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**Tectonic evolution of the Standfast Creek fault and
Clachnacudainn terrane, southern Omineca Belt, Canadian Cordillera**

by

JAMES L. CROWLEY, B.A.

**A thesis submitted to the Faculty of
Graduate Studies and Research in partial fulfillment
of the requirements for the degree of
Master of Science, Department of Earth Sciences**

Carleton University

Ottawa, Ontario

September, 1992

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in partial fulfillment of the requirements for
the degree of Master of Science.

Chairman, Department of Earth Sciences

Thesis supervisor

Carleton University
September, 1992

ABSTRACT

The Standfast Creek fault, the upper boundary of the Clachnacudainn terrane in the southern Omineca Belt, is interpreted as a minor tectonic feature based on stratigraphic, metamorphic, and structural links between the terrane and the overlying Selkirk allochthon. This interpretation is the basis for a new tectonic model for the Clachnacudainn terrane and base of the overlying Selkirk allochthon. Correlation of metasedimentary rocks on both sides of the Standfast Creek fault with the lower Paleozoic Hamill Group, Badshot Formation, and Lardeau Group is consistent with a few tens to hundreds of metres of fault attenuation. Identical peak regional metamorphic (M2) garnet-bearing assemblages across the fault require that the fault is not associated with appressed or omitted isograds; this evidence is consistent with minor post-M2 displacement. Thermobarometric estimates of peak M2 P-T conditions corroborate the observed lower to middle amphibolite facies assemblages in the highest to lowest levels of the terrane, respectively. M2 in the Clachnacudainn terrane post-dated the first phase of deformation (D1) and was diachronous with respect to the second phase of deformation (D2); in the lowest level of the terrane M2 occurred syn-D2 and in the highest level M2 occurred post-D2. The 104 ± 1 Ma Albert stock (U-Pb zircon method) cross-cuts the Standfast Creek fault and places an upper age limit on M2 and penetrative ductile deformation in adjacent rocks. Late-tectonic granitoids in the Clachnacudainn igneous complex that occur at all levels of the terrane are also interpreted as mid-Cretaceous and younger. Three identical phases of ductile deformation occurred on both sides of the Standfast Creek fault, suggesting that the fault is not a boundary between structural levels. First phase folds are interpreted as major, tight to isoclinal west-verging structures. The main foliation is axial planar to smaller scale second phase folds. Upright third phase folds gently warp all older structures, including the Standfast Creek fault. There are no ductile high strain zones (such as mylonites) within the Standfast Creek fault. The strong fabric in the Devonian orthogneiss in the terrane is interpreted as a result of regional deformation rather than a result of shearing on the Standfast Creek fault because there is no strain gradient observed from the fault to the base of the terrane.

I use this evidence along with the location of the Standfast Creek fault (in the limbs of major west-verging F1 folds) to suggest that most displacement on the fault is a result of fold limb attenuation during F1 folding. However, distributed pre-mid-Cretaceous extensional shear throughout the Clachnacudainn terrane and overlying Selkirk allochthon cannot be ruled out; that possibility is suggested by the barometric calculations and attenuation of the distance between the major structures to the south. The Clachnacudainn terrane is therefore an integral part of the overlying Selkirk allochthon and shares a similar tectonic history. The aspects that differentiate the terrane from the rest of the allochthon are a slightly higher regional metamorphic grade and the presence of numerous granitoid phases in the Clachnacudainn igneous complex.

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ORIGINAL CONTRIBUTION

During the three and a half months of fieldwork in the summers of 1990 and 1991 I mapped in the following 1:50000 sheets: Revelstoke (82 L/16), Camborne (82 K/13), Illecillewaet (82 N/4), and Mount Revelstoke (82 M/1). Since most of the geological contacts were already mapped by previous fieldworkers, my work mostly consisted of making lithologic descriptions, measuring structural elements, and collecting samples for laboratory study. Because the Standfast Creek fault crosses areas mapped in detail by four previous fieldworkers, my work also included comparing lithologies, structures, and metamorphic assemblages between the four areas. I examined about 120 thin sections with a petrographic microscope. I determined the compositions of minerals in six samples with an electron microprobe at the University of Calgary, upon which I applied thermobarometric calculations using the GEO-CALC computer program. Zircons from the Albert stock were dated utilizing the U-Pb method. I performed all stages of sample collection, crushing, and grinding, and mineral separation and selection. The chemistry and mass spectrometry were performed by technicians at the Geological Survey of Canada in Ottawa. I compiled metamorphic and structural data from previous studies and my own study onto maps produced in AutocadTM.

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CHAPTER 1. Regional geology, previous work, and recent tectonic models

Introduction

The purpose of this chapter is to outline the regional geology, previous fieldworkers' numerous (and often confusing) definitions of the Clachnacudainn terrane and Standfast Creek fault, and three recently proposed tectonic models.

Regional geology

Mesozoic to Paleocene compression and crustal thickening in the southern Omineca Belt began with accretion of the allochthonous Intermontane Superterrane to the peri-cratonic Kootenay terrane and continental margin of the North American craton in the Early to Middle Jurassic (Monger et al., 1982; Brown et al., 1986). Tertiary extension followed shortly after the cessation of the compression (Carr, 1990), resulting in exhumation of deeply buried crust (Parrish et al., 1988). These episodes of compression and extension are briefly outlined below. Figure 1-1 is a map showing the major tectonic features of the southern Omineca Belt.

Compression

The Omineca Belt is the metamorphic and plutonic welt that developed during accretion of the allochthonous terranes to the North American margin (Monger et al., 1982). The Omineca Belt is therefore the hinterland to the Rocky

Mountain Foreland Belt. The Monashee decollement, an east-directed ductile thrust fault, accommodated much of the compression in the southern Omineca Belt by displacing its hanging wall, the Selkirk allochthon, at least 80 km (Read and Brown, 1981; Brown et al., 1986; Journeay, 1986; Journeay and Brown, 1986). The decollement was active possibly as early as the Jurassic (Read and Brown, 1981), with latest motion occurring in the Late Cretaceous and Paleocene (Journeay, 1986; Journeay and Brown, 1986; Parrish et al, 1988; Carr, 1990). The decollement is arched over the Monashee complex, which is cored by Early Proterozoic ortho- and paragneiss that is interpreted as North American basement (Wanless and Reesor, 1975; Parrish and Armstrong, 1983). Monashee complex cover metasedimentary rocks that are infolded with the core gneiss are thought to be Upper Proterozoic or Cambrian (Wheeler, 1965; Brown and Psutka, 1979; Read and Brown, 1981; Hoy and Godwin, 1988; Parrish and Scammell, 1988).

The Selkirk allochthon, the hanging wall of the Monashee decollement, is composed of Upper Proterozoic and lower Paleozoic North American continental margin deposits and more distal, possibly marginal-basin deposits of the Kootenay terrane (Brown and Lane, 1988). The Clachnacudainn terrane (also termed Clachnacudainn slice and salient) lies within the Selkirk allochthon, above and east of the Monashee decollement and superposed Columbia River fault; these faults form its lower western boundary. The upper boundary of the terrane is defined by the Standfast Creek fault. This fault outlines a semicircle east of the Monashee complex that is centred on Revelstoke and has a radius of about 25-30 km. The timing, direction, and style of motion on the Standfast Creek fault are enigmatic, as the fault has been interpreted in a variety of ways (see section below on previous work).

Major plutonic pulses in the Selkirk allochthon east of the Monashee

complex occurred in the Middle Jurassic and mid-Cretaceous (Read and Wheeler, 1976; Archibald et al., 1983, 1984; Parrish and Wheeler, 1983; Brown et al., 1991). Devonian and Late Cretaceous intrusions, which occur in the Selkirk allochthon west of the Monashee complex (Parrish, 1992), occur only in the Clachnacudainn terrane east of the Monashee complex. First and second phase large-scale folds in the Selkirk allochthon are west-verging (Simony et al., 1980; Raeside and Simony, 1983; Brown et al., 1983; Brown, 1991) and are interpreted to have been formed during accretion of the accreted terranes and therefore prior to initial eastward displacement on the Monashee decollement (Brown and Lane, 1988). In the Kootenay Arc, southeast of the Clachnacudainn terrane, there is evidence of middle Paleozoic deformation (Read and Wheeler, 1976). In the northern Selkirk Mountains, major deformation and associated regional metamorphism have been ascribed to Middle Jurassic (Brown et al., 1986) and Cretaceous (Price, 1986) tectonic events. In the Monashee Mountains north and south of the Monashee complex, sillimanite grade metamorphism is Late Cretaceous to Paleocene (Parrish et al., 1988; Sevigny et al., 1990).

Extension

The Shuswap metamorphic complex in the southern Omineca Belt has been defined by most workers as the rocks bounded by the sillimanite isograd. The high grade culminations in the Shuswap metamorphic complex were interpreted as Cordilleran metamorphic core complexes by Coney (1980) and Armstrong (1982). It was subsequently recognized that the high grade culminations within the Shuswap metamorphic complex were exhumed on low- to moderate-angle ductile-brittle normal faults during Eocene crustal extension (Journeay and Brown, 1986; Carr et al., 1987; Parrish et al., 1988). Parrish et al.

estimated that the normal fault systems which bound the metamorphic complexes are fundamental crustal breaks which accommodated about 30% extension across the 300 km width of the southern Omineca Belt. These normal faults, which were active within a few million years after cessation on the compressional shear zones (Carr, 1990), juxtapose lower crustal levels characterized by Eocene mica cooling dates against upper levels with Mesozoic dates (Parrish et al., 1988).

The Columbia River fault is an east-dipping Eocene normal fault that bounds the eastern margin of the Monashee complex (Read and Wheeler, 1976; Read and Brown, 1981; Lane, 1984, Parrish and Armstrong, 1987). Displacement along the fault south of the Clachnacudainn terrane is estimated at 20-30 km by Parrish et al. (1988), contrasting significantly with the 1-10 km of brittle displacement proposed by Lane (1984) for the segment which overprinted the Monashee decollement north of Revelstoke. This segment of the Columbia River fault is also peculiar because it contains Eocene mica cooling dates in its hanging wall (Read and Brown, 1981; Parrish et al., 1988), part of which is the Clachnacudainn terrane.

The tectonic history of the Clachnacudainn terrane has been the subject of considerable debate recently because it is bounded above and below by enigmatic faults. The goal of my thesis is to better constrain the tectonic evolution of the terrane by focusing on its upper boundary, the Standfast Creek fault.

Studies within and along the flanks of the Clachnacudainn terrane

The nature of the contact in the vicinity of Albert Canyon (Maps 1 and 2, in back pocket) between a variety of intrusive rocks, orthogneiss, and metasedimentary rocks to the west and lower Paleozoic metasedimentary rocks to the east has been enigmatic to geologists for nearly 80 years. This contact, which

has received considerable attention because it crosses the Canadian Pacific Railway and the Trans-Canada Highway, has been interpreted as unconformable, intrusive, and faulted. Daly (1915) suggested that metasedimentary rocks lie with marked unconformity on highly metamorphosed basement orthogneiss of probable Archean age in Albert Canyon. Gunning (1929) examined the contact over a much greater north-south distance than Daly and concluded that the intimate manner in which the orthogneiss is interbedded with metasedimentary rocks indicates an intrusive relationship. He termed the area west of Albert Canyon a "granite-gneiss-sediment complex", recognizing that it includes highly metamorphosed sediment of the Lardeau Group in addition to granitoids. Okulitch (1949) examined the same Albert Canyon section and also argued against an unconformable basement-cover contact by concluding that the high metamorphic grade of the granite-gneiss-sediment complex resulted from its proximity to the Shuswap complex.

Wheeler (1963, 1965) mapped the granite-gneiss-sediment complex northwest of Albert Canyon into the Clachnacudainn Range, where he recognized that hornblende biotite granodiorite gneiss is dominant, with granite, leucogranite, pegmatite, metasedimentary rocks, and hornblende diorite also composing significant portions. He termed the granite-gneiss-sediment complex a "migmatite complex in the Clachnacudainn salient", using the term "salient" because he considered these rocks to be a structural and metamorphic culmination that protrudes eastward from the sillimanite grade rocks of the Shuswap complex. Metasedimentary rocks in the complex were correlated with the lower Paleozoic Hamill Group, Badshot Formation, and Lardeau Group. He also correlated the Hamill Group with quartzite in the core of the Albert Canyon anticline (a west-verging structure overlying the salient) and the Badshot Formation with marble in

its limbs. Wheeler (1963) noted that the lower limb of the Albert Canyon anticline was sheared off and suggested that it may lie along a fault that separates structural levels. Wheeler (1965) suggested that this highly sheared zone is part of a similar, if not the same, mylonitic fault zone that underlies much of the Columbia River valley north of LaForme Creek. This correlation was made on the basis that both fault zones apparently juxtapose high grade rocks to the west against low grade rocks to the east. This suggestion prompted later workers to map within and along the flank of the Clachnacudainn salient. The locations of the following fieldworkers' map areas are shown in Figure 1-2.

Based on mapping in the Clachnacudainn Range, Ross (1968) postulated that a slice of Precambrian basement orthogneiss was tectonically wedged upward into cover metasedimentary rocks, forming a large east-verging nappe along the northeast flank of the salient. He suggested that a west-dipping mylonitic fault formed at the base of the wedge because of a ductility contrast between basement and cover. Gilman (1972) investigated the nature of the salient west of Albert Canyon and concluded that a sheet of concordant, folded orthogneiss was emplaced between two contrasting panels of metasedimentary rocks. He observed the same evidence that Gunning (1929) had used to interpret the contact between the orthogneiss and metasedimentary rocks as intrusive, but suggested that infolding and transposition of the orthogneiss and metasedimentary rocks could account for the interlayering relationships.

Subsequent mappers (Thompson, 1972; Zwanzig, 1973; Sears, 1979) contended that a transition between contrasting tectono-stratigraphic, structural, and metamorphic regimes occurs near the flank of the Clachnacudainn salient. The deeper tectonic level, termed the infrastructure, is characterized by high grade metamorphic rocks and structures that resulted mostly from vertical

flattening. The Clachnacudainn salient was included within the infrastructure, in agreement with Wheeler's (1963, 1965) contention that it is a metamorphic and structural culmination that protrudes eastward from the Shuswap complex. In contrast, the upper level, termed the superstructure, was characterized by lower grade metamorphic rocks and structures that resulted from horizontal compression. Metasedimentary rocks overlying the salient were included within the superstructure because they are mostly low grade rocks that were deformed by thrust faults and upright folds.

Thompson (1972) mapped a low-angle, southwest-directed extensional fault as the boundary between the infrastructure and superstructure on the southern flank of the Clachnacudainn salient, in the Akolkolex River area. He named this fault the "Standfast Creek fault", which has also been termed the "Standfast Creek slide" by later workers. Thompson considered the Standfast Creek fault to be a significant boundary because he interpreted the hanging wall and footwall as having "major differences in the character and succession of the rock units, as well as in the degree of metamorphism and structural fabric" (p. 13). No physical features of the fault were described and the timing and amount of displacement were not estimated. The Hamill Group, Badshot Formation, and Lardeau Group were correlated by Thompson with metasedimentary rocks in the hanging wall, whereas he concluded that the lack of marker units in the footwall made correlation largely indeterminable. Thompson observed footwall metasedimentary rocks tightly folded within the orthogneiss that forms a nearly conformable, subhorizontal underlying unit; the contact was described as sharp with no evidence of mylonitization. Another notable conclusion by Thompson is that although the Clachnacudainn salient borders his map area to the north, it had apparently minor influence on metamorphic grade, fold orientation, and fold style.

