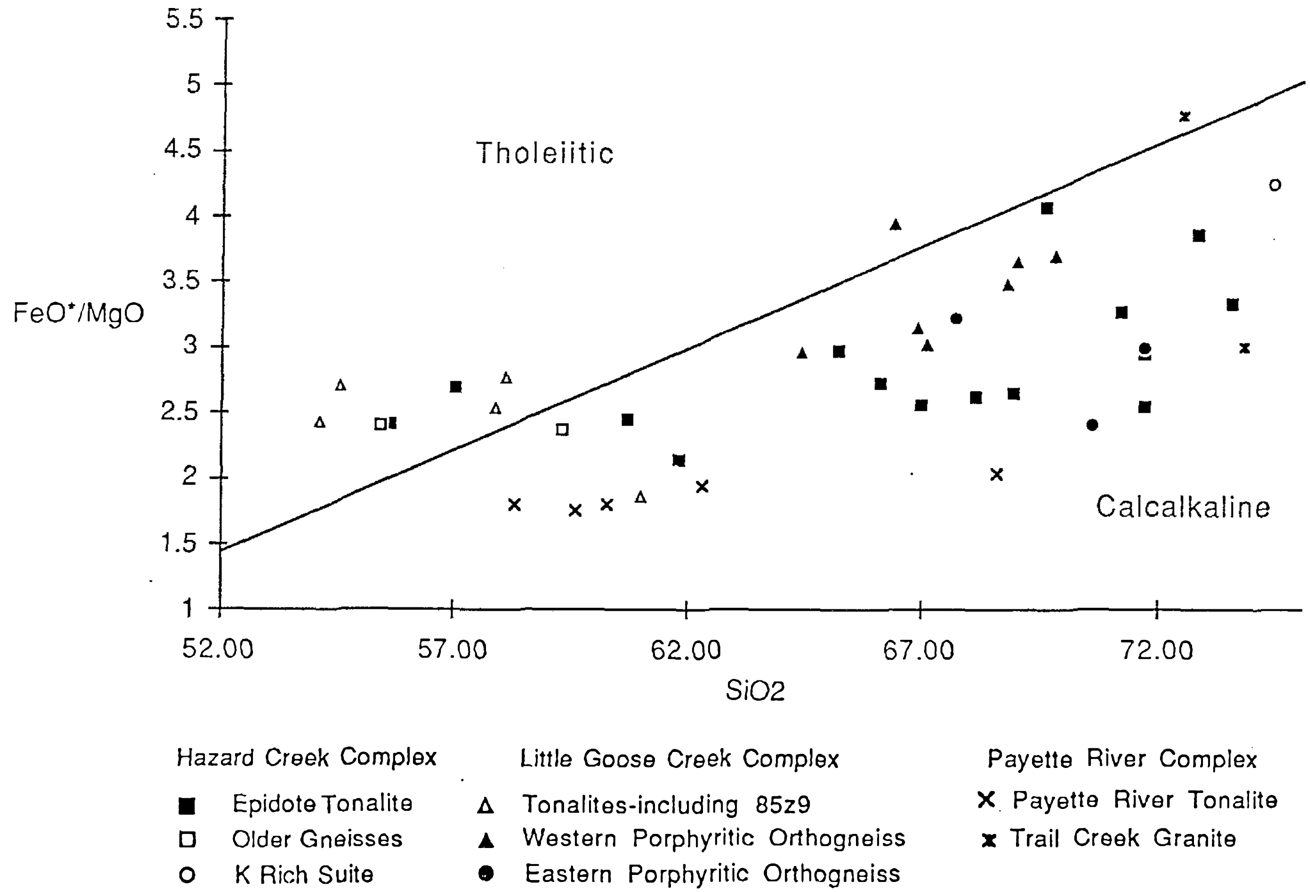
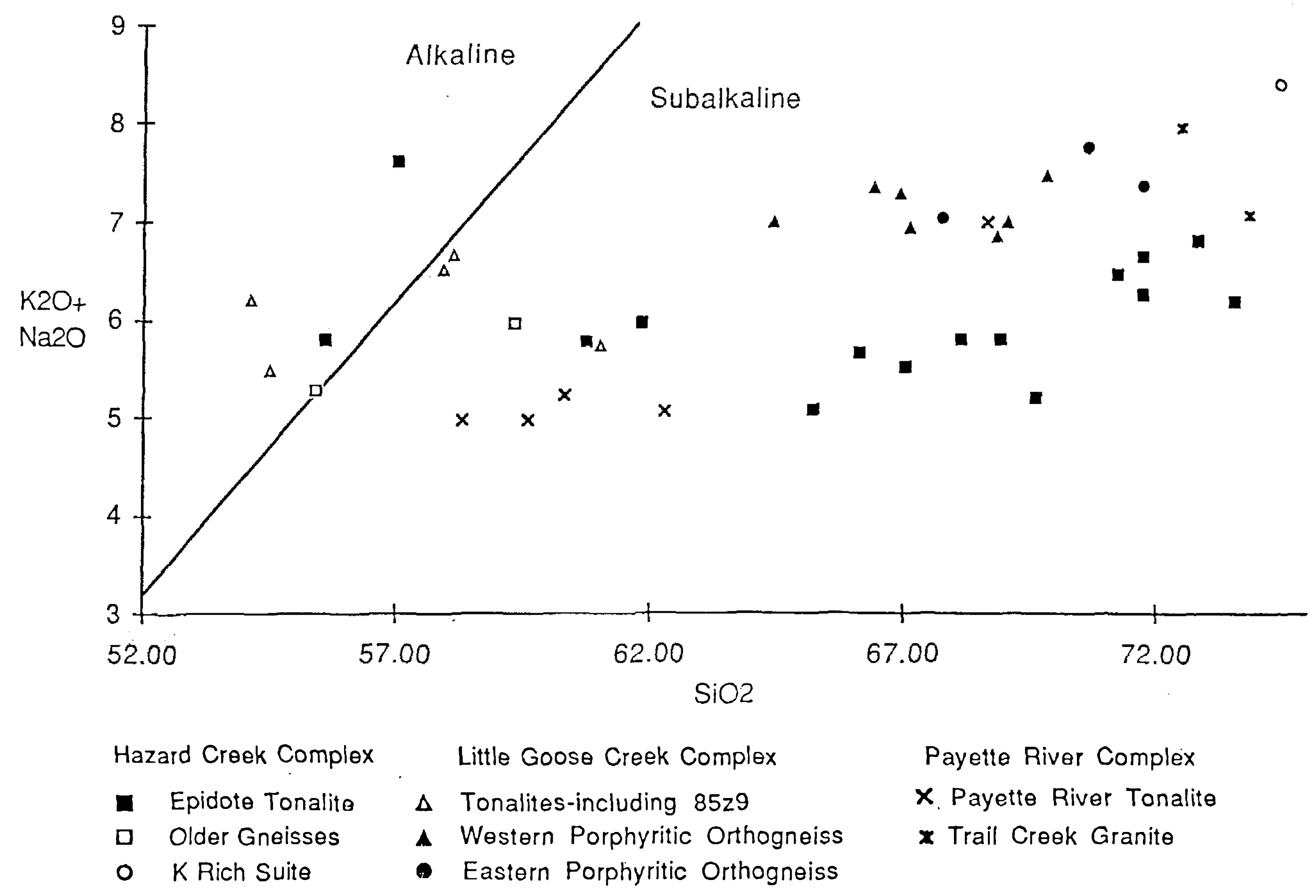


Figure 4.9 Total Fe as FeO/Mgo versus SiO_2; solid line separates calcalkaline from tholeiitic suites of Miyashiro (1974).



<u>Figure 4.10</u> Total alkalies versus SiO_2; solid line separates alkaline from subalkaline fields of Irvine and Baragar (1971).

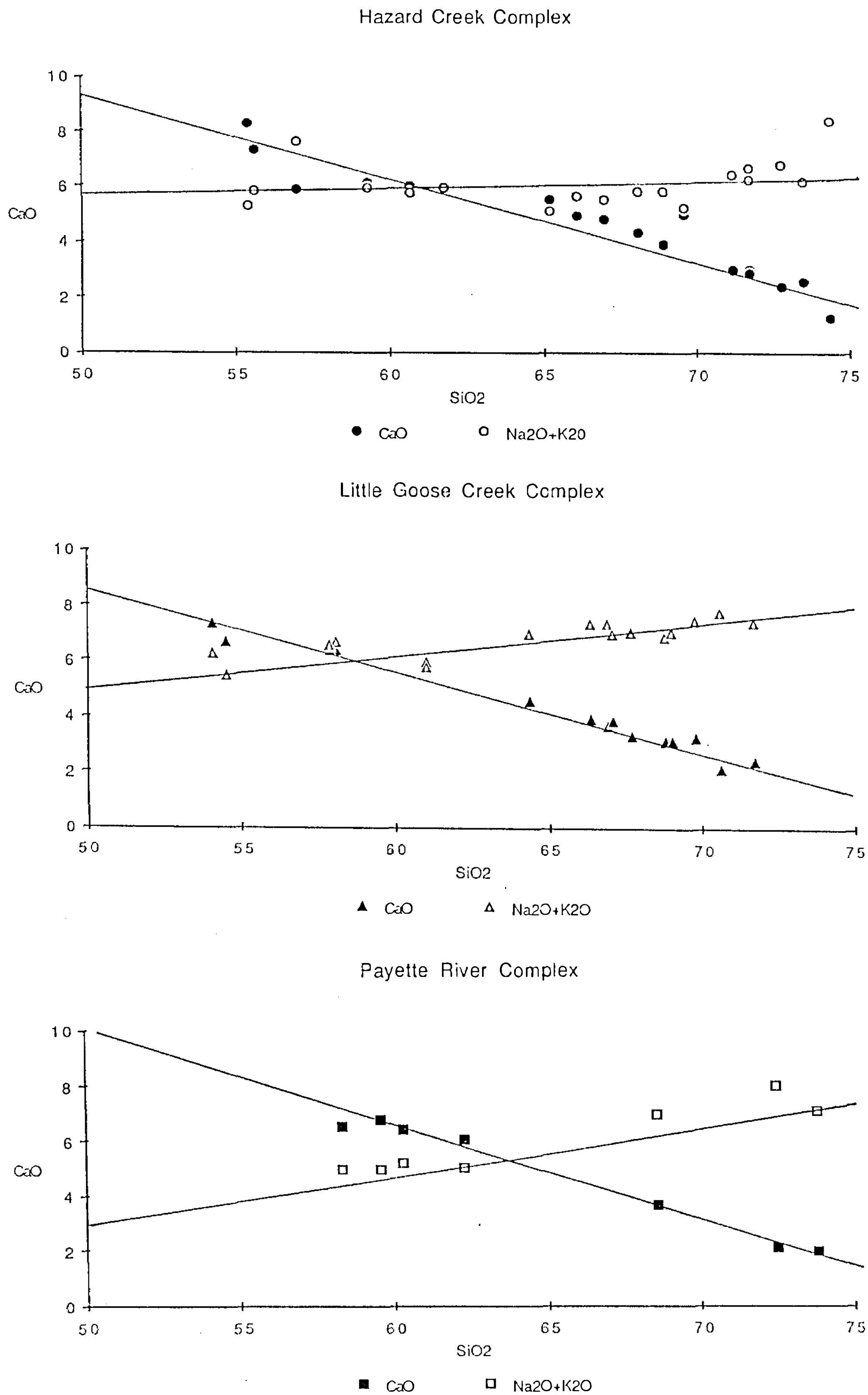


Figure 4.11 Peacock diagram for samples from a) Hazard Creek complex, b) Little Goose Creek complex, c) Payette River complex.

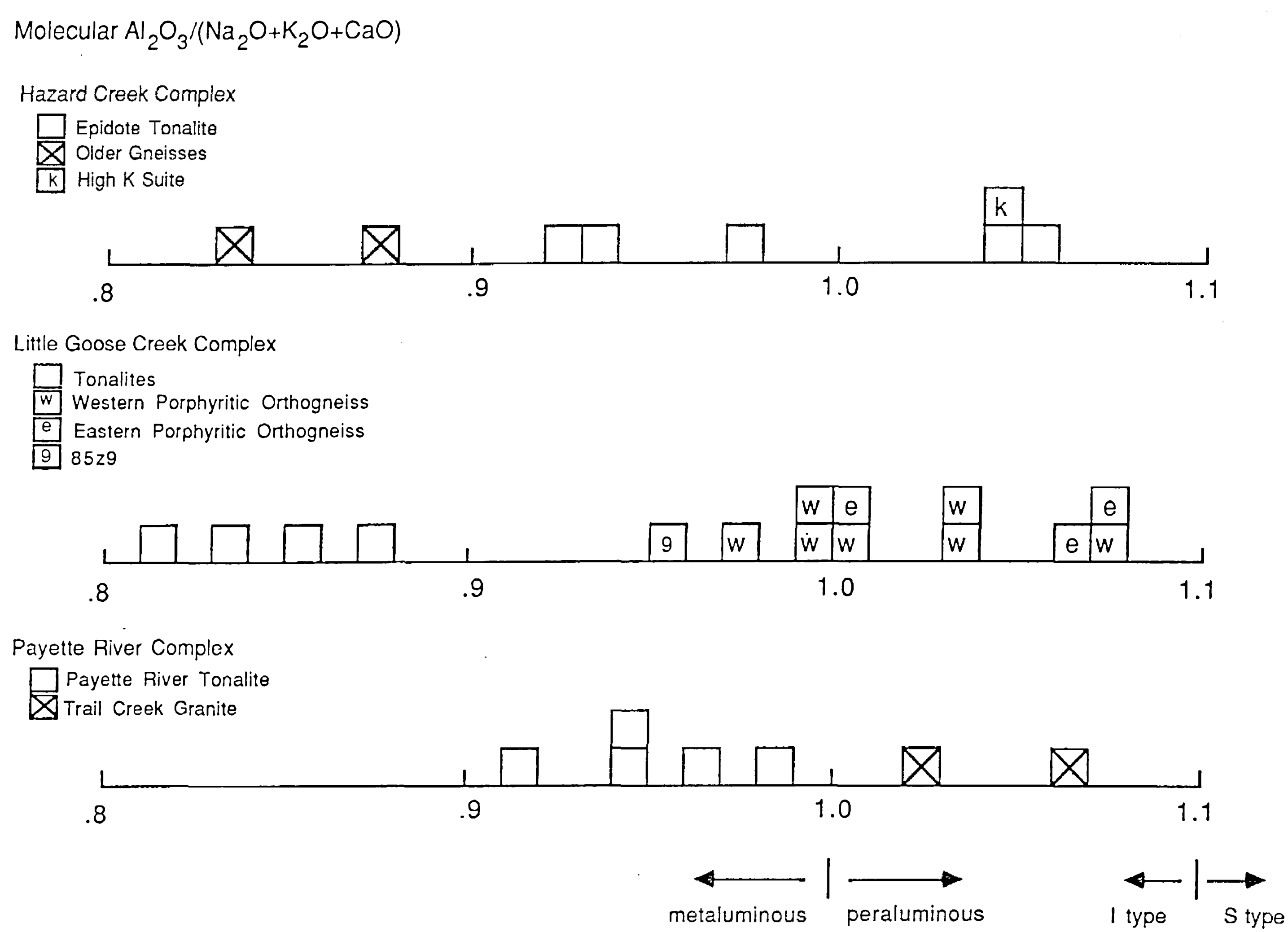
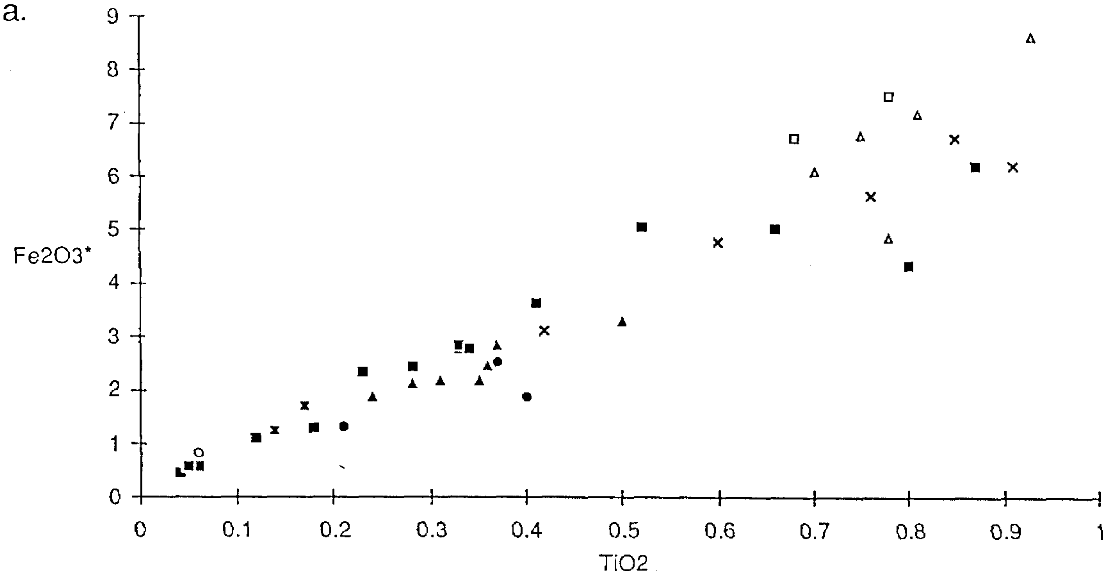
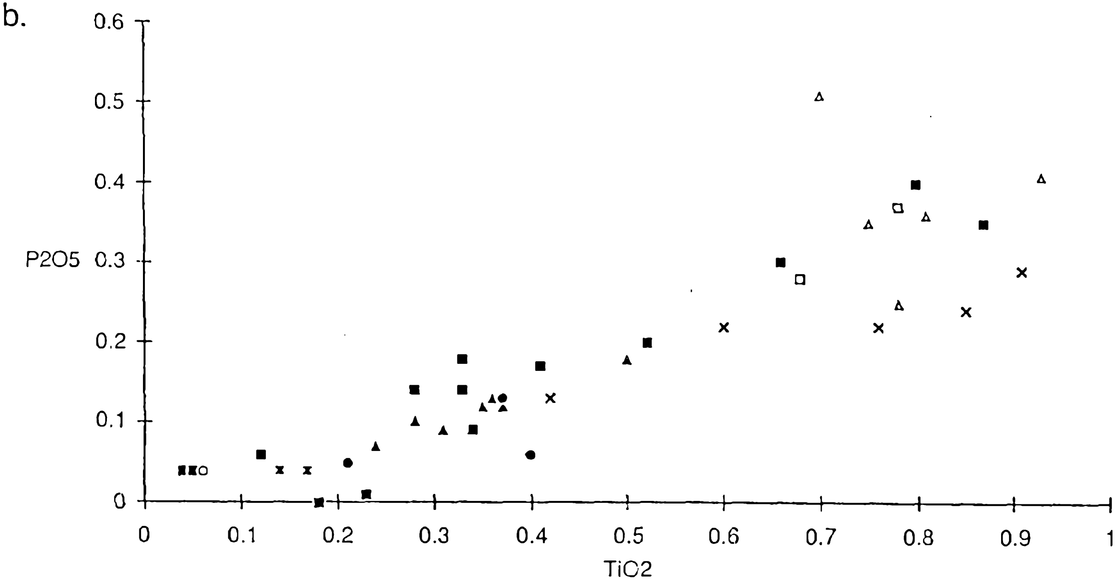


Figure 4.12 Histogram of molecular $Al_2O_3/(Na_2O+K_2O+CaO)$; I and S type suites as defined by Chappell and White (1974).

a.



b.



Hazard Creek Complex	Little Goose Creek Complex	Payette River Complex
■ Epidote Tonalite	△ Tonalites-including 85z9	× Payette River Tonalite
□ Older Gneisses	▲ Western Porphyritic Orthogneiss	⁎ Trail Creek Granite
o K Rich Suite	● Eastern Porphyritic Orthogneiss	

Figure 4.13 Total Fe as Fe_2O_3 (a) and P_2O_5 (b) versus TiO_2.

destructive plate margins (Gill, 1981). Total alkalies for most samples fall within the subalkaline field of Irvine and Baragar (1971; Figure 4.10). Most samples also have low Y concentrations, less than 15 ppm, which suggests emplacement through thick continental crust (Gill, 1981). This is consistent with the abundance of felsic rocks in the area. Trace element data are similar to volcanic arc suites as defined by Pearce et al. (1984).

The Peacock Index calculated for the data set as a whole is 62, which is calcic. Peacock Indices calculated for each complex are 61 for the Hazard Creek complex, 59 forthe Little Goose Creek complex, and 63 for the Payette River complex (Figure 4.11). These values are in the calc-alkaline to calcic range and are most similar to values found in batholiths emplaced into eugeoclinal sequences. They are much higher than values calculated for the central portion of the batholith to the east (Criss and Fleck, 1987).

All samples are chemically similar to I type suites as defined by Chappell and White (1974). The molecular ratio $Al_2O_3/(K_2O+Na_2O+CaO)$ is less than 1.1 (Figure 4.12); Na_2O is generally high; normative diopside or corundum less than one percent is calculated for most samples (Table 4.2).

These comparisions show that in general the data is similar to that from bodies formed at destructive margins along continental margins. The data set as a whole appears to be calc-alkaline and fairly calcic. These features suggest that the plutonic rocks formed in response to west-dipping subduction outboard of the Blue Mountain Arc.

4.3.2.b Characterization of Each Complex and Comparison with Other Igneous Suites

In this section, the chemistry of each complex is characterized. In order to understand which features of the chemistry are unusual and which are typical of igneous rocks in different settings, the chemistry of the samples in the study area was compared to that of a variety of other igneous rocks formed at destructive plate margins: orogenic andesites, the rest of the Idaho Batholith, and other Cordilleran batholiths.

Data for orogenic andesites are taken from Gill (1981), who summarized representative values and typical variation diagrams for high, medium and low K suites. These rocks have SiO_2 similar to tonalites from the study area. Data for the rest of the Idaho Batholith are from Lewis et al. (1987), Toth (1987) and Hyndman (1984). Lewis et al. studied the Atlanta lobe to the east and southeast of the study area, which is composed primarily of biotite granodiorite and two-mica granite. Tonalites occur along both the eastern and western margins of the batholith; they are more abundant on the west. Their data includes samples inferred to be correlative with the Trail Creek Granite and Payette River Tonalite. Hyndman and Toth studied the Bitterroot lobe, to the northeast of the study area, which consists primarily of granite and granodiorite but also contains less abundant tonalite. All samples from both lobes are from areas characterized by high R_i, and are believed to have intruded into thick continental crust. Most workers advocate source materials that are dominantly crustal (Fleck and Criss, 1985; Lewis et al., 1987; Hyndman, 1983; Schuster and Bickford, 1984). In the following discussion, data from Lewis et al. will be referred to as from the Atlanta lobe, data from Toth and Hyndman will be referred to as from the Bitterroot lobe. This is not meant to imply that rocks from this study are are not part of the Atlanta lobe.

Data for other Cordilleran style batholiths are from the Coastal Batholith of Peru (Atherton and Sanderson, 1985), the Peninsular Ranges Batholith of southern and Baja California (Larsen, 1948; Silver et. al, 1979), the Sierra Nevada Batholith of California (Bateman and Dodge, 1970) and the central gneiss complex of the Coast Plutonic Complex, British Columbia, Canada (Barker and Arth, 1984). These batholiths are believed to have incorporated less continental crust than the Idaho Batholith (Hill et al., 1986; Gromet and Silver, 1987).

Major Element Chemical Data

The largest variations between complexes are in K_2O (Figure 4.7a). The Hazard Creek complex has the lowest K_2O with values falling in the low-K field of Gill (1981). A notable exception is the K-rich suite that has much higher K_2O falling in the extension of Gill's high-K field. The Little Goose Creek complex has the highest K_2O for a given SiO_2. Eastern porphyritic orthogneiss has higher K_2O for a given SiO_2 than western porphyritic orthogneiss; Little Goose Creek complex tonalites and western porphyritic orthogneiss have similar K_2O concentrations. Payette River Tonalite and Coarse Grained Tonalite Orthogneiss have intermediate values falling in the medium K field. Trail Creek Granite has high K_2O similar to eastern porphyritic orthogneiss. K_2O is poorly correlated with SiO_2 (Figure 4.7a).

Little Goose Creek complex and Trail Creek Granite K_2O contents are comparable to data from the Atlanta and Bitterroot lobes and are as high in K_2O as any samples from the core of the batholith. Data from the batholith form a loose trend of decreasing K_2O with decreasing SiO_2. Payette River Tonalite falls along the extension of this trend while Little Goose Creek complex tonalites fall above the trend.

K_2O contents vary substantially in Cordilleran batholiths. Not only are variations among batholiths large, but variations across individual batholiths are equally well known (Bateman and Dodge, 1970; Silver et al., 1979). The intermediate K_2O content of Payette River Tonalite and western porphyritic orthogneiss is similar to that seen in the western Sierra Nevada Batholith (Shaver sequence and Western foothills rocks) for rocks of similar SiO_2. Trail Creek Granite and eastern porphyritic orthogneiss have K_2O contents similar to those in the eastern Sierra Nevada Batholith (John Muir and Palisade Crest sequences), the central gneiss complex of the Coast Plutonic Complex, and the Arequipa segment of the Coastal Batholith for rocks with similar SiO_2. The Central Gneiss Assemblage and the Arequipa segment both contain many rocks with higher K_2O for a given SiO_2 than those seen in the Little Goose Creek complex. The Arequipa segment and the eastern Sierra Nevada Batholith are underlain by thick crust and intruded into continentally derived material. The central gneiss complex, in contrast, is intruded into a eugeosynclinal section. Rocks with low K_2O similar to that in the Hazard Creek complex are reported only in parts of the western foothills in the Sierra Nevada and in the Santa Rosa complex of the Coastal Batholith.

Na_2O contents (Figure 4.7b) in the Hazard Creek complex epidote tonalites, western porphyritic orthogneiss, and Little Goose Creek complex tonalites, are unusually high for orogenic andesites or Cordilleran batholiths. The highest concentrations of Na_2O, in the Hazard Creek complex epidote tonalites, are similar only to those seen in trondhjemitic suites. Concentrations in the Payette River Tonalite, Trail Creek Granite, eastern porphyritic orthogneiss and the Hazard Creek complex K-rich suite are more typical of Cordilleran batholiths and similar to contents in the Atlanta and Bitterroot lobes. Na_2O shows no variation with SiO_2.

Al_2O_3 (Figure 4.7c) is also enriched in the Hazard Creek complex epidote tonalites, western porphyritic orthogneiss and Little Goose Creek complex tonalites. The highest values are again from Hazard Creek complex epidote tonalites. The Payette River complex and eastern Little Goose Creek complex are again more typical of Cordilleran batholiths. Most arc andesites have Al_2O_3 between 16 and 18% (Gill, 1981). Hazard Creek complex epidote tonalites fall above the range; Little Goose Creek complex tonalites fall within or above this range; Payette River Tonalite falls at the high end of this range. Na_2O in the western Little Goose Creek complex and the central gneiss complex of the Coast Plutonic complex are similar. Eastern porphyritic orthogneiss and Trail Creek Granite are comparable to samples from the Bitterroot and Atlanta lobes with similar SiO_2.

Payette River Tonalite, Trail Creek Granite and eastern porphyritic orthogneiss have MgO similar to Atlanta and Bitterroot data and other Cordilleran batholiths. MgO is depleted in the other groups relative to this (Figure 4.7d) raising FeO*/MgO and causing many of these samples to fall in the tholeiitic field on a plot of FeO*/MgO against SiO_2 (Figure 4.9; FeO* = total Fe as FeO). The low MgO content is also evident on an AFM diagram (Figure 4.8). MgO in the Payette River Tonalite is lower than average for medium-K andesites.

The variation in CaO between suites is small compared to the variation with SiO_2 (Figure 4.7f). Payette River Tonalite appears to contain slightly more CaO than other tonalitic rocks. The Hazard Creek complex epidote tonalites are less rapidly depleted in CaO with increasing SiO_2; thus, at high SiO_2, Hazard Creek complex epidote tonalites contain more CaO than samples from other suites. CaO in all suites is comparable to that in batholithic rocks. Values for Hazard Creek complex and Little Goose Creek complex tonalites with SiO_2 less than 56% are lower than average for orogenic andesites.

The Payette River complex and eastern porphyritic orthogneiss have slightly higher TiO_2 for a given SiO_2 than the Hazard Creek complex and western Little Goose Creek complex. Values are lower than average for orogenic andesites but are comparable with other batholithic data. TiO_2 correlates well with SiO_2, FeO and P_2O_5 as in orogenic andesites (Figures 4.7g, 4.13). P_2O_5 does not discriminate between suites (Figure 4.7h). Most orogenic andesites have P_2O_5 less than 0.3%, similar to the Payette River Tonalite but less than many of the Hazard Creek complex and Little Goose Creek complex samples. P_2O_5 in granites from the Atlanta lobe is higher for a given SiO_2 than that from the study area, while data from the Bitterroot lobe and other batholiths is similar.

<u>Trace Element Data</u>

Rb concentrations (Figure 4.14a) are lowest in the Hazard Creek complex and highest in the eastern porphyritic orthogneiss and Trail Creek Granite. Other suites are intermediate and variable. The range of data is typical of Cordilleran batholiths, including the Idaho Batholith; however, values as low as those in the Hazard Creek

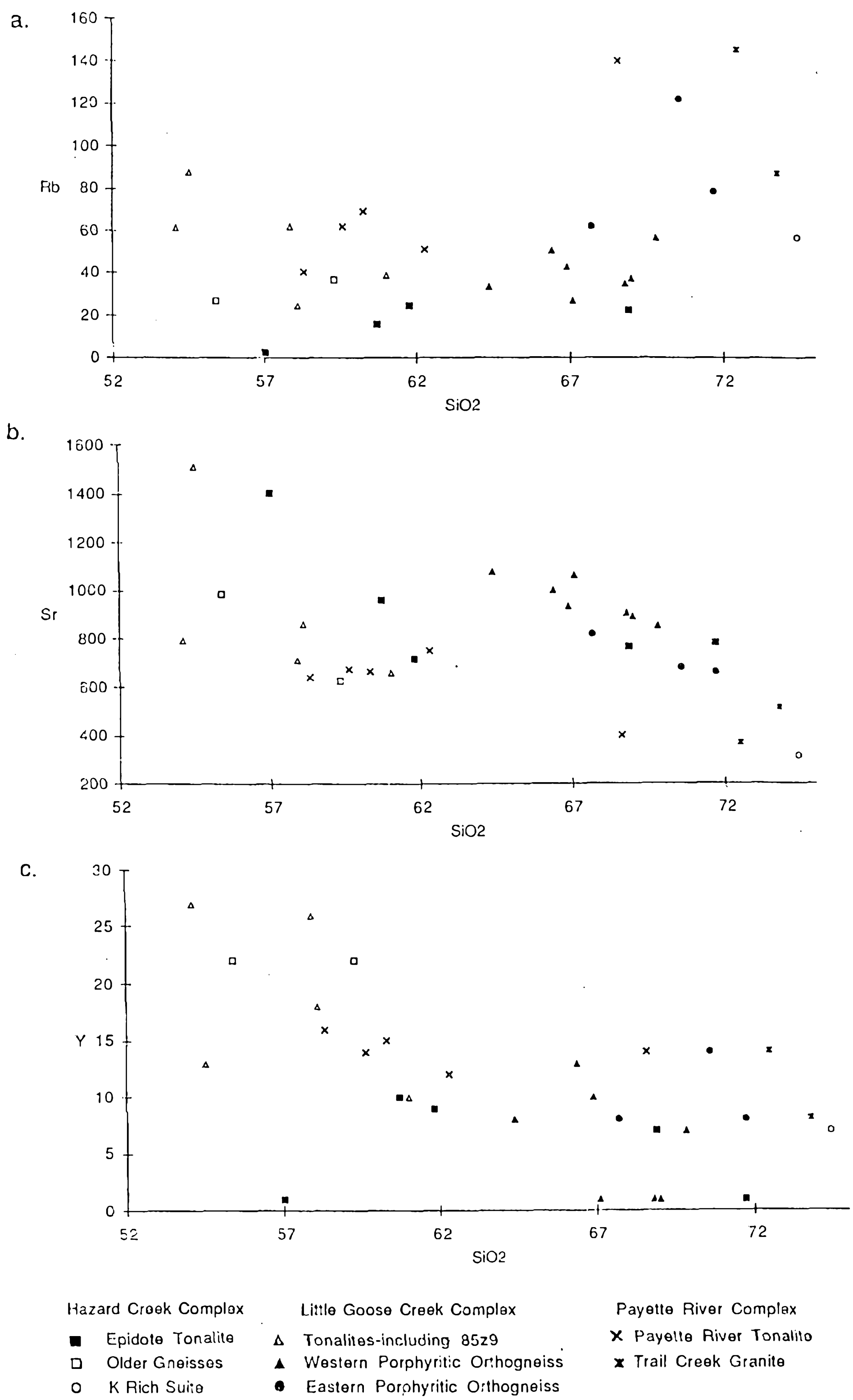
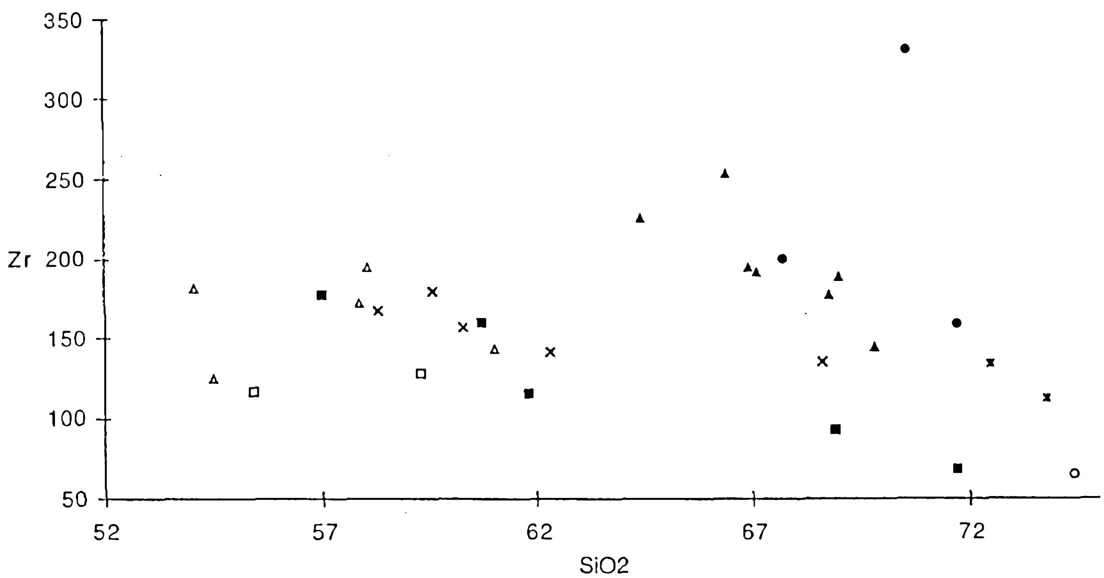
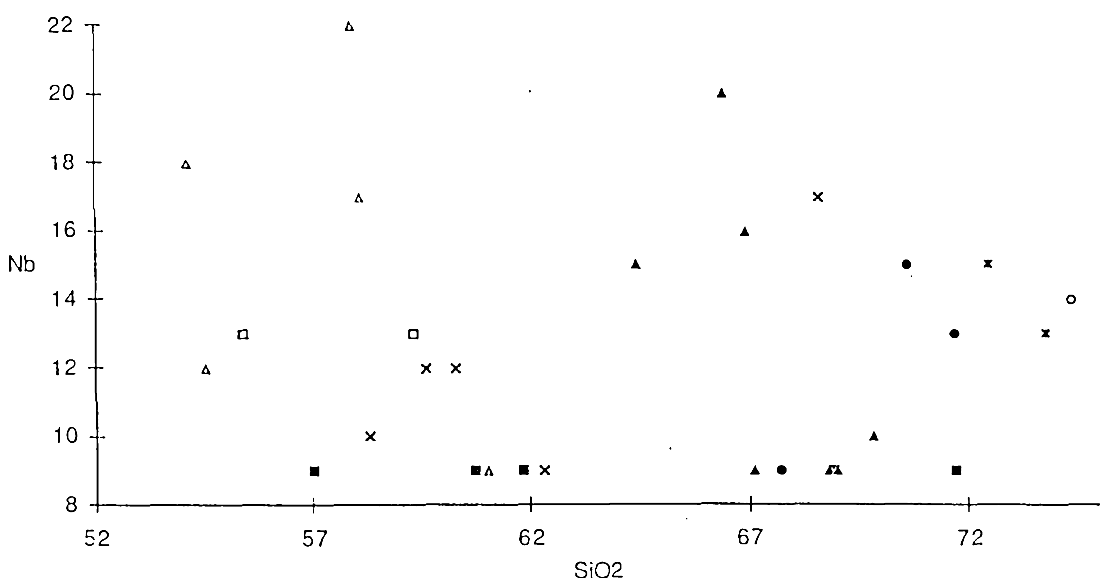


Figure 4.14 Variation diagrams for trace elements.

d.



e.



Hazard Creek Complex
- Epidote Tonalite
- Older Gneisses
- K Rich Suite

Little Goose Creek Complex
- Tonalites-including 85z9
- Western Porphyritic Orthogneiss
- Eastern Porphyritic Orthogneiss

Payette River Complex
- Payette River Tonalite
- Trail Creek Granite

Figure 4.14 continued

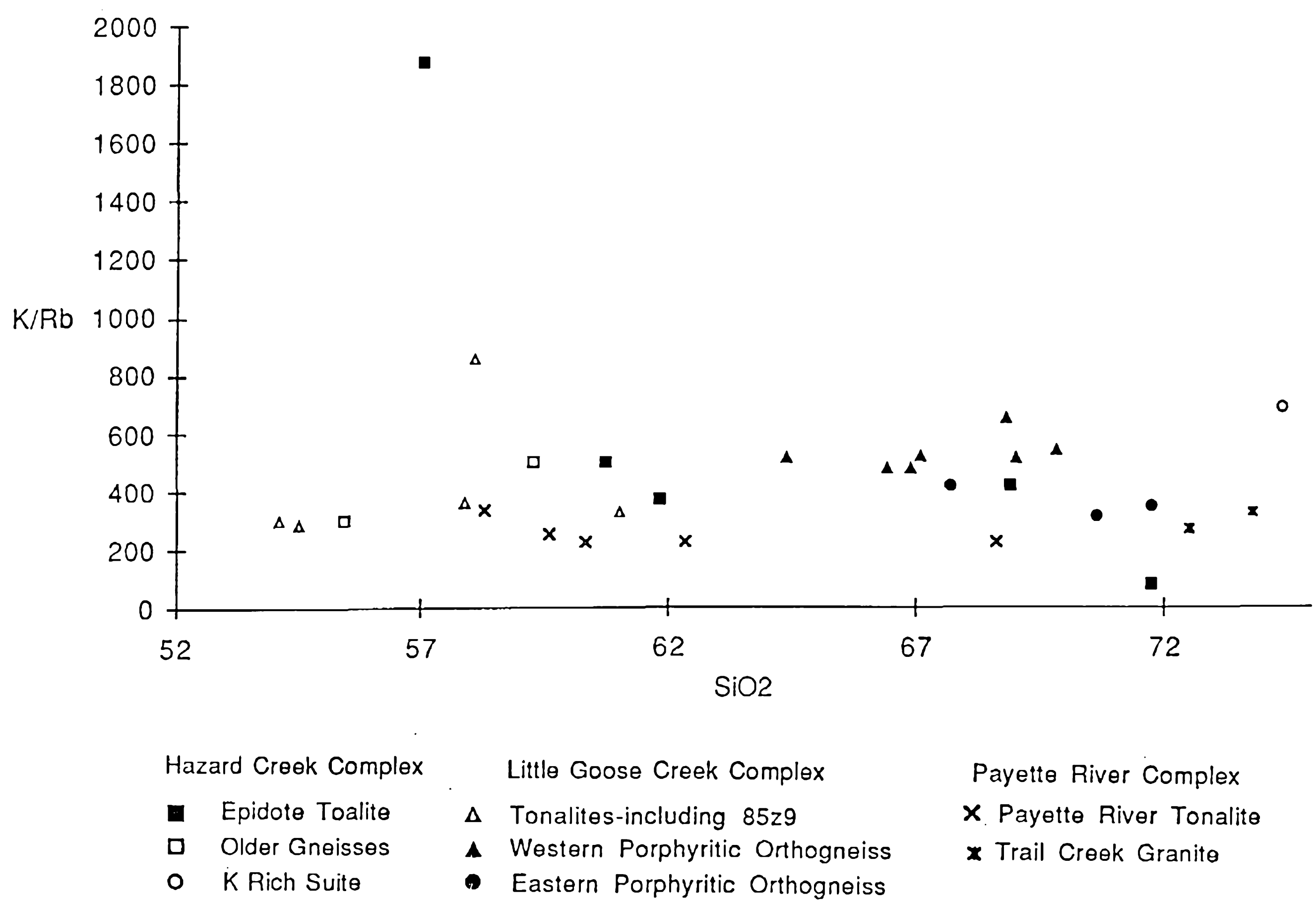
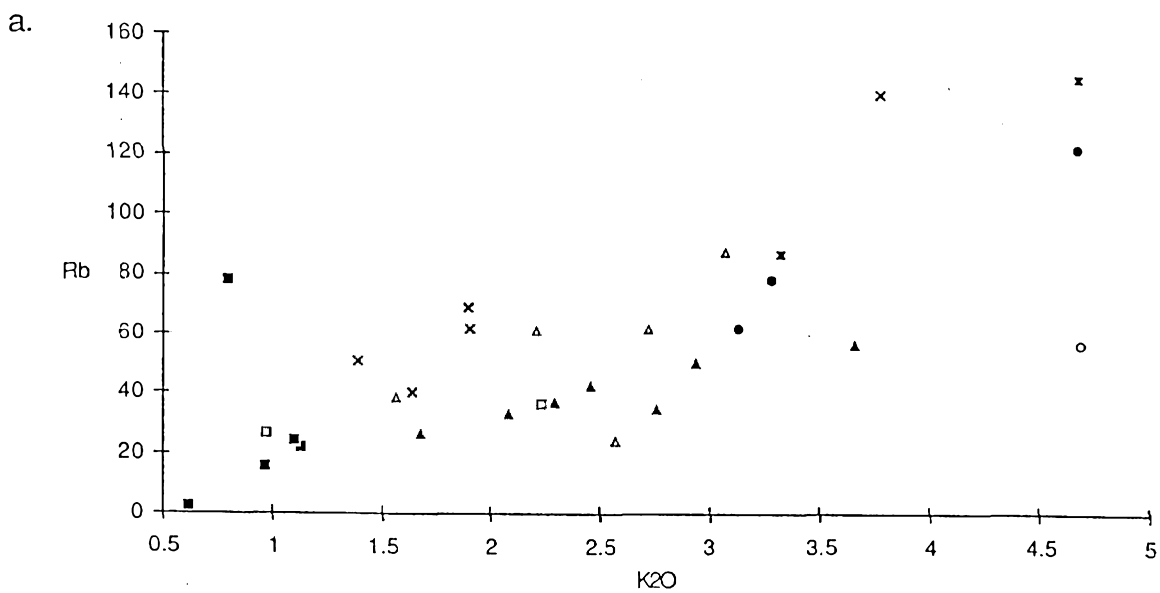


Figure 4.15 K/Rb versus SiO_2.



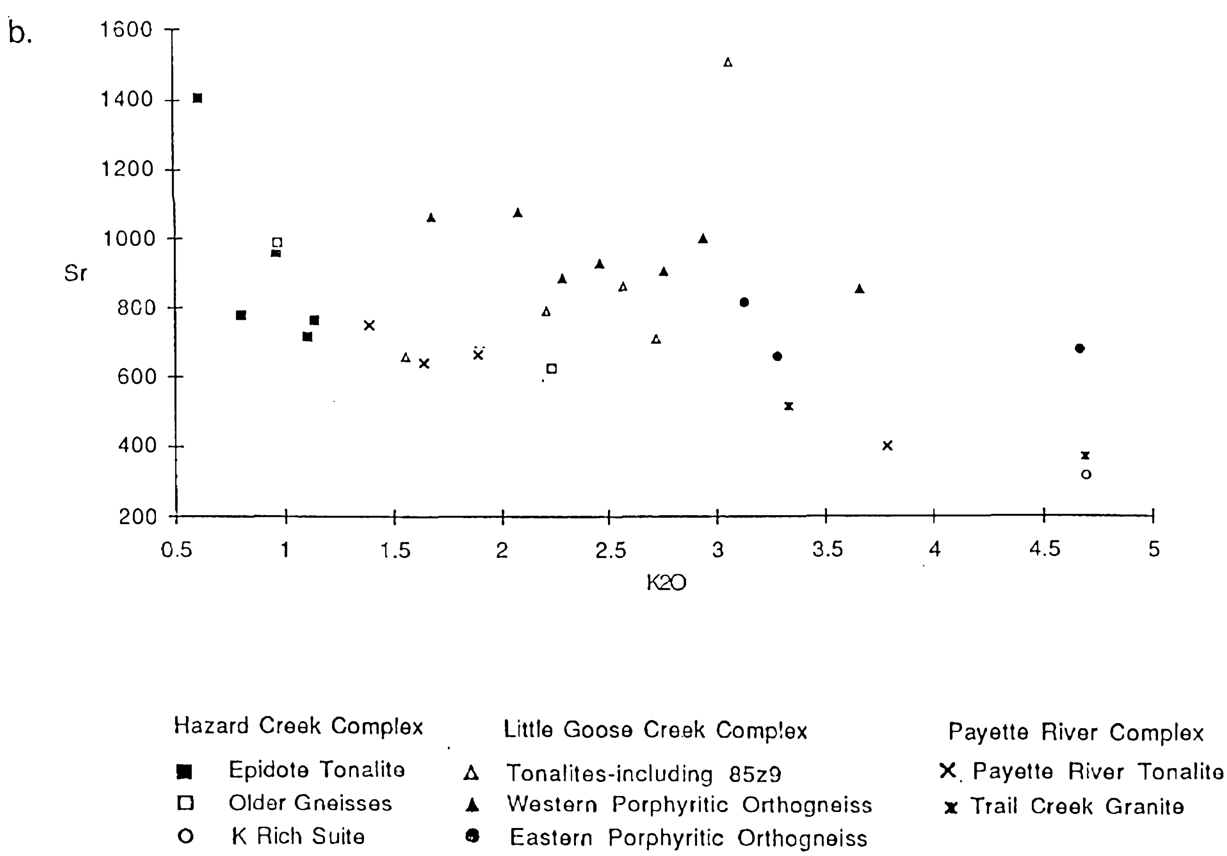


Figure 4.16 Rb (a) and Sr (b) versus K_2O.

complex are relatively rare (DePaolo, 1981; Hill et al., 1986; Kistler and Peterman, 1973; Silver et al., 1975) and are more typical of oceanic-arcs (Gill, 1981).

K/Rb (Figure 4.15) is lowest in the Trail Creek Granite and Payette River Tonalite, with values between 225 and 340. Eastern porphyritic orthogneiss and most Little Goose Creek complex tonalites have slightly higher ratios (250 to 420). The highest ratios are from the western porphyritic orthogneiss and Hazard Creek complex (300 to 690). Samples 83z3 and 83z4 have high-K/Rb due to unusually low Rb. 85z17 has unusually low K/Rb due to high Rb, possibly due to incorporation of a biotite clot. The correlation between Rb, SiO_2 and K_2O is weak (Figures 4.14, 4.16), consistent with the weak correlation between SiO_2 and K_2O, suggesting that K and Rb were variable in the magma or have been mobile.

K/Rb is often used to descriminate between evolved and primitive source materials. N type MORB has K/Rb near 1000 (Gill, 1981). Andesites have a large range of K/Rb between 250 and 1000. High-K suites commonly have lower K/Rb than low K suites. Samples from the Sierra Nevada Batholith with R_i less than .704 have K/Rb between 400 and 500, while samples with R_i greater than .704 have K/Rb between 200 and 400 (Kistler and Peterman, 1973). Samples from the Peninsular Ranges Batholith have K/Rb between 100 and 900 (DePaolo, 1981) with values increasing to the west. Values for the Atlanta and Bitterroot lobes are similar to those in the eastern Little Goose Creek complex and Payette River complex. The Hazard Creek complex and western porphyritic orthogneiss have K/Rb similar to the low R_i portions of batholiths while the eastern porphyritic orthogneiss, Little Goose Creek complex tonalites, and Payette River complex are similar to the high R_i portions of batholiths. All values are in the range of andesites and none are as primitive as N type MORB.

Sr concentrations (Figure 4.14b) are highest in the Hazard Creek complex and western Little Goose Creek complex, intermediate in the Payette River Tonalite and eastern porphyritic orthogneiss, and lowest in the Trail Creek Granite and the Hazard

Creek complex K-rich suite. Concentrations in the Hazard Creek complex and western Little Goose Creek complex are unusually high (700 to 1400 ppm). These values are near the upper end of those reported for arc magmas. Similar high concentrations are found in the central gneiss complex of the Coast Plutonic Complex, and in granite samples from the Bitterroot Lobe. Payette River Tonalite, eastern porphyritic orthogneiss, and Trail Creek Granite have Sr concentrations similar to most other batholithic data, and to data from the Atlanta Lobe.