Zwanzig (1973) contended that the west-verging Albert Canyon anticline is the transition zone between the infrastructure and superstructure on the east flank of the Clachnacudainn salient. He could not confirm whether the orthogneiss was intrusive into or faulted against the overlying metasedimentary rocks because its protolith and contacts were highly altered during metamorphism. Although Zwanzig mapped the same fault that Wheeler (1963, 1965) noted in the lower limb of the Albert Canyon anticline, he termed it an "unsubstantiated fault" (p. 84) and suggested that the lower limb is composed of a series of unfaulted, appressed isoclinal folds. Mapping by Zwanzig of the Hamill Group, Badshot Formation, and Lardeau Group in both the immediate hanging wall and footwall of the fault suggested that displacement across it is minimal.

J.W. Hakkinen (unpublished data, 1976, 1977) mapped the metasedimentary screens and intrusive rocks in the Clachnacudainn Range. He also mapped two subparallel faults in the eastern Clachnacudainn Range. The western fault is the contact between the Devonian orthogneiss and overlying metasedimentary rocks; this contact was also mapped as a fault by Ross (1968). The eastern fault is a northwestward continuation of the fault mapped by Wheeler (1963, 1965) and Zwanzig (1973) in Albert Canyon. According to Hakkinen's correlations, the eastern fault is extensional because it juxtaposes the Hamill Group against the base of the overlying Lardeau Group.

Sears (1979) mapped a fault in the Albert Peak area that has an early history of east-directed extension, a later episode of west-directed compression, and was post-dated by intrusion of the Albert stock. According to Sears' correlations, the fault juxtaposes the Hamill Group against the base of the overlying Lardeau Group. A boulder of brecciated schist in a calcareous matrix that Sears assumed was derived from the fault zone is the only physical feature of

the fault which he described. Sears postulated that the Standfast Creek fault is a composite of the fault that he mapped in the Albert Peak area plus those mapped i) by Wheeler (1963, 1965), Zwanzig (1973), and Hakkinen (unpublished data, 1976, 1977) in the lower limb of the Albert Canyon anticline and ii) by Thompson (1972) in the metasedimentary rocks south of the Illecillewaet River. These faults were correlated by Sears because each is apparently marked by an omission of stratigraphy and/or by a steep metamorphic gradient. The Akolkolex River fault, located east of the southern portion of the Standfast Creek fault and joined to it 1 km south of the Albert stock, was interpreted as the boundary between the infrastructure and superstructure. The Clachnacudainn salient was therefore defined by Sears as the entire footwall of the Akolkolex River fault and the Standfast Creek fault north of where it joins the Akolkolex River fault. This definition of the salient grouped together the greenschist to amphibolite grade metasedimentary rocks mapped by Thompson south of the Illecillewaet River with the metamorphic and structural culmination (Clachnacudainn salient) mapped by Wheeler to the north. This definition of the salient has not been used by later workers. Instead, the salient has been defined as the footwall of the entire Standfast Creek fault. Some later workers renamed the salient the "Clachnacudainn slice" or "Clachnacudainn terrane" (the name that I use).

Read (1977) suggested that the Standfast Creek slide continues 90 km south along the Columbia River valley to connect with faults in the Nakusp area. This segment was later mapped as part of the Columbia River fault by Read and Brown (1981). Lower Paleozoic metasedimentary rocks were mapped by Read and Wheeler (1976) south of the Clachnacudainn terrane in the northern Kootenay Arc (Lardeau west-half map area) and by Brown (1991) north of the terrane in the northern Selkirk Mountains (Downie Creek map area). Read and

Thompson (1980) correlated metasedimentary rocks in the terrane in Thompson's (1972) map area with regional lower Paleozoic stratigraphy. This recognition of stratigraphic continuity across the Standfast Creek fault suggested to Read and Thompson that displacement between the salient and the Kootenay Arc is not major. Murphy (1980) and Brown and Murphy (1982) described granitic mylonite in the Monashee decollement along Sale Creek in the Columbia River valley north of Revelstoke. Based on mapping north of Revelstoke, Lane (1984) concluded that the Columbia River fault accommodated 1-10 km of brittle displacement, overprinted mylonite associated with the Monashee decollement, and was the lower boundary of the Clachnacudainn terrane. Seismic reflection data recorded along the Akolkolex River imaged a weak, east-dipping reflection that Cook (1986) tentatively interpreted as the Standfast Creek fault and a reflection in the terrane that he interpreted as an amphibolite layer mapped by Read and Thompson. Gillett (1990) documented east-directed kinematic indicators along the fault by studying microfabrics south of the Albert stock.

Geochronological investigations compose most of the recent work in the vicinity of the Clachnacudainn terrane. Eocene K-Ar mica cooling dates from the footwall and Mesozoic dates from the hanging wall of the Standfast Creek fault were reported in Read and Brown (1981), Lane (1984), and Parrish et al. (1988). Mid-Cretaceous K-Ar hornblende and biotite cooling dates from the Albert stock were reported in Price et al. (1985). U-Pb zircon dates from granitoids in the terrane were obtained by Lane (1984), Parrish (1992), and R.L. Armstrong (unpublished data).

Recent tectonic models

After completion of the extensive field mapping outlined above, there was

renewed interest in the Clachnacudainn terrane and the Standfast Creek fault due to the revelation by Read and Brown (1981) that they occur in the hanging wall of a regionally significant compressional fault (Monashee decollement) and extensional fault (Columbia River fault). This recent interest spawned three tectonic models for the Clachnacudainn terrane and Standfast Creek fault. These models are shown in schematic west-to-east cross-sections in Figure 1-3.

Read and Brown (1981) model

Read and Brown (1981) interpreted the Standfast Creek fault as an east-directed Middle Jurassic thrust fault which was active during initial stages of east-directed transport of the Selkirk allochthon on the Monashee decollement (Fig. 1-3). The Clachnacudainn terrane in this model is a structural horse that was accreted to the hanging wall of the Monashee decollement. This model is compatible with the foreland-dipping duplex model for the Monashee complex proposed by Brown et al. (1986). Brown and Lane (1988) provided corroborating evidence for Read and Brown's model, such as outcrop and hand sample kinematic indicators and mapping along the northern margin of the terrane. Read and Thompson's (1980) correlations of metasedimentary rocks in the terrane were used as evidence that the terrane has stratigraphic continuity with the remainder of Selkirk allochthon. Read and Brown therefore considered the terrane as a "slice" of the Selkirk allochthon east of the Monashee complex; the terrane is overlain to the east and south by the Illecillewaet slice and to the north by the Goldstream slice. Henceforth, I will refer to the entire hanging wall of the Standfast Creek fault as the Illecillewaet slice.

Price (1986) model

Price (1986) interpreted the Standfast Creek fault as a west-directed Cretaceous thrust fault (Fig. 1-3). This model contends that the crust was delaminated during northeastward insertion of a tectonic wedge by synchronous motion on the Standfast Creek fault and an east-directed sole thrust. The Clachnacudainn terrane in this model is the leading edge of the tectonic wedge, which Price termed a "suspect terrane" because of its uncertain stratigraphic affinity. The main evidence for the model is the southwest-verging folds in the base of the Illecillewaet slice, such as the Albert Canyon anticline (Wheeler, 1963, 1965; Zwanzig, 1973), which Price contended were formed during insertion of the wedge beneath them. Price et al. (1985) ascribed the pressure reduction between the Barrovian- and Buchan-style metamorphic events in the Albert Peak area to uplift and erosional unroofing that occurred during wedging. Price was the first to suggest that the Standfast Creek fault was bifurcated north of the Albert stock into two subparallel branches. The western branch lies at the contact between mixed granitoids and orthogneiss in the Clachnacudainn Range and metasedimentary rocks to the east, along the fault mapped by Ross (1968) and Hakkinen (unpublished data, 1976, 1977). I refer to this branch as the "lower Standfast Creek fault". The eastern fault branch (referred to herein as the "upper Standfast Creek fault") includes the fault mapped by Wheeler (1963, 1965), Zwanzig (1973), and Hakkinen in the lower limb of the Albert Canyon anticline.

Price's (1986) wedge model differs drastically from the Read and Brown (1981) model because it requires that displacement on the Standfast Creek fault was i) responsible for underthrusting the Monashee complex and Clachnacudainn terrane beneath the Selkirk allochthon, ii) synchronous with motion on the Monashee decollement, and iii) synchronous with and postdates regional metamorphism. Although the Read and Brown (1981) and Price models disagree

on the major sense and timing of Standfast Creek fault displacement, they agree that minor, brittle, east-directed extensional reactivation on the Standfast Creek fault occurred in the Eocene.

Parrish et al. (1988) model

Parrish et al. (1988) inferred that the Standfast Creek fault is an east-directed Eocene normal fault (Fig. 1-3). In this model the Standfast Creek fault is a ductile to ductile-brittle splay from the extensional Columbia River fault that tectonically denuded and rapidly cooled the Monashee complex and Clachnacudainn terrane. The terrane is therefore interpreted as a high grade salient from the Shuswap metamorphic complex. Evidence for this model is a discrepancy in K-Ar mica and hornblende cooling dates across the Standfast Creek fault; the footwall has Eocene mica cooling dates and the hanging wall has Mesozoic dates (Read and Brown, 1981; Lane, 1984; Parrish et al., 1988). Parrish et al. interpreted these data to imply that in the Eocene the Standfast Creek fault unroofed the Monashee complex and Clachnacudainn terrane, which were at high temperatures, and juxtaposed them against the overlying Illecillewaet slice, which had cooled in the Mesozoic. Parrish et al. therefore considered the Standfast Creek fault to be a crucial part of Tertiary extension at 51°N, accommodating a large amount of displacement that is apparently absent along the brittle normal fault in the Columbia River valley (Lane, 1984). In order to comply with the mid-Cretaceous K-Ar hornblende cooling date from the Albert stock (reported in Price et al., 1985), which post-dated the Standfast Creek fault (Sears, 1979), Parrish et al. speculated that the ductile extensional branch of the fault skirts the stock along the contact between the orthogneiss and the overlying lower Paleozoic metasedimentary rocks; this is the same contact observed by Daly (1915) and

subsequently reinterpreted by Gunning (1929).

Figure 1-1. Tectonic map of southeastern British Columbia, modified after Parrish et al. (1988). Box in inset outlines the map location in British Columbia with respect to the Omineca Belt and allochthonous Superterranes I and II of Monger et al. (1982). Thrust faults are the heavy lines with teeth on the hanging wall, including the Rocky Mountain Trench (RMT) and the Monashee decollement (MD). Hanging wall of the Monashee decollement is the Selkirk allochthon, which includes the Clachnacudainn terrane (C), Gold Stream slice (G), and Illecillewaet slice (I). The Standfast Creek fault (SCF) is the upper boundary of the Clachnacudainn terrane. On the east side of the Monashee complex (MC), the Monashee decollement is coincident with the east-dipping Columbia River normal fault (CRF), which has pegs on its hanging wall. Okanagan valley fault (OVF) is a normal fault west of the Monashee complex. Map units are 1) Aphebian core gneiss of the Monashee complex, including the Frenchman Cap dome (FCD) and Thor-Odin dome (TOD); 2) mid-Late Proterozoic Belt-Purcell Supergroup; 3) mantling gneiss of the Monashee complex; 4) Late Proterozoic Windermere Supergroup and Paleozoic metasedimentary rocks of North American affinity; 5) Paleozoic granitoid; 6) Middle Jurassic granitoid, including the Adamant batholith (AB), Fang stock (FS), Galena Bay stock (GBS), and Kuskanax batholith (KB); 7) mid-Cretaceous granitoid, including the Albert stock (AS) and Battle Range batholith (BRB); 8) early Middle Eocene Ladybird granitoid; 9) Miocene-Pliocene basalt; 10) late Paleozoic-early Mesozoic allochthonous terrane and Paleozoic allochthonous ultramafic rocks; 11) intrusive rocks of unknown age in the Clachnacudainn terrane.

Figure 1-3. Recent tectonic models shown in schematic cross-sections

ACA - Albert Canyon anticline	MC - Monashee complex
AS - Albert stock	MD - Monashee decollement
CRF - Columbia River fault	OVF - Okanagan valley fault
CT - Clachnacudainn terrane	SCF - Standfast Creek fault
IS - Illecillewaet slice	

CHAPTER 2. Lithologic description and correlation

Introduction

Previous fieldworkers readily correlated metasedimentary rocks in the immediate hanging wall of the Standfast Creek fault with lower Paleozoic stratigraphy in the Kootenay Arc (Fyles and Eastwood, 1962; Fyles, 1964) because the fossiliferous Lower Cambrian Badshot Formation was identified in many localities (Thompson, 1972; Read and Wheeler, 1976; Brown, 1991). Once this age control was established, lithologic marker units above and below the Badshot Formation were identified and used to correlate metasedimentary rocks in unfossiliferous sections. Previous fieldworkers generally did not firmly correlate metasedimentary rocks in the Clachnacudainn terrane with regional stratigraphy because neither the Badshot Formation nor any other marker unit was identified. Read and Thompson (1980) were an exception, as they identified unfossiliferous marker units in the terrane that were correlated with stratigraphy to the south (Read and Wheeler, 1976), east (Thompson, 1972; Zwanzig, 1973; Sears, 1979), and north (Brown, 1991).

I mapped the same marker units that were identified by Read and Thompson (1980) and additional marker units in the Clachnacudainn terrane, and therefore agree with their correlation between terrane and Illecillewaet slice stratigraphy. Although terrane stratigraphy can be traced continuously across the Standfast Creek fault in only one locality, it is interpreted as continuing upward into the base of the Illecillewaet slice over major west-verging anticlines that

occur at the level of the fault (see Chapter 5). Stratigraphy also outlines a major west-verging fold in the terrane. The proposed correlations require that the terrane has stratigraphic continuity with the Illecillewaet slice, suggesting that displacement between them on the Standfast Creek fault is not major.

In this chapter, lithologic descriptions by previous workers of formations in the base of the Illecillewaet slice are followed by description and discussion of the units in the Clachnacudainn terrane with which I correlate them. Table 2-1 presents these correlations in a condensed form.

Hamill Group

The unfossiliferous Hamill Group in the Kootenay Arc overlies the Upper Proterozoic Windermere Supergroup with local conformity and is overlain conformably by the fossiliferous Lower Cambrian Badshot Formation (Fyles and Eastwood, 1962; Fyles, 1964; Read and Wheeler, 1976). The Hamill Group has been divided by previous workers into the following conformable sequence: blocky and massive quartzite of the basal Mount Gainer Formation, grey and brown micaceous quartzite and schist of the Marsh Adams Formation, and siliceous and calcareous phyllite and limestone of the Mohican Formation (Fyles and Eastwood, 1962; Fyles, 1964; Read and Wheeler, 1976). Greenstone layers were also documented in the Hamill Group by many workers.

The Hamill Group, which is thought to be the oldest unit in the map area, crops out in the cores of major anticlines. Thompson (1972) correlated the Marsh Adams Formation with fine-grained, laminated tan and grey quartzite, grey and brown micaceous quartzite, and subordinate grey and brown quartzose phyllite in the core of the Akolkolex anticline (Plate 2-1). The lack of conspicuous thick white quartzite markers and the appressed nature of the core of the anticline

suggested to Thompson that the underlying Mount Gainer Formation is absent. The Mohican Formation is correlated with intercalated light grey quartzite, greenish grey phyllite, and dark grey calcareous phyllite in the limbs of the Akolkolex anticline (Thompson, 1972). Thompson noted that the Mohican Formation is commonly indistinguishable from the Marsh Adams Formation. Wheeler (1963) and Zwanzig (1973) correlated the Hamill Group with quartzite, phyllite, schist, marble, and greenstone in the core of the Albert Canyon anticline. Gilman (1972) correlated the Hamill Group with thick beds of white or light grey quartzite in the Clachnacudainn Range.