Y contents (Figure 4.14c) are very low in the Hazard Creek complex epidote tonalites and are low in the Payette River Tonalite, Trail Creek Granite, and porphyritic orthogneiss. Similar low values in andesites are seen only where they intrude thick crust (Gill, 1981). Values in the Little Goose Creek complex tonalites and Hazard Creek complex older gneisses are similar to those seen in orogenic andesites. Y behaves chemically like a HREE. Depletions in Y may be due to hornblende fractionation, to hornblende or garnet bearing residues during partial melting, or to zircon fractionation (Hanson, 1978; Pearce and Norry, 1979; Gromet and Silver, 1987). Y contents in samples from the Idaho Batholith are comparable to those in rocks from the study area.

Zr is generally lowest in the Hazard Creek complex and highest in the Little Goose Creek complex (Figure 4.14d). There may be some correlation with SiO_2 in high SiO_2 rocks. These values are similar to those in the Idaho Batholith. Nb shows much within-suite variation (Figure 4.14e). Hazard Creek complex epidote tonalites are depleted in Nb while older gneisses have higher Nb. All samples fall within the arc range of Pearce et al. (1984), which is characterized by lower Nb than observed in within plate granites. Zr/Nb for tonalites from all suites is less than 20 in the range of andesites intruding continental margins. Nb in the Bitterroot lobe is similar to that in the Trail Creek Granite and Nb-rich samples from the porphyritic orthogneiss.

Ni for all samples is within the range of orogenic andesites. Cu is severely depleted. Zn concentrations, in contrast, are within or slightly higher than the typical range for orogenic andesites.

REE patterns for samples from the field area (Table 4.4), normalized to the chondritic values of Anders and Ebihara (1982), are shown in Figure 4.17. Sample 83z9 is Hazard Creek complex epidote tonalite; 84z3 is a Little Goose Creek complex tonalite; 83z2 and 83z11 are Payette River Tonalite and 83z15 is Trail Creek Granite. There is no analysis of porphyritic orthogneiss. Three samples of tonalite and trondhjemite from unknown locations in the Hazard Creek complex are also shown (Barker et al., 1979).

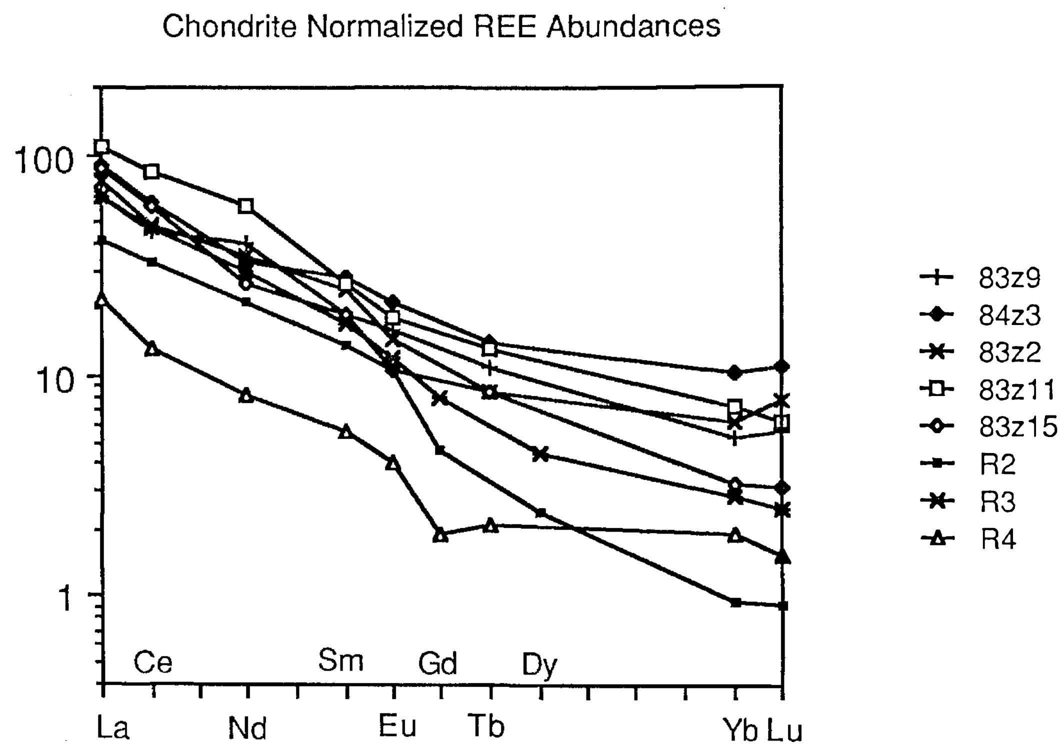
Samples from all complexes are enriched in light REE (LREE), depleted in heavy REE (HREE), and show little variation in total REE. Hazard Creek complex samples have the least LREE enrichment, greatest HREE depletion, and lack Eu anomalies. Little Goose Creek complex tonalites (84z3), Payette River Tonalite, and Trail Creek Granite have similar LREE enrichments. Little Goose Creek complex tonalites have less HREE depletion, consistent with the relatively high Y in these rocks, and a small negative Eu anomaly (Eu/Eu* of .91). Payette River Tonalite has moderate negative Eu anomalies (Eu/Eu* from .78 to .84). Trail Creek Granite has the greatest HREE depletion and largest negative Eu anomaly (Eu/Eu* of .7). All samples are within the range found in the eastern portion of the Peninsular Ranges Batholith (Gromet and Silver, 1987). Samples from the central Peninsular Ranges Batholith have smaller LREE enrichments but are otherwise similar. REE patterns in these rocks are attributed to melting of LREE-enriched material at depth with garnet in the residuum (Gromet and Silver, 1987). Similar patterns are also observed in batholithic rocks farther north in Idaho from the South Fork of the Clearwater River (Hoover, 1986).

Values for several other trace elements were also obtained for the five samples analyzed by INAA (Table 4.4). These analyses show that the Hazard Creek complex is

Table 4.4: INAA analyses of REE and other trace elements

	Study area: analyses by A. Hoover*					Riggins region (Barker, 1979)			
Sample	83Z9	84Z3	83Z2	83Z11	83Z15	R2	R3	R4	R5
Sc	7.6	11	10.7	15	10.2	1.9	4.6	2.1	10
Cr	54	74	109	83	23	4.4	7		3
Co	79	70	67.5	71	106	3.4	3.3	1.2	0.1
Zn	92	93	74	98					
Rb		54	73	71	94	26	9	17	23
Sr						815	113	711	92
Zr	149		147	177	114				
Cs		0.7	1.6	0.76	0.8				
Ba	650	1340	830	955	1348	798	262	477	713
La	20.2	28	24	34	27	13	20	7	4
Ce	37	50	40	68	48	27	38	11	9
Nd	24	20	21	36	16	13	18	5	5
Sm	3.7	5.6	4.8	5.2	3.7	2.7	3.4	1.1	
Eu	1.21	1.6	1.1	1.37	0.8	0.8	0.9	0.3	0.5
Gd						1.2	2.1	0.5	2.7
Tb	0.52	0.68	0.40	0.62	0.4			0.1	
Dy						0.6	1.1		
Tm									
Yb	1.1	2.2	1.3	1.5	0.69	0.2	0.6	0.4	2.4
Lu	0.18	0.36	0.25	0.20	0.10	0.03	0.08	0.05	0.3
Hf	3.7			4.4	2.9	2.9	3.8	2.2	1.1
Th	1.7	7.3	3.5	4.9	9.3	1.2	1.1	0.8	0.7
U						0.45	0.44	0.36	2
Zr						107	156	87	64

*estimated errors at one standard deviation: ± 1-3% for Sc, Co, and La; ± 2-5% for Sm, Eu, Yb, and Lu; ± 4-7% for Zn, Ce, Tb, Hf, and Th; ± 10-15% for Cr, Rb, and Ta; and ± 15-25% for Cs, Ba, Nd, and Zr



*Normalized to chondritic values of Anders and Ebihara (1982)

Figure 4.17 Rare earth element patterns for five samples analyzed by A. Hoover and three samples from Barker et al. (1979), normalized to the chondritic values of Anders and Ebihara (1982). Samples are described in the text.

depleted in Th. The Little Goose Creek complex tonalites and Hazard Creek complex have higher Ba/Rb than the Payette River Tonalite and Trail Creek Granite.

Summary of Complex Characteristics

The three complexes have distinct chemical characteristics. The Hazard Creek and Payette River complexes plot in distinct fields on most variation diagrams, while the Hazard Creek complex has higher Na_2O, Al_2O_3, and Sr and lower K_2O, MgO, FeO*,TiO_2, Rb, Y, Zr and Nb for a given SiO_2 than the Payette River complex. The Little Goose Creek complex contains a remarkable chemical discontinuity: western and eastern porphyritic orthogneisses, as defined on the basis of R_i, plot in distinct fields on every diagram. Western porphyritic orthogneiss is chemically more similar to the Hazard Creek complex: it is characterized by high Na_2O, Al_2O_3, and Sr and low MgO and TiO_2 though these characteristics are generally less extreme than in the Hazard Creek complex. K_2O and Rb are higher in the western porphyritic orthogneiss than in the Hazard Creek complex. Eastern porphyritic orthogneiss is similar to Payette River complex Trail Creek Granite. Within the Payette River complex, Trail Creek Granite is distinguished from Payette River Tonalite only by the difference in SiO_2 and K_2O; most other elements form a continuous variation with SiO_2. Coarse Grained Tonalite Orthogneiss is chemically distinct from Little Goose Creek complex tonalites and is more similar to Payette River Tonalite. This is particularly clear in Harker diagrams for K_2O, TiO_2, and Na_20 (Figure 4.7).

Summary of Comparison with Other Igneous Suites

The Payette River Tonalite is a medium K suite similar to other tonalites from the Idaho Batholith, the western Sierra Nevada, and the Peninsular Ranges Batholith. Al_2O_3 is enriched relative to orogenic andesites and to the rest of the Idaho Batholith but similar to that in the Peninsular Ranges Batholith, Sierra Nevada Batholith and Coastal Batholith. Na_2O, MgO and CaO concentrations are all similar to values in Cordilleran batholiths. These rocks appear to be typical of medium-K tonalite and granodiorite in Cordilleran batholiths.

The Trail Creek Granite and eastern porphyritic orthogneiss are a higher-K suite, similar to the eastern Sierra Nevada Batholith, the central gneiss complex of the Coast Plutonic Complex, and the portion of the Coastal Batholith that intrudes continental crust (Arequipa section). Na_2O, Al_2O_3, MgO and CaO concentrations are all similar to those in other Cordilleran batholiths. These rocks appear to be typical of high-K granodiorites and granites in Cordilleran batholiths. Many higher K_2O rocks are known from other batholiths.

The western Little Goose Creek complex contains medium-K western porphyritic orthogneiss and high-K Little Goose Creek complex tonalites. Concentrations of K_2O in tonalites are most similar to samples from the central gneiss complex of the Coast Plutonic Complex; the porphyritic orthogneiss is similar to the western Sierra Nevada Batholith and Peninsular Ranges Batholith. The rocks of the western Little Goose Creek complex contain less MgO and more Al_2O_3 and Na_2O than most Cordilleran batholithic rocks. Other elements have concentrations that are more typical, and correlations between elements are not unusual. The suite is most similar to the high-K, high Al_2O_3 central gneiss complex of the Coast Plutonic Complex, but is distinguished by somewhat higher Na_2O.

The epidote tonalites of the Hazard Creek complex are not chemically similar to andesites or Cordilleran batholiths. They have very low K similar to that found in the western foothills of the Sierra Nevada. Na_2O, Al_2O_3, and Sr are unusually high, and MgO uncommonly low, similar to the western Little Goose Creek complex. The older gneisses are generally similar to the epidote tonalites; however, much more data is needed to characterize them fully. They are tentatively assigned to the early portions of the epidote tonalite intrusive episode. The K-rich suite of the Hazard Creek complex is markedly higher in K and Rb and lower in Na_2O, Al_2O_3 and Sr. These fundamental differences suggest that these rocks are not related to the epidote tonalite intrusive episode. They may be related to younger K-rich magmatism in the Little Goose Creek complex.

The epidote tonalites have chemistry that is typical of tonalite-trondhjemite suites, consistent with the presence of samples identified as trondhjemite on the basis of mineralogy (Chapter 2). The tonalite-trondhjemite suite forms a small percentage of intrusions related to subduction. In the Mesozoic and Cenozoic, these intrusions usually occur on the oceanward side of larger calc-alkaline intrusives (Barker, 1979). The Hazard Creek complex is this type of occurrence. It is not known why some tonalite suites differentiate into trondhjemite while others evolve toward granite. In the western Cordillera, many tonalites intrude the eugeosynclinal section; however, only a small percentage evolve toward trondhjemite rather than granodiorite. These bodies appear to be of varying age.

Barker subdivides trondhjemites chemically on the basis of Al_2O_3. Rocks with greater than 15 percent Al_2O_3 at 70 percent SiO_2 are classified as high Al_2O_3 trondhjemites (Barker, 1979). These rocks are generally characterized by high concentrations of Al_2O_3 and low concentrations of HREE and are more common in continental settings (Arth, 1979). Low Al_2O_3 trondhjemites are more common in oceanic settings and are characterized by flat REE patterns, lower Sr concentrations and marked Eu anomalies. The difference between the trondhjemite suites is proposed to be due to different conditions during melting (Barker, 1979). Low Al_2O_3 trondhjemite suites are proposed to form by fractional crystallization or partial melting of basalt at relatively shallow depths where plagioclase is a residual phase. High Al_2O_3 trondhjemite suites are proposed to form by fractional crystallization of basalt at intermediate depths removing hornblende and plagioclase, or by partial melting of basalt or eclogite, leaving hornblende or garnet in the residue (Arth, 1979). The Hazard

Creek complex is a high Al_2O_3 suite with the characteristic high Sr concentration and HREE depletion.

Comparison of the Hazard Creek complex data to several trondhjemite suites of varying ages and origins given in Barker (1979) show that the Hazard Creek complex is unusually rich in Al_2O_3, Na_2O, and Sr while MgO and FeO* are relatively low. All elements except CaO vary with SiO_2 in a fashion similar to other high Al_2O_3 trondhjemite suites, suggesting that similar processes may have operated in all cases. The Hazard Creek complex is most similar to the trondhjemites of Norway but differs from these rocks in trace elements, containing markedly higher concentrations of Ba, Sr and REE (Barker and Millard, 1979).

The Hazard Creek complex epidote tonalites are also unusual because they contain large epidote crystals that may be a primary igneous phase. The Hazard Creek complex is one of a series of epidote-bearing occurrences extending the length of the western Cordillera (Zen and Hammarstrom, 1984). Two of these occurrences are tonalite-trondhjemite suites; the remainder appear to be tonalite-granodiorite suites. Chemical data for comparison is available for only three areas: the Ahsahka area in northern Idaho (Hietanen, 1962), the Tenpeak-Sulfur Mountain area in north-central Washington (Cater, 1982), and the Feather River Region in northern California (Hietanen, 1973, 1976). These rocks all have high Al_2O_3 and Sr and low Y concentrations; however, those in the Hazard Creek complex are more extreme.

Comparison with other trondhjemites and with other epidote-bearing intrusives show that the Hazard Creek complex epidote tonalites are more similar to these suites than to more typical Cordilleran batholithic rocks. However, Al_2O_3, Na_2O and Sr are uncommonly high and MgO uncommonly low even relative to these rocks. This suggests that these may be uncommon features of the source material.

4.4 DISCUSSION

4.4.1 Discussion of Chemical Mobility

Isotopic data suggested that deformation in the Hazard Creek complex and Little Goose Creek complex did not involve large scale chemical mobility. This is generally corroborated by the regular chemical variation of most elements with SiO_2. K_2O and Na_2O show considerable scatter, suggesting that they may have been mobile during deformation or during cooling. Within the Little Goose Creek complex, the gross correlation between K_2O, Sr concentration, R_i, and $\delta^{18}O$ suggests the variations are primary. It is unlikely that a metasomatic fluid would alter all four parameters in unison. In addition, the medium K_2O, Coarse Grained Tonalite Orthogneiss in the center of the Little Goose Creek complex is unlikely to have retained its relatively low K_2O signature, if large scale K metasomatism occurred during deformation.

The high-K nature of some rocks may be due to metasomatism of country rocks during the emplacement of porphyritic orthogneiss in the Little Goose Creek complex, and the high-K garnet granodiorite in the Hazard Creek complex. Sample 83z8 of Quartz Diorite Orthogneiss in the Hazard Creek complex has unusually high K_2O. This sample is from the Granite Twin Lakes area where the gneiss is cross-cut by numerous alkali feldspar-bearing pegmatites derived from the high-K garnet granodiorite. In this area, the gneiss locally contains alkali feldspar that is not abundant elsewhere and may represent a metasomatic addition to the rock. Similarly, in Little Goose Creek Canyon, where Eastern Tonalite Gneiss is adjacent to the porphyritic orthogneiss of the Little Goose Creek complex, the eastern tonalite gneiss contains alkali feldspar. Deformed alkali feldspar-bearing dikes and sills are common in the Eastern Tonalite Gneiss at this locality. Farther to the north, the gneiss is juxtaposed against tonalite of the Little Goose Creek complex and lacks alkali feldspar.

The high-K nature of the Little Goose Creek complex tonalites may be due to the addition of K_2O from the surrounding granodiorite during intrusion. These samples have K_2O and Rb concentrations similar to those in the surrounding porphyritic orthogneiss. Alkali feldspar-bearing dikes and sills are abundant throughout the Tonalite Orthogneiss.

4.4.2 Role of Fractional Crystallization in Generation of Chemical Characteristics

The effects of fractional crystallization are often difficult to distinguish because trace element patterns formed by removal of a phase during crystallization are similar to those formed by retention of that phase in the source during partial melting. Hornblende, plagioclase, magnetite or ilmenite, and pyroxene are the most likely early crystallizing phases that should be considered as possible fractionates. Early crystallizing accessory phases can also have a large effect on the trace element composition of the magmas, as they incorporate large amounts of trace elements (Gromet and Silver, 1987). The Hazard Creek complex and Little Goose Creek complex each contain a variety of rock types in suites that may be related by fractional crystallization. The role of fractional crystallization in the evolution of Payette River Tonalite and associated granodiorites and granites cannot be evaluated due to restricted sampling. The homogeneous composition of the Trail Creek Granite and similar granites from the core of the batholith suggests that little fractionation has occurred in this suite (Hyndman 1984; Lewis et al. 1987).

Plagioclase fractionation would remove Sr, Ca, Al and Si from the melt. Within the porphyritic orthogneiss, Sr decreases with increasing SiO_2 (Figure 4.14b); Sr and CaO are also strongly correlated (Figure 4.18). These features are consistent with either plagioclase fractionation or mixing of a high Sr, Ca, low Si melt with low Sr, Ca, high Si material. The good correlation between Sr and R_i suggests that mixing has been important. Within the Hazard Creek complex, Sr concentration is highly variable and is

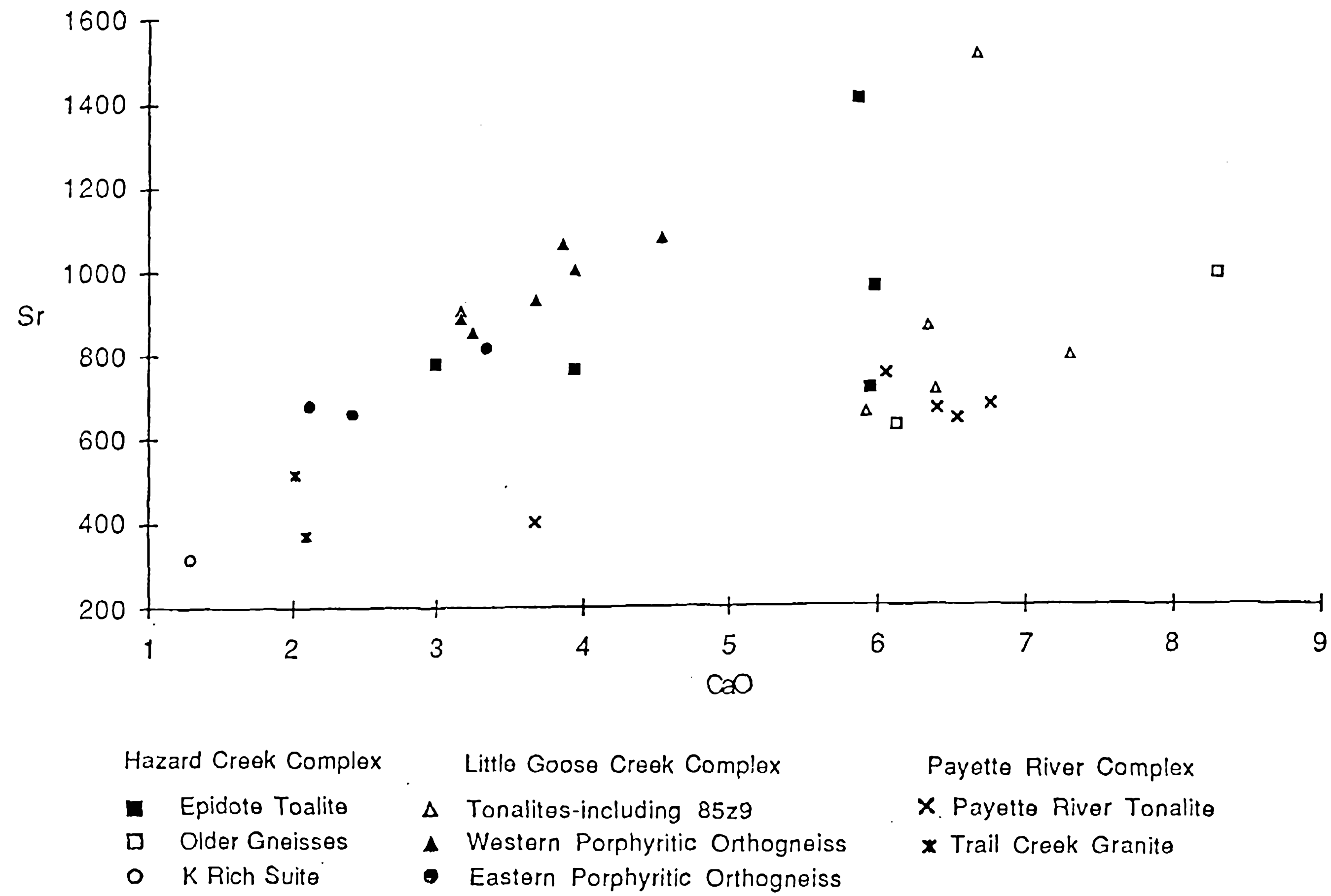


Figure 4.18 Sr versus CaO.

uncorrelated with SiO_2 or CaO. This suggests that source variations in Sr have overwhelmed any effect of plagioclase fractionation.

Hb fractionation removes Y, Nb and lesser amounts of Zr from the melt (Pearce and Norry, 1979; Gromet and Silver, 1987; Hanson, 1978) as well as Ca, Al, Fe, Mg and Si. LREE are enriched, HREE depleted, and positive Eu anomalies are formed by the removal of hornblende (Arth and Barker, 1976; Hanson, 1978; Gromet and Silver, 1987). Within the Hazard Creek complex Y, Zr, Nb, Hf and Yb generally decrease as SiO_2 increases, consistent with the removal of hornblende. REE patterns have LREE enrichment, HREE depletion and no Eu anomaly. Total abundance generally decreases as rocks become more siliceous. These patterns may have formed by the fractionation of hornblende and lesser plagioclase. Removal of plagioclase and hornblende from dacite produces parallel patterns with decreasing abundances and positive Eu anomalies (Arth and Barker,1976). These anomalies could be suppressed with increased plagioclase fractionation. Hornblendite and gabbro inferred to be deformed cumulate rocks (Table 4.5) contain abundant Y, as is required to form the observed pattern. They do not contain detectable Nb, suggesting that the magmas contained little Nb prior to fractionation.

Fractionation of constant proportions of plagioclase and hornblende alone could not create the chemical patterns observed because there is a lack of one-to-one correspondence between SiO2 and REE depletion and a lack of complete parallelism in the HREE patterns. However, the presence of hornblende-plagioclase cumulates within the complex suggests some aspects of the chemistry may be due to this process. It is not clear if hornblende fractionation could enrich the LREEs to the observed levels.

The paucity of cumulate material and the abundance of tonalite make it unlikely that these rocks formed by fractional crystallization of a basaltic melt. Partial melting of amphibolite, leaving garnet, hornblende, or both in the residue, seems more plausible. This magma may then have been modified by fractional crystallization of

Table 4.5: XRF analyses of wall rocks and cumulates

	LAYERED MAFIC GNIESSES		HORNBLENDITE AND GABBRO***		
	84z9*	85z10**	84z25a	84z24	84z23
SiO2	63.6	62.20	41.9	50.1	44.3
Al2O3	12.5	16.10	9.2	11.5	13.9
FeTo3	7.6	6.75	17.6	11.0	13.3
MgO	3.61	3.38	11.9	11.4	10.8
CaO	7.65	5.85	14.9	11.2	10.6
Na2O	2.68	3.04	1.13	1.86	2.53
K2O	1.01	1.31	0.55	0.88	0.73
TiO2	0.71	0.27	2.30	1.46	2.03
P2O5	0.17	<0.05	0.11	<0.05	<0.05
MnO	0.11	0.07	0.20	0.22	0.16
Rb	34	23	3	9	13
Sr	346	90	202	217	384
Nb	10	<10	<10	11	<10
Zr	85	59	42	44	48
Y	24	20	16	23	24
Cu	82	<2	<2	<2	<2
Ni	53	20	32	26	36
Zn	80	75	106	106	97

* from Little Goose Creek Complex
**from Hazard Creek Complex, Layered Mafic Gneiss

***from Granite Twin Lakes area of Hazard Creek Complex hornblende and plagioclase to yield the range of compositions observed. The trace element variations may largely reflect patterns established during partial melting.

Within the Little Goose Creek complex Zr, Y, and Nb are highly variable and not correlated with SiO_2. If hornblende has been fractionated, the effect has been masked by variable initial concentrations or by mixing processes. Y and Nb are often more abundant in the Little Goose Creek complex than in the Hazard Creek complex, suggesting that the two complexes are not related by hornblende fractionation. High Ba concentrations indicate neither biotite nor alkali feldspar have been fractionated.

4.4.3 Models of Magma Evolution

4.4.3.a Hazard Creek Complex

The Hazard Creek complex epidote tonalites form a tonalite-trondhjemite suite characterized by unusually high Al_2O_3, Na_2O and Sr, and low MgO. Low R_i, typical of oceanic-arc magmas, precludes interaction with evolved crustal material. $\delta^{18}O$, in contrast, is elevated above typical arc values. High Sr and Al_2O_3 concentrations indicate that plagioclase was not a residual phase during melting. HREE depletion as described above may be due in part to hornblende fractionation; however, the presence of amphibolite with low Nb concentrations suggests that much of the HREE depletion is due to partial melting at depth with garnet, hornblende, or both as residual phases. The absence of plagioclase and presence of garnet or hornblende as residual phases suggests melting at depths where plagioclase is unstable.

$\delta^{18}O$ is higher in the Hazard Creek complex than is typical for modern oceanic-arcs. Similar values in oceanic-arc rocks (up to 9.2) have been attributed to anomalous mantle material but have been observed only in conjunction with high R_i, Rb and K (Margaritz et al., 1978). The low R_i, Rb and K in the Hazard Creek complex suggests that the anomalous values are more easily explained by the incorporation of oxygen from sedimentary or hydorthermally altered material into the melt. This may be accomplished by melting the material in the source area, assimilating it during emplacement, or by fluids that have exchanged with this material, metasomatically altering the source material.

Oceanic-arc material with low R_i is the most plausible source for this oxygen. Mafic layered gneisses in the Hazard Creek and Little Goose Creek complexes (Table 4.5), Pollock Mountain Amphibolite, Riggins Group (Hamilton, 1963a) and Seven Devils Group (Vallier and Batiza, 1978) are all comprised of appropriate material. These rocks are of broadly basaltic to andesitic composition providing a plausible source material (Arth, 1979) for the Hazard Creek complex magmas, if carried down to appropriate depth. K_2O, Na_2O and Rb concentrations in these rocks are similar to those in the epidote tonalites; however, CaO, FeO*, MgO and TiO_2 are higher and Al_2O_3 lower. Sr and Ba concentrations are much lower and Zr, Cu and Y concentrations are higher. REE in similar rocks from the South Fork of the Clearwater River have slightly depleted to moderately enriched LREE and undepleted to slightly depleted HREE (Hoover, 1986; Figure 4.19).

Large amounts of partial melting of material similar to this might produce a melt with appropriately high Al_2O_3 and Na_2O and low K_2O; however, low Sr concentrations in the metavolcanic rocks (20 to 300 ppm) would not yield unusually high Sr (700 to 1400 ppm) concentrations as seen in the epidote tonalites. Assimilation of this material would enrich the melt in Na_2O and possibly Al_2O_3 but not K_2O . LREE enrichment would occur if the material had LREE enrichment similar to that seen in some samples to the north (Hoover, 1986).

To test the plausibility of assimilated metavolcanic material causing elevated $\delta^{18}O$ in the Hazard Creek complex epidote tonalites, an isotopic simple mixing model has been calculated. The magma that is generated at depth, "arc magma," is assumed to have low $\delta^{18}O$ and R_i similar to that seen in oceanic island arcs. Typical oceanic island arc

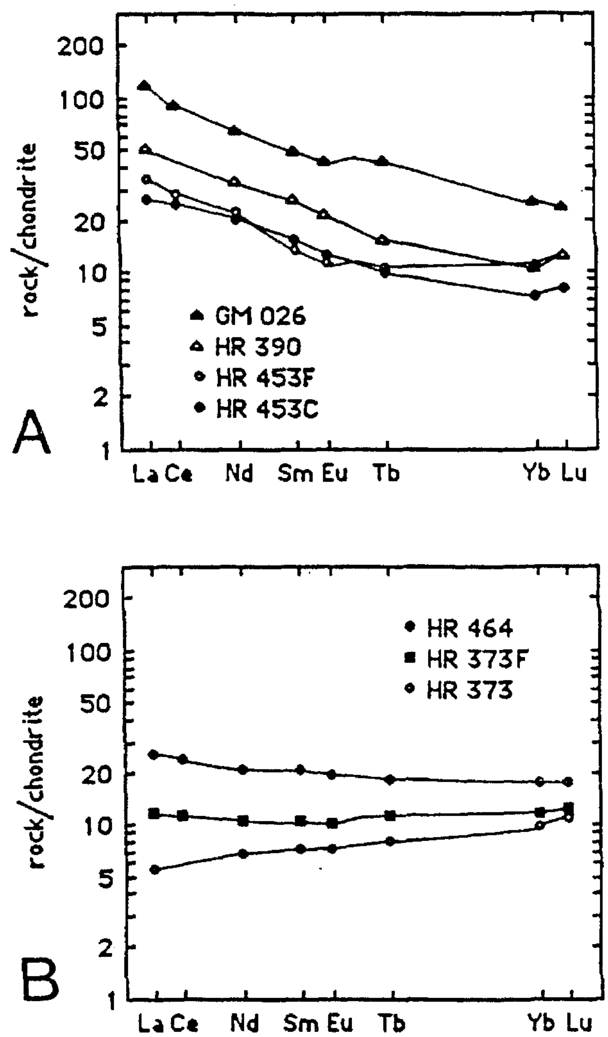


Figure 4.19 Rare earth element patterns normalized to the chondritic values of Anders and Ebihara (1982) of metamorphosed arc rocks exposed along the South Fork of the Clearwater River. Rocks in group A are hornblende gneiss and biotite gneiss; rocks in group B are hornblende gneiss. Figure and data from Hoover, 1985.

magma is presumed to represent uncontaminated magma generated in response to subduction.

Isotopic data are available only for the mafic layered gneiss, which at the time of formation of the epidote tonalites (approximately 100 Ma) had $^{87}Sr/^{86}Sr$ ($R_{(100)}$) of .7067 and .7052 and $\delta^{18}O$ of 9.7 and 10.2. Pristine arc volcanics generally have R_i between .703 and .704. Interaction with seawater (.7067 and .7078 in Permian to Triassic time; Burke et al., 1982) should drive $^{87}Sr/^{86}Sr$ (R) to higher values. Based on these data the metavolcanic contaminant is estimated to have $R_{(100)}$ of .705. This is sufficiently low that the amount of possible assimilation is limited by Sr concentration and $\delta^{18}O$.

It is difficult to estimate $\delta^{18}O$ in spilitized volcanic rocks, because fractionation is strongly temperature-dependent and because the isotopic properties of the exchanging fluid may evolve with time. Low temperatures alteration in ophiolites increases $\delta^{18}O$ (McCulloch et al., 1980), consistent with estimates of the fractionation factor (Taylor, 1974). Metamorphism prior to melting may also have altered $\delta^{18}O$. The values measured for the mafic layered gneisses are taken as the best estimates; however, these values may have been altered during the emplacement of the intrusives. $\delta^{18}O$ of 10 is assumed for the metavolcanic contaminant.

A simple mixture of 50% metavolcanic material ($\delta^{18}O$ =10, 200 ppm Sr) with "arc magma" ($\delta^{18}O$=7, 1200 ppm Sr) has $\delta^{18}O$ of 8.5 and contains 700 ppm Sr. The range of values seen in the epidote tonalites of the Hazard Creek complex is produced in mixtures with varying proportions only if the Sr concentration in the metavolcanic component is allowed to vary between 100 and 300 ppm. Fractional crystallization coupled with assimilation would remove Sr, requiring higher Sr concentrations in the "arc magma". Variable Sr concentrations in the contaminant would still be required.

This model indicates that incorporation of significant amounts of metavolcanic material like that exposed as wall rocks in the complex could raise the $\delta^{18}O$ of the melt without raising R_i above the observed values. The presence of this component throughout the complex suggests incorporation occurred below the level of exposure. The "arc magma" must have had very high concentrations of Sr (1200 ppm or higher). These very high Sr concentrations are inferred to be a property of the source material at depth. The variable Sr concentration in the Hazard Creek complex epidote tonalites is not correlated with R_i, CaO, or mineralogy. This variation cannot be explained by assimilation of a contaminant with constant Sr concentration. It must reflect either variations in the source material or variations in the contaminant. Complex combinations of fractional crystallization and variability in magma composition are also possible.

In summary, the Hazard Creek complex epidote tonalites appear to have formed by partial melting at depth, leaving garnet or hornblende but not plagioclase as a residual phase. The unusually high Sr concentrations in the rocks appear to be a reflection of high concentrations in the "arc magma" generated at depth. This material may also have been enriched in Al_2O_3, Na_2O and LREE and depleted in MgO. $\delta^{18}O$ in the epidote tonalites is unusually high for an uncontaminated arc magma; assimilation of up to 50% spilitized arc rocks similar to those exposed locally would raise $\delta^{18}O$ to the appropriate value without raising R_i or lowering Sr concentration below the observed value. Either the assimilant, the "arc magma," or both, must have had variable Sr concentration. The assimilant may have also contributed to the high Na_2O, Al_2O_3, and LREE observed in the epidote tonalites. Fractionation of hornblende and plagioclase may have generated the diversity of rock types seen in the epidote tonalites.

Sample 83z8 of older gneiss has $R_{(120)}$ of .7044, which is the highest in the complex. This high ratio suggests the rock may be much older; however, an unreasonable age of 300 Ma is required to reduce R_i to .7039. It is unlikely that Rb has been lost from this sample as it may have been enriched in K_2O during garnet granodiorite emplacement. The high ratio appears to reflect an unusually high initial ratio due to variability in a source or contaminant. However, more data from this unit is needed.

4.4.3.b Little Goose Creek Complex

The Little Goose Creek complex contains four distinct geochemical suites: Little Goose Creek complex tonalites, western porphyritic orthogneiss, eastern porphyritic orthogneiss, and Coarse Grained Tonalite Orthogneiss (correlative with Payette River Tonalite). Little Goose Creek complex tonalites and western porphyritic orthogneiss both contain unusually high Na_2O, Al_2O_3 and Sr and low MgO similar to Hazard Creek complex epidote tonalites. As discussed above, the concentrations of these elements are unusual for batholithic rocks and for tonalite-trondhjemite suites. The presence of this unusual signature in all three suites suggests that they were derived in part from similar source materials. In contrast, eastern porphyritic orthogneiss and Coarse Grained Tonalite Orthogneiss contain abundances of Al_2O_3, Na_2O, Sr and MgO typical of Cordilleran batholithic rocks. The porphyritic orthogneiss, inferred to be emplaced in a single intrusive episode, contains rocks from both suites.

The general correlation in the Little Goose Creek complex between Sr concentration, R_i, and $\delta^{18}O$ (Figure 4.3b,4.4b,4.5b) suggests that mixing of a high R_i, high $\delta^{18}O$, low Sr, high Rb material with a low R_i, low $\delta^{18}O$, high Sr, low Rb material was important in the formation of the porphyritic orthogneiss. The chemical similarity between Hazard Creek complex epidote tonalites and western porphyritic orthogneiss suggests that the low R_i component of the mixture is a high Sr, high Na_2O, high Al_2O_3, low MgO, low R_i, low $\delta^{18}O$ magma similar to that which formed the Hazard Creek complex epidote tonalites. As discussed above, the unusual properties of this magma may be due to an unusually high Sr magma generated at depth, which has incorporated metavolcanic rocks. Western porphyritic orthogneiss is distinguished from the Hazard Creek complex by higher SiO_2, K_2O, Rb and Zr. Y is not as severely depleted and Nb has a larger range than in the Hazard Creek complex epidote tonalites. R_i and $\delta^{18}O$ are also higher than observed in the Hazard Creek complex. Eastern porphyritic orthogneiss is characterized by generally higher SiO_2, K_2O and Rb than western porphyritic orthogneiss. These relationships suggest that the second major component in the formation of the porphyritic orthogneiss is a high SiO_2, K_2O, Rb, R_i, $\delta^{18}O$ low Sr material. Plausible materials include Precambrian Belt or Salmon supergroups, Paleozoic sedimentary strata and Proterozoic or Archean basement. Metamorphosed Precambrian or Paleozoic sedimentary material may have different isotopic character and should also be considered.

Belt supergroup, which crops out to the north and east of the Idaho Batholith, has been isotopically well-characterized. Sr isotopic studies by Obradovich and Petterman (1968) and by Fleck and Criss (1985) show that R varies from .7141 to 1.18. Unmetamorphosed samples with R less than .80 are rare. Sr concentrations are low: generally less than 100 ppm. Rb concentrations are high: generally between 150 and 250. Both higher and lower values were measured. $\delta^{18}O$ in unmetamorphosed Belt supergroup was studied by Eslinger and Savin (1973), who measured values between 12.5 and 19. Fleck and Criss measured values between 12 and 16 for rocks that were at grades below the garnet isograd. $\delta^{18}O$ was lowered to 10 with increasing metamorphic grade. These data suggest that unmetamorphosed Belt supergroup is characterized by R between .8 and 1.0, Sr concentrations less than 100, Rb between 150 and 250 and $\delta^{18}O$ between 12 and 20. Metamorphosed Belt supergroup near the batholith may have somewhat lower R, .72-.86, and $\delta^{18}O$ between 10 and 12. Sr and Rb concentrations are similar. Material from the Salmon supergroup has not been studied but is assumed to be similar to material from the Belt supergroup.

Paleozoic miogeoclinal sedimentary rocks crop out in southern Idaho in the Albion Range (Armstrong and Hills, 1967), in southeastern Idaho in the Bayhorse Region (Hobbs et al., 1968), and in northern Idaho and northeastern Washington

(Harrison and Jobin, 1965; Greenman et al., 1977). Lund (1984) has proposed that they also occur as pendants in the western Idaho Batholith. These rocks are not isotopically well-characterized. They would be expected to have high $\delta^{18}O$ similar to the Belt supergroup because these values are typical for sedimentary rocks. R measured in the Albion Range (Armstrong and Hills, 1967) on metasedimentary rocks metamorphosed is between .71 and .75, consistent with their Cambrian to Pennsylvanian age. Samples of Paleozoic wall rocks in the Sierra Nevada (Kistler and Peterman, 1973) have similar values. Sr and Rb concentrations in miogeoclinal sedimentary rocks are expected to vary substantially. Sr concentrations in the Albion Range are between 130 and 150 ppm; those in the Sierra are between 10 and 170, suggesting that Paleozoic sedimentary rocks may contain more Sr in general than is present in the Belt supergroup. Rb concentrations are between 89 to 220 in the Albion Range and 12 to 96 in the Sierra, similar to those in the Belt supergroup. These data suggest that Paleozoic sedimentary rocks may be characterized by high $\delta^{18}O$ between 15 and 20, and R between .71 and .75. Sr concentrations may be 100 or greater, higher than in the Belt supergroup, while Rb concentrations are variable but similar to the Belt supergroup.

The isotopic composition of basement rocks in central Idaho is not well-constrained. Archean gneisses are present in the Albion range and in southern Montana. Rocks in the Albion Range are characterized by R between .8 and 1.5 (Armstrong and Hills, 1967), 50 to 180 ppm Sr, and 114 to 415 ppm Rb. $\delta^{18}O$ of the gneiss is unknown; however, it is comprised of granitic gneiss and overlying metasedimentary rock. Due to the presence of granitic gneiss, $\delta^{18}O$ is expected to be somewhat lower on average than in sedimentary rocks.

The boundary between Proterozoic and Archean basement preserved in the block uplifts of southern Montana trends southwest toward the southern edge of the Idaho Batholith, suggesting that the batholith may be underlain by Proterozoic basement.

There are no documented occurrences of this basement in Idaho. However, schists, gneiss and amphibolite, which structurally underlie the Salmon supergroup along the Salmon River (Cater et al., 1973; Weis et al., 1972), are good candidates for this material. These rocks are described as 20% intrusive material interlayered with quartzofeldspathic paragneiss. Isotopic information is unavailable; however, $\delta^{18}O$ is expected to be lower than in the Belt supergroup due to the abundance of igneous material, and theprobability that $\delta^{18}O$ of sedimentary material was lowered during high-grade metamorphism. R is expected to be high, because the quartzofeldspathic composition suggests that Rb is enriched. These rocks may be isotopically similar to the wall rocks measured by Fleck and Criss (1985).

Fleck and Criss (1985) proposed that samples from the South Fork of the Clearwater River formed by mixing a "primitive magma" with low R_i, and high Sr concentration with Precambrian wall rocks with high R_i and low Sr concentration. The wall rocks they analyized have $R_{(100)}$ .73 to .825 and Sr concentrations between 85 and 300. Sr concentration correlates well with $R_{(100)}$. These Sr concentrations are higher than those observed for Belt supergroup or metamorphosed Belt supergroup. Rb concentration is generalized as 135 ppm in rocks with $R_{(100)}$ of .75. $\delta^{18}O$ varies from 9.5 to 12. These rocks are described as biotite schist and gneiss, and are correlated with a similar unit dated as 1705 Ma by U-Pb on zircon (reported in Fleck and Criss, 1985). The schist and gneiss may be similar to schist and gneiss described by Lund (1984), which crops out in pendants within the central Idaho Batholith. These rocks may be Proterozoic basement rocks.

Fleck and Criss showed that an endmember with Sr isotopic composition and Rb and Sr concentrations similar to these rocks but containing higher $\delta^{18}O$, generated a model that fit their data. They suggested that $\delta^{18}O$ in the wall rocks had been lowered during metamorphism caused by the emplacement of the batholith. In their calculations, they used $\delta^{18}O$ of 12.5; this is too low for the porphyritic orthogneiss data. A hypothetical endmember with $\delta^{18}O$ of 15 is modeled for this data and is similar to the "early orogenic" endmember they discuss.

One continentally derived metamorphosed sedimentary rock from the study area was analyzed (Table 4.1). This rock is a silimanite schist, which crops out in a pendant of metamorphic rocks within the Payette River Tonalite. This sample has $\delta^{18}O$ and R similar to wall rocks studied by Fleck and Criss. Sr concentration is lower. Sr and Rb concentrations within these pendants are expected to be extremely variable, because rock types include carbonate, biotite schist and quartzite. The unit as a whole might be expected to have somewhat more Sr and less Rb than the sample analyzed. This rock is generally similar to those studied by Fleck and Criss.

All of the rocks described above are of appropriate composition to generate melts enriched in K_2O, SiO_2 and Rb. Calculated simple mixtures of these materials and low $R_{(100)}$, low $\delta^{18}O$, high Sr material similar to Hazard Creek complex epidote tonalites are shown in Figure 4.20. Endmember compositions used are given in Table 4.6

The porphyritic orthogneiss data are best modeled by mixtures containing 10% to 40% unmetamorphosed Belt sedimentary rock. Mixtures with the hypothetical "early orogenic" endmember proposed by Fleck and Criss also lie close to the data set. The estimated Paleozoic sedimentary rock does not contain enough radiogenic Sr. Measured wall rocks, as observed by Fleck and Criss, have $\delta^{18}O$ values that are too low, as do estimated Archean or Proterozoic gneisses.

This model and the work of Fleck and Criss both demonstrate the importance of a component with high $\delta^{18}O$. This material must have $\delta^{18}O$ near 15 to model the porphyritic orthogneiss. Values similar to this are typical of argillaceous sedimentary rocks. This material must also have high R, requiring that it either be an old material with enriched Rb or be derived from such a material. A range of $R_{(100)}$ values is allowable because Sr concentration and $R_{(100)}$ can both be adjusted. Therefore, both Belt sedimentary rocks with high R and low Sr and material similar to measured wall

Table 4.6: Endmembers used in mixing models, sources discussed in text.

	Rb	Sr	R(100)	O18
Estimated Hazard Creek magma	15	1000	.70350	8.0
Paleozoic sedimentary rock	200	140	.72400	15.0
Hypothetical endmember, after Fleck & Criss	135	185	.75200	15.0
Precambrian Belt sediments	165	65	.83800	15.0
Archean Basement	274	120	.98000	11.0
Metamorphosed Belt sedimentary rocks	165	65	.79800	11.0
Batholithic wall rocks	135	185	.75200	10.5

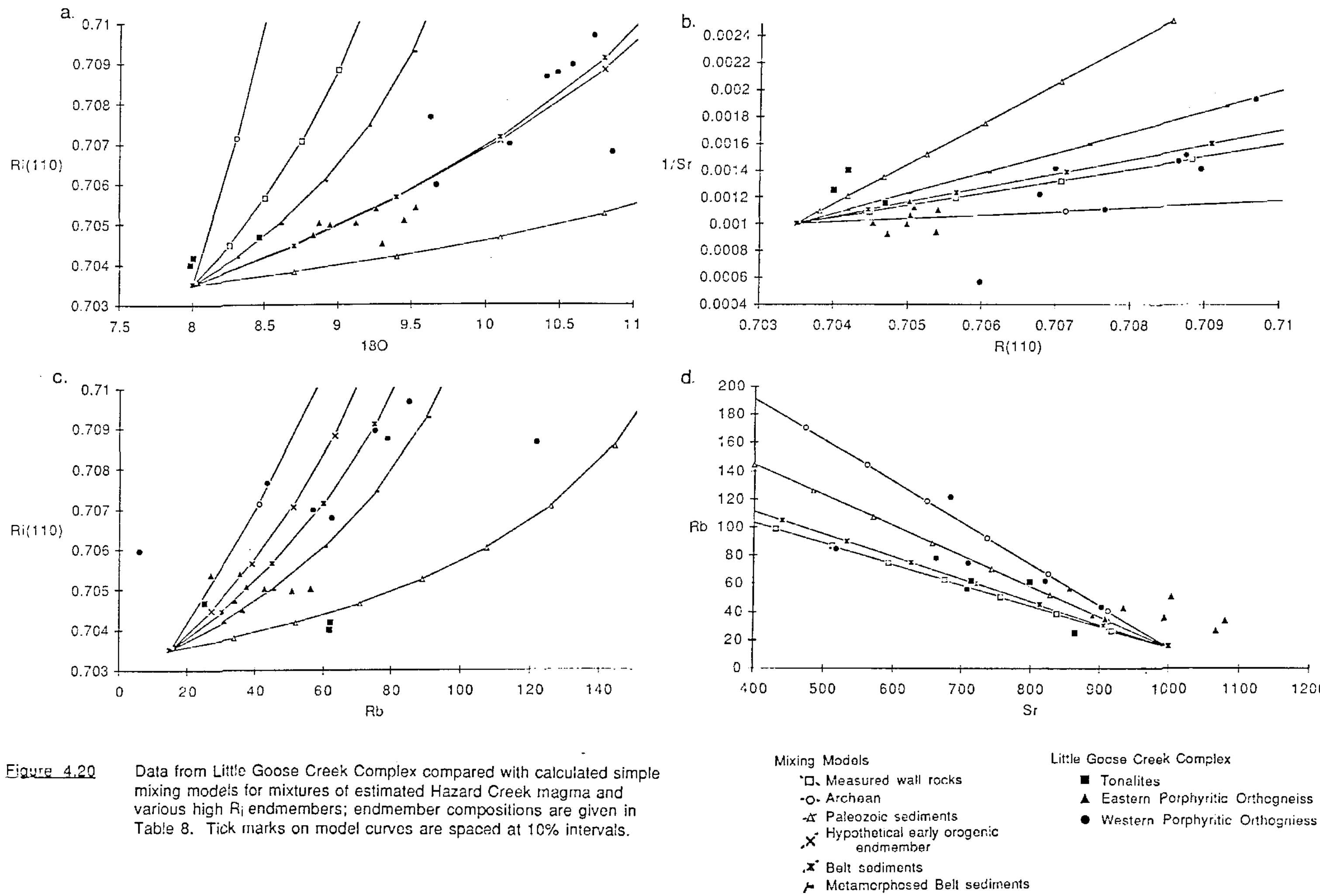


Figure 4.20 Data from Little Goose Creek Complex compared with calculated simple mixing models for mixtures of estimated Hazard Creek magma and various high R_i endmembers; endmember compositions are given in Table 8. Tick marks on model curves are spaced at 10% intervals.

rocks with lower R and higher Sr fit the data. This latter material must have had higher $\delta^{18}O$ than is preserved in the wall rocks studied by Fleck and Criss or wall rocks in this study.