Thick layers of marble and siliceous clastics in the uppermost level of the Clachnacudainn terrane form the ridge between Ghost and Albert peaks, which is the divide between the Illecillewaet and Akolkolex rivers (Plates 2-2 and 2-3). This section is thickest at the head of East Twin Creek and thins considerably southwestward toward Ghost Peak and northeastward toward Albert Peak, where it is heavily intruded by the Albert stock. I agree with Sears' (1979) and Read and Thompson's (1980) correlation of the Hamill Group with the uppermost unit in this section because it is composed mostly of centimetre- to metre-thick beds of buff, grey, and tan micaceous and massive quartzite. The unit is thickest (1 km) at the head of East Twin Creek (Sears, 1979). On the east slope of upper Siegfried Creek, the unit is composed of medium (up to 1 m) beds of white and light blue quartzite with thin (usually less than 20 cm) intercalations of brown garnet biotite schist (Plate 2-4). Although Thompson (1972) assigned it an informal name, he noted that the unit closely resembles the Marsh Adams Formation. Since graded bedding on the ridge south of Albert Canyon suggests the that unit is upright (Sears, 1979), then the Mohican Formation of the Hamill Group may be represented by light green biotite chlorite siliceous phyllite that

locally lies adjacent to the Standfast Creek fault.

Sears mapped the quartzite-dominated unit northeastward from Standfast Creek for over 10 km, through the heavily intruded Albert Peak area, to South Albert Creek. At South Albert Creek, the unit abuts the Standfast Creek fault mapped by Sears (1979). Since Hamill quartzite crops out in the immediate hanging wall of the Standfast Creek fault at this locality (Zwanzig, 1973; Sears, 1979), I suggest that the Hamill quartzite mapped by Sears in the footwall is continuous with the Hamill quartzite in the hanging wall. This relationship requires that the Hamill quartzite crosses the Standfast Creek fault in South Albert Creek with little or no disruption. My hypothesis is corroborated by Gilman's (1972) mapping, which shows that the unit (composed of quartzofeldspathic schist with minor quartzite and marble) that underlies the quartzite-dominated unit is continuous across South Albert Creek around a fold closure. My hypothesis can only be disproved by mapping that reveals that either a shear zone in South Albert Creek disrupts the apparent continuity or Sears' correlation of the Hamill Group with the footwall quartzite is erroneous.

Below the uppermost, quartzite-dominated unit on the Illecillewaet-Akolkolex divide are thick layers of marble (up to 100 m) and siliceous clastics (up to 300 m). There are between three and seven layers of each lithology and the section is up to 800 m thick at the head of East Twin Creek (Plates 2-2 and 2-3). These marble and siliceous clastics layers were correlated with the Badshot Formation and the Hamill Group, respectively, by Read and Thompson (1980). Sears termed this section the "marble-pelite unit" and suggested that it may correlate with the middle marble member of the Horsethief Creek Group. I, however, correlate the Hamill Group with the lighter coloured siliceous clastics and the Index Formation of the Lardeau Group with the darker coloured siliceous

clastics. The Hamill Group at this locality is composed of light brown semipelitic and pelitic schist, whitish grey micaceous quartzite, thick-bedded grey to white quartzite, and minor amphibolite, whereas the Index Formation mostly consists of dark grey graphitic phyllite (see section on the Index Formation below for more detail). These layers are considerably less siliceous and more pelitic than the overlying quartzite-dominated unit. Thompson (1972) also noted that these lithologies are more pelitic than the Marsh Adams Formation and suggested that they represent the Mohican Formation or part of the Hamill not previously separated. Below the lowest marble layer in the Albert Peak area are semipelitic gneiss and micaceous quartzite that occupy the stratigraphic position of the Index Formation, but the distinctive markers necessary for firm correlation with the Index were not observed (see below).

Thompson (1972) noted that the axial surface of the west-closing, antiformal Drimmie Creek syncline is outlined directly southeast of Ghost Peak by a thin (15-35 m) band of siliceous clastics. This east-dipping band is definitely the core of a west-closing antiform because it is sandwiched between layers of Badshot Formation marble that converge westward and close within Ghost Peak (Plates 2-5, 2-6, and 2-7). The Badshot marble in the lower limb of the antiform is everywhere underlain by the Standfast Creek fault mapped by Thompson (1972; Plate 2-8). The sense of vergence of the antiform therefore depends solely on the correlation of the core. Thompson mapped the antiform as an east-verging syncline because he correlated the core with the Index Formation, which regionally overlies the Badshot Formation. On the contrary, I interpret the antiform as a west-verging anticline because I correlate the core with the Hamill Group, which regionally underlies the Badshot Formation. I rename the structure the Ghost Peak anticline, after the rugged horn in which it closes (Plate 2-8). The

basis for my correlation of the core is the predominance of medium- to thick-bedded (generally more than 1 m), non-calcareous, clean, tan to light blue quartzite and micaceous quartzite (Plate 2-7) that bear strong resemblance to the Hamill Group in the core of the Akolkolex anticline (Plate 2-1). The Mohican Formation may be represented by minor interlayers of light green garnet biotite chlorite schist and impure marble. Moreover, dark grey graphitic phyllite that typifies the Index Formation where it is in contact with the underlying Badshot Formation marble in nearby localities is entirely absent from this section. The Index Formation includes quartzite, but it invariably occurs stratigraphically higher in the formation and does not crop out adjacent to Badshot marble in the Ghost Peak area. Massive and micaceous quartzite in the cliffs directly east of Ghost Peak are also readily correlated with the Hamill Group and therefore outline the axial surface of the Ghost Peak anticline.

Badshot Formation

The Lower Cambrian Badshot Formation is an archaeocyathid-bearing shelf carbonate unit in the Kootenay Arc that conformably overlies the Hamill Group and underlies the Lardeau Group (Fyles and Eastwood, 1962; Fyles, 1964; Read and Wheeler, 1976). The four archaeocyathid localities reported in Thompson (1972) firmly correlate the limestone in the Akolkolex River valley with the Badshot Formation. The Badshot Formation in the hanging wall of the Standfast Creek fault is finely bedded, grey and dark grey, fine-grained homogeneous limestone intercalated with bands and lenses of coarsely crystalline white marble (Thompson, 1972). Badshot marble in the Ghost Peak area forms the limbs and thickened hinge zone of the west-verging Ghost Peak anticline (Plates 2-5, 2-6, and 2-7), the lower limb of which is underlain by the Standfast

Creek fault mapped by Thompson (1972; Plate 2-8). Wheeler (1963) and Zwanzig (1973) correlated the Badshot Formation with limestone in the limbs of the Albert Canyon anticline. The Badshot marble in the upper limb is overlain by black phyllitic limestone of the Lade Peak Formation and the lower limb is underlain by undifferentiated Lardeau Group (Zwanzig, 1973). Gilman (1972) correlated the Badshot Formation with marble in the Clachnacudainn Range.

I agree with the correlation by Wheeler (1963) and Read and Thompson (1980) of the Badshot Formation with marble layers that crop out in the upper level of the Clachnacudainn terrane along the Illecillewaet-Akolkoex divide. The three to seven marble layers form impressive palisades on the northwest side of the ridge between Ghost and Albert peaks and dip slopes on the southeast side (Plates 2-2 and 2-3). The thickness of each layer generally varies between 25 and 50 m with a maximum of 100 m (Sears, 1979). Thompson (1972) informally named the marble layers but noted that their composition, colour, and texture suggest that they are correlative with the Badshot Formation. Unfortunately, there is no fossil evidence to prove this correlation. Sears (1979) suggested that these layers are part of the middle marble member of the Horsethief Creek Group. He mapped the marble layers from the ridge between Ghost and Albert peaks northeastward to the Standfast Creek fault in South Albert Creek. I suggest that these marble beds are correlatable across the Standfast Creek fault with discontinuous beds of Badshot marble in the limbs of the Albert Canyon anticline (Zwanzig, 1973). The Hamill Group at this locality is also apparently continuous across the fault (see above). This hypothesis requires that Badshot marble layers in the Clachnacudainn terrane cross the Standfast Creek fault into the Illecillewaet slice without significant offset.

I suggest that the numerous layers of Hamill Group, Badshot marble, and

Index Formation along the Illecillewaet-Akolkolex divide (Plates 2-2 and 2-3) were duplicated by isoclinal folding related to the west-verging Ghost Peak anticline, which is interpreted as continuing northward as the west-verging Albert Canyon anticline. Dark grey lithologies that I correlate with the Index Formation probably are the core of an appressed west-verging syncline noted by Thompson (1972). Thompson also found stratigraphic evidence for a west-verging anticline at the head of Standfast Creek. However, he concluded that the remaining clastic layers do not exhibit the consistent alternation that is expected to result from folding, possibly a result of high variability in clastic lithology both along and across strike. Sears (1979) correlated the Hamill Group with the uppermost, quartzite-dominated unit and the Horsethief Creek Group with the underlying marble and siliceous clastic layers. Sears' alternative correlation, that the sequence is largely inverted and isoclinally folded repetitions of Hamill, Badshot, and Lardeau strata, was based on the realization that the downward progression from massive quartzite to marble to black phyllite which occurs along the southeast side of the Illecillewaet-Akolkolex divide is similar to the Hamill-Badshot-Lardeau sequence. However, Sears considered this hypothesis unlikely because the symmetrical repetition of units and major fold hinges were not detected.

Lardeau Group

The unfossiliferous lower Paleozoic Lardeau Group was divided into the conformable Lade Peak, Index, Jowett, and Broadview formations by some workers in the Illecillewaet slice (Fyles and Eastwood, 1962; Fyles, 1964; Zwanzig, 1973; Sears, 1979). Other workers did not distinguish the Lade Peak Formation (Thompson, 1972; Read and Wheeler, 1976; Brown, 1991). It has generally been

accepted that the Lardeau Group conformably overlies the Lower Cambrian Badshot Formation and is unconformably overlain by the upper Mississippian to Pennsylvanian or Permian Milford Group. Gehrels and Smith (1987) and Smith and Gehrels (1991) recently suggested that there is a fault between the Badshot Formation and the Lardeau Group, which they contended is inverted. I suggest that there is a depositional contact between the Badshot Formation and Lardeau Group in my map area because clasts of marble in the base of the Index Formation are interpreted to be derived from the Badshot Formation by sedimentary processes rather than shearing (Plate 2-9). Whether the contact is faulted or conformable does not affect the following correlations.

Lade Peak Formation

Fyles and Eastwood (1962) suggested that grey and black calcareous phyllite and limestone of the Lade Peak Formation are laterally equivalent to the Badshot Formation in the Ferguson area. Zwanzig (1973) tentatively correlated the Lade Peak Formation with black argillaceous limestone and phyllite in the Illecillewaet synclinorium. He interpreted the Lade Peak Formation as the base of the Lardeau Group because it gradationally overlies Badshot marble in the upper limb of the Albert Canyon anticline. Since the Lade Peak Formation does not continue over the anticline to underlie Badshot marble in the lower limb (amphibolitic mica schist correlated with undifferentiated Lardeau Group occupies this position), there must be a shear zone along the lower limb or a facies change in order to account for its omission. Sears (1979) also mapped the Lade Peak Formation as the base of the Lardeau Group, describing it as fine-grained carbonaceous marble with fine-grained siliceous and pelitic bands a few centimetres thick. In the Albert stock contact aureole, the marble layers are

interbedded on a larger scale with fine-grained, rusty weathering siliceous rocks, which Sears interpreted as hornfels equivalents of siltite and siliceous pelitic schist. The Lade Peak Formation is mapped by Sears as detached from the underlying Badshot marble by bedding plane faults (Standfast Creek and Akolkolex River faults) and overlain by the Index Formation along its entire outcrop length. The Lade Peak Formation is not laterally extensive, as it is absent in the Ghost Peak area (Thompson, 1972).

I found that the Lade Peak Formation in the east slope of upper Siegfried Creek valley is characterized by alternating 3-20 m thick calcareous and siliceous layers (Plates 2-10 and 3-7). The calcareous layers are composed of thinly banded, light grey and white, fine-grained carbonaceous siliceous diopside marble that weathers white to buff. The rusty weathering, dark grey siliceous layers are composed mostly of fine- to medium-grained graphite diopside quartzite. The distinctive light grey and rusty weathering of these layers can be used to identify the Lade Peak Formation from considerable distances (Plate 2-10). In this locality, the Lade Peak Formation overlies medium-bedded quartzite and pelitic schist which are correlated with the Hamill Group by Sears (1979) and this study (see above). This contact, which is poorly exposed and heavily intruded by numerous late-tectonic (post-dates all penetrative ductile deformation) phases of the Albert stock, is the Standfast Creek fault mapped by Sears (1979).

Index Formation

The Index Formation is characterized by the predominance of dark grey graphitic phyllite and a multitude of subordinate heterogeneous lithologies. The Index is the dominant formation in the map area. Thompson (1972) described a basal light grey phyllite in the base of the Illecillewaet slice that grades upward

into the dark grey and black phyllite which constitutes most of the formation. Distinctive lithologies in the middle to top of the formation are dark grey fine-grained silty marble, quartzite, homogeneous greenstone, and greenish grey phyllite with albite porphyroblasts (Thompson, 1972). Zwanzig (1973), Read and Wheeler (1976), Sears (1979), and Brown (1991) mapped the Index as mostly dark grey graphitic phyllite with locally abundant limy phyllite, light green phyllite, siliceous phyllite, phyllitic limestone, quartz grit, and greenstone.

I agree with Read and Thompson's (1980) correlation of the Index Formation with the variable lithologies in the uppermost level of the Clachnacudainn terrane in the Ghost Peak area. This section, which I name the Ghost Peak section, overlies the amphibolite layer that is correlated with the Jowett Formation (see below) and underlies the Badshot marble in the hanging wall of the Standfast Creek fault (Plate 2-8). The Ghost Peak section is therefore in the inverted lower limb of the west-verging Ghost Peak anticline (see above and Chapter 5 on deformation for description of this fold). Thompson (1972) noted a variation within the section (which he informally named because correlatable marker layers were not identified) from mostly dark coloured phyllite in the upper part to a more siliceous lower part. The upper part, the immediate footwall of the Standfast Creek fault, is predominantly composed of dark grey to black graphitic phyllite that often contains thin (0.1-3 cm) layers of quartz and calcite, and partially oxidized pyrite cubes (Plate 2-11). This lithology is highly fractured and silicified in a brittle fault zone that overprints the Standfast Creek fault within Ghost Peak and cuts down section to the south (Plate 5-2). Light green garnet chlorite phyllite and schist also constitute a significant portion of the upper part. The boundary between the upper and lower parts roughly occurs in the vicinity of two to seven marble layers that are 1-15 m thick and vary from

banded light blue and white marble, to brown impure carbonaceous marble, to light blue and silver micaceous marble. These marble layers are seen in Plate 2-8. Although dark grey graphitic phyllite is locally abundant beneath the marble layers, the lower part is significantly more siliceous than the upper. The lower part is composed mostly of light grey and reddish brown semipelitic and pelitic schist that is thinly interbedded (typical beds are a few centimetres thick) with tan to light grey quartzite and micaceous quartzite. Thick-bedded (2-4 m) tan quartzite, interbedded dark grey graphitic garnet schist and grey quartzite, and minor marble and amphibolite also overlie the amphibolite layer correlated with the Jowett Formation (see below).

I also correlate the Index Formation with lithologies that occur along the Illecillewaet-Akolkolex divide and in the hanging wall of the lower Standfast Creek fault north of the Illecillewaet River. These correlations are based on lithologic similarity with the Index Formation observed in the immediate hanging wall of the Standfast Creek fault and with the Index described by Read and Wheeler (1976) and Brown (1991). In particular, the significant amount of dark grey graphitic phyllite present in these sections is only reported in the Index Formation. Lithologies that occupy the assumed stratigraphic position of the Index Formation in the Albert Peak area are less firmly correlated with the Index because they lack dark grey phyllite.