The high $\delta^{18}O$ rocks must have been metamorphosed and melted without a significant reduction in $\delta^{18}O$. This is unlikely to happen if a large external source of fluid is available, as appears to be the case at shallow depth. Increased temperature will cause the establishment of large scale hydrothermal systems, and interaction with high temperature fluids will reduce the $\delta^{18}O$ of the rocks (Wickham and Taylor, 1985; Fleck and Criss, 1985). Wickham and Taylor demonstrated that the melting of sedimentary rocks in these circumstances produces melts with $\delta^{18}O$, similar to the exchanged sedimentary rocks. These data suggest that incorporation of metasedimentary material occurred at depths below the penetration of ground water. Abundant connate water is also unlikely.

The wall rocks exposed at the level of emplacement appear to have interacted with high temperature fluids, thereby reducing their $\delta^{18}O$. This may have occurred during the emplacement of the plutonic material or during metamorphism prior to batholith emplacement. The low $\delta^{18}O$ of exposed wall rock material suggests it is not a major component in the magmas. The lack of detailed correlation between the occurrence of arc-related and continentally derived wall rocks and the isotopic characteristics of the rocks further support this conclusion. This is in contrast to the observation by Fleck and Criss (1985) for the core of the batholith where R_i and $\delta^{18}O$ increase toward metamorphic pendants.

The origin of the sedimentary material at depth is unclear. Belt supergroup is permissible, as are sedimentary rocks with $R_{(100)}$ near .75 and Sr concentrations near 200. The abundance of Belt supergroup to the north and east suggests it may be the most plausible material.

The calculated mixing model is less sensitive to the isotopic composition of the high Sr endmember. A mantle-derived "arc magma" with $R_{(100)}$ near .7035 and $\delta^{18}O$ near 7 or a material similar to Hazard Creek complex epidote tonalties with $R_{(100)}$ of .704 and $\delta^{18}O$ near 8 are equally acceptable. Thus, it is not possible to tell if metavolcanic material may have been involved in the porphyritic orthogneiss magma. Variations in the Sr concentration of this endmember would cause significant variation in the Sr concentration of the mixtures, particularly in mixtures containing relatively small amounts of the high $R_{(100)}$ endmember. Porphyritic orthogneiss samples with low R_i show a range in Sr concentration which is not correlated with SiO_2, rock type, or plagioclase content, suggesting variations in the Sr concentration of the low R_i endmember between 700 and 1000 ppm. The Rb-Sr model suggests that Sr concentration in the uncontaminated magma may have been as high as 1200 ppm. This variability is similar to that seen in the Hazard Creek complex epidote tonalites. It may represent variable concentrations in the source material, assimilation of material with variable Sr concentration or a combination of processes.

The higher R_i and K_2O content of the western porphyritic orthogneiss relative to the Hazard Creek complex suggest that some high $R_{(100)}$ component is present in all samples. The high Na_2O, Al_2O_3 and low MgO seen in the Hazard Creek complex and western porphyritic orthogneiss are absent from the eastern porphyritic orthogneiss. These features may be masked by the abundance of high R_i material, or they may reflect an absence of metavolcanic component in the eastern porphyritic orthogneiss. Alternately, the geochemical nature of the low R_i component may be different in the eastern porphyritic orthogneiss than in the western porphyritic orthogneiss, reflecting a geochemical discontinuity at depth, in the source material.

Incomplete mixing of isotopically primitive and evolved material is the most characteristic feature of the porphyritic orthogneiss. The chemical similarity between the western Little Goose Creek complex and the Hazard Creek complex suggest that the low R_i magmas formed from similar source materials in both complexes. This material contained variable concentrations of Sr, possibly due to assimilation of metavolcanic material. Low R_i magma in the eastern Little Goose Creek complex may have had lower Na_2O, Al_2O_3 and higher MgO than in the western Little Goose Creek complex. The evolved endmember is required to have $\delta^{18}O$ near 15, suggesting it was a sedimentary rock incorporated at depth. Precambrian sedimentary rocks present in the middle crust would provide an appropriate material. Wall rocks do not appear to have contributed largely to the melts at the level of exposure.

It is difficult to know the original distribution of high and low R_i rocks in the pluton that formed the porphyritic orthogneiss. Deformation may have transposed chemical patterns or juxtaposed rocks from different depths in the pluton. The pluton may have been vertically zoned with high R_i rocks near the top. Whatever the cause, a remarkable chemical discontinuity has been preserved in the middle of this intrusive unit. In some fashion rocks that are dominated by different source materials have been juxtaposed. Either an abrupt boundary in the source materials or a discontinuity between regions of a pluton has been preserved.

The chemical relationship between Little Goose Creek complex tonalites and porphyritic orthogneiss is not clear. Field evidence indicates that the tonalites are older than porphyritic orthogneiss. The high Na_2O and Sr and low MgO that distinguish the Hazard Creek complex and western porphyritic orthogneiss also distinguish these tonalites, indicating they formed from similar magmas. This suggests they are part of the same magma generating event and are probably of Cretaceous age. The tonalites have higher R_i and $\delta^{18}O$ and lower Sr than is typical of samples in the Hazard Creek complex or is appropriate for simple mixtures with an evolved material. This may be an artifact of the data: if Rb has been metasomatically added to this material, calculated R_i is lower than true R_i. Alternately, these features are compatible with the incorporation of a larger component of spilitized metavolcanic material. Higher Y, Nb and HREE concentrations than observed elsewhere may be from this material. The low SiO_2 contents of the tonalites suggest no Precambrian sedimentary material was involved in their generation; however, this cannot be ruled out isotopically.

4.4.3.c Payette River Complex

The Payette River Tonalite is characterized by a lack of correlation between R_i, $\delta^{18}O$ and Sr concentration and by samples with high R_i and low $\delta^{18}O$. These features suggest mixing of three or more isotopically distinct components. The Payette River Tonalite has generally lower R_i and $\delta^{18}O$ than the eastern Little Goose Creek complex, even though screens of metamorphosed continental sedimentary rocks are more abundant. The Payette River Tonalite also has lower SiO_2 and K_2O than the eastern Little Goose Creek complex. These features suggest that a component low in SiO_2, K_2O and $\delta^{18}O$ but with moderate $R_{(100)}$ was involved in its formation. An ancient (> 1Ga) basaltic material present in the lower crust or upper mantle was proposed by Hill et al. (1986) to be the third component for tonalites with similar isotopic variations in the Peninsular Ranges Batholith, Mojave, and southern Sierra Nevada Batholith. This "subcontinental lithosphere" (Hill et al., 1986) would also be an appropriate third component for the Payette River Tonalite (Figure 4.21). The Payette River Tonalite data do not require R_i greater than .708. While Sr concentration in the Payette River Tonalite is not as anomalous as that in the Hazard Creek complex or Little Goose Creek complex, it is still higher than that seen in the Peninsular Ranges or Sierra Nevada batholiths in rocks with similar R_i. This suggests that the Sr rich source material inferred for the Hazard Creek and Little Goose Creek magmas was also involved in the generation of the low R_i, high Sr "arc magma" component in the Payette River Tonalite. The final endmember must again have high $\delta^{18}O$, suggesting the involvement of sedimentary rocks at depth. The Payette River Tonalite is suggested to form from a

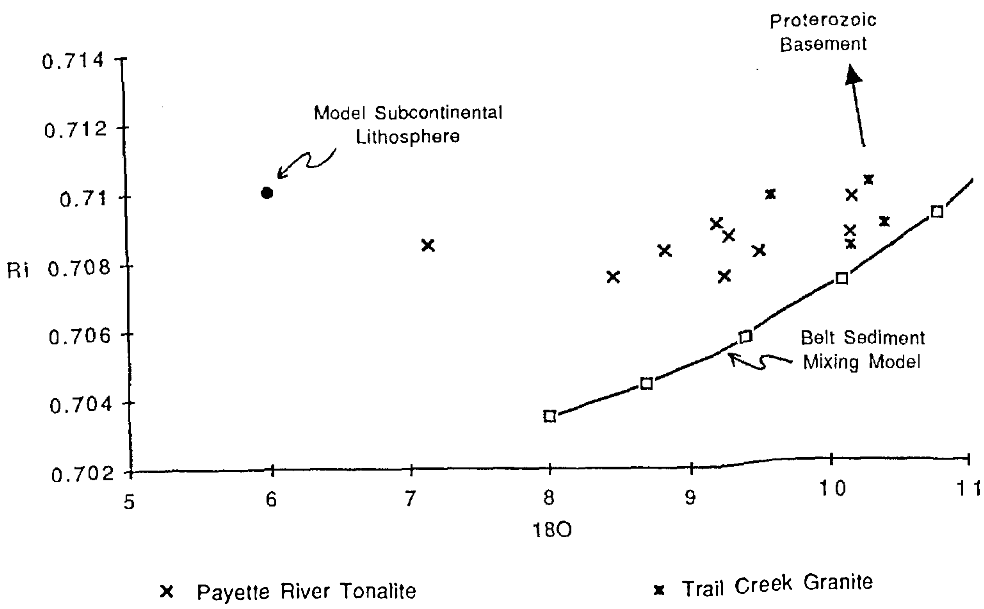


Figure 4.21 Data from the Payette River complex compared with calculated simple mixture of Belt sediment and estimated Hazard Creek Model. Data falls above model mixure in an array extending toward Model Subcontinental Lithosphere of Hill et al., 1986.

mixture of high Sr "arc magma", "subcontinental lithosphere", and high $\delta^{18}O$, high $R_{(100)}$ sedimentary material (Figure 4.21).

The Trail Creek Granite granites are chemically very similar to the eastern porphyritic orthogneiss and to other samples from the Idaho Batholith (Fleck and Criss, 1985). They are richer in K_2O and SiO_2 than Payette River Tonalite. R_i is sufficiently low to require a significant contribution of low $R_{(100)}$ material. Lower Sr concentration than observed in eastern porphyritic orthogneiss suggests this material may not be the high Sr "arc magma" seen to the west. The poor correlation between R_i and $\delta^{18}O$ in the Trail Creek Granite samples suggests that "subcontinental lithosphere" or other high $R_{(100)}$ intermediate $\delta^{18}O$ basement rocks may also be involved in the generation of the melts. Most samples from the interior of the Idaho Batholith (Fleck and Criss, 1985) fall to the left of the mixing line defined by the Little Goose Creek porphyritic orthogneiss (Figure 4.22, the "early orogenic" model of Fleck and Criss). The involvement of a third component with high R_i and low $\delta^{18}O$ would pull the data into this field. This component could be "subcontinental lithosphere." However, Proterozoic basement rocks of granitic composition are equally appropriate, isotopically. The change in rock composition from tonalite to granite at the western margin of the Payette River Tonalite may reflect a change from "subcontinental lithosphere" of basaltic composition to granitic lower crustal material, due to a change in the locus of magmatism, the crustal thickness, or the crustal character. The study of tonalites along the eastern margin of the batholith would help to evaluate these options.

4.4.4 Discussion of the Distribution of Rock Types at Depth

These data provide several constraints on the distribution of rocks at depth. There is no evidence of isotopically evolved material in the Hazard Creek complex. This suggests a complete absence of cratonal basement and cratonally derived sedimentary material from the rock column traversed by these magmas. Elevated $\delta^{18}O$ in the Hazard

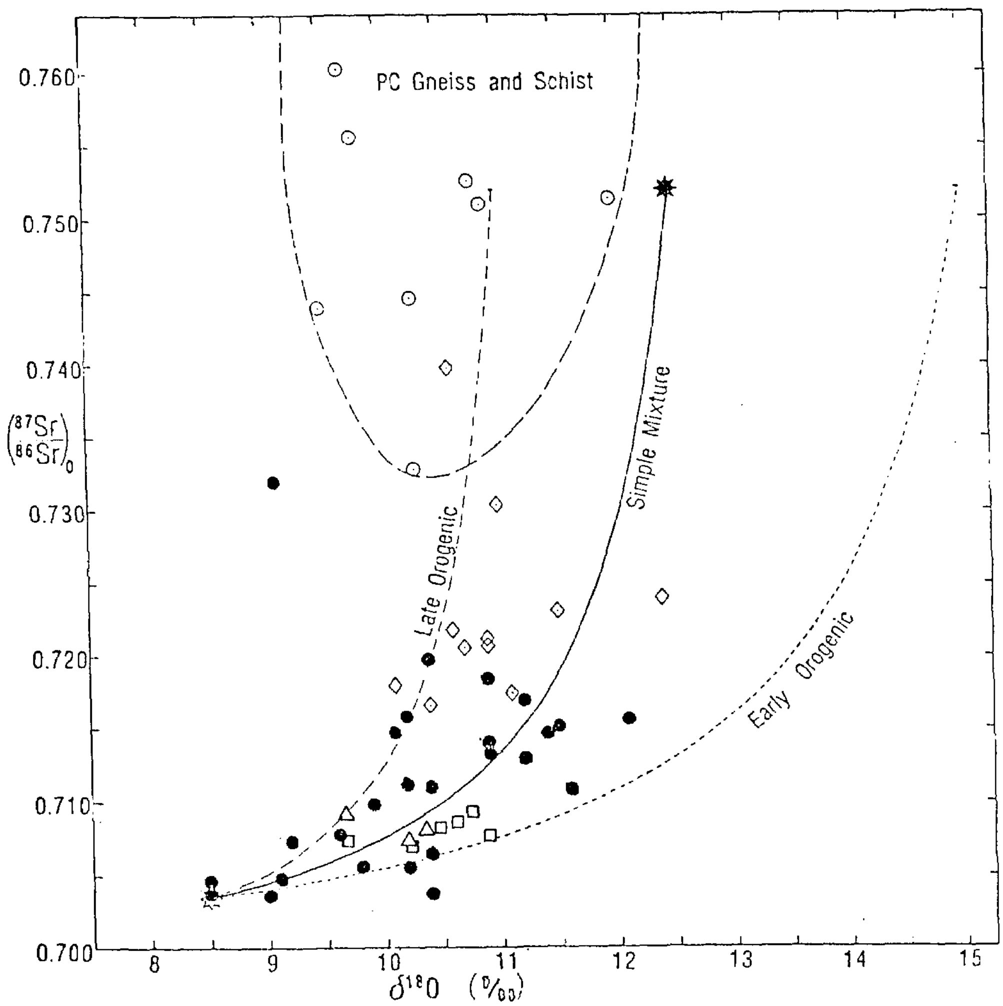


Figure 4.22 Data for eastern porphyritic orthogneiss (open squares) and Trail Creek Granite (open triangles) compared with pluton data for South Fork of the Clearwater River (solid dots, Fleck and Criss, 1985) showing range of granite samples from the core of the batholith. Solid line is mixing model of Fleck and Criss, 1985. Open circles are high grade wall rocks measured by Fleck and Criss. Diagram modified from Fleck and Criss, 1985.

Creek complex suggests the incorporation of spilitized arc volcanics similar to those exposed locally. Variable Sr concentration and high Na_2O, Al_2O_3 and LREE in the Hazard Creek epidote tonalites may reflect variable source material or the incorporation of spilitized arc volcanics. Hazard Creek epidote tonalites have a distinctive high Na_2O, high Al_2O_3, high Sr, and low MgO character. A similar component appears to be involved in the formation of the western Little Goose Creek complex, suggesting that the Little Goose Creek complex formed adjacent to the Hazard Creek complex.

The westernmost signature of evolved cratonal material is in the Little Goose Creek complex. The porphyritic orthogneiss contains both isotopically evolved and primitive material. It is modeled as a mixture of Precambrian sedimentary rock with high $\delta^{18}O$ and R_i and "arc magma" with low R_i, and $\delta^{18}O$ similar to the Hazard Creek complex epidote tonalites. The high $\delta^{18}O$ required for the evolved material suggests that the sedimentary rocks were incorporated at depth indicating a sharp boundary in the middle-to-lower crust between metamorphosed Precambrian sedimentary rocks and isotopically unevolved crust. The boundary must lie between the Hazard Creek and Little Goose Creek complexes, or within the Little Goose Creek complex. The abrupt change in geochemistry between the western and eastern porphyritic orthogneiss can be modeled as an abrupt change in the proportions of the endmembers. This may reflect a very abrupt boundary in the crust or lithosphere beneath the complex. Alternately, it may reflect the juxtaposition of different portions of a chemically stratified pluton.

The Payette River Tonalite is inferred to represent a mixture of "arc magma," Precambrian sedimentary material, and a third component derived from ancient basaltic material with isotopically evolved Sr and low $\delta^{18}O$. The low $\delta^{18}O$ required for this component suggests that it is at depth and has not interacted with surficial water. This component, which is also recognized in the northeastern Peninsular Ranges Batholith, the Mojave and the central Sierra Nevada Batholith, has been proposed to be ancient basaltic material that resides in the lower crust or upper mantle of the craton (Hill et al., 1986). The absence of its signature in the porphyritic orthogneiss suggests that either this material is missing at depth beneath the Little Goose Creek complex or that it was imported into the area between the time of Little Goose Creek complex and Payette River Tonalite emplacement. The former hypothesis is preferred as the component appears to be associated with the craton which is known to have been present during the emplacement of the Little Goose Creek complex. Thus a second boundary at greater depths is inferred between "subcontinental lithosphere" and isotopically primitive lithosphere. This boundary is between the Little Goose Creek complex and the Payette River complex. There is no evidence of Archean or Proterozoic basement in the isotopic signature of either the Little Goose Creek complex or the Payette River Tonalite. The cratonal section present in the area may consist of metamorphosed sedimentary rocks deposited on a thinned cratonal margin in a Precambrian miogeocline.

The composition of the Trail Creek Granite is consistent with mixtures of sedimentary material, "arc magma" and granitic lower crustal material. The change from tonalitic to granitic composition may mark a change in the composition of the lower crustal material or may reflect changes in the conditions or loci of melting.

4.5 SUMMARY

The major and trace element geochemical data and isotopic data all show that the intrusive complexes recognized on the basis of field mapping are geochemically distinct. Data from each complex define distinct fields on variation diagrams for both major and trace elements, and isotopic systematics in each complex are distinct. All magmas appear to be arc related and have formed in response to subduction processes occurring west of the Blue Mountain Arc.

The porphyritic orthogneiss of the Little Goose Creek complex contains a fundamental geochemical discontinuity across which the geochemical character of the rocks changes dramatically. Rocks west of this boundary, in both the Hazard Creek complex and Little Goose Creek complex, are characterized by relatively low R_i, $\delta^{18}O$, and MgO, and high Sr, Na_2O and Al_2O_3. Rocks east of this boundary in both the Payette River complex and the Little Goose Creek complex have generally high R_i and more moderate MgO, Sr, Na_2O and Al_2O_3. This boundary marks a change in the relative importance of material associated with the continent. It occurs within the same complex that contains the boundary between oceanic-arc and continental supracrustal rocks.

Isotopic modeling suggests that the geochemical differences between the complexes are due to differences in the lithospheric column beneath them. The Hazard Creek complex is modeled as an "arc magma" generated at depth in the mantle, which has interacted with spilitized oceanic-arc volcanics at lower crustal depths. The Little Goose Creek complex is inferred to have high $\delta^{18}O$ Precambrian sedimentary rocks at mid-to-lower crustal depths beneath some or all of the complex. Payette River Tonalite is inferred to have Precambrian sedimentary material and ancient basaltic "subcontinental lithosphere" beneath it. Payette River Granite may have Precambrian sedimentary rock, and Precambrian granitic basement, "subcontinental basement," or both beneath it. These relationships are depicted in Figure 4.23. These models are consistent with the variations in major element chemistry of the complexes.

Both the abrupt change in geochemical character of the intrusives and the isotopic modeling indicate that the lithospheric character changes dramatically and abruptly in the 10 km interval occupied by the Little Goose Creek complex. The correspondence between this abrupt geochemical change and the change in supracrustal rocks exposed as sheets within the intrusives indicate that the oceanic arc-continent boundary is a steeply-dipping structure juxtaposing an oceanic-arc and its relatively primitive lithosphere against continental sedimentary rocks and their more evolved lithosphere. There is no evidence of Precambrian cratonal basement in rocks west of the Trail Creek granite.

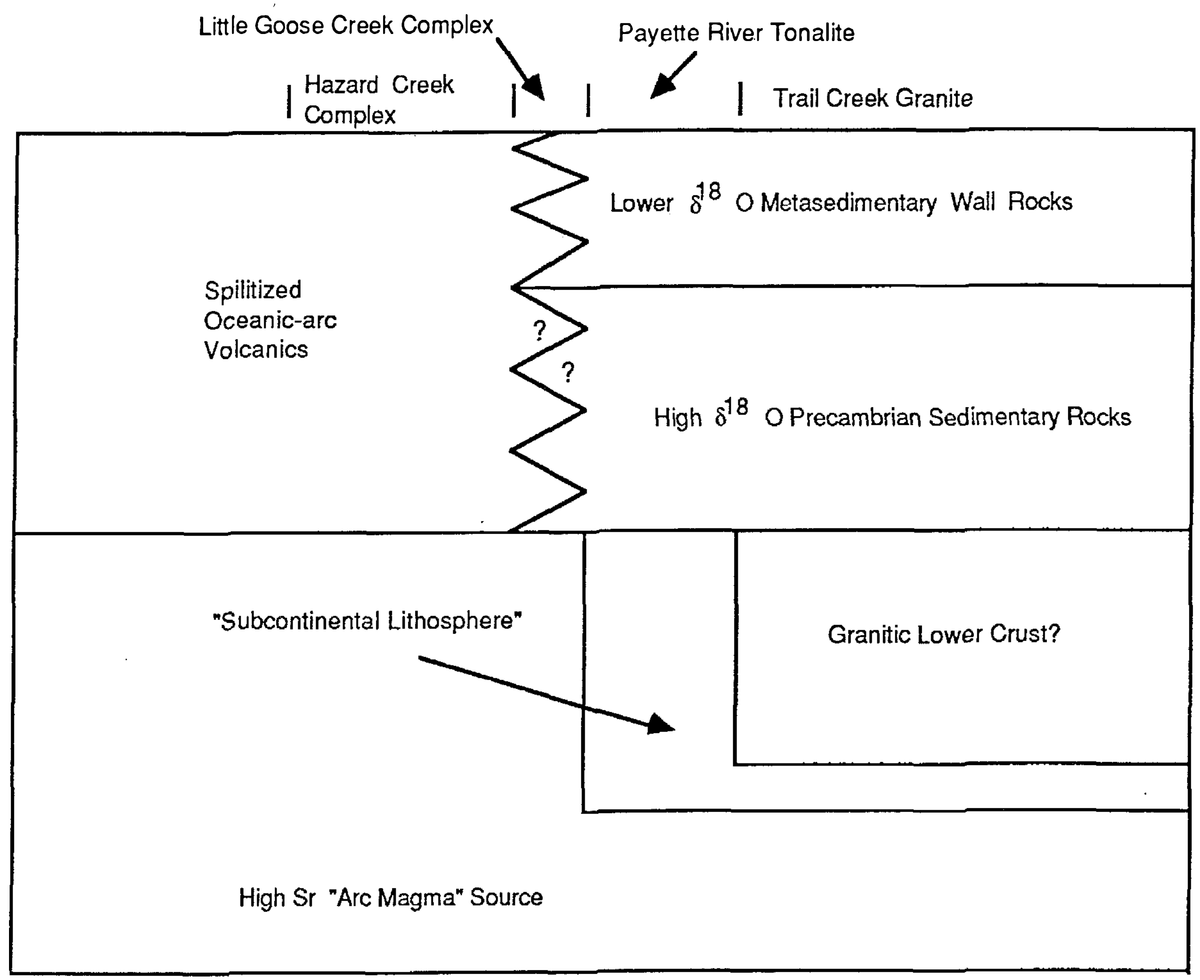


Figure 4.23 Schematic diagram showing suggested distribution of components in lithosphere.

CHAPTER 5: CONCLUSIONS

In the previous chapters, the geology and geochemistry of the oceanic arc-continent boundary in the western Idaho Batholith near McCall have been described in detail. The supracrustal boundary is preserved as a change in the lithology of pendants and screens within Cretaceous intrusives, from mafic layered gneiss inferred to be metamorphosed oceanic-arc material in the west, to biotite and biotite-sillimanite schist, quartzite and calc-silicate gneiss inferred to be metamorphosed cratonally-derived sedimentary rocks in the east. No structures related to the initial juxtaposition of these lithologies prior to intrusion in the Cretaceous have been found.

During Cretaceous time the boundary was intruded by subduction related magmas and multiply deformed. Intrusion began west of the boundary with the emplacement of the Hazard Creek complex. The oldest intrusives in the complex, ranging in composition from diorite to trondhjemite, were metamorphosed to amphibolite facies and isoclinally folded. This is the oldest deformation preserved in the area.

More voluminous magmatism began with the emplacement of the epidote tonalites of the Hazard Creek complex beginning prior to 118 Ma. These intrusives, which range in composition from diorite to trondhjemite with lesser granite, were generated at depth leaving garnet or hornblende but no plagioclase in the residue. The presence of primary epidote suggests crystallization at pressures of 8 kb or more (Zen and Hammarstrom, 1984). The elevated $\delta^{18}O$ of the rocks may reflect incorporation at depth of spilitized oceanic-arc rocks, similar to those that now form the wall rocks to the intrusions. The intrusives are typical of the tonalite-trondhjemite series seen in the western Cordillera within the eugeosyncline, and contain no component of isotopically evolved crustal material. Subduction must have begun west of the Blue Mountain Arc by this time in order to generate these magmas.

The Hazard Creek complex was deformed during a period broadly coeval with emplacement of the plutonic rocks. Deformation, which was localized on the eastern side of the complex, is characterized by steep north-south striking foliations, down-dip lineations and steeply-plunging ductile fold axes. These structures are inferred to have formed in response to east-west compression, which was accommodated by flattening and vertical flow. The localization of deformation on the eastern side of the complex adjacent to the current location of the oceanic arc-continent boundary suggests this boundary was represented by a discrete structure during deformation and that the magmas were emplaced adjacent to this structure. Localization of stresses may have been the result of buttressing against the thicker continental crust or due to the existence of an older structural zone of weakness.

The boundary between oceanic-arc and continental supracrustal rocks was intruded by the porphyritic orthogneiss protolith at approximately 111 Ma. This unit, inferred to have been emplaced in a single intrusive episode, contains an abrupt chemical discontinuity. Rocks in the western portion of the complex are dominated by a component similar to the Hazard Creek complex but have incorporated some higher SiO_2, K_2O, Rb, R_i, $\delta^{18}O$ material. Rocks in the eastern portion of the complex are dominated by a component with high R_i and $\delta^{18}O$. Isotopic modeling suggests these two components are "arc magma" generated at depth and Precambrian sedimentary rocks. The incorporation of Precambrian sedimentary material probably occurred at middle-to-lower crustal levels, indicating that a boundary between isotopically unevolved and Precambrian sedimentary materials lies at depth either within or to the west of the Little Goose Creek complex. Compressional deformation localized along the arc-continent boundary may have continued through the time of emplacement of the porphyritic orthogneiss.

The contact between the Hazard Creek complex and Little Goose Creek complex is presently strongly deformed; however, three lines of evidence suggest that the complexes formed adjacent to one another during an extended interval of intrusion and deformation. Preliminary age data suggest that the porphyritic orthogneiss is only slightly younger than the younger members of the Hazard Creek complex. Chemical composition suggests that similar high Sr, high Al, high Na, low MgO "arc magma" contributed largely to intrusives in both complexes. Deformation characterized by similar structures is recognized in both the Little Goose Creek complex tonalites and the eastern Hazard Creek complex.

Intrusion of the porphyritic orthogneiss may have been followed by a hiatus in magmatism which resumed with the emplacement of the Payette River Tonalite and Granite approximately 90 Ma. The Payette River Tonalite is tentatively correlated with the Coarse Grained Tonalite Orthogneiss, indicating the Payette River Tonalite was emplaced adjacent to the Little Goose Creek complex. The Payette River Tonalite is similar to rocks from calc-alkaline Cordilleran batholiths formed in response to subduction. Isotopic systematics suggest three components were involved in its formation: (1) high Sr, low R_i, low $\delta^{18}O$ "arc magma", (2) low Sr, high R_i, high $\delta^{18}O$ Precambrian sedimentary material and (3) high R_i, low $\delta^{18}O$, "subcontinental lithosphere." "Subcontinental lithosphere" is present in tonalites from the eastern Peninsular Ranges and central Sierra Nevada batholiths and appears to be characteristic of magmas emplaced near the craton (Hill et al., 1986). The inferred presence of this component only in the Payette River Tonalite indicates that a boundary, between "subcontinental lithosphere" and isotopically unevolved lithosphere at lower crustal or upper mantle depths, is present between the Little Goose Creek complex and the Payette River complex.

During emplacement, a strong north-south-striking steeply-dipping igneous foliation and steeply-plunging igneous lineation developed in the Payette River Tonalite. Metasedimentary wall rocks were transposed and folded. These features again suggest that the crust near the oceanic arc-continent boundary was accomodating east-west compressive stresses by flattening and vertical flow.

Following emplacement of the Payette River Tonalite, a mylonitic deformation centered on the oceanic arc-continent boundary affected rocks in all three complexes. Steep north-south striking foliations, down-dip lineations and steeply-plunging fold axes indicate that during this deformation east-west shortening was again accommodated by flattening and vertical flow. The lack of a consistent vergence to structures in the zone and the similarity in metamorphic grade of rocks on either side of the zone suggests that, on the large scale, flattening may have been more important than simple shear. A $^{40}Ar/^{39}Ar$ age of 87 Ma (Snee, per. comm.) may suggest that deformation occurred shortly after the emplacement of the Payette River Tonalite.

Magmatism continued to the east with the emplacement of the voluminous core of the Idaho Batholith, represented locally by the Trail Creek Granite around 60 to 75 Ma. These intrusives were passively emplaced across the regional grain and contain little or no foliation. While this change in intrusive style may be partly attributable to the higher viscosity of the granitic melt, it also suggests that deformation in the area had waned. Isotopic systematics suggest these magmas also had three major components: "arc magma", Precambrian sedimentary material, and a third component with high R_i and intermediate $\delta^{18}O$ - possibly granitic Precambrian crystalline basement material. Evidence for Precambrian crystalline basement at depth is absent from the other intrusives in the study area.

During Cenozoic time, brittle faulting led to the formation of prominent north-south striking valleys. These were filled with Columbia River Basalt during Miocene time while they were actively deforming (Hooper, per. comm.). Deformation continued after the cessation of volcanism but ceased prior to the Pleistocene glaciation of the area.

5.1 THE OCEANIC ARC-CONTINENT BOUNDARY

This study shows that the oceanic arc-continent boundary is a steeply dipping boundary juxtaposing two lithospheric blocks. Intrusive rocks change dramatically in geochemical character across the Little Goose Creek complex, which contains the supracrustal boundary. Isotopic modeling suggests that the boundary between oceanic-arc and continental supracrustal rocks coincides closely with a boundary at middle-to-lower crustal levels between metamorphosed Precambrian sedimentary material and isotopically primitive crust and is within 10 km of a boundary at deeper levels between more evolved "subcontinental lithosphere" and isotopically primitive lithosphere. The juxtaposition of lithospheric blocks must have occurred prior to emplacement of the Little Goose Creek porphyritic orthogneiss protolith approximately 111 Ma. The geochemical similarity and the closeness in age between the Hazard Creek complex and the Little Goose Creek complex suggest that they formed above the same subducting slab. This indicates that the oceanic arc-continent boundary had formed and subduction west of the Blue Mountain Arc had begun prior to 118 Ma. This conclusion is supported by the localization of deformation in the eastern Hazard Creek complex, apparently along a pre-exisiting structure, during emplacement of the complex beginning prior to 118 Ma.

No structural evidence of the initial formation of the boundary is recognized. It seems most probable that the boundary was formed by a lithospheric scale transform fault similar to the San Andreas Fault or the Mojave-Sonora megashear (Silver and Anderson, 1974, 1983). This fault could have been active during any of the several periods of pre-Cretaceous strike-slip faulting proposed for the western Cordillera (Saleeby and Gehrels, 1987; Oldow et al., 1984; Silver and Anderson, 1974, 1983). A transpressive structure (Lund, 1984; Lund and Snee, 1988) might form a boundary of the appropriate shape; however, it must have been active prior to 118 Ma.

Alternatively, rifting might form a steep edge to the continent such that subsequent accretion of the Blue Mountain arc would form the abrupt boundary observed today. As described in Chapter 1, the absence of a subduction complex suggests the Blue Mountain Arc was not accreted to the continent by westward dipping subduction under the arc (Lund, 1984). However, the Blue Mountain Arc may have formed as a fringing arc to North America. Collapse of the back-arc basin separating it from the continent might have been accomplished in such a manner that little oceanic crust was preserved along the boundary. The Jurassic deformation and amalgamation of the Blue Mountain Arc, described in Chapter 1, might record such an event.

5.2 CRETACEOUS DEFORMATION

The oceanic arc-continent boundary in the study area was intruded and deformed, perhaps episodically, during a period that began prior to 118 Ma and continued until after 90 Ma. Deformation during this interval was characterized by north-south striking, steeply dipping foliations, and steeply plunging lineations and fold axes, suggesting response to east-west compression by flattening and vertical flow.

The oldest deformation in the study area is the metamorphism and folding of layered mafic gneisses and crystalloblastic orthogneisses. The layered mafic gneisses within the Hazard Creek complex and Little Goose Creek complex are tentatively correlated with the Pollock Mountain Amphobolite. Post-kinematic garnets in this unit record burial to depths of 25 to 30 km (Selverstone et al., 1987). Similar garnets within the Hazard Creek complex suggest that some deformation and metamorphism took place during the burial event. Older crystalloblastic orthogneisses in the Hazard Creek complex may have been emplaced during or after this event. Primary magmatic epidote in the Hazard Creek complex suggests emplacement at 25 to 30 km depths (Zen and Hammarstrom, 1984) similar to those recorded in the garnets.

Deformation of the Hazard Creek and Little Goose Creek complexes, broadly coeval with plutonism, may also be broadly coeval with metamorphism and upthrusting seen in the oceanic-arc rocks along the oceanic arc-continent boundary (Hamilton, 1963a; Onasch, 1987; Aliberti and Wernicke, 1986a,b; Lund and Snee, 1988; Myers, 1982; Figure 1.3). Vertical structures formed in the study area may represent the deeper roots of thrust zones which brought deep seated oceanic-arc rocks toward the surface (e.g., the Pollock Mountain Fault, Aliberti and Wernicke, 1986a,b; Figure 1.3). The relative age of these structures and those in the study area is unconstrained; however, they both reflect response to broadly east-west compression by vertical movement (Aliberti and Wernicke, 1986a,b). Metamorphic minerals with Ar^{40}/Ar^{39} ages between 118 and 101 Ma in the oceanic-arc rocks have been interpreted to indicate that metamorphism and deformation prior to thrusting occurred during this interval (Lund and Snee, 1988). However, the inclusion of metamorphosed layered mafic gneisses in rocks 118 Ma and older indicates that deformation and metamorphism began prior to 118 Ma in the study area. Metamorphic mineral ages may reflect metamorphism and cooling during uplift. Metamorphic mineral ages younger than the emplacement age of the porphyritic orthogneiss protolith suggests that deformation continued through much of the interval between the emplacement of the Little Goose Creek and Payette River complexes.

Mylonitic deformation similar to that centered on the Little Goose Creek complex is found in a narrow zone as far north as Orofino (Fig. 1.3; Lund, 1984; Lund and Snee, 1988; Myers, 1982; Hoover, 1986; Strayer, 1987). This zone everywhere contains an abrupt change in isotopic properties of the intrusive rocks (Criss and Fleck, 1987). The change in supracrustal rocks occurs within a mylonitized porphyritic orthogneiss similar to that in the Little Goose Creek complex on the South Fork of the Clearwater River (Hoover, 1986). These data suggest a similar steep lithospheric boundary has localized deformation characterized by vertical movement along the entire length of the oceanic arc-continent boundary. Ar^{40}/Ar^{39} data from intrusives in this zone are interpreted to indicate deformation between 89 and 78 Ma (Lund and Snee, 1988) similar to that in the study area.

The Cretaceous deformation along the oceanic arc-continent boundary has been interpreted as recording the juxtaposition of oceanic-arc against continent by a transpressive structure (Lund, 1984; Lund and Snee, 1988). However, the data presented in this work indicate that the boundary must have existed prior to most if not all of this deformation. All of the Cretaceous deformation is here interpreted as the response to compression of a pre-existing vertical lithospheric boundary. All of the deformation is inferred to be localized along this boundary, as was clearly the case for the late Cretaceous mylonitic deformation. The well-defined vertical boundary juxtaposing lithospheric blocks of different character is envisioned as responding to compression by vertical upthrusting and flow rather than by thrusting and nappe development. The response is similar to that of a strike-slip fault to a compressional component of stress, in which case, blocks along the fault are thrown up once a through-going fault system is established (Lowell, 1972; Sylvester and Smith, 1976). Cretaceous deformation may reflect east-west compressive stresses generated by subduction to the west and by the emplacement of plutonic material near the boundary. The presence of plutonic material may have facilitated deformation by acting as a lubricant along the boundary (Hollister and Crawford, 1986).

The Cretaceous deformation in the study area is not interpreted as transpressional because there is no evidence for a horizontal component to the lineation. This is generally true along the north-south trending segment of the oceanic arc-continent boundary (Aliberti and Wernicke, 1986a; Lund, 1984; Myers, 1982). In cases where transcurrent or transpressional motion are documented, lineations have a horizontal component (e.g., Spitsbergen: Mahre, 1984; Harland, 1971; Alpine Fault:

Sibson et al., 1981). In addition, plutonism is not a characteristic of transcurrent faulting.

While no other lower-to-middle Cretaceous deformation is known in the Blue Mountain Arc, deformation during this period was widespread within the continent. As described in Chapter 1, during this time, prior to the intrusion of the core of the Idaho Batholith, Precambrian sedimentary rocks were metamorphosed to amphibolite facies and deformed along the entire northern margin of the batholith. West-directed thrusting of Precambrian and Paleozoic sedimentary rocks commenced along the eastern margin of the batholith in south-central Idaho and western Montana. The Lewis and Clark line (Figure 1.1) was an active zone of strike-slip faulting. Sedimentation recording uplift of central Idaho also commences at this time (Scholten, 1973). The compressional deformation recorded along the arc-craton boundary was part of a larger regional episode of compressive deformation. This regional deformation continued through the Late Cretaceous and may have continued into the Eocene (Hyndman et al., 1975; Ryder and Scholten, 1973; Ruppel et al., 1981). It ceased prior to the Eocene Challis igneous event (Ruppel, 1978).

A deformation of this scale may have resulted from changes in plate interactions. The compressional regime may have been caused by a change in the rate of plate convergence or the accretion of an allocthonous terrane west of the Blue Mountain Arc may have upset plate interactions (this terrane must be removed by later transcurrent faulting). The compressional stresses are not due to the accretion of the Wallowa/Seven Devils arc to the Blue Mountain Arc, or to the accretion of the Blue Mountain Arc to the continent, because the entire Blue Mountain Arc must have lain between the subduction zone and the zone of plutonism along the boundary at this time.

5.3 PETROGENESIS

The preservation of the oceanic arc-continent boundary within intrusive rocks provides an ideal situation for evaluating some aspects of magma evolution and emplacement. The occurrence of an abrupt change in the geochemistry of intrusives across the oceanic arc-continent boundary indicates that the magmas were generated and transported from source regions that were not much wider than the dimensions of the intrusive bodies. All of these bodies contain components derived from the crust through which they ascended; isotopic systematics suggest that even the Hazard Creek intrusives, emplaced through relatively primitive crust, contain a component derived from that material. The preservation of an abrupt discontinuity in the chemical composition of the porphyritic orthogneiss indicates that it was not a well-mixed magmatic body. The abrupt discontinuity may be a reflection of an abrupt discontinuity in the crustal material at depth or may reflect chemical stratification of the magma body. Assimilation of local wall rocks does not appear to have been an important process anywhere in the study area.

5.4 REPRISE

Combined field mapping and geochemical studies have provided new insight to the nature, formation and subsequent history of the oceanic arc-continent boundary. This boundary appears to reflect the juxtaposition of lithospheric scale blocks along a steeply dipping structure. No structural evidence for the formation of this boundary was found. Plutonism related to subduction under the amalgamated Blue Mountain Arc and continent, and deformation of broadly Cretaceous age, have modified the boundary. This deformation is interpreted to be the response of a pre-exisiting lithospheric boundary to compressive stresses. The presence of the lithospheric boundary is recorded in the geochemistry of the plutonic rocks, which reflect the interaction of "arc magma" with the crustal section traversed.

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APPENDIX A: TABULATED ESTIMATED MODAL DATA

The following tables list modes estimated from thin sections for samples shown on Plate 3. The modal mineralogy was estimated from thin sections using a spot chart; all samples except those with names containing "z" were stained for alkali feldspar. Comparison of estimated modes, with norms for those samples that were analyzed chemically, suggest that quartz was often overestimated and plagioclase underestimated by ten to twenty percent, particularly in samples from the Little Goose Creek complex where plagioclase is untwinned. Alkali feldspar appears to be overestimated in samples from the Trail Creek Granite. Most alkali feldspar in the Little Goose Creek complex is in large megacrysts, thus modes estimated from thin sections tend to underestimate the alkali feldspar present in the unit. Accessory phases are occasionally overestimated: an estimate of 2% or 3% was used to indicate a very abundant phase, and may not truly reflect the volume percent. These tables are included because in spite of the difficulties in estimating the modes, they contain valuable information on the mineralogy and texture of the samples, and the distribution of these features throughout the units.

Rock types are formed from the following abbreviations with minerals listed in decreasing order of abundance.

b:biotite	h:hornblende	e:epidote	m:muscovite	g:garnet
t:tonalite	gd:granodiorite	gr:granite	qd:quartz diorite	
dio:diorite	gab:gabbro	gn:gneiss	my:mylonite	
amphib:amphibolite				

The following abbreviations are used to give textural information.

Q,P,MA:	mylonitic texture with (Q) quartz, (P) plagioclase, (MA) mafic minerals deformed. Mineral abbreviation is capitalized if all grains of that type are deformed, lower case if some are deformed.
hypi:	hypidiomorphic texture
allo:	allotrimorphic texture

xtal:	crystalloblastic texture
mi:	modified igneous texture as defined in text
mmi:	extensively modified igneous texture as defined in text
mmi/xtal:	extensively modified igneous texture where more than 50% of the grains have recrystallized.
und p:	undulatory extinction in plagioclase grains
xtal ma:	some mafic minerals have crystalloblastic habit
myr:	abundant myrmekite
relict p:	porphyroclasts of plagioclase

The following other abbreviations are also used.

oss p:	oscillatorily zoned plagioclase
nor p:	normally zoned plagioclase
cg:	coarse-grained
fg:	fine-grained

Table A-1
Hazard Creek Complex
Jack's Creek Pluton

		Modal %										
Sample	Rock Type	hb	bio	plag	ksp	qtz	ep	op	Accessories	An	Texture	Comments
83c48	bhet	10	10	55		25	1	1	al,ap,zir	35	Q,p,ma	
85k13	bhet	3	7	60		25	5	tr	al,sp,zir	28+	Q,p	
83c42	bht	5	10	60		20		tr	al,sp 3%,zir	18?	Q	
83c50	bhtmy	5	10	60		20		tr	al,sp 2%,ap,zir	25?	Q,P,MA	
83c51	hbdio	15	5	75		3		tr	al,sp 1%,ap,zir	26	Q,p	
83z9	hbet	10	8	45		30	4	tr	sp,ap,zir,op	28	Q,p	
85k29	hbet	8	7	67		15	3	1	al,ap,zir	32	Q,p	
83c43	hbetgn	15	5	55		20	4		sp,ap,zir	30	Q,p,ma	
83c44	hbetgn	6	6	55		30	4	1	sp,ap,zir	32	Q,p	
84c24	hbetmy	7	3	71		18	1	tr	al,sp,ap,zir	30	Q,p,ma	ig and my fabrics parallel
84c25	hbetmy	5	3	75		12	5	tr	al,ap	32	Q,P,MA	cg ma augen
83z3	hbt	8	7	83		2		tr	al,sp,ap,zir	28?	allo	
85k34	begd		7	50	15	25	3	tr	al,sp,ap	and?	Q,P,MA	
84j14	betmy		15	60		22	3	tr	ap	30	Q,p,ma	
85k33	bgdmy	1	10	55	9	25		tr	al,sp,ap,zir		Q,P,MA	
85c31	hbetmy	12	8	70		8	tr	2	sl,sp,ap,zir	37	Q,P,MA	much annealing
85c30	hbtmy	7	11	50	tr	30	2	tr	ap,zir	28	Q,P,MA	
84z24	hbgab	80	3	13		2			sp 2%	34		gabbro pod
84z25-b	hbgab	25	15	60		tr	1	tr	sp,zir,ru	35		gabbro pod
84z25-a	hgabgn	70		15		15		tr	al,sp 2%	35		gabbro pod, cpx cores hb

Table A-2
Hazard Creek Complex
Epidote-Biotite Tonalite

		Modal %											
Sample	Rock Type	hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An	Texture	Comments
81k228	bemt		5	70	5	20	2	3			30	mmi	nor p
83k10	bemt		7	77	tr	15	tr	1		ap,zir	30	mi, Q	oss p
81k243	bet		5	77		15	tr	3			30	mi	oss p
81k244	bet		12	62		23		3			30?	mi	oss p
81k270-a	bet		10	80		10	tr	1		ap,zir	30	hypi	oss p
83k11	bet	tr	8	60		30		2		al,th	35	hypi, Q	nor p
85k35	bet		7	60	3	25		4		al,sp 1%,ap,zir,th	28	Q,p,ma	
81k210	bmet		10	66		20	1	3	tr		38	hypi	
82k59	bmet		7	40		45	2	3		al,sp,ap,zir	32	mmi	nor p
82c32	bmet		7	47		40	2	3	tr	sp,ap,zir	30	hypi, Q	nor p
83k26	bmet		3	70		20	3	3	tr	ap	36	mmi	nor p
83c77	bmet		12	50		35	2	3	tr	al,ap,zir	29	mi	nor p
83c64	bmet		7	50		40	1	2		ap	28	hypi	
83c63	bmet		5	45		45	2	4		ap,zir	35	mi	oss p
83z1	bmetmy		10	60		35	1	4	tr	al,ap	34	Q,p,ma	
84c6	mbt		2	55	tr	40	3	tr		ap,zir	20?	hypi	oss p
82k60	bhet	1	10	50	35	35		4	tr	sp,ap,zir	32	hypi	oss p
84c7	bhetgn	3	7	70		15		5		sp,zir	38	mmi	oss p
81k272	bmetgn		3	60	4	20	2	1			28	mmi	nor p
81k130	bgdmy		3	45	17	35		tr	tr	ap,gar	and	Q,P,MA	
85c25	bmet		5	50		42	1	1	tr		35	mi,Q	
86c5	bmet		5	60		32	1	2			28	mmi	
86z5	bmet		10	42		45	1	2	tr	al	28	mi	
85k27	bmet		5	60	tr	30	1	3		gar	31	mmi	
86k19	bmet		7	70		20	tr	3			28	mmi	oss plag
86c6	hbet	8	2	77		10		3	1	al	34	mmi/xtal	oss plag
86c7	het	7	tr	50		40		1	2		?	mmi	
Inclusions and screens													
86c3	bgsqpgn		10	35		50	1			gar 2%	40	xtal	3%sil?
86z4	hbetgn	15	12	55		20		3		al,sp,ap	28	xtal	
Dikes													
81k270-b	bmtgn		3	75	2	23	2	tr		al,ap	20?	mmi	oss p
81k124	mbgr		3	30	30	30	5				30?	mmi/xtal	
81k128	mgdgn			35	15	47	3				22?	xtal	
81k271	mt			40		50	10			gar,zir	15?	hypi	

Table A-3
Hazard Creek Complex
Round Valley Granite

Sample	Rock Type	Modal % hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An	Texture	Comments
84z10	mbgr		1	35	25	35	4	tr		ap	27	mmi	nor p
81k223	mgd		tr	58	15	25	2	tr			25	mi?, hypi	oss p
81k226	mgr		tr	40	23	30	7	tr		gar	28	mmi, und p	nor p, secondary mica
81k246-b	mgr			50	20	27	3	tr		gar	26?	mmi	nor p, secondary mica
81k246-a	bhetgn	2	10	56		17		5	tr	ap	38	mmi, und p	nor p, intruded by bmg