Along the Illecillewaet-Akolkolex divide, Thompson (1972) mapped the following dark grey to black lithologies: calcareous phyllite, siliceous phyllite, slate, pelitic schist, calcareous phyllitic quartzite, and coarsely crystalline limestone. Although he noted that these lithologies are typical of the Index Formation, he informally named them. Read and Thompson (1980) correlated the Hamill Group with these dark and lighter lithologies, and the Badshot

Formation with the intervening light grey to white marble (see above). Along strike to the northeast of the section described by Thompson, Sears (1979) mapped a 100 m thick jet black siltite unit that pinches out or becomes more chloritic to the north, on the flank of Albert Peak. Based on Thompson's and Sears' descriptions and my observations of the unit at the head of Standfast and East Twin Creeks, I correlate the Index Formation with this unit because of its lithologic similarity. This unit is discernable in Plates 2-2 and 2-3 as a dark grey layer that lies adjacent to the light grey Badshot marble and tan Hamill quartzite.

Zwanzig (1973) correlated the Hamill Group, Badshot Formation, and Lardeau Group with rocks immediately above the orthogneiss in Albert Canyon, in the lower limb of the Albert Canyon anticline. Price (1986) placed part of this section between the lower and upper Standfast Creek faults as metasedimentary rocks of unknown age. Ross (1968) correlated the Horsethief Creek and Hamill groups with this section. Zwanzig (1973) placed it in the lower limb of the Albert Canyon anticline by noting that "about 2000 feet of dark schist, amphibolite and impure marble which lie structurally underneath the Badshot marble along Woolsey Creek and Albert Creek appear to represent an inverted succession of the Lardeau Group" (p. 85). Hakkinen (unpublished data, 1976, 1977) correlated the section on the ridge north of West Woolsey Creek (Plate 3-9) with (from top to bottom) the Badshot Formation, Lade Peak Formation, the Hamill Group, and marble and amphibolite that he did not correlate. He placed a fault between the Lade Peak Formation and Hamill Group. This fault was interpreted as the upper Standfast Creek fault by Price.

I agree with Zwanzig's interpretation of the section as inverted Lardeau Group by correlating the Jowett Formation with the basal amphibolite + marble + calc-silicate unit on the ridge north West Woolsey Creek (Hakkinen

(unpublished data, 1976, 1977) mapped this unit as uncorrelatable amphibolite and marble) and by correlating the Index Formation with variable lithologies that overlie the Jowett Formation and underlie the Badshot marble exposed farther east on the ridge (mapped as the Hamill Group and Lade Peak Formation by Hakkinen).

This section, which I name the West Woolsey Creek section, is continuous from the ridge north of West Woolsey Creek northwestward into the Mount LaForme area. The Index Formation in the West Woolsey Creek section is divided into a lower part that is characterized by quartzite and schist and a lithologically heterogeneous upper part. The lower part, which directly overlies the Jowett Formation, is composed predominantly of light grey and silver garnet chlorite biotite schist and quartz biotite schist, light green chlorite schist with coarse albite porphyroblasts, grey quartzite, and grey garnetiferous micaceous quartzite. The siliceous nature of this part may have been the reason why Hakkinen (unpublished data, 1976, 1977) correlated it with the Hamill Group. On the ridge south of Mount LaForme, the top of the lower part is marked by the lowest occurrence of dark grey graphitic phyllite and coincides with the upper Standfast Creek fault. On the ridge north of West Woolsey Creek, the top of the lower part is marked by the lowest and simultaneous occurrence of dark grey graphitic phyllite, amphibolite, and marble. This boundary is in the footwall of the upper Standfast Creek fault. The upper part of the section is differentiated from the lower by the common occurrence of calcareous and non-calcareous dark grey graphitic phyllite, amphibolite, pure and impure marble, and light green chlorite phyllite. Dark green actinolite calc-silicate and actinolite chlorite schist are minor components. Light green schist with coarse albite porphyroblasts and quartzite that occur in the lower part also crop out in the base of the upper part.

Amphibolite layers are less than 4 m thick and marble layers are generally less than 7 m thick, except for the 35 m thick, highly fractured marble layer at the locality of the upper Standfast Creek fault on the ridge north of West Woolsey Creek. Nearly identical lithologies outcrop on both sides of the fault. The exception is a higher percentage of clean white quartzite with pelitic schist and micaceous quartzite interbeds (2-10 cm thick) east of the fault. The reason why Hakkinen correlated these rocks with the Lade Peak Formation is unknown.

The Index Formation is less firmly correlated with lithologies that occupy the assumed stratigraphic position of the Index in the Albert Peak area because the distinctive dark grey graphitic phyllite was not observed. One possibility for the lack of graphitic phyllite is that contact metamorphism altered the phyllite beyond recognition (high temperature may have obliterated the graphite). The rocks that occupy the stratigraphic position of the Index are grey and light blue micaceous quartzofeldspathic gneiss, light grey micaceous quartzite, thick-bedded quartzite, and light purple and green garnet biotite schist with albite porphyroblasts. Sears (1979) described the section as grey and brown pelitic quartzite with partings and thin interlayers of muscovite biotite schist and interbeds of metre-thick light grey garnetiferous quartzite. This lithology can only be tentatively correlated with the Index Formation until marker units are identified.

Jowett Formation

The Jowett Formation in the base of the Illecillewaet slice south of the Akolkolex River is a dominantly volcanic unit composed of green phyllite, limy green phyllite, and greenstone (Read and Wheeler, 1976). Read and Wheeler mapped it as separating the grit unit of the Index Formation from the overlying

Broadview Formation. Zwanzig (1973) mapped grey phyllite that contains thin interbeds of limestone, local greenstone layers, and widely scattered intercalations of green phyllite. Although he noted that this unit occupies a similar stratigraphic position as the Jowett Formation in the Ferguson area (Fyles and Eastwood, 1962), he termed it simply the "phyllite unit". Brown (1991) mapped two units of the Jowett Formation in the Columbia River valley: a green phyllite, limy green phyllite, and greenstone unit dominates over a white and grey limestone or dolomite unit.

Thompson (1972) informally named a 10-30 m thick coarse-grained amphibolite layer with local lenses of marble and calcareous biotite schist in the upper level of the Clachnacudainn terrane, on the east slope of the Columbia River valley. This layer directly underlies the inverted Ghost Peak section that Read and Thompson (1980) and I correlate with the Index Formation (see above). The amphibolite, biotite chlorite calcite schist, and spatially associated marble of the Jowett Formation south of the Akolkolex River thins rapidly westward from about 400 m near Comaplix Mountain to 30 m on the south slope of the Akolkolex River valley, approximately 3 km from the amphibolite layer mapped by Thompson (1972) in the Clachnacudainn terrane (Read and Thompson, 1980). Since the thickness and lithology of these layers agree with those of the amphibolite layer mapped by Thompson (1972), Read and Thompson (1980) suggested that the Jowett Formation in the Illecillewaet slice extends northward across the Standfast Creek fault into the terrane. Read and Thompson (1980) also correlated the Jowett Formation with a similar-looking amphibolite layer mapped by Read (1980) that outcrops about 300 m below the layer mapped by Thompson (1972). They interpreted the two amphibolite layers as limbs of an overturned synform rather than faulted repetitions. Read (unpublished data,

1982) mapped the east-dipping lower limb around a fold closure (1.5 km south of Mount Mackenzie) to the upper limb, thereby corroborating Read and Thompson's (1980) interpretation of the two amphibolite layers as a single folded unit. Read and Thompson's (1980) correlation of the surrounding lithologies with the Index and Broadview formations hinges on this correlation of the amphibolite layers and requires that the fold is a west-verging syncline. I name this structure the Mount Cartier syncline, after the peak that dominates its core.

I agree with Read and Thompson's (1980) correlation of the Jowett Formation with the amphibolite + marble + calc-silicate units in the Clachnacudainn terrane that are described above. In addition, I correlate the Jowett Formation with the following amphibolite + marble + calc-silicate units in the terrane based on lithologic similarity. Numerous thin (2-15 m) amphibolite layers occur structurally beneath the lower Jowett layer identified by Read (1980) on the east slope of the Columbia River valley. A unit of light grey crystalline marble with associated calc-silicate and amphibolite outcrops in layers a few tens to about 100 of metres thick near the base of the metasedimentary section on the south through northwest flanks of North Albert Peak (Plate 2-12). Sears suggested that this unit may be equivalent to the amphibolite + marble + calc-silicate unit mapped by Thompson (1972) on the upper east slope of the Columbia River valley, which was later correlated with the Jowett Formation by Read and Thompson (1980). On the basis of a possible lateral equivalence and lithologic similarity, I correlate the Jowett Formation with this amphibolite-dominated unit on the flank of North Albert Peak. Zwanzig (1973) correlated the Lardeau Group and Badshot Formation with amphibolitic mica schist interbedded with thin impure marble layers and thick-bedded marble on the ridge north of West Woolsey Creek (Plate 3-9). Hakkinen (unpublished data, 1976, 1977)

mapped the marble and amphibolite but did not correlate them. I correlate the Jowett Formation with the marble and amphibolite based on lithologic similarity. This correlation is made stronger by the lithologic position of the layer; they underlie rocks to the east that I correlate with the Index Formation (see above). Late-tectonic granodiorite intrudes this Jowett Formation layer to the west.

The Jowett Formation in the Clachnacudainn terrane is characterized by amphibolite that is typically intercalated with marble and often with micaceous schist. The scale of intercalation varies greatly, as the common rock types are i) calcareous amphibolite that could also be termed an amphibole-rich calc-silicate schist; ii) garnet-bearing amphibolite schist with biotite laminations; iii) very thin-bedded (0.5-3 cm) amphibolite and marble; iv) metre-wide lenses and boudins of marble in amphibolite layers, and vice-versa; v) thin- to medium-bedded (0.1-10 m) amphibolite and marble (Plate 2-13); and vi) thick-bedded (up to 100 m) amphibolite and marble (Plate 2-12). The thickest amphibolite layers are typically massive, non-calcareous, coarse-grained, dark green plagioclase amphibolite and massive to foliated, non-calcareous, medium- to coarse-grained, green and white plagioclase amphibolite. The amphibole is hornblende and/or actinolite, with local tremolite. Marble beds are coarse, crystalline, and white, light grey, and buff coloured. Calcareous garnet biotite actinolite schist and chlorite phyllite are locally interbedded with amphibolite and marble. Granoblastic green calc-silicate that weathers rusty brown crops out on the ridge north of West Woolsey Creek.

The amphibolite + marble + calc-silicate association is interpreted as a combination of metamorphosed i) limestone; ii) mafic flows, sills, transposed dykes, and tuff; and iii) sediments derived from these calcareous and mafic sources. The Jowett Formation therefore most likely represents syn-sedimentary mafic volcanism on a carbonate platform. A rare metatuff unit may be

represented by thinly (less than 1 cm) laminated light green, grey, and tan layers composed predominantly of very fine-grained quartz with minor biotite (Plate 2-14). The metatuff, which was only observed in the southern part of the terrane, is intercalated on a metre scale with amphibolite.

The siliceous clastics that are interlayered with amphibolite + marble + calc-silicate of the Jowett Formation on the heavily timbered east slope of the Columbia River valley are correlated with the Broadview Formation (described below). Although uncertain without further mapping, I suggest that isoclinal folding is responsible for the duplication of the Jowett and Broadview formations. The thick (30-50 m) amphibolite layers above the lake at the head of East Twin Creek are not correlated with the Jowett Formation because they are not interbedded with marble or calc-silicate and they crosscut stratigraphy. These layers are probably post-sedimentary mafic dykes.

Broadview Formation

Thompson (1972) mapped the Broadview Formation above the Index Formation on the south slope of the Akolkolex River valley as green and dark grey phyllite with abundant recrystallized quartz pods and interbedded light brown feldspathic grit with conspicuous quartz "eyes". I found that the "eyes" at one locality are composed strictly of feldspar. Read and Wheeler (1976) correlated these same rocks with the quartz grit unit of the Index Formation that is continuous from its type area near Ferguson (Fyles and Eastwood, 1962) northward to the Illecillewaet synclinorium. Read and Wheeler (1976) mapped the Broadview Formation as mostly grey and green phyllitic grit and phyllite with minor calcareous interlayers. Zwanzig (1973) correlated the Broadview Formation with the grit unit, describing it as distinctive greenish grey quartz

wacke and feldspathic quartz wacke interbedded with green or grey phyllite. Sears (1979), working between the areas mapped by Zwanzig (1973) and Read and Wheeler (1976), suggested that the grit unit gradationally overlies the main body of the Index Formation phyllite and is distinct from and older than the Broadview Formation of the Ferguson area. Brown (1991) correlated the Broadview Formation with grey and green phyllitic grit and phyllite.

Thompson (1972) assigned informal names to the dominantly quartz-rich and pelitic units in the Clachnacudainn terrane in the east slope of the Columbia River valley but suggested possible correlations with the Hamill Group and part of the Upper Proterozoic Horsethief Creek Group. I agree with Read and Thompson's (1980) correlation of the Broadview Formation with some of these lithologies, which form the core of a west-verging syncline. This correlation hinges largely on the correlation of the Jowett and Index formations (see above) rather than independent identification of the Broadview Formation because it contains few marker layers.

I also correlate the Broadview Formation with the quartz- and mica-dominated rocks on the east slope of the Columbia River valley on the basis of lithologic similarity. In contrast to sections to the north and south, grit is rare in this section; the only grit observed is the few layers of light green gritty schist with feldspar granules up to 2 mm in diameter. Micaceous quartzite, quartz mica schist, and minor quartzite form the quartz-dominated rocks. Micaceous quartzite is characterized by planar, fine-grained, thin-bedded (1-5 cm) light blue, tan, and grey quartzite that is interlayered with thin (0.1-3 mm) micaceous laminae (chiefly muscovite and biotite). Micaceous quartzite is typified by layers that remain at constant thickness for up to 1 m along strike. In contrast to the micaceous quartzite, the thickness of the coarse-grained quartzite and micaceous layers in

the quartz mica schist varies drastically along strike; the layers vary from a 1 mm to few centimetres in thickness over a few centimetres along strike. The anastomosing micaceous layers are numerous and thick enough to impart a strong schistose fabric. The fresh foliation surfaces are brown, silver, light green, and greenish purple. Mustard yellow and orange are the typical colours on weathered surfaces. Medium-bedded (0.3-2 m) tan, light grey, and blue quartzite constitutes a minor portion of the quartz-dominated lithologies.

Mica-dominated schist composes about half of the Broadview Formation on the east slope of the Columbia River valley. The semipelitic and pelitic schist contain a large percentage of quartz, but it is interspersed with the abundant mica and therefore rarely forms monomineralic layers. The metamorphic mineral assemblage in the semipelitic schist is biotite-muscovite and the pelitic schist contains biotite-muscovite-garnet $\pm$ chlorite $\pm$ staurolite $\pm$ kyanite $\pm$ fibrolitic sillimanite. The schist is light green, brown, and silver, with dark grey graphitic, "Index-looking" schist also present in significant portions. Light green biotite chlorite phyllite and light grey quartz biotite garnet chlorite schist dominate the Broadview Formation in the upper level of the terrane, northwest of Ghost Peak.

I also correlate the Broadview Formation with lithologies on the west flank of North Albert Peak. The section underlies an amphibolite + marble + calc-silicate unit to the east that is correlated with the Jowett Formation (see above) and is juxtaposed against granodiorite gneiss to the west along a brittle, steeply east-dipping normal fault (Plates 5-7 and 5-8). Although Sears (1979) did not correlate this section, he noted that it underlies amphibolite + marble + calc-silicate layers that he suggested may be the northeast extension of a lithologically similar unit mapped by Thompson (1972) high on the east slope of the Columbia River valley. If true, then the Broadview Formation on the west flank of North

Albert Peak is inverted and laterally equivalent to the Broadview Formation on the east slope of the Columbia River valley described above. Determining through further mapping whether a continuity exists between the two sections may be difficult because of the lack of accessible exposure and the possibility that the entire formation is absent over this interval due to intrusion of orthogneiss.

This inverted section of Broadview Formation on the west flank of North Albert Peak is roughly divisible into upper and lower parts. Stratigraphic details of the section are hindered by structural complications due to third phase folds and by large gaps in the section due to the intrusion of diorite and granodiorite dykes from the Albert stock. The upper part, which is adjacent to the overlying Jowett Formation, is dominated by thinly interbedded semipelitic schist (1-2 cm) and quartzite (1-5 cm) and medium-bedded (10-50 cm) tan and white quartzite. Pelitic schist composes a minor amount of the upper part. The lower part of the section is dominated by green chlorite phyllite and schist, and micaceous quartzite. This part is also distinguished from the upper by the presence of minor grit, marble, and amphibolite. The 3 m thick grit layer has mono- and polycrystalline feldspar granules that are 1-3 mm in diameter and embedded in a fine-grained matrix of quartz, feldspar, and mica. The grit layer is underlain by metre-thick light grey marble layers that are intercalated with chlorite schist. Rare amphibolite ± marble layers are less than a few metres thick.

Table 2-1. Summary of stratigraphic correlations

	Formation	Akolkotlex River area Thompson (1972)	Albert Canyon Zwanzig (1973)	Lardeau west 1/2 map area Read and Wheeler (1976)
Lardeau Group	Broadview	Green and dark grey phyllite, local brown feldspathic grit	Grey and green sandstone and phyllite	Limestone and grey phyllitic limestone Grey phyllite, grey and green phyllitic grit and phyllite
	Jowett		Grey and black phyllite Minor meta-volcanic (dark green schist, light and dark green laminated phyllite, green soapstone, pink talc schist, meta-basalt)	Green phyllite, limy green phyllite, greenstone
	Index	Grey and black phyllite and slate; thin bands of blue-grey and black limestone and silty limestone Green epidote-chlorite-feldspar phyllite Grey and whitish grey quartzite, grey and black graphitic phyllite and slate, calcareous and siliceous phyllite and slate Greenish grey and grey lustrous phyllite	Lustrous grey phyllite with thin beds of limestone	Green phyllite, limy green phyllite, and greenstone Phyllitic and arenaceous limestone, minor grey phyllite Grey and light green phyllite, minor phyllitic limestone and quartz grit Quartz grit and minor gritty phyllite
	Lade Pk		Black and grey argillaceous limestone and phyllite	
	Badshot	Grey and white crystalline limestone, siliceous limestone and chert	Grey limestone and marble	Grey and white limestone
Hamill Group	Mohican	Siliceous green and greenish brown phyllite, calcareous brown and grey phyllite, brown and grey micaceous quartzite, thin grey and buff crystalline limestone	Grey and green quartz phyllite, grey phyllitic marble, marble	Grey and brown phyllite, micaceous quartzite, minor limestone Green phyllite and minor grey phyllitic limestone White to light grey limestone
	Marsh Adams	Grey and tan quartzite, grey and brown micaceous quartzite, subordinate grey and brown siliceous phyllite	Thinly interbedded white quartzite and brown mica schist, light green and light brown quartzite, grey micaceous quartzite, blocky white quartzite, minor marble, white and green quartzite	White, grey and brown quartzite, phyllitic quartzite, minor grey and black phyllite
	Mt Gainer	Flaggy white quartzite	White and green quartzite	White quartzite

Table 2-1 continued. Summary of stratigraphic correlations

Formation	Albert Peak area Sears (1979)	Downie Creek map area Brown (1991)	This study
Lardeau Group	Broadview	Grey and green phyllitic gnt and phyllite	Light green phyllite and minor gnt Light blue, tan, and grey quartzite and micaceous quartzite Quartz mica schist Light green, brown, silver, and dark grey semipelite and pelitic schist
	Jowett	Green phyllite, limy green phyllite, and greenstone White and grey limestone or dolomite	Well-laminated fine-grained siliceous metabuff Light green phyllite Grey and white marble Amphibole biotite calc-silicate schist Dark green amphibolite
	Index	Light green siliceous phyllite, minor phyllitic limestone, and quartz gnt Quartz gnt and minor gnt phyllite Green basic metavolcanic flows and minor green phyllite Grey limestone and phyllitic limestone Light grey siliceous phyllite Dark grey graphitic phyllite Quartz pebble conglomerate, quartzite, and limestone breccia	Green feldspathic gnt phyllite Amphibolite and amphibolitic schist Tan to light grey quartzite and micaceous quartzite Light grey and reddish brown pelitic and semipelite schist Grey pure and micaceous marble Light green siliceous phyllite Dark grey calcareous and noncalcareous graphitic phyllite
	Lade Pk	Fine-grained grey calcareous marble with fine-grained siliceous and pelitic interbeds	Thinly banded fine-grained light grey to white graphitic marble interbedded with fine- to medium-grained dark grey graphitic siliceous siltstone
Hamill Group	Badshot	Grey marble	Grey and white marble
	Mohican		Green and brown phyllite and micaceous quartzite, minor limestone
	Marsh Adams	Buff, light grey, and brown platy and massive micaceous quartzite; orthoquartzite; thin pelitic interbeds	White, grey, light blue, and tan quartzite interbedded with tan semipelite and pelitic schist
	Mt Gainer		

Plate 2-1. Thompson (1972) correlated the Hamill Group with this thick-bedded clean tan quartzite in the core of the Akolkolex anticline, cirque at head of southwestmost tributary of Standfast Creek.

Plate 2-2. Eastward view of stratigraphy on the Illecillewaet-Akolkolex divide, upper level of the Clachnacudainn terrane. Prominent peaks are Mount Llewelyn (ML), Albert Peak (AP), and North Albert Peak (NAP). Light grey marble layers are correlated with the Badshot Formation (Bd) by Read and Thompson (1980) and this study. Tan siliceous clastics layers are correlated with the Hamill Group (Ha) by Read and Thompson and this study. I correlate the dark grey graphitic clastics layer and the clastics below the lowest Badshot marble layer with the Index Formation (In). I interpret these layers as isoclinal fold limbs within the west-verging Ghost Peak anticline; the Hamill Group cores anticlines and the Index Formation cores a syncline. This interpretation is depicted in cross-section C-C', which is drawn through Mount Llewelyn's north-south ridge. The Albert stock (AS) intrudes metasedimentary rocks on the ridge between North Albert Peak (NAP) and Albert Peak (AP). Devonian orthogneiss (Dogn) is juxtaposed against metasedimentary rocks on the west ridge of North Albert Peak by an east-dipping high-angle brittle fault (see plates 5-7 and 5-8).

Plate 2-3. Southwestward view of south-dipping stratigraphy on Mount Llewelyn (ML) from a logging road on the north slope of the West Twin Creek valley. This is a view of the opposite side of the ridge shown in Plate 2-2. Light grey marble layers are correlated with the Badshot Formation (Bd), tan siliceous clastics layer are correlated with the Hamill Group (Ha), and the dark grey graphitic clastics layer is correlated with the Index Formation (In). My interpretation of these layers as isoclinal fold limbs in the west-verging Ghost Peak anticline is depicted in cross-section C-C'.

Plate 2-4. Sears (1979) and I correlate the Hamill Group with this white quartzite and grey micaceous quartzite in the footwall of the Standfast Creek fault (Clachnacudainn terrane) in the Albert Peak area, east slope of upper Siegfried Creek valley (station 203).

Plate 2-5. Northwestward view from a helicopter of tan siliceous clastics in the core of the west-verging Ghost Peak anticline (line GPA is its axial surface trace) that I correlate with the Hamill Group, east of Ghost Peak, west slope of upper Drimmie Creek valley. Hamill Group is sandwiched between limbs of Badshot marble (Bd) that close westward within Ghost Peak (GP). See text for discussion of Thompson's (1972) interpretation of this structure as the Drimmie Creek syncline. Cross-section A-A' is drawn through this part of the Ghost Peak anticline. Arrow points to position of geologist in Plate 2-6.

Plate 2-6. Northwestward view of tan quartzite and micaceous quartzite of the Hamill Group (Ha) in the core of the Ghost Peak anticline (GPA). Hamill Group is underlain by white Badshot marble (Bd, contact is beneath geologist), overlain by grey Badshot marble, and continues to the skyline (toward Ghost Peak) the Badshot marble closes around it in the hinge of the Ghost Peak anticline.

Plate 2-7. Tan, white, and grey quartzite in the core of the Ghost Peak anticline east of Ghost Peak (located near site of Plate 2-6) is lithologically similar to Hamill Group quartzite in the core of the Akolkolex anticline (Plate 2-1). I correlate the core of the Ghost Peak anticline with the Hamill Group based on this lithologic similarity. Thompson (1972) correlated these rocks with the Index Formation, yet note the absence of dark grey phyllite which typifies the basal Index Formation that lies adjacent to the Badshot marble (Plates 2-9 and 2-11).

Plate 2-8. The Standfast Creek fault (SCF) mapped by Thompson (1972) is well-displayed in this view of the south face of Ghost Peak (GP); Badshot marble (Bd) lies in a klippe in the hanging wall (Illecillewaet slice) and dark grey graphitic phyllite (Plate 2-11) and marble in the footwall (Clachnacudainn terrane) are correlated with the Index Formation (In) by Read and Thompson (1980) and this study. Since the Badshot and Index formations are in conformable contact in the hanging wall of the Standfast Creek fault (Plate 2-9; all previous workers), then I interpret the fault as an insignificant tectonic feature. This section of Index Formation is termed the Ghost Peak section in the text. Cross-section A-A' (drawn through Ghost Peak) depicts my interpretation of the Badshot marble as lying within the thickened hinge zone of the Ghost Peak anticline and the Index Formation as lying in its lower limb.

Plate 2-9. Light grey and white marble clasts in dark grey graphitic phyllite at the contact between the Badshot and Index formations are interpreted as evidence of a depositional rather than a faulted contact, Illecillewaet slice, ridge along east slope of upper Drimmie Creek valley, lower limb of Akolkolex anticline.

Plate 2-10. Alternating weathered grey graphitic marble and rusty coloured graphitic siliceous siltstone layers of the Lade Peak Formation (correlated by Sears, 1979), hanging wall of the Standfast Creek fault on the east slope of upper Siegfried Creek valley. The unique alternating grey and rusty weathering of the Lade Peak Formation was used to identify this formation in cliffs from considerable distances. Note the white dyke from the Albert stock that crosscuts the well-foliated metasedimentary rocks.

Plate 2-11. Dark grey graphitic phyllite in the footwall of the Standfast Creek fault beneath Ghost Peak (Plate 2-8) is lithologically similar to the Index Formation in the hanging wall. This similarity is the reason why I agreed with Read and Thompson's (1980) correlation of this footwall section (termed the Ghost Peak section in the text) with the Index Formation.

Plate 2-12. Thick-bedded dark green amphibolite (am) and white marble (mb) that I correlate with the Jowett Formation, Clachnacudainn terrane, north ridge of North Albert Peak, just above the contact between Devonian orthogneiss and metasedimentary rocks shown in Plate 3-3 (near station 211). These are the thickest beds that I observed in the Jowett Formation.

Plate 2-13. Thin- to medium-bedded amphibolite (am) and marble (mb) that I correlate with the Jowett Formation, ridge north of West Woolsey Creek (Plate 3-9). This scale of interbedding of marble and amphibolite is typical of the Jowett Formation.

Plate 2-14. Coarse-grained green and white amphibolite (am) overlies a very fine-grained, laminated light green, grey, and tan siliceous rock that I interpret as metatuff (mt), Columbia River valley, west of McCrae Peak, south of Akolkolex River (station 233). The lithologic similarity of these rocks with those north of the Akolkolex River that Read and Thompson (1980) correlated with the Jowett Formation is evidence that the Jowett Formation also crops out in the Clachnacudainn terrane south of the Akolkolex River.

CHAPTER 3. Clachnacudainn igneous complex

Introduction

The Clachnacudainn igneous complex (my terminology) is composed of the intrusives that dominate the Clachnacudainn Range, the Albert Peak area, and the base of the Clachnacudainn terrane south of Revelstoke. Most of the complex lies within the Clachnacudainn salient as mapped by Wheeler (1963, 1965). The numerous phases (with variable states of deformation) in the complex differentiate the Clachnacudainn terrane from the remainder of the Selkirk allochthon east of the Monashee complex; other portions of the eastern Selkirk allochthon have a greater abundance of intrusives (such as the Battle Range batholith to the east and Kuskanax batholith to the south), but few have as much variety as the Clachnacudainn igneous complex.

Although four intrusives have been investigated geochronologically, most were neither fully described nor accurately mapped by previous fieldworkers. Considerably more fieldwork and geochronological investigation are required in order to fully elucidate the magmatic history of the complex. I examined a few phases in some detail and dated zircon from the Albert stock because their ages and deformational states are important to understanding the tectonic evolution of the Clachnacudainn terrane. Each phase that was observed is described in this chapter from the oldest, strongly deformed phase to the youngest phases that crosscut all deformational features. The state of ductile deformation and inferred age of each phase are presented in Table 3-1 in a condensed form.

Geochronological results from the Albert stock are also reported in this chapter (Table 3-2 and Fig. 3-1). The locations of the intrusive bodies and geochronology samples discussed in this chapter are shown in Map 2 (in back pocket).

**Phase strongly deformed by ductile deformation - Devonian
hornblende biotite granodiorite gneiss**

Hornblende biotite granodiorite gneiss (Dogn in Map 2) in the Clachnacudainn terrane crops out along the Illecillewaet River west of Albert Canyon, south of LaForme Creek, and at the base of the terrane south of Revelstoke. The orthogneiss in Albert Canyon was dated at 356 ± 6 Ma by the U-Pb zircon method (Parrish, 1992). Although the terrane contains the only Devonian orthogneiss in the Selkirk allochthon east of the Monashee complex, there is Devonian orthogneiss west of the Monashee complex that has similar composition and fabric, such as the Mount Fowler orthogneiss near Shuswap Lake (Okulitch et al., 1975) and orthogneiss in the Seymour Range (Parrish, 1992). Figure 1-1 shows the approximate locations of these bodies.

Orthogneiss in the Clachnacudainn terrane varies in appearance and mineral composition. The gneissosity is defined by well foliated, thin (less than 1 cm), and discontinuous (over a few centimetres) light and dark compositional layers (Plate 3-1). The typically strong foliation is formed by the alignment of biotite and hornblende, which define the dark layers, and flattened quartz and plagioclase, which define the light layers. A lineation formed by stretched clots of hornblende and biotite is locally developed. The compositional layers are often so discontinuous, thin, and poorly developed that the orthogneiss could also be termed a strongly foliated granitoid. Zoisite and epidote locally compose up to 5%, with epidote commonly forming centimetre-wide layers. Lithologic variations

include a greyish white quartz-rich gneiss (possible granite protolith) that lacks hornblende and a dark grey hornblende- and biotite-rich gneiss (possible diorite protolith).