Table A-4
Hazard Creek Complex
Eastern Tonalite Gneiss

		Modal %										
Sample	Rock Type	hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An	Texture
82c21	bhetgn	1	20	50		20		10		al,sp,ap,zir	30	Q,P,MA
82k37	bhetmy	7	17	35		30		10		al,ap,zir	35	Q,P,MA
84c35	bhetmy	2	10	53	tr	35		tr		ap		Q,P,MA
82c23	bhetmy	5	15	40	tr	35		5		sp,ap,zir	28	Q,P,MA
84c22	hbediomy	27	3	63	tr	3		2		al,sp 2%,ap,zir	29	mmi, und p
82c20	hbetmy	25	7	45	tr	15		5		al,sp 2%,ap 1%,zir	30	Q,P,ma,mmi
83c47	hbetmy	17	8	50		25		1		al,sp 1%,zir	30?	Q,p,ma
85c33	hbetmy	10	10	50	50	15		7		sp 2%,ap	35	Q,P,ma
83c54	hbtmy	20	8	50		20		tr	tr	sp 2%,ap,zir	32	xtal, und p
86c2	amphib	50	0	45		5			2	sp,ap	40	Q,P,MA
85c32	begdmy		10	55	10	25		1	tr	sp,ap	25?	Q,P,MA
84j6	betmy	tr	10	65	1	20		3	tr	al,sp,ap,zir	and?	Q,P,MA
84j7	betmy		5	25		65		2		al,zir	35	Q,p,ma,mmi
84j8	betmy		7	65	tr	25	tr	1			28	Q,P,ma,mmi
86z2	bmetgn		3	50		44	2	1		ap	29	Q,P,MA
85k28	hbdiomy	23	7	62	3	2		2		al,sp	27+	Q,P,ma
86z1	hbetgn	25	5	57		10		3	tr	sp,ap		Q,P,MA
81k109-b	hbtmy	20	10	55		15			1	sp,ap,zir,ru		Q,P,MA
86z3	heqdgn	40		51		7		2	tr	sp	30+	xtal, Q,P
84z11-b	betgn		3	58		37		2		ap,zir	30	Q,p,ma
84z12	betmy		7	60		32		1		ap	35	Q,p,ma
84c54	betmy		15	50		33		2		ap,zir	40?	Q,P,MA
84c53	bt		15	42		43	tr	tr		al,sp,ap	38	mi
84c8	hbegdmy	20	15	40	7	13		5		al,sp,ru?	.34	Q,P,MA
84c9	hbegdmy	15	15	43	8	12		7		sp,ap,zir	30	Q,P,MA
84z11-a	hbetgn	15	6	55		20		4	tr	al,sp,ap,zir	35	xtal?, und p
84z13	hbetmy	15	10	40	3	25		7		sp 1%,zir	30	xtal, und p
84c50	hbetmy	18	7	55		10		10		sp 1%	35?	Q,P,ma
84c51-a	hbetmy	13	12	58		12		4		al,ap,zir	30?	Q,P,ma
84c51-b	hbetmy	27	8	57		5		1		sp 1%	30?	Q,P,MA
84c52	hbetmy	8	7	33	tr	50		2		al,sp,zir	26?	Q,P,MA
81k109-a	hbtmy	5	5	70		20			tr	al,sp 1%,ap	30	mmi, und p

Table A-5
Hazard Creek Complex
Quartz Diorite Gneiss

Sample	Rock Type	Modal % hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An	Texture	Comments
Southern outcrops													
82k41	hbetgn	30	10	45		10		5		sp,ap,zir	32	xtal	
82c25	hbetgn	30	10	45		20		5	tr	al,sp,ap 1%,zir	40	xtal?	
82c27	hbetgn	20	15	40		15		7		sp 1%,ap 1%,zir	35	xtal	
85z16	hbetmy	15	15	40	3	17		10		sp,ap	30+	Q,p,ma,xtal ma	
83z8	behtmy	5	23	27	5	18		22		ap,zir	and?	Q,p,ma,xtal ma	
82c26	betgn		20	40	tr	35		5	tr	al,ap,zir	35	Q,P,xtal ma	
84j13	bpqgn		30	50		20		tr			35	Q,P,MA	
81k259	hbeqd	20	10	57		5		7		sp 1%,ap	40+	xtal	
84j12	bet		5	70		25		1		ap	34	mi,Q	light layer in gneiss
81k229	betmy		7	65	1	25		2		ap	and?	Q,P,MA	light layer in gneiss
84c33	hbetgn	20	11	50		15		3		al,sp 1%,ru,ap	34	Q,p,xtal ma	intrusive margin of Hhbet
83z13	pegmy		tr	40	5	55	tr		tr		and	Q,P,ma	pegmatite dike
Northern outcrops													
85k9	hbeqd	30	3	60		5		2	tr	sp,ap,zir	32	xtal	nor p
85c11	hbetgn	15	15	50		10		10			30	xtal	
85c27	hbetmy	20	12	55		21		2	tr	sp 1%,ap	45?	Q,P,ma	
85c28	hbetmy	10	10	65		5		7	tr	sp 3%,ap	35?	Q,P,ma	
84c12	bhetmy	7	15	60		8		15	tr	sp,ap	35	Q,P,ma	

Table A-6
Hazard Creek Complex
Light and Dark Gneiss

Sample	Rock Type	hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An	Texture	Comments
		Modal %											
Dark gneiss													
85k26	hbdio	20	10	70		tr		tr	tr	sp,ap	32+	mmi?	oss p
83k25	hbeqd	25	5	57		10		3	tr	sp,ap	30	xtal	
83c41	hbet	28	7	45		10		7		sp 2%,ap,zir	32	fg layers	nor p
84j10	hbetgn											mmi/xtal	
Light gneiss													
84j9	betmy		7	75	tr	15		2		sp 1%,ap,zir	28	Q,P,MA	
83c62-a	bhedio	5	10	6	3	5		4		al,sp 3%,ap,zir	35?	mmi, p und	
83c62-b	bhet	8	12	40	2	30		5	tr	al,sp 2%,ap,zir	35?	Q,P,ma	
83k24	bmgrmy		3	30	20	45	2	tr		al,ap,zir	28	Q,P,MA	
84c38	btmy		10	65	3	20		1		al,ap	30	Q,P,ma	
83c40	hbt	15	5	50		30		tr	tr	al,sp,ap,zir	25	mmi, p und	

Table A-7
Hazard Creek Complex
Garnet Granodiorite Mylonite

Sample	Rock Type	Modal % hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An	Texture	Comments
Garnet granodiorite													
82c22	alkmy		2	35	40	25	tr			gar 1%	25	Q,P,MA,myr	
83z12	bgdmy		7	45	5	43				al,sp	and?	Q,P,MA,myr	
82c29	bgdmy		5	40	15	40		tr	tr	al,ap,zir	35		
84c34	bgrmy		2	30	25	43			tr	al	34?	Q,P,MA,myr	
82c24	bgrmy		5	40	15	40	tr	tr		al,ap,zir,gar	30	Q,P,MA,myr	
83c81	bgrmy		7	23	30	40			tr	ap,zir,gar	30?	Q,P,MA,myr	
81k240	bgtmy		2	50	1	47		tr	tr	gar	and	Q,P,MA	nor p
Associated epidote biotite tonalite													
82c30	betmy		20	40		25		15		sp,ap,zir	35?	Q,P,MA	
82c31-b	betmy		10	45	tr	40		1		sp 2%,ap,zir	40?	Q,P,MA	
84c36	betmy		8	55	3	33		1		ap,zir	26?	Q,P,MA	
81k221	betgn		10	64	1	25		1		zir,gar	28		chlor alt
Gabroic and ultramafic pods and layers in garnet granodiorite mylonite													
83c83	hbgab	75	2	20					1	al,sp,ap,zir	40		
84z23	hbite	90	1	10						sp	34		

Table A-8
Hazard Creek Complex
Epidote-Biotite Tonalite Gneiss

Sample	Rock Type	opx	cpx	hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An	Texture	Comments
							Modal %								
85c29	betmygn			1	15	60		15		10	tr	sp	32	Q, und p, xtal ma	
85c13	bmetmy				5	55		35	3	2		sp	30	Q,P,ma	
85c26	betmygn				5	54		40		1		ap	30	mmi,xtal ma	
85c34	bgtmygn				9	60	tr	30				gar 1%	35?	Q,xtal ma	
Gabroic and ultramafic pods and layers															
85c14	hblite			97		tr?					3				
85k31	hbweb	10	50	30						3	7	sp			*op is il*
85k30	2pxton	tr	30	10		45		15				sp 1%	and?		

Table A-9
Hazard Creek Complex
Dark Gneiss Assemblage

Sample	Rock Type	Modal % hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An	Texture	Comments
Variably Deformed Tonalites													
85k36	bet		7	61		30	tr	2		ap	32	Q,p,ma	
81k116	bhet	1	10	65		20		4	tr	ap	30	Q	
85z11	bhet	7	8	67		15		3	tr	ap	32	mmi	nor p
81k231	bhet	3	10	45		35		7	tr	zir	32	mmi	alt-cal,chl,wm
83c91-a	bhet	7	23	45		30		5	tr	ap	32	mmi	chl 1%
83c91-c	bhetgn	8	12	58		20		2		al,ap	35	mmi	nor p
83c91-b	hbet	10	10	60		15		2		al,ap	30	mmi	nor p
83c91-d	hbetgn	14	8	46		30		2	tr	ap	35	mmi	oss p
81k235	heqd	7		66		5		9	2	al cores ep	65?	mmi	alt-chl,wm,scap
81k278	hbet	5	7	60		25		3	tr	al,ap	34	mmi	alt-q, chl, wm
85k4	hbet	10	10	62		15		3			31	mmi	oss plag
85k5	hbet	15	10	55		16		4	tr		32	mmi	
85k10	hbet	10	11	60		15		4			35	mmi	
Crystalloblastic Orthogneisses													
83c90	hbetgn	15	12	59		10		4	tr		32	mmi/xtal	nor p
83c65	hbetgn	30	2	40		10		15		al,sp,ap,zir	39	xtal-polygonal	banded
85k6	hbeqdgn	10	15	60		10		5			35	xtal	
84c11	hgab	15	tr	60		0		18	1	ap,zir	50	xtal	oss p,sieved ep,alt-cc,chl,wm
85k37	hbetgn	12	13	58		12		5		al,sp,ap	33	xtal	nor p

Table A-9 (cont.)
Hazard Creek Complex
Dark Gneiss Assemblage

		Modal %											
Sample	Rock Type	hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An	Texture	Comments
Corral Lake Area													
84c21	amphib	30	10	60		tr				al,ap	36	xtal	
82k38	betmy		13	60	tr	25		2		ap,zir	30	Q,P,ma	
82k39	bhetmy	15	20	40		20		5		sp	?	p	
82k40	hbetgn	15	5	60		20		3	tr	al,sp,ap,zir	32	mmi, p und	
84c20	hbt	8	2	50		40			tr	al 1%,sp,ap,zir,gar 1%	28	xtal, p und	
Areas south of Hard Creek													
81k135-a	amphib	60		40						sp 1%,ap	40	xtal	
81k135-b	dunite								tr				ol 22%,talc 30%, serp 45%
81k129	epam	20		75		tr		5	tr		42	xtal	
81k135-c	qtzite		2	5		80	2		7	sp,ap	20?	xtal	anthoph 4%
82c52	bgmtgn		10	30		45	3			zir	32	xtal	gar 10%
82c50	bhetgn	15	17	45		15		8		al,sp,ap,zir	35	xtal	
82c51	btgn		20	30		50						xtal	

Table A-10
Hazard Creek Complex
Light Gneiss Assemblage

Sample	Subunit	Rock Type	Modal % hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An	Texture	Comments
83c92-a	xtal gn	amphib	40	5	40		15		tr		sp 1%,ap,zir	32	xtal	
83c92-f	xtal gn	betgn		5	65		25		4		ap	34	xtal, relict p	oss p
83c92-e	xtal gn	btgn		10	63		25		1	tr	ap,gar 1%	34	xtal, relict p	oss p
83c92-h	xtal gn	hbediogn	20	10	62		3		3		al,ap 2%,ap,zir,gar	32	xtal, relict p	
83c92-b	xtal gn	hbetgn	17	8	50		20		5	tr	al,sp 2%,ap,zir	35	xtal, relict p	
84c10	xtal gn	hbetgn	15	1	65		10		10		al,ap,zir	36	xtal, relict p	oss p
85z15	xtalgn	hbetgn	23	7	62		7		1		sp,ap,zir	34	xtal, relict p	nor p
84j5	bet	bmet		5	50	3	40	1	1			30	mi	oss p
85z17	bet	bet		3	60	1	35	tr	1			and?	mmi	dike, oss p
81k280	bet	bmgd		10	37	10	43	1	tr		ap	26?	mmi	oss p
83c92-d	bet	btgn		3	57	5	35	tr	tr		ap	40?	mmi	dike, oss p
83c92-c	bet	mbtgn		3	50	3	40	4	tr			30	mmi	dike, oss p
81k279	hbet	bhetgn	8	12	61		15		3		al,sp 1%,ap,zir,ru?	32	mi	oss p
81k224	hbet	hbet	6	6	66		20		2	tr	al,sp,ap	32	mmi	oss p

Table A-11
Hazard Creek Complex
Tonalite with Gneiss Pods

		Modal %											
Sample	Rock Type	cal	dio	hb	bio	plag	qtz	ep	gar	op	Accessories	An	Comments
Gneiss pods													
84c28	amphib	15		30	tr	70		1		tr	al,ap,zir		scap 15%
84c32	amphib			70		10		20		tr		38?	
84c27-a	cs gn			10			7	40	40		sp 1%,cal		
84c27-b	cs gn		15				5	tr	50		sp,zir		
84c30	hbite			95		2		tr	tr	tr	rut	40	anth 1%, scap 3%
85k12-a	marble	90	3				1		7				
85k12-b	qf gn			4	11	50	35			tr	ap,rut	32	
Associated igneous rocks													
84c29	hbdmy			15	10	75	tr			tr	al,ap,zir	40	
84c39	hbetmy			2	8	20	67	2		tr	al, sp 1%	38?	

Table A-12
Hazard Creek Complex
Mafic Layered Gneiss

		Modal %												
Sample	Rock Type	opx	cpx	hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An	Comments
Andesite gneiss-suite 1														
85z10	hbepqgn			18	10	30		35		2	tr	ap	40	
84c5	amphib			30	tr	65		3		1	tr		40	
86c12	hbepqgn			25	1	55		15		2	tr		32	
86c8-b	amphib			25	1	71				1	2	gar,ap	38	
Andesite gneiss-suite 2														
81k274	hbgqdgn			30	10	55		3		tr		gar 2%	38	
83c88-b	hbeqdgn			25	15	45	15	15		1	tr	sp 1%,ap,zir	32	
83c89	hbtgn			40	3	28		30		tr		ap	40?	
Amphibolite														
86k20	amphib			40		55		5			tr		52	
Quartzite and calcsilicate														
81k277	cs		7	35	2	40		10		10		sp,zir,gar	36	hb->cumm
84c2	cs			45		15				30		sp 1%,zir		cumm 10%
86c13	qtzite							70	15					15% alteration
Light orthogneiss														
81k273	betgn				7	64	2	25	1-sec	2		al,ap	28	
83c88-a	bgetgn				7	65		25	tr	1		gar 2%	35	
86c8-c	hbpqggn			7	8	50		35		tr	1	gar,ap	32	
Dark orthogneiss														
86c10	hbepqgn			20	1	60		15		2	1		34	
86c9	epam			35	1	50		25		1	tr		38	
85c12	hpxpqgn	15		45		37		3				sp,zir	40?	
Dikes														
83c88-c	bgdgn				3	45	15	37	tr	tr		gar	30	
84c4	mbetgn				3	45		43	7	2		al	40?	
84c3	mbgdgn				1	60	7	30	3			ap	olig	
81k269	mtgn					52	tr	30	18			gar	15?	
86c8-a	hbtgn			4	1	50		45			tr		28	

Table A-13
Little Goose Creek Complex
Porphyritic Orthogneiss

Sample	Rock Type	Modal % hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An
83z7	bgdmy		7	50	5	38		tr	tr	al,ap,zir	34
83z10	bgdmy		13	45	2	35			tr	al,ap,zir	
85z12	bgdmy		7	53	10	30		tr		al,zir	ol-and?
86z52	bgdmy		4	32	25	33	tr		tr	al,ap,zir	ol?
81k119-b	bgdmy		10	55	7	28			tr	al,sp,ap,zir	28
81k123	bgdmy		10	30	15	45				al,sp,ap,zir	28
81k132	bgdmy		10	35	10	45				al,sp,ap,zir	30
83c72	bgdmy	tr	5	38	20	38			tr	al,sp,ap,zir	33
86z11	bgdmy	tr	7	48	25	20	tr		tr	al,sp	32
85z13	bgrmy		5	35	30	30	tr			ap,zir	28
86z20	bgrmy		6	30	25	40			tr	al,ap	30
83c22	bgrmy		10	30	10	50	tr			sp,ap,zir	30
83c25	bgrmy		10	32	25	33	tr			al,ap,zir	34
83c61	bgrmy		10	25	30	35	tr-sec			al,ap,zir	30
85z7	bhgdmy	6	9	50	5	30			tr	al,sp,ap	28
85z8	bhgdmy	1	5	70	7	20				sp,ap,zir	30?
83c53	bhgdmy	2	12	40	15	30	tr	tr	tr	al,sp 1%,ap,zir	25
85z6	bhgrmy	2	8	37	25	30	tr	tr	tr	al 1%,sp 2%,ap,zir	olig
84z20	bhmtmy	1	5	50	tr	40	2		1	al,sp 1%,ap,zir	35
86z53	bhtmy	7	8	70		15			tr	sp,ap	32
83c16	bhtmy	2	10	25	1	60				al,ap,zir	38
83c17	bhtmy	1	15	40	tr	44				ap,zir	30
82k53	bmgd		7	55	10	20	5	tr-sec	tr	al	
85z2	bmgrmy		5	32	30	30	3			zir	30
84z15	btmy		7	65	5	25			tr	al,sp,ap,zir	35?
86z6	btmy		7	55	3	35				al, sp,ap	28
82c4	btmy		10	60		30	tr		tr	ap,zir	30
83c39	btmy		10	43	3	42	tr		2	al,ap,zir	30
81k133-a	hbgdmy	13	7	35	5	27				sp 3%,ap,zir	26
82c6	hbgdmy	10	10	55	10	15	tr		tr	al,sp,ap,zir	
83c35	hbgdmy	20	55	30	30	5			tr	al,sp 2%,ap,zir	26
83c85	hbgdmy	12	8	50	10	20			tr	al,sp 1%,ap	32
83c34	hbgdmy	3	12	38	15	32			tr	al,sp,ap,zir	30
81k133-b	hbpqgn	40	45	25		20				sp,ap,zir	35
81k119-a	hbtmy	15	15	65	tr	5				al,sp,ap,zir	28
84z8-b	pbgdmy		7	50	7	35			tr	al,ap,zir	30
83z14-1	pbhgdmy	5	7	44	8	36			tr	sp,ap,zir	
83z14-2	pbhgdmy	4	7	46	17	26	tr-sec			al,sp,ap,zir	28
84z2-b	pbhgrmy	2	3	30	25	40			tr	sp,ap,zir	26
84z14	pbtmy		3	48	tr	48	tr-sec		tr	al	30
84z2-a	phbgdmy	7	5	50	8	30			tr	al,sp,ap,zir,gar	32
84z6-d	phbtmy	7	7	55	tr	30	tr		2	al,sp	36
85z3	qpbmy		5	30		65	tr-sec				32
85z1	btmy	1	15	60	1	25	tr-sec		0	al,ap,zir	?

Table A-14
Little Goose Creek Complex
Orthogneiss layers, pods and dikes in Porphyritic Orthogneiss

Sample	Subunit	Rock Type	Modal % hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An	Comments
84z19	ortho layer	bgdmy		5	53	10	32			tr	al,ap,zir	and?	
84z6-c	ortho layer	bgrmy		2	33	20	45			tr	zir	30	
82k17	ortho layer	anorthosite		3	95	tr		tr-sec		tr	ap,zir	?	
81k131-b	ortho layer	bgdmy		15	40	20	35				al,zir,gar	26?	
82c47	ortho layer	bgdmy		15	40	5	40				sp,ap,zir	35?	
83c37	ortho layer	bgdmy		10	33	15	32	tr		tr	al,sp,ap,zir	26	
81k161	ortho layer	bhgdmy	6	9	40	5	50				al,sp,ap,zir	35	
81k33	ortho layer	bhtmy	3	20	40		37	tr-sec?			sp,ap,zir	37	
85z1	ortho layer	btmy	tr	15	60	tr	25	tr-sec?			al,ap,zir	?	
81k131-c	ortho layer	btmy		10	55	tr	35	tr			al,ap	25	
81k34	ortho layer	btmy		10	40		50				al,ap,zir	34	
82k46	ortho layer	hbqmdmy	8	8	55	20	10	tr-sec	tr-sec	tr	al,sp,ap zir		
82k18	ortho layer	hbdio	25	5	55				tr		sp 1%,ap 1%		
82c45-a	ortho layer	hbdiomy	20	1	50	15	10			3	sp 1%,ap,zir		
82k45	ortho layer	hbgd	35	5	50	7	2			tr	al,sp 1%,ap 1%,zir		
84z4	ortho layer	hbgdmy	17	8	55	5	15				sp,ap,zir	33	
81k131-a	ortho layer	hbgdmy	8	7	35	10	40				ap	25	
84z8-a	ortho layer	hbgdmy	20	5	55	15	5			tr	al,sp,ap,zir	32	
81k131-d	ortho layer	hbggdmy	5	5	40	10	40				al,sp,ap,zir,gar 2%	25	
84z3	ortho layer	hbtmy	15	5	50	10	20				al,sp,ap,zir	30	
84z6-b	ortho layer	hbtmy	25	10	50	5	10				sp,ap	30	
84z7-a	ortho layer	hbtmy	25	10	45		20		tr-sec	tr	sp,ap	32	
84z7-b	ortho layer	hbtmy	23	7	55	tr	15			tr	al,sp,ap,zir	34	
84z7-c	ortho layer	hbtmy	25	10	60					tr	al,zir	37	5% scapolite
84z17	ortho layer	hbtmy	10	7	48	tr	35		tr	tr	al,sp 1%,ap,zir	and?	ep rims al
84z18	ortho layer	hbtmy	25	5	55	tr	15		tr		sp 1%,ap,zir	?	ep rims al
83c73	ortho layer	hbtmy	12	5	70	tr	10				sp 2%	35	
83c76	ortho layer	hbtmy	18	12	35	1	25				sp,ap,zir	35	
85c18	ortho layer	hbtmy	28	7	55	tr	10			tr	sp 2%,ap,zir	30	
82k10	ortho layer	leucogr		1	30	30	40	tr		1	zir		
84z6-a	hb gab pod	amphib	40	5	50		tr			tr	al,ep,sp,ap,zir	32	cpx cores hb
84z6-e	hb gab pod	amphibmy	60	tr	40						sp 1%,ap,zir	40	
84z6-f	hb gab pod	amphibmy	35	10	55						al,sp 1%,ap	32	
84z5	hb gab pod	hbite	88		tr				2	7	sp 1%,ap,zir	olig	
84z8-c	dike	anorthmy	4	5	89		tr			tr	al,sp,ap,zir	34	2% cummingtonite
85c35	inclusion	btgn		20	50	tr	30			tr	th,ap,zir	28	

Table A-15
Little Goose Creek Complex
Mafic layered gneisses and associated calcsilicate gneisses interlayered with porphyritic orthogneiss

		Modal %									
Sample	Rock Type	dio	hb	bio	plag	ksp	qtz	op	Accessories	An	Comments
84z16	amphib		40		40		20	tr	al,ap,zir	and?	
81k81-b	amphib		20	8	71			1			alt-cc,ep,chl
81k81-c	amphib		23	7	67			3	sp,ap,zir,cc		zoned relict plag
81k84	amphib		25	5	60		5	2	ap		
84z9	cs gn	25	10	tr	30	5	30	tr	sp,ap,zir	34	
86z41	cs gn	7	10	10	70		3	tr	sp,ap	35?	
81k78	qfgn	15	15	15	45		8	2			
85c10	qfgn		12	12	65		20		al,ap	35	
Metasedimentary Rocks											
82k35	gar ton			3	40		40	tr	ap,zir	40	gar 15%,sil 2%
85c8	silgargn			10	2	28	35	tr		30?	sil 5%,gar 20%