The nature of the contact between orthogneiss and overlying lower Paleozoic metasedimentary rocks was enigmatic to previous workers, as they interpreted this contact as unconformable, intrusive, and faulted. However, proper interpretation of this contact is crucial to understanding the tectonic history of the Clachnacudainn terrane. Mapping along the Canadian Pacific Railway in Albert Canyon, Daly (1915) suggested that the orthogneiss is highly metamorphosed basement of probable Archean age which unconformably underlies metasedimentary rocks. After more extensive mapping in Albert Canyon, Gunning (1929) disagreed with Daly's unconformity hypothesis by concluding that an intrusive contact is shown by the intimate manner in which orthogneiss is interbedded with metasedimentary rocks. Wheeler (1963, 1965) suggested that a fault lies immediately above orthogneiss in Albert Canyon, in the lower limb of the west-verging Albert Canyon anticline. Ross (1968) postulated that the orthogneiss is a slice of Precambrian basement that was wedged upward into cover metasedimentary rocks, forming the core of a large east-verging nappe along the northeast flank of the terrane. Ross interpreted the orthogneiss as Precambrian basement based on evidence which suggests it was metamorphosed and deformed prior to deposition of cover metasedimentary rocks; a disoriented fragment of orthogneiss occurs within an amphibolite dyke which Ross contended was intruded during deposition of the cover rocks. Gilman (1972) mapped orthogneiss along the Illecillewaet River as a concordant, folded sheet that was emplaced between two contrasting panels of metasedimentary rocks. He observed the same evidence, the interlayering of orthogneiss and metasedimentary rocks,

that allowed Gunning (1929) to interpret the contact as intrusive. However, having not observed conclusive evidence of an intrusive contact, such as metasedimentary screens or dykes of orthogneiss penetrating metasedimentary rocks, Gilman suggested that infolding and transposition of the two units could account for the interlayering relationships. Thompson (1972) also noted that metasedimentary rocks are tightly folded within orthogneiss, which he interpreted as a nearly conformable, subhorizontal lower unit; the contact is sharp with no evidence of mylonitization. Zwanzig (1973) observed i) interlayering of orthogneiss and metasedimentary rocks, ii) a sheared metasediment-orthogneiss contact that is partly altered to chlorite schist, and iii) "cataclastic zones" within the upper part of the orthogneiss along Woolsey and Albert creeks. The contact that is altered to chlorite schist may be a brittle fault. It is likely that the "cataclastic zones" in the orthogneiss are mylonitic shear zones because Zwanzig described the penetrative foliation within them as parallel with the gneissosity and formed by highly strained quartz and polygonized feldspar. The occurrence of orthogneiss on both sides of these shear zones constrains the amount of displacement. Zwanzig concluded that he could not determine whether orthogneiss was intrusive into or faulted against overlying metasedimentary rocks because its protolith and contacts were too severely altered during metamorphism.

Price (1986) inferred that a western branch from the Standfast Creek fault mapped by Wheeler (1963, 1965), Zwanzig (1973), and Hakkinen (unpublished data, 1976, 1977) lies at the fault mapped by Ross (1968) and Hakkinen between orthogneiss and overlying metasedimentary rocks. The evidence for this fault (here termed the lower Standfast Creek fault), although not discussed by Price (1986), is that footwall orthogneiss in Albert Canyon is highly fractured, cut by pseudotachylyte, and mylonitic according to Price et al. (1985). Parrish et al.

(1988) used these observations and personal communication with R.A. Price to infer that an extensional Eocene ductile to ductile-brittle branch of the Standfast Creek fault skirts the late-tectonic Albert stock (Sears, 1979) along the orthogneiss-metasediment contact. This inference requires that mylonitic footwall orthogneiss is not crosscut by the Albert stock. Parrish (1992) termed the orthogneiss in Albert Canyon that he investigated geochronologically an annealed mylonite.

I observed a sharp orthogneiss-metasediment contact on the ridge north of North Albert Peak and west of upper Siegfried Creek (Plate 3-2) and at the base of the terrane south of Revelstoke. The nature of this contact in the col south of Mount LaForme (at the lower Standfast Creek fault) is uncertain because it is not exposed. The only evidence of displacement near the contact is a 3 m wide brittle strike-slip fault within the uppermost exposure of the orthogneiss (gneiss crops out on both sides of the fault, which is about 5 m from the metasedimentary rocks). Brittle strike-slip faults also occur within the uppermost part of the orthogneiss on the ridge west of North Albert Peak (2 m wide fault) and in Albert Canyon (10-15 m wide fault). These faults are described in detail in Chapter 5.

In all localities in which I observed the orthogneiss-metasediment contact, the gneissic fabric, the orthogneiss-metasediment contact and the dominant schistosity in the adjacent metasedimentary rocks are parallel (Plate 3-3 is an example). Other than local low strain zones in which the gneiss fabric is less pervasive, there is no strain gradient in the orthogneiss from the contact with the metasedimentary rocks to the base of the terrane; gneissic fabric at the contact looks identical to that at the base. South of the lower Standfast Creek fault in the col south of Mount LaForme, orthogneiss is continuously exposed for 1.25 km, from the fault to an intrusive contact with late-tectonic (Cretaceous?) granite.

Over this distance, the orientation and appearance of the gneissic fabric do not vary.

Discussion

The question that has puzzled geologists for nearly 80 years remains: what is the nature of the contact between the orthogneiss and overlying metasedimentary rocks? As outlined above, previous workers put forward three hypotheses - unconformable, intrusive, and faulted. The Devonian U-Pb zircon age from the orthogneiss conflicts with hypotheses that require it to be Precambrian basement in unconformable (Daly, 1915) or fault (Ross, 1968) contact with metasedimentary rocks. The orthogneiss could be termed a mylonite, as was done by Price et al. (1985) and Parrish (1992), because the grain size was most likely reduced from its original size by dynamic recrystallization. However, the amount of grain size reduction cannot be estimated because its undeformed protolith is not exposed. Deformation of the orthogneiss after intrusion of the Albert stock is required in the Parrish et al. (1988) model, which infers that an Eocene ductile to ductile-brittle branch of the Standfast Creek fault skirts the Albert stock along the contact between the orthogneiss (footwall) and metasedimentary rocks (hanging wall). This model implies that the fault mapped by Sears (1979) on the west ridge of North Albert Peak as a brittle, steeply east-dipping normal fault at the contact between the orthogneiss and metasedimentary rocks overprints a ductile normal fault. I confirm that the contact is overprinted by a 10 m wide, steeply east-dipping brittle fault and suggest that east-directed dip-slip extensional displacement between the orthogneiss and metasedimentary rocks is unlikely because ductile fabrics on both sides of the contact dip moderately south-southeast outside of the brittle fault zone (Plates 5-7 and 5-8)

and there is no strain gradient away from the contact.

The parallelism of the gneissic fabric, main schistosity in the metasedimentary rocks, and orthogneiss-metasediment contact suggests that the orthogneiss was emplaced either i) as an igneous intrusive prior to or during the main phase of regional deformation and the contact was subsequently transposed into parallelism with the fabric; or ii) as the footwall of a thrust or normal fault and the gneissosity is a recrystallized mylonitic fabric that resulted from the shearing on the Standfast Creek fault. Four fabric and intrusive relationships constrain the age and origin of the gneissic fabric and therefore assist in revealing its tectonic significance. First, xenoliths of the orthogneiss within late-tectonic (shows no evidence of ductile deformation) granite on the west ridge of North Albert Peak require that the gneissic fabric was developed prior to intrusion of the granite (Plate 3-2). Second, the 15 m wide late-tectonic leucogranite dyke that crosscuts the contact between the orthogneiss and metasedimentary rocks on the north ridge of North Albert Peak requires that intrusion of the dyke post-dated any displacement along this segment of the contact. The granite and leucogranite are interpreted to be coeval with the main body of the Albert stock that is dated at 104 ± 1 Ma by the U-Pb zircon method (see section below on the Albert stock). Third, Devonian orthogneiss along Highway 23 north of Revelstoke (between Hathaway and Sale creeks) is intruded by mid-Cretaceous (R.L. Armstrong, unpublished data; R.R. Parrish, personal communication, 1990) biotite K-feldspar megacrystic granite that lacks any ductile deformation fabrics at the contact; this relationship requires that the gneissic fabric was developed prior to the mid-Cretaceous. Fourth and most important, the lack of a strain gradient in the orthogneiss from the contact with the metasedimentary rocks to the base of the Clachnacudainn terrane suggests that the contact is not a zone along which

higher ductile strain was accommodated. There are lower strain zones in the orthogneiss in which the fabric is less pervasive, yet these zones are underlain by highly strained orthogneiss that continues to the base of the terrane. The gneissic fabric is therefore interpreted to have developed during folding associated with regional deformation and is therefore not directly related to displacement on the Standfast Creek fault. Since identical-looking fabric occurs in the Devonian Mount Fowler orthogneiss on the west side of the Shuswap metamorphic complex (B. Johnson, personal communication, 1991), the deformation responsible for the gneissic fabric may have been penetrative at the scale of the southern Omineca Belt.

Among the three hypotheses, I consider an intrusive contact between orthogneiss and metasedimentary rocks as the most reasonable given the available data. Orthogneiss must have intruded into the then unmetamorphosed sediment as granodiorite (locally granite and diorite), and the intrusive contact was then transposed into subparallelism with compositional layering in the metasedimentary rocks by major isoclinal folds during regional deformation. All contact metamorphic effects must have been obliterated by recrystallization during regional metamorphism because there is no discernible contact aureole. This hypothesis should be tested by more fieldwork since conclusive evidence of an intrusive contact, such as metasedimentary screens or dykes of orthogneiss penetrating the surrounding metasedimentary rocks, has not been observed (Gilman, 1972). It is possible that screens and dykes could be recognized during further fieldwork around the "infolds" and "interlayers" of metasedimentary rocks within the orthogneiss that were observed by Gilman (1972) and Thompson (1972).

The timing (and therefore tectonic setting) of orthogneiss fabric

development could be constrained by geochronological investigation of the variably deformed leucogranite dykes that intrude orthogneiss at Albert Canyon, near the lower Standfast Creek fault (Plate 3-4). Pre- to syn-deformational dykes are folded (axial surfaces parallel the gneissic fabric) and boudinaged (stretched parallel to the gneissic fabric), whereas post-deformational dykes crosscut the gneissic fabric and deformed dykes. If the deformed dykes are Devonian (related to intrusion of the gneiss) and the undeformed dykes are Cretaceous or younger (related to intrusion of late-tectonic granitoids), then my hypothesis of fabric development during regional deformation would be supported. If the deformed dykes are instead younger than Cretaceous, then the gneissic fabric near the Standfast Creek fault at Albert Canyon would be younger than the gneissic fabric elsewhere in the terrane (which must be older than mid-Cretaceous for reasons given above). This age scenario would support the contention by Price (1986) that the orthogneiss in Albert Canyon is a mylonite that accommodated motion on the Standfast Creek fault.

Phases that are variably deformed by ductile deformation

Biotite K-feldspar megacrystic granite

Biotite K-feldspar megacrystic granite (Kmgr in Map 2) that crops out along Sale Creek and the Columbia River north of Revelstoke was dated as mid-Cretaceous by the U-Pb zircon method (R.L. Armstrong, unpublished data; R.R. Parrish, personal communication, 1990). It is generally only deformed by ductile deformation in an east-dipping, 600 m thick section of mylonitic granite with east directed kinematic indicators at the base of the Clachnaculainn terrane (Brown and Murphy, 1982). There is debate whether the mylonite is mostly related to

compression along the Monashee decollement (Lane et al., 1989) or to extension along the Columbia River fault (Parrish et al., 1988). The megacrystic granite is mildly deformed in only one locality (Mount Revelstoke road) above the mylonite zone at the base of the terrane. It has a completely igneous texture where it crosscuts Devonian orthogneiss and it is intruded by an equigranular two-mica granite which forms a large pluton to the east that was mapped by Ross (1968) and Hakkinen (unpublished data, 1976, 1977).

Biotite hornblende diorite

Nonfoliated to moderately foliated fine- to medium-grained bluish green biotite hornblende diorite (?d in Map 2) outcrops along the Trans-Canada Highway, Highway 23, and in the Clachnacudainn Range. The diorite is heavily intruded by foliated, transposed leucogranite dykes and undeformed leucogranite and pegmatite dykes (Plate 3-5). The diorite composes a substantial portion of the Clachnacudainn Range, where Hakkinen (unpublished data, 1976, 1977) mapped large bodies of diorite and hornblende gneiss.

It is possible that the diorite which I observed is similar to the hornblende biotite diorite that Lane (1984) mapped in the Columbia River valley north of Revelstoke as "essentially undeformed except at its margins" (p. 178). The 127 Ma K-Ar hornblende cooling date obtained by Lane from this phase suggests that it could be Middle Jurassic, which is the age of quartz monzonite in the Pass Creek and Fang plutons about 35 km to the north (Brown et al., 1991). However, Lane used incomplete U-Pb zircon data to postulate that the diorite is possibly mid-Cretaceous, attributing the Early Cretaceous cooling date to excess argon in the hornblende. Since the deformed margin of diorite that Lane mapped and the locally strong foliation in the diorite that I mapped indicate ductile strain occurred

after their emplacement, then I interpret the diorites to be older than mid-Cretaceous because the mid-Cretaceous K-feldspar megacrystic granite along Sale Creek (see above) is only locally mildly deformed by ductile deformation above the base of the Clachnacudainn terrane (this deformed base of the terrane could be the Monashee decollement or Columbia River fault).

Leucogranite and leucopegmatite

Syn- and late-tectonic leucogranite and leucopegmatite intrude the granitoids and surrounding metasedimentary rocks as dykes and form large bodies in the Clachnacudainn Range (Klgr in Map 2). Devonian orthogneiss and biotite hornblende diorite are locally heavily intruded (Plates 3-4 and 3-5). Syn-tectonic dykes are foliated, folded, boudinaged, and transposed whereas undeformed dykes crosscut all ductile fabrics. Geochronological investigation of these dykes would obviously constrain the timing of deformation. Some of the large bodies of undeformed leucogranite mapped by Hakkinen (unpublished data, 1976, 1977) in the Clachnacudainn Range may be Late Cretaceous because Parrish (1992) dated a leucogranite from Mount Revelstoke at 71 ± 1 Ma by the U-Pb zircon method. However, the sample dated by Parrish is distinct from most of the large bodies of leucogranite in the Clachnacudainn terrane because, according to Parrish, this sample "has a weak to moderate developed foliation", the origin of which "was not investigated in detail" (p. 147). My fieldwork suggests that the leucogranite has a well developed lineation and lacks a planar fabric (it is a L-tectonite). Other relatively "young" intrusives are deformed with the leucogranite along the Mount Revelstoke road, including pegmatite of unknown age and a phase that looks like sheared mid-Cretaceous megacrystic K-feldspar granite. I suggest that these "young" intrusives have undergone relatively minor ductile shearing (compared to

the surrounding hornblende diorite gneiss and metasedimentary rocks) because their intrusive contacts are at a high angle to their tectonic fabrics and those in the country rocks; presumably these contacts would have been transposed into parallelism with the structural fabrics upon only a moderate amount of shearing. Nevertheless, the tectonic fabrics in these "young" intrusives requires that some ductile deformation continued at this locality (Mount Revelstoke) later than the regional deformation in the upper structural level of the terrane (near the late-tectonic Albert stock) and in the low to middle structural levels of the terrane to the north (near the undeformed late-tectonic megacrystic K-feldspar granite along Highway 23 and Sale Creek). Further fieldwork and geochronological investigation are required in order to fully elucidate the nature of this young ductile deformation.

Scammell (1991) reported an age of 70-71 Ma for weakly foliated leucogranite in the northern Monashee Mountains, and leucogranite in the Seymour Range was dated at 63 Ma (R.R. Parrish, unpublished data). Parrish (1992) noted that these age similarities strengthen the correlation between rocks in the Selkirk allochthon north and west of the Monashee complex with those in the Clachnacudainn terrane.

Late-tectonic phases

Granitoids that show no evidence of ductile deformation compose a significant portion of the hanging wall and footwall of the Standfast Creek fault in the Albert Peak area and Clachnacudainn Range.

Albert stock

Sears (1979) mapped the boundaries of the approximately 35 km² Albert

stock north and east of Albert Peak. The eastern edge of the stock is separated from the western edge of the much larger Battle Range batholith by 2 km of metasedimentary rocks. I examined the stock in the well exposed cirques at the head of the Akolkolex River, Siegfried Creek, and East Twin Creek. I refer to the "Albert stock" as consisting of the main body, satellite plugs, and associated dykes because all were most likely part of the same intrusive event and therefore emplaced within a short time interval, perhaps over a few million years.

The dominant, coarse-grained biotite hornblende quartz monzonite phase of the Albert stock (Kqm in Map 2) was sampled for U-Pb zircon geochronology in order to confirm a mid-Cretaceous age that was suggested for this phase by K-Ar data reported in Price et al. (1985); these data indicate that it cooled through the hornblende closure ($530 \pm 40^\circ\text{C}$, reported in Ghent et al., 1988) at 100 Ma and through the biotite closure ($280 \pm 40^\circ\text{C}$, reported in Ghent et al., 1988) at 83 Ma. The age determination of this quartz monzonite is crucial to the tectonic interpretation of the map area because it forms the main body of the stock that Sears mapped as intruding the Standfast Creek fault. A phase that also forms much of the main body and appears to be coeval with the quartz monzonite is fine- to medium-grained green biotite hornblende diorite. These phases were intruded by many finer grained, lighter coloured felsic phases, such as biotite garnet leucogranite, medium-grained muscovite biotite granite, and fine-grained chlorite muscovite granite. A felsic dyke on the north side of the west ridge of North Albert Peak crosscuts the ductile normal fault branch from the Standfast Creek fault that was inferred by Parrish et al. (1988). Sears' mapping places the intrusive body at the head of Siegfried Creek (mostly medium-grained biotite hornblende quartz diorite, K?qd in Map 2) in the footwall of the Standfast Creek fault and separate from the main body of stock by a few hundred metres of

metasedimentary rocks. I suggest that the Siegfried Creek body is coeval with phases in the main body of the stock because the two are compositionally similar, in the same state of deformation, and continuous at the level of erosion around large screens of metasedimentary rocks (Plate 3-6). Other intrusives in the footwall of the Standfast Creek fault are fine- to medium-grained green biotite hornblende diorite and fine- to medium-grained muscovite biotite granite that intrude metasedimentary rocks and Devonian orthogneiss on the southwest flank of North Albert Peak. These phases are not shown in Map 2 because they form small bodies that are dispersed amongst the metasedimentary rocks.

My fieldwork confirms Sears' (1979) observation that the Albert stock post-dated the penetrative ductile deformation which is evident in the adjacent metasedimentary rocks. All phases in the Albert stock, including those that crosscut the Standfast Creek fault mapped by Sears and the fault inferred by Parrish et al. (1988), have completely igneous textures; the strong foliations in the surrounding metasedimentary rocks and metasedimentary xenoliths terminate abruptly against large granitoid bodies and narrow dykes (Plates 3-6, 3-7, and 3-8). Completely igneous textures and well defined, unfolded contact metamorphic isograds (see Chapter 4 on metamorphism) require that crystallization of the stock post-dated i) development of all penetrative ductile fabrics seen within the adjacent metasedimentary rocks, ii) displacement along the Standfast Creek faults that were mapped by Sears and inferred by Parrish et al., and iii) peak regional metamorphism. The only deformational features that post-dated the stock are decimetre- to metre-wide normal and strike-slip faults that are characterized by clay gouge and chlorite-filled slickenside surfaces. The amount of displacement on these brittle faults, which are continuous from the surrounding metasedimentary rocks into the intrusives, must be minimal since identical-looking

rocks occur on both sides.

U-Pb data

Sample AS-1 is a coarse-grained biotite hornblende quartz monzonite that is the dominant phase in the main body of the Albert stock. AS-1 was sampled in the cirque at the head of the Akolkolex River, about 100 m from the contact with foliated metasedimentary rocks mapped in the hanging wall of the Standfast Creek fault by Sears (see Table 3-2 for exact coordinates). Like all phases in the stock, AS-1 has a completely igneous texture that shows no evidence of ductile deformation. Standard techniques of crushing, grinding, separation on the Rogers GoldTM table, and heavy liquid separation were performed at Carleton University. FrantzTM magnetic separation, zircon selection, chemistry, and mass spectrometry, error propagation, and data reduction were completed at the geochronological laboratory at the Geological Survey of Canada, Ottawa. Three fractions of clear, colourless, sharply terminated, euhedral, crack-free, slightly to moderately elongate (1.5:1 to 4:1 length to width ratios), generally tablet-shaped zircons with numerous small inclusions were analyzed. These zircons are all greater than 149 microns in size, contain no visible cores, and except for the small inclusions, are considered high quality igneous grains. Zircons in fractions A (8 grains) and B (10 grains) were nonmagnetic at 1.8 amps and 5° side slope on the FrantzTM magnetic separator, whereas those in fraction C (8 grains) were magnetic. All analyzed zircons were air abraded with pyrite (Krough, 1982). U-Pb analytical methods utilized are those outlined by Parrish et al. (1987). Table 3-2 briefly outlines these analytical procedures and summarizes the data. The concordia diagram in Figure 3-1 displays the results.

Figure 3-1 shows that the 2 standard error ellipses from the three zircon

fractions are concordant (error ellipses intersect concordia) and overlap each other at about 104 Ma. The rather large errors result from high Pb blanks (42, 85, and 33 pg for fractions A, B, and C, respectively, Table 3-2), which are a result of accidental contamination introduced during chemistry or possibly by the pyrite that was residual from the abrasion process (R.R. Parrish, personal communication, 1992). An age of 104 ± 1 Ma is interpreted as the crystallization age of AS-1 because the three fractions closely agree with each other and are concordant at 104 Ma. The 104 ± 1 Ma crystallization age is consistent with the 100 Ma K-Ar hornblende and 83 Ma biotite cooling dates reported in Price et al. (1985) from a sample located about 6 km northwest of AS-1, on the east side of upper Siegfried Creek valley (M. Colpron, personal communication, 1992). The 104 ± 1 Ma age of AS-1 is interpreted as representing the age of all phases in the Albert stock within a few million years because there is no evidence which suggests that they were intruded during different magmatic episodes.

As outlined above, field relationships mapped by Sears (1979) and this study clearly require that the 104 ± 1 Ma crystallization age of the Albert stock is an upper age limit on the ductile deformation and regional metamorphism in the vicinity of the stock (upper structural level of the Clachnacudainn terrane and base of the overlying Illecillewaet slice). I interpret the Albert stock to also place an upper age limit on most of the regional ductile deformation and metamorphism in the middle to lower structural levels of the Clachnacudainn terrane based on evidence which suggests that the tectonothermal events are correlatable throughout the terrane (see Chapters 4 and 5). The amount of time that elapsed between the main phase of deformation (D2) and crystallization of the stock is presently unconstrained by geochronology. However, since D2 pre-dates or is synchronous with regional metamorphism (M2) and the stock has a

contact metamorphic aureole (M3), then the amount of time that elapsed between D2 and crystallization can be estimated by the amount of time necessary to exhume the country rocks to the depths of burial during M2 and M3. As Chapter 4 on metamorphism outlines, the minimum amount of exhumation between M2 and M3 is about 5.3 km and the maximum is about 16.8 km. Although an accurate timing estimate is dependent on exhumation rates, it can confidently be assumed that at least a few million years to tens of millions of years elapsed between D2 and crystallization of the 104 ± 1 Ma Albert stock. The ages of regional deformation and metamorphism are discussed further in the tectonic model that is proposed in the Chapter 7.

Cretaceous(?) granitoids in the Clachnacudainn Range

Medium-grained epidote biotite granodiorite (K?gd in Map 2) on the well exposed ridge north of West Woolsey Creek intrudes interbedded marble and amphibolite that I correlate with the Jowett Formation (Plates 3-9, 3-10, and 3-11). Price et al. (1985) placed the lower Standfast Creek fault along this intrusive contact, probably because Ross (1968) mapped the granodiorite as granite gneiss and the contact as a mylonitic, east-directed thrust fault. However, Ross' mapping is suspect because of this misidentification and the fact that he mapped granodiorite and diorite farther west on the ridge as metasedimentary rocks. I interpret this portion of the lower Standfast Creek fault as an intrusive contact because the granodiorite which intrudes the Jowett Formation has igneous textures only (no evidence for ductile deformation); the strong foliation in the Jowett Formation and uncorrelated metasedimentary screens is abruptly truncated at the main body of the granodiorite (Plate 3-10) and undeformed dykes crosscut the foliation in the Jowett Formation at a high angle (Plate 3-11). This intrusive

relationship requires that any motion along the lower Standfast Creek fault at this locality must have pre-dated intrusion of the granodiorite. Since there is no evidence suggesting that the granodiorite intrudes a fault, I consider any hypothesis involving major displacement on the lower Standfast Creek fault to be conjecture.

Other phases associated with the granodiorite are coarse-grained biotite granite (K?gr in Map 2), and farther west along the ridge, medium-grained green and white hornblende diorite (K?d) crops out. The extent of this granodiorite- and diorite-dominated phase beyond the ridge north of West Woolsey Creek is uncertain. Devonian granodiorite gneiss underlies metasedimentary rocks on the next ridge to the northwest (northwest of Maunder Creek and south of Mount LaForme) and Ross (1968) mapped late-tectonic granite farther west on the ridge north of West Woolsey Creek. However, since Ross' mapping on the ridge north of West Woolsey Creek is suspect, his map is probably not a useful guide.

Muscovite biotite granite (K?gr in Map 2) intrudes mid-Cretaceous biotite K-feldspar megacrystic granite along Sale Creek and Devonian orthogneiss on the ridge north of Mount St. Cyr. The granite at these localities has a completely igneous texture throughout. The granite is probably the same phase as the large body of granite mapped by Ross (1968) and Hakkinen (unpublished data, 1976, 1977) in the headwaters of West Woolsey Creek. It is locally associated with a very coarse-grained dark green biotite hornblende diorite (K?d).

I interpret these granitoids in the Clachnacudainn Range to be coeval with the 104 ± 1 Ma Albert stock because they have similar compositions and deformational states. If this inference is true, then all ductile deformation in the vicinity of these granitoids (including any motion on the lower Standfast Creek fault) must have ceased by mid-Cretaceous. This age relationship can obviously

only be proven through further geochronological investigation in the Clachnacudainn Range.

The late-tectonic granitoids of probable mid-Cretaceous age are most likely part of the same intrusive event as the Battle Range batholith, the western edge of which is separated by 2 km of metasedimentary rocks from the eastern edge of the Albert stock. According to Read and Wheeler (1976), the Battle Range batholith has the following similarities with the late-tectonic granitoids in the Clachnacudainn igneous complex: i) it is discordantly intrusive into the Hamill to Lardeau groups; ii) it is a composite batholith composed of biotite hornblende quartz monzonite, muscovite biotite granodiorite, and a late finer grained felsic phase; and iii) it is probably Cretaceous.

Eocene(?) lamprophyre and siliceous dykes

Lamprophyre dykes on the ridge north of West Woolsey Creek intrude metasedimentary rocks and crosscut late-tectonic leucogranite dykes. Lare (1984) obtained Early Eocene K-Ar whole rock cooling dates from lamprophyre dykes that post-date the Columbia River fault north of the Clachnacudainn terrane. A fine-grained garnet biotite siliceous dyke that intrudes Devonian orthogneiss and crosscuts a brittle fault in Albert Canyon is the only intrusive I observed that post-dates brittle deformation in the terrane. It is not known whether the brittle fault in Albert Canyon is coeval with the brittle deformation in the Columbia River fault.

Table 3'. Summary of intrusives in the Clachnacudainn igneous complex (CIC)

state of ductile deformation

	strongly deformed	variably deformed	undeformed
Devonian	hornblende biotite orthogneiss 356 ± 6 Ma (U-Pb zircon from Albert Canyon; Parrish, 1992) throughout CIC		
Jurassic		?bluish green biotite hornblende diorite Clachnacudainn Range ?leucogranite and pegmatite Clachnacudainn Range	
mid-Cretaceous		biotite K-feldspar megacrystic granite mid-Cretaceous (U-Pb zircon from along Sale Creek; R.L. Armstrong, unpublished data, R.L. Parrish, pers. comm., 1990) along Sale Creek and Columbia River north of Revelstoke	Albert stock, mostly quartz monzonite 104 ± 1 Ma (U-Pb zircon from biotite hornblende quartz monzonite in main body of stock) crosscuts Standfast Ck fault (SCF) in Albert Pk area ?biotite granodiorite and granite, hornblende diorite ridge N of W Woolsey Creek, crosscuts lower SCF ?muscovite biotite granite along Sale Creek, throughout Clachnacudainn Range
Late Cretaceous		leucogranite 71 ± 1 Ma (U-Pb zircon from Mt Revelstoke; Parrish, 1992) throughout CIC	?leucogranite and pegmatite throughout CIC, crosscuts SCF in Albert Pk area
Early Eocene			lamprophyre dykes 52, 55 Ma (K-Ar whole rock cooling dates from Columbia River valley N of terrane, Lane, 1994) ridge north of West Woolsey Creek

? = age is uncertain

age of intrusion

Table 3-2. U-Pb analytical data from the Altent stock, Selkirk Mountains, British Columbia¹

sample analysis	wt. [#] (mg)	U (ppm)	Pb. ⁺ (ppm)	$^{206}\text{Pb}/^{238}\text{U}$ [*]	Pb _c [#] (pg)	$^{206}\text{Pb}/^{208}\text{Pb}$ ⁺	$^{206}\text{Pb}/^{238}\text{U}$ ⁺⁺	$^{206}\text{Pb}/^{238}\text{U}$ ⁺⁺ (Ma)	$^{207}\text{Pb}/^{235}\text{U}$ ⁺⁺	$^{207}\text{Pb}/^{235}\text{U}$ ⁺⁺ (Ma)	$^{207}\text{Pb}/^{206}\text{Pb}$ ⁺⁺	$^{207}\text{Pb}/^{206}\text{Pb}$ ⁺⁺ (Ma)
Altent stock AS-1												
A+149,cl,nm	0.035	279.8	5.33	250.9	42.4	0.313	0.016252 ± 0.22	103.9 ± 0.5	0.10904 ± 0.96	105.1 ± 1.9	0.04866 ± 0.84	131.6 ± 41.4
B+149,cl,nm	0.040	305.8	6.06	163.7	84.6	0.370	0.016222 ± 0.30	103.9 ± 0.6	0.10778 ± 1.35	103.9 ± 2.7	0.04818 ± 1.18	108.3 ± 56.1
C+149,cl,nm	0.034	287.5	5.32	340.7	32.5	0.370	0.016275 ± 0.20	104.1 ± 0.4	0.10904 ± 0.98	105.1 ± 2.0	0.04859 ± 0.89	126.3 ± 41.4

⁺⁺ A=Zircon fraction; +149=age of zircon in zircons, all zircons were abraded, cl=clear, m=magnetic, nm=nonmagnetic
[#] Weighing error ± 0.001 mg.
⁺ Radiogenic Pb.
^{*} Measured ratio, corrected for spike, and Pb fractionation 0.08% ± 0.03%/AMU
[#] Total common Pb in analysis corrected for fractionation and spike
⁺⁺ Corrected for blank Pb and U, and common Pb (Stacey-Kramers model Pb composition equivalent to the interpreted age of the rock), errors are 1 standard error of the mean in percent for ratios and 2 standard errors of the mean when expressed in Ma.
⁺⁺⁺ Corrected for blank and common Pb, errors are 2 standard errors of the mean in Ma.

¹ U-Pb analytical methods are those outlined in Parrish et al. (1987). Techniques included air abrasion (Krogh, 1982), mineral dissolution in microcapsules (Parrish, 1987), a micro step-234U-235U isotopic tracer (Parrish and Krogh, 1987), multicollector mass spectrometry (Roddick et al., 1987), and estimation of errors by numerical error propagation (Roddick, 1987).