Table A-16
Little Goose Creek Complex

Sample	Rock Type	hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An	Comments
				Modal %								
Western Orthogneiss												
83c38	hbgdmy	15	10	40	8	25				al,sp 2%,ap,zir	35?	
83c45	hbgmy	13	7	45	30	15				al,sp,ap	16	
83c46	hbgrmy	10	5	20	30	30	tr		tr	al,sp	28	
84z1	hbtmy	20	10	55	5	10				al,sp,ap,zir	34	
85z5	hbtmy	10	10	55	3	20				sp 2%,ap,zir	37	
81k120	hbtmy	15	10	60		15				al,sp 2%,ap,zir	30	
81k134	hbtmy	20	8	55		15			tr	al,sp 2%,ap,zir	32	
82k36-b	bgrmy		3	45	30	30	tr		tr	al	30	
82k36-a	hbgdmy	28	7	30	10	23				al,sp 2%,ap,zir	30?	
81k192	btmy		7	60	5	25	tr		tr	sp 1%,ap,zir	28+	
81k191	hbgdmy	14	10	50	7	18				al, sp 1%,ap	28+	
81k195	hbgdmy	5	5	35	15	40				al,sp	30	
84c23	hbsgdmy	20	10	40	15	15				al,ap	30	
81k188	hbtgn	20	10	60		10				al,sp 1%,ap	38	schist layer
81k62	hbtmy	5	10	65	2	15				al,sp 1%,ap	28	alt-chl, ep
Spotted Orthogneiss												
81k127-a	bmtmy		13	50	tr	35	2		tr		30	
81k127-d	hbdiogn	25	15	60						ap,zir	28	
81k127-b	hbdiomy	20	5	72	3	tr			tr	sp,ap,zir	25	
85z18	hbgdmy	20	10	45	15	10			tr	sp,ap	30	
85c7	hbmdgn	20	5	52	15	tr			tr	sp 2%,ap 1%	24-45?	
81k121	hbqdgn	25	15	53	5	2				sp,ap,zir	30	
Coarse-Grained Tonalite Orthogneiss												
83k22	bgr		15	30	20	35				ap	31	
83c70	bhtmy	5	7	65	tr	20			tr	al,sp,ap,zir	38	
85z9	hbgdmy	10	10	55	5	25			tr	sp 1%,ap	32	
83c36	hbtmy	8	7	45		40			tr	al,sp,ap,zir	35	
83c52	hbtmy	12	8	50		30			tr	al,ap,zir	37	
85c9	hbtmy	10	10	60	tr	20		1	tr	sp,ap	30	
83k8	bhtmy	7	13	45	3	30			tr	sp 1%,ap		alt-chl,ep

Table A-17
Little Goose Creek Complex
Mafic Orthogneiss

Sample	Subunit	Rock Type	Modal % opx	cpx	hb	bio	plag	qtz	op	Accessories	An	Comments
82k56	BR	cpxite		80		10	10		tr		30?	
81k74	BR	hbdgn			23	2	72		3		40+	
83c87-d	BR	hbdiogn			35	2	55	3	5	ap	35?	
82k55	BR	hbdiogn			30	7	45	15	3	ap,zir	40	
82k57-a	BR	hbdiogn			8	1	80	10	1	zir	40	
81k76	BR	hbtgn			15	10	55	20	tr		35	
83c87-b	BR	hbtgn			10	15	65	10	1	sp,ap	35	
83c87-f	BR	hbtgn			20	15	50	15		ap	30?	
82k54	BR	hbtgn			12	8	50	30	2	ap,zir	33	
83c87-a	BR	hpxgab	10		15	1	65	5	3		40+	
83c87-c	BR	hpxgab	20		10	3	85		2	ap	38	
83c87-e	BR	hpxgab		10	15	5	52	10	3	ap	30?	alt
83c87-g	BR	hpxite	20	20	30		5	2	3			alt 20%
83c59	LR	bht			2	12	50	25		al,sp	38	
85k20	LR	hbdgn			20	10	70		2	ap	40+	
83c60-b	LR	hbgab	15		30	tr	60		tr		65	
83c60-a	LR	hbtgn			20	15	55	10	tr	ap,zir	40	
83c24	MC	hbgab	35	35	15		15				65	
82k34-a	SB	amphib			45	tr	47	5	3	ap,zir	35	
86z7	SB	hbdgn			25	5	70		2	sp,ap,zir	35	
86z8	SB	hbdgn			18	7	75		1	sp,ap	38	
86z9	SB	hbdgn			20	5	75		2	ap	38	alt-chl,ep
82c46	SB	hbdiogn			7	8	80	5	tr	ap,zir		alt-ser
82k34-b	SB	hbgad		5	75	tr	10	10	tr	sp	32	
82c45	SB	hbqdmy			10	3	73	15	tr	ap,zir	35	
82c44	SB	hbtgn			20	5	45	30	tr	sp,ap,gar	38	
86c4	SB	hpxgabgn		20	25		40	15		sp	40	

Subunits: BR=Brundage Resevoir, LR=Lava Ridge, MC=McCall, SB=Slab Butte

Table A-18
Payette River Complex
Payette River Tonalite

Sample	Rock Type	hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An	Comments
				Modal %								
86z32	bgd		12	38	17	33				thor,ap	35	
82k15	bgd	1	15	40	15	30				sp,ap,zir	32	
82c12	bgd	1	15	40	25	20		tr	tr	sp 1%,ap	31	
82k31	bgd	1	20	35	15	30				al,sp 1%,ap,zir	33	
85z4	bhgd	2	8	40	20	30				al,sp,ap	33	
81k31-a	bhgd	5	10	32	15	38			tr	sp,ap,zir	34	
81k43	bhgd	4	6	50	10	30		tr	tr	al,sp,ap,ziir	32	
86z18	bht	3	12	50		35			tr	al,sp,ap	33	
86z21	bht	1	14	45		40			tr	al,sp,ap	35	
83z6	bht	2	15	55		28				sp,ap	36	
82k11	bht	10	10	50		30		tr	tr	sp 1%,ap,zir	35	
82k30-a	bht	3	15	55	3	25				sp 1%,ap,zir	38	
83c22	bht	5	15	43	2	40			tr	al,sp,ap,zir	35	
85c5	bht	5	15	55	2	23		tr	tr	al,sp 1%,ap	34	
82k24	bht	7	15	45	tr	30			tr	sp,ap,zir	48	
82k50	bt		25	45	tr	30		1		al 1%,sp 1%,ap,zir	34	
83c18	bt		1	45	2	50	tr			al,ap,zir	28	
83c28	bt		20	38		43				al,ap,zir,cpx	36	
83c29	bt		20	40		40				al,ap,zir	35	
83c80	bt	1	7	45		47				al,ap,zir	36	
83c58	btmy	3	15	65		15				ap,zir	38	
86z47	hbgd	10	5	45	10	30			tr	al,sp,ap	35	
82k3	hbgd	20	10	35	20	15				al,sp,ap,zir	38	
82k13	hbgd	10	10	30	10	40				sp,ap	30	
82c1	hbgd	17	13	40	10	20		tr-sec	tr	al,sp,ap	34	
83c79	hbgd	5	11	45	4	35			tr	al,sp,ap,zir	30	
83z2	hbt	12	8	40		40				al,ap,zir	40	
83z11-1	hbt	15	5	50		30			tr	sp,ap,zir	44	
83z11-2	hbt	15	5	45		35			tr	sp,ap,zir		
85z14	hbt	10	10	50		30		tr	tr	al,sp,ap,zir	40	
82c3	hbt	20	10	55	tr	15				sp,zir	52	
82k4	hbtgn	30	5	45		20			tr	ap,zir		
83c57	hbtgn	15	10	50		25			tr	ap	35	
Dikes, inclusions and granites in Payette River tonalite												
83c19	bgr		3	30	30	30	tr				35?	dike
83c20	bgr		10	25	20	45				ap,zir,gar	30?	dike
81k31-b	bmgr		3	25	40	30	2			ap,zir	25	dike
82c40	bgr		7	30	30	30			tr	al,ap,zir	35	gr
82k5	bgr		20	25	20	35			tr	al,sp,ap,zir	32	gr
82k32	bgr		2	35	35	30			tr	ap,zir	33	gr
82k12	leuco gr		2	18	40	40			tr	al,ap	30	gr
82k25	bt	1	30	30		40		tr	1	sp,ap,zir	30	inc
82k30-b	hbdio	25	15	60		tr		tr	tr	al,sp,ap,zir	40	inc
82k16	hbtmy	10	10	30	tr	50		tr	tr	ap	40	inc
Porphyritic Granodiorite												
85c16	bhgr	5	15	30	25	25				al,ap,zir	30	
83c67	hbgdmy	13	7	30	20	30	tr			al,ap,zir	34	
85k14	pbhgr	3	7	35	15	40				al,ap	30	

Table A-19
Payette River Complex
Payette River Granite

				Modal %							
Sample	Rock Type	hb	bio	plag	ksp	qtz	mus	op	Accessories	An	Comments
81k17	alaskite			10	45	45	tr		gar 1%	28	
82c41	bgr		3	30	25	40		tr	ap,zir	30	
83z5	bgr		5	25	25	45		tr	ap	26	
83c14	bgd	tr	12	32	10	45	tr		sp 1%,ap,zir	27	
85c36	bgd		15	40	5	40	tr	tr	ap,zir	28	
84z21	bggr		5	35	25	35	tr		gar,zir	30	
81k18	bgr		7	35	25	35	tr		ap,zir	30	
84j15	bgr		7	28	25	40		tr	al,ap,zir	30	
82k52	bgr		3	20	30	50		tr	al,ap,zir	29	
85c1	bgrgn		10	27	35	27		tr	al,ap	25	
85c6	bgrgn		10	35	35	20			ap	32	
83c55	bgrmy		5	30	20	50	tr-sec		ap	38?	
82k2	bgrmy		7	30	30	30		tr	al,ap,zir	26	
83c26	bgrmy		5	30	30	35	tr		ap,zir	15?	
82c9	bhgr	5	10	20	25	40			al,ap,zir	28	
81k58	bmgr		2	20	30	50	tr		ap,zir	?	
82c8	btmy		7	25	tr	65?	tr		al,ap,zir	26	
82c11	pbgr		10	30	30	30		tr	ap,zir	30	
81k30	bsch		45	55				tr	ap,zir	32	inclusion

Table A-20
Payette River Complex
Trail Creek Granite

		Modal %									
Sample	Rock Type	hb	bio	plag	ksp	qtz	mus	ep	op	Accessories	An
83z4	bgr		7	45	20	30				ap,zir	28?
83z15	bgr		5	30	35	35	tr			ap,zir	
83z16-a	bgr		3	25	40	25			tr	al,ap,zir	30
83z16-b	bgr		3	20	35	45				al,zp,zir	28
83c33	bgr		5	20	27	27				ap,zir	
83c69	bgr		12	30	23	30	1			ap,zir	35
83c86	bgr		7	33	30	30				ap,zir	30
82c2	bgr	1	7	7	40	45		tr		al,ap,zir	
86z17	bgr		8	35	25	30	1		tr	ap,zir	32
82k23	bgr		10	35	25	30	1			ap,zir	35?
83c82	bgrgn		15	35	10	45				ap,zir	35
83c75	bmgr		5	25	25	40	2		tr	zir	31
83c21	bmgr		7	30	25	43	5			zir	35?
82k57-b	bmgr		10	25	15	45	5		tr	ap,zir	40?
83c32	hbgr	5	5	35	20	35		tr		al,sp,ap,zir	30
83c74	mbgr		2	25	25	40	7			ap,zir	30

Table A-21
Payette River Complex
Metasedimentary rocks and associated hybrid igneous rocks

							Modal %								
Sample	Rock Type	dio	hb	bio	plag	ksp	qtz	mus	sil	ep	gar	op	Accessories	An	Comments
85c17	amphib		20		40							tr	sp 3%,ap	50?	pale hb 35%,scap 2%
82c14-b	bpgn			35	65		tr	tr				tr	ap,zir		
82c15	bqmy			5	10	5	70	sec				tr	zir	35	
86z28	bsch		12	23	20		40			tr		1	ap,zir	30-50	
86z29	bsch		1	18	25		60					1	al	30-50	
82c14-h	bt			15	43	tr	43						al,sp,ap,zir	36	
82c14-i	bt			15	45		40	tr					zir	30	
83z17-1	calqtz	10					90						ap,zir		
83z17-2	calqtz	15					55					tr	ap		scap 20%,
82k1	cs gn	8					3				15	3			cc 85%
82c14-a	cs gn				70								woll		
82c14-j	cs gn	50	3		10		35					2	sp 2%,ap,zir	70	
82c16-a	cs gn	20			50		20						sp,ap,zir,scap?		woll? 10%
82c16-b	cs gn	25			30		45						sp,ap,zir,cal		act? 3%
82c16-c	cs gn	30					35				35		sp,ap,zir		
82k19-b	cs gn				75					5		tr	al,sp 5%		act 15%
82k19-c	cs gn	35			45						20		sp 2%, zir		
82k19-e	cs gn	45			20						20		sp,ap,zir,cal 2%		
82k19-f	cs gn				65		tr						sp 1%, cal 1%		
86z30	gr			7	40	25	25					tr	al	28	mmi
82c14-d	hbgn		15		85					tr			al,sp 1%,ap,zir	58	
85c2	hbpqkgn		12	12	35	15	25			tr			al,sp,ap,zir	35	
82c14-g	pqgn			tr	60	tr	40					tr			
82c5-a	sil sch			40			25	tr	35			tr	zir		

Table A-21 (cont.)

Sample	Rock Type	dio	hb	bio	plag	ksp	qtz	mus	sil	ep	gar	op	Accessories	An	Comments
							Modal %								
82c5-b	sil sch			30			45	tr	30			tr	zir		
82c10-a	sil sch			20	tr	42	25	tr	15				ap,zir		
82k19-d	sil sch			25	3	30	15	1	25			tr	ap		
82k21	sil sch			25	2	25	30	5	10			tr	ap,zir		
86z31	sil sch			7	10		75	5	3			tr	gar	32	
85c3	silgargn			7	2	20	63		5		3		ap	50?	
Associated igneous rocks															
82k22	alaskite			1	15	45	40	tr				tr	al,ap,zir	olig?	
82c13	bgr			7	40	25	25	1					al,ap,zir	26	
82k58	bgr			4	30	33	33						al,zir	30?	
82k33-a	bgrmy			10	28	25	33						ap,zir		
82c14-f	bmt			7	60		30	2					al,ap,zir	35	
82c14-c	hbgdmy		20	5	40	20	15						al,sp 1%,ap,zir	30	
82c10-b	bgd			12	35	10	40						al,ap,zir	46	
86z24	bht		7	13	50		30					tr	sp,zir	38	
85c4	hbtgn		12	12	48	3	15						al,sp	35	
82c7	bhggn		5	20	15	25	35			tr			al,sp,ap,zir	34	inclusion
82c10-c	hbtgn		25	5	50		20					tr	al,ap,zir	36	inclusion
82k19-a	btgn			28	45		28					tr	al,ap,zir	45	inclusion

APPENDIX B: ANALYTIC PROCEDURES AND ERRORS

RB, SR AND $^{87}SR/^{86}SR$ ANALYSES

Samples were analyzed for Rb and Sr concentration and $^{87}Sr/^{86}Sr$ using isotope dilution techniques in the laboratories of Prof. L.T. Silver in the Division of Earth and Planetary Sciences, California Institute of Technology, Pasadena. The technique was modified from that used by Robert I. Hill, which is reported in his doctoral thesis (Hill, 1984). 200 to 400 grams of rock powder were produced by grinding in a tungsten carbide lined SPEX "Shatterbox." 60 to 120 mg of this powder was washed into a clean Teflon dish and spiked with Rb and Sr. Four ml HF and 1 ml $HClO_4$ were added, and the whole thing was allowed to stand in a nitrogen-flushed teflon tank for six to eight hours. The sample was then dried slowly over four to six hours. Four ml HF and 1 ml $HClO_4$ were added for a second attack and the sample was allowed to stand in a nitrogen-flushed teflon tank for six to eight more hours. After slow drying, a small amount of 2.5N HCl was added to the residue to assure complete conversion to chloride salts.

The chloride salt residue was disolved in 3 ml 1N HCl and loaded with an additional 2 ml 1N HCl into a 20 x 1 cm column of Dowex AG50W-8X 100-200 cation exchange resin equilibrated to 2.5N HCl. The sample was eluted with 2.5N HCl and Rb and Sr fractions were collected in teflon beakers. Samples were dried in nitrogen-flushed tanks. This procedure was repeated a second time for the Sr fraction to further purify it of Rb.

Samples were dissolved in a small amount of distilled water and evaporated onto the side filaments of a degassed rhenium triple-filament assembly. All analyses were performed on a 12-inch radius curvature solid-source mass spectrometer with magnetic peak switching and on-line data processing. For Sr analysis, the center filament was heated to 1710°C and the current in the side filaments was incrementally raised to 0.7 amp to ionize any Rb present. ^{85}Rb was monitored and ionization under these conditions proceeded until the current was negligible. The center filament was then lowered to 890°C and current in the side filaments increased incrementally until a stable ^{88}Sr beam of approximately 1.5×10^{-11} amp was obtained. This usually required approximately two amps of current in the side filaments. Data for 100 sets of peaks were then collected. A delay of 15 seconds between the time of peak switching and data collection was necessary for correct readings from the collector. For Rb analyses the center filament was heated to 890°C and the current in the side filaments was raised incrementally until a current of aproximately 1.5×10^{-11} amp was obtained for the larger Rb peak. Data for 30 sets of peaks were then collected.

Ratios were calculated using linear interpolation and were corrected for machine fractionation using a power model. Rb and Sr concentrations were calculated using the atomic ratios recommended by Stieger and Jaeger (1977).

The E&A $SrCO_3$ standard was run frequently to monitor machine drift and to determine analytical precision (Table B-1). During a run of samples, $^{87}Sr/^{86}Sr$ for this standard varied between .00007 and .00017, with two standard errors of the mean between .00003 and .00004. All analyses are reported corrected to a value of .70800 for this standard. The correction was determined by averaging the standard analyses during the run of samples (Table B-1). These data suggest that errors due to machine precision are slightly less than ± .0001. This is much larger than the error induced by counting statistics, which is in the range ± .00002-.00004 from most samples.

On several occasions a standard run yielded an unusually low $^{87}Sr/^{86}Sr$. These values were discarded. The runs could not be distinguished from runs that gave normal values. The problem is believed to be related to focusing the beam but has not been resolved.

Samples 83z9 and 83z11 were analyzed multiple times to estimate the precision of the total laboratory procedure (Table B-2). $^{87}Sr/^{86}Sr$ has a range of only .00004 in three replicate analyses of sample 83z9; this is much higher precision than observed in the standard E&A data. $^{87}Sr/^{86}Sr$ varies by .00023 in the two measurements of 83z11. This large error may also reflect difficulties in focusing. Based on these data and the standard data, $^{87}Sr/^{86}Sr$ measurements are believed to be reproducible to $\pm$.0001. It appears that reproducibility may often be much better. Larger errors may be present in some samples due to unrecognizable machine errors, possibly related to focusing.

Rb concentration in samples 83z9 was reproducable to 1.2 %. Sr concentrations for two analyses of 83z9 agree within .09%. The third sample gave a much lower value, which is within 2.1% of the other measurements. Rb concentrations for the two analyses of 83z11 agree within 0.5% and Sr concentrations within 1.1%. Measurements of a spiked standard was reproducible to 0.3% (Table B-2), suggesting that the lack of precision may be due to variability in the sample. These data suggest that measurements of concentration have a precision better than $\pm$ 1%. Difficulties with spiking may sporadically produce larger errors.

$\delta\,^{18}O$ ANALYSES

The procedures used in this study for oxygen isotope analysis follow closely those of Taylor and Epstein (1962), with the exception that pure F_2 gas was used for sample attack. Samples were analyzed in the laboratories of Dr. H.P. Taylor, Jr. in the Division of Earth and Planetary Sciences, California Institute of Technology, Pasadena by the author and Jack Colson. Mass spectrometery was done on a McKinney-Nier, 60^{o} sector double-collecting mass spectrometer, with a working standard of Harding Iceland Spar by Joop Goris. Analyses were corrected to SMOW using the value measured for the Rose Quartz intra-laboratory standard run with each manifold of five samples. The correction formula used was

$$\delta_{SMOW}=(\delta_{raw}-\delta_{RQS})[(1+(background/2520))(1.02354)+8.45]$$

Eleven analyses of the Rose Quartz standard have a mean value of 9.17 ± .32 (2) Duplicate analyses of several samples (Table B-3) indicate that precision is ± 0.2 per mil.

MAJOR AND TRACE ELEMENT CHEMICAL ANALYSES

Major element chemical analyses were performed by the U.S. Geological Survey, Office of Mineral Resources, Branch of Analytical Chemistry in Lakewood, Colorado, under the direction of J. E. Taggart. Trace element analyses were performed by the U.S. Geological Survey, Branch of Geochemistry in Reston, Virginia, under the direction of R. Johnson. Duplicate analyses are shown in Table B-4. Most duplicate analyses of the same powder have precision of ± 0.5% for major elements. Trace element data have much lower precision with variations as large as 40%. Values are generally similar to those obtained by INAA by Amy Hoover at Oregon State University.

Table B-1
E&A analyses

Date	Value	Error*		Average	2 standard errors of mean	Range
3-Mar-85	.70795	3				
11-Mar-85	.70811	4				
13-Mar-85	.70810	14				
15-Mar-85	.70801	4				
18-Mar-85	.70804	4				
3-May-85	.70793	4				
10-May-85	.70795	4				
15-May-85	.70801	4				
21-May-85	.70804	3		.708014	4.1E-05	.00018
24-Oct-85	.70803	4				
29-Oct-85	.70791	3				
31-Oct-85	.70806	4				
1-Nov-85	.70796	3				
5-Dec-85	.70804	3				
9-Dec-85	.70799	13				
12-Dec-85	.70803	5				
12-Dec-85	.70801	10				
14-Dec-85	.70802	3		.708005	3.0E-05	.00014
2-Dec-85	.70771	2	discarded			
3-Oct-86	.70813	2				
6-Oct-86	.70812	2				
10-Oct-86	.70803	3				
13-Oct-86	.70802	3				
21-Oct-86	.70810	3				
3-Nov-86	.70806	2				
13-Nov-86	.70803	2		.708068	3.3E-05	.00011
14-Jan-87	.70811	3				
15-Jan-87	.70809	2				
23-Jan-87	.70811	3				
30-Jan-87	.70804	3		.708075**	3.0E-05	.00011
13-Nov-86	.70770	3	discarded			

* two standard errors of mean
** values for all 1986 and 1987 data

Table B-2
Replicate Sr analyses

	Rb	Sr	87Sr/86Sr	Error*	Date
83z9	16.3	968.6	.703948	3.2E-05	28-Oct-85
	16.2	967.7	.703932	3.1E-05	1-Nov-86
	16.1	947.3	.703968	3.7E-05	7-Oct-86
83z11	62.8	684.3	.708719	2.0E-05	20-Jan-87
	63.1	677.1	.708952	2.7E-05	28-Jan-87

Spiked Standard

Sr	Date
8.961	21-Nov-86
8.962	13-Oct-86
8.966	30-Jan-87
8.938	23-Jan-87

*two standard errors of the mean

Table B-3
Replicate 18O Analyses

Sample		Sample	
83z14	8.55	83z9	8.02
	8.74		7.65
	9.19		7.87
84z3	8.27	85z17	10.09
	8.64		9.99
85z1	10.79	83z11	8.85
	10.92		8.94
			7.59
86z20	10.13		
	10.18	83z2	9.41
			9.11
86z6	9.49		
	9.74	83z16a	9.48
			9.71
86z52	10.72		
	10.44	85z10	10.31
			10.17

Table B-4: Replicate XRF analyses and comparison with data from other techniques

	XRF			ID	NA	XRF		ID	NA			ID
Sample	83z9					83z11				83z14		
SiO2	60.7	60.8	61.6			59.6	59.8			64.4	64.3	
Al2O3	18.4	18.5	18.9			17.5	18.0			18.0	18.0	
Fe2O3*	5.04	5.04	4.74		5	6.20	5.73		6.233	3.32	3.32	
MgO	1.85	1.85	1.69			3.18	2.94			1.01	1.02	
CaO	5.98	6.00	6.14			6.76	7.08			4.54	4.55	
Na2O	4.82	4.75	5.08		4.85	3.08	3.37		3.22	4.91	4.96	
K2O	0.96	0.99	0.84			1.90	1.68			2.08	2.06	
TiO2	0.66	0.67	0.62			0.91	0.81			0.50	0.50	
P2O5	0.30	0.30	0.29			0.29	0.30			0.18	0.18	
MnO	0.1	0.1	0.1			0.09	0.08			0.06	0.06	
Nb	<10	<10	<10			12	12			15	14	
Rb	35	33	32	16		72	65	62	71	53	61	34
Sr	850	1050	1050	963		717	758	675		1190	1210	1079
Zr	93	135	121		149	180	144		177	226	232	
Y	7	11	10			14	16			8	9	
Cu	<2	<2	<2			<2	<2			<2	<2	
Ni	<5	9	<5			10	8			6	8	
Zn	61	88	74		92	96	97			97	100	

	XRF	NA	ID	XRF	NA	ID
Sample	83z15			84z3		
Na2O	3.72	3.73		4.09	4.1	
Fe2O3*	1.23	1.278		6.8	7.111	
Rb	106	94	87	67	54	25
Zr	112	114		195		
Zn	38			101	93	

XRF = X-ray fluorescence analyses from U.S. Geological Survey
NA = induced neutron activation analyses by Hoover
ID = isotope dilution measurement this study

BIG MAPS NOT INCLUDED---- PLEASE REFER TO HARD COPY.