Sample locality: coarse-grained biotite hornblende quartz monzonite, collected from the southwest side of cirque at head of Aludior River at an elevation of about 5900 ft. (Geological map sheet 82N/4), southeastern British Columbia, 117°47'25" W - 51°01'26" N, UTM 11u, 444700E - 5652200N.

Figure 3-1. Concordia diagram of U-Pb zircon data from the Albert stock

Plate 3-1. Typical appearance of Devonian orthogneiss in the Clachnacudainn terrane, 1 km south of the lower Standfast Creek fault in the col south of Mount LaForme (near station 107).

Plate 3-2. Grey xenolith of Devonian orthogneiss (Dogn, beneath ice ax) lies within late-tectonic granite (gr), Clachnacudainn terrane, south side of the west ridge of North Albert Peak. This proves that the gneissic fabric developed prior to intrusion of the granite, which I interpret as coeval with the 104 ± 2.5 Ma Albert stock. This and other fabric relationships suggest that the gneissic fabric is not related to ductile extensional Eocene displacement on the Standfast Creek fault that was inferred by Parrish et al. (1988).

Plate 3-3. Sharp contact (at head of ice ax) between underlying Devonian orthogneiss (Dogn) and overlying metasedimentary rocks that I correlate with the Lardeau Group (LG), Clachnacudainn terrane, north side of the west ridge of North Albert Peak (near station 211). Note the parallelism of the contact, gneissic fabric, and bedding and schistosity in the metasediment. Parrish et al. (1988) inferred this contact to be a ductile extensional Eocene branch of the Standfast Creek fault. I interpret it as a transposed intrusive contact.

Plate 3-4. Z-folds (viewed to the southeast) of leucogranite dykes within Devonian orthogneiss, 2 km west of Albert Canyon station along the Canadian Pacific Railway. The gneissic fabric is axial planar to the Z-folds and is crosscut by a second generation of late-tectonic leucogranite dykes. Geochronologic investigation of these dykes could better constrain the age of regional deformation in the Clachnacuda terrane and the origin of the fabric in the Devonian orthogneiss.

Plate 3-5. Foliated biotite hornblende diorite in the Clachnacuda igneous complex is intruded by i) late-tectonic leucogranite dykes that were transposed into parallelism with the foliation in the diorite and ii) late-tectonic leucopegmatite, about 13.5 km east of Revelstoke on the Trans-Canada Highway. The ages of these intrusives are unknown.

Plate 3-6. A rusty coloured screen of metasedimentary rocks (ms; possibly Lade Peak Formation) lies within late-tectonic granitoids (gr) that occur between the main body of the Albert stock (104 ± 2.5 Ma) and the body at the head of Siegfried Creek, above icefield 1 km north of Albert Peak. Since these granitoids are in the footwall of the Standfast Creek fault mapped by Sears (1979), then I agree with Sears' interpretation that intrusion of the Albert stock post-dates motion on the fault.

Plate 3-7. A late-tectonic undeformed dyke from the Albert stock (104 ± 2.5 Ma) crosscuts all ductile structures within the adjacent metasedimentary rocks in the footwall of the Standfast Creek fault (Clachnacuda terrane), 3.5 km north of Albert Peak. Relationships such as this suggest that regional ductile deformation in the vicinity of the Albert stock ceased by mid-Cretaceous (also see Plate 3-8).

Plate 3-8. Late-tectonic granodiorite in the Albert stock (104 ± 1.0 Ma) crosscuts the foliation in the adjacent metasedimentary rocks, head of Siegfried Creek (near station 203). Relationships such as this suggest that regional ductile deformation in the vicinity of the Albert stock ceased by mid-Cretaceous (also see Plates 3-6 and 3-7).

Plate 3-9. The south face of the ridge north of West Woolsey Creek that contains two branches of the Standfast Creek fault inferred by Price (1986) and an inverted stratigraphic section (termed West Woolsey Creek section in text) of Lardeau Group (according to my correlations) in the lower limb of the west-verging Albert Canyon anticline. The lower Standfast Creek fault (LSCF) is an intrusive contact between granodiorite (gd) to the west and the Jowett Formation (Jt) to the east. The Index Formation (In) overlies the Jowett Formation to the east. The upper Standfast Creek fault (USCF) lies at a brittle strike-slip fault within marble that has the Index Formation on both sides. The Badshot marble (Bd) in the core of the Albert Canyon anticline (ACA) overlies the Index Formation farther to the east. Mount LaForme (ML) is peak to the north. Cross-section D-D' is drawn through this ridge.

Plate 3-10. Late-tectonic (Cretaceous?) granodiorite crosscuts the foliation in the adjacent metasedimentary rocks on the ridge north of West Woolsey Creek. Relationships such as this suggest that the lower Standfast Creek fault is an intrusive contact at this locality and that regional ductile deformation ceased prior to intrusion of the (Cretaceous?) granodiorite (also see Plate 3-11).

Plate 3-11. Late-tectonic dyke (d) from the (Cretaceous?) granodiorite on the ridge north of West Woolsey Creek crosscuts well-foliated marble and amphibolite that I correlate with the Jowett Formation (Jt), about 150 m east of the lower Standfast Creek fault (near station 219). Relationships such as these require that any motion on the lower Standfast Creek fault must have pre-dated intrusion of the (Cretaceous?) granodiorite.

CHAPTER 4. Metamorphism

Introduction

The contrast between the sillimanite grade rocks in the Clachnacudainn terrane and the significantly lower grade rocks in the overlying Illecillewaet slice suggested to Sears (1979) and Price et al. (1985) that a steep metamorphic gradient occurs in the vicinity of the Standfast Creek fault. Parrish et al. (1988) interpreted this suggestion to mean that the Standfast Creek fault is associated with an appressed or omitted series of isograds, which they attributed to Eocene ductile extension. Evidence for a steep metamorphic gradient in the vicinity of the Standfast Creek fault is examined in the first section of this chapter by presenting the metamorphic mineral assemblages documented by previous fieldworkers and by this study. These assemblages are compiled in Map 2 (in back pocket). The second section of this chapter uses textural data to place constraints on the timing of metamorphism relative to deformation. The peak pressure and temperature conditions of metamorphism are estimated in the third section through comparison of mineral assemblages with a petrogenetic grid and through thermobarometry.

Metamorphic mineral assemblages

Metamorphic assemblages documented by previous fieldworkers

The high metamorphic grade of migmatite- and sillimanite-bearing schists in the Clachnacudainn Range suggested to Wheeler (1965) that they lie within a salient of the sillimanite grade Shuswap metamorphic complex to the west.

Wheeler gave no indication whether sillimanite grew during regional metamorphism or during contact metamorphism associated with the intrusion of the abundant late-tectonic granitoids (which I include within the Clachnacudainn igneous complex, Chapter 3). In the eastern Clachnacudainn Range, Gilman (1972) documented lower to middle amphibolite facies assemblages: plagioclase-biotite-garnet $\pm$ muscovite $\pm$ K-feldspar in pelitic schist and plagioclase-hornblende $\pm$ biotite $\pm$ epidote in amphibolite.

Thompson (1972) mapped the progressive downward appearance of syn-tectonic biotite, garnet, staurolite, kyanite, and sillimanite-muscovite from the highest to lowest structural level in the Clachnacudainn terrane in the Akolkolex River area. The garnet-in isograd closely follows the Standfast Creek fault, the kyanite-staurolite assemblage is confined to a narrow zone along the contact with Devonian orthogneiss (which underlies the metasedimentary rocks, see Chapter 3), and the sillimanite-muscovite assemblage is restricted to metasedimentary infolds within the underlying Devonian orthogneiss. Therefore, Thompson documented that the upper structural level of the terrane attained a maximum of garnet grade during regional metamorphism and the overlying Illecillewaet slice attained a maximum of biotite grade. The simple, downward, prograde variation of mineral assemblages toward the underlying orthogneiss suggested to Thompson that it was an important local heat source.

Zwanzig (1973) mapped three metamorphic zones in Albert Canyon that record a decrease in grade upward from the orthogneiss, from amphibolite to micaschist facies over a distance of 3 km. The staurolite zone (assemblage

staurolite-biotite-garnet-muscovite), which occurs in pelitic metasedimentary infolds within the orthogneiss and in overlying pelitic rocks, is overlain by the biotite-epidote zone, which is overlain by the muscovite-chlorite-calcite zone. The latter two zones occur mostly in calcareous graphitic phyllite. The isograd between the staurolite and biotite zones lies entirely within the hanging wall of the Standfast Creek fault and cuts upward into the hanging wall from north to south; metasedimentary rocks on both sides of the Standfast Creek fault in Albert Canyon attained a maximum of staurolite grade. Although isograds overprinted the major structures, Zwanzig suggested that a final phase of deformation must have outlasted peak metamorphism because porphyroblasts are deformed and the main foliation is flattened around them. The rapid decrease in grade upward from the orthogneiss suggested to Zwanzig that it may have been emplaced as a heat source during metamorphism. He suggested retrogression to explain why the upper part of the orthogneiss contains amphibolite facies assemblages and not higher grade assemblages. The reason why the orthogneiss would have retrogressed to amphibolite facies is not clear because Zwanzig presented no evidence which suggests it was at a higher grade anywhere in the terrane at any time.

Sears (1979) documented two distinct episodes of metamorphism in the Albert Peak area. The older episode is a syn-tectonic, intermediate pressure, Barrovian-style regional metamorphism that is characterized by a simple downward increase in grade from lower greenschist to amphibolite facies. The highest grade regional assemblage in the Clachnacudainn terrane documented by Sears is staurolite-garnet-biotite, and the garnet zone of regional metamorphism extends into the hanging wall of the Standfast Creek fault south of Albert Peak. The younger episode of metamorphism is a post-tectonic, lower pressure, Buchan-

style contact metamorphism that resulted from intrusion of the Albert stock. Sears documented distinct biotite, andalusite, sillimanite, and sillimanite - K-feldspar zones in the Albert stock contact aureole. A chlorite zone may be represented by the retrogradation of garnet and biotite that grew during regional metamorphism. Contact metamorphic isograds crosscut major structural features (including the Standfast Creek fault) and regional metamorphic isograds. Porphyroblasts which grew during contact metamorphism post-date the youngest deformational feature, crenulation of the main fabric, except locally (Sears, 1979).

The difference between the conditions of equilibration of the highest grade regional assemblage at the base of the terrane (sillimanite-kyanite-muscovite) and contact metamorphic assemblages indicated to Sears (1979) that pressure was reduced by more than 1 kbar between the two episodes of metamorphism. Sears ascribed the pressure decrease to attenuation of the overlying pile by "tectonic sliding" and to erosional denudation which resulted from tectonic uplift. Price et al. (1985) ascribed the pressure reduction between the two episodes of metamorphism to uplift and erosional unroofing that occurred as the terrane was driven as a wedge beneath the overlying Illecillewaet slice.

Read and Brown (1981), Lane (1984), and Parrish et al. (1988) reported Eocene K-Ar biotite cooling dates from granitoids in the Clachnacudainn Range. Parrish et al. (1988) interpreted the suggestion by Price et al. (1985) that the Standfast Creek fault is associated with a steep gradient in regional metamorphism to mean that the fault is associated with an appressed or omitted series of isograds. Parrish et al. (1988) combined this interpretation with the mineral cooling data (which implied there was a temperature mismatch between the Clachnacudainn terrane and overlying Illecillewaet slice in the Eocene) to suggest that the terrane is part of the Shuswap metamorphic complex which was

denuded by the Eocene ductile Standfast Creek fault.

Documentation of metamorphic assemblages by this study

My mapping of the following three regional metamorphic zones generally agrees with mapping by Thompson (1972), Zwanzig (1973), and Sears (1979).

Map 2 shows the AFM (Thompson, 1957) assemblages mapped by previous fieldworkers and myself that define the zones. Map 2 also shows the location of thin sections with complete mineral assemblage documentation in Appendix 1.

The kyanite zone is defined by the assemblage kyanite-staurolite-muscovite-biotite $\pm$ fibrolitic sillimanite that occurs in the base of the southern Clachnacudainn terrane, in metasedimentary rocks less than 100 m above various intrusives (mostly pegmatite and leucogranite with minor Devonian orthogneiss). The kyanite zone is overlain by the staurolite zone assemblage staurolite-garnet-muscovite-biotite. The staurolite zone in the southern part of the terrane may be continuous with the staurolite zone mapped by Zwanzig (1973) in Albert Canyon. This zone is less than a few hundred metres thick, and it is overlain by the garnet zone assemblage garnet-muscovite $\pm$ biotite $\pm$ chlorite. Staurolite-bearing pelitic schist has also been documented at the head of the uppermost west branch of the Akolkolex River (Sears, 1979) and staurolite inclusions within andalusite occur in the Albert Peak area. The garnet zone occurs in the hanging wall of the Standfast Creek fault in Albert Canyon (Zwanzig, 1973; this study, Sample 21 in Map 2), south of Albert Peak (Sears, 1979), and in at least two localities in the Ghost Peak area (this study). The two localities in the Ghost Peak area that I mapped occur less than 200 m above the Standfast Creek fault; one locality is in the Index Formation that crops out low on the west slope of upper Greeley Creek valley (Samples 60 and 65) and the other locality is in the thin band of Hamill

Group that crops out southeast of Ghost Peak (Sample 182) in the core of the Ghost Peak anticline. Since garnets in the hanging wall have the same relationship to deformation as those in the footwall (post-D2/pre-D3, see below), I interpret them to be coeval. The extent of garnet grade metamorphism in the hanging wall of the Standfast Creek fault may be larger than the area containing garnet because the hanging wall is dominated by non-pelitic lithologies, such as marble, calcareous graphitic phyllite, and clean quartzite.

Other notable, previously undocumented pelitic schist assemblages are i) garnet-chlorite-biotite-muscovite $\pm$ andalusite in the hanging wall of the lower Standfast Creek fault on the ridge north of West Woolsey Creek, ii) garnet-chlorite-biotite-muscovite in the hanging wall of the lower Standfast Creek fault south of Mount LaForme, iii) sillimanite (prismatic and fibrolitic)-biotite-muscovite in metasedimentary screens within late-tectonic granite along Sale Creek, and iv) biotite-chlorite in the hanging wall of the fault along Standfast Creek.

Mineral assemblages in the non-pelitic lithologies in the terrane also suggest that the maximum regional metamorphic grade is amphibolite facies. Devonian orthogneiss contains the assemblage hornblende/actinolite-biotite-plagioclase-quartz $\pm$ muscovite $\pm$ tremolite $\pm$ epidote. Calc-silicate schist contains the assemblage hornblende/actinolite-plagioclase-chlorite-biotite-calcite-quartz. Amphibolite contains the assemblage hornblende/actinolite-plagioclase-quartz $\pm$ biotite $\pm$ garnet $\pm$ calcite.

I term this lower to middle amphibolite facies regional metamorphism (the oldest episode of Sears, 1979) M2 because it overprints an older greenschist grade metamorphism (M1). Evidence for M1 is the assemblage muscovite-chlorite-quartz that forms the locally preserved, oldest fabric (S1, see below for details).

2

PM-1 3½"x4" PHOTOGRAPHIC MICROCOPY TARGET
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Biotite and staurolite grown during M2 were preserved by andalusite porphyroblasts that overgrew them during contact metamorphism associated with intrusion of the Albert stock (the younger episode of Sears and termed M3 by this study). This relationship, along with the data given by Sears, is evidence that M3 is younger than and distinct from M2. The sillimanite- and andalusite-bearing pelite in the Clachnacudainn Range is invariably engulfed by or proximal to late-tectonic granitoids in the Clachnacudainn igneous complex. Mineralogical evidence for these aureoles includes post-tectonic andalusite in pelite that is intruded by late-tectonic granodiorite on the ridge north of West Woolsey Creek and randomly oriented prismatic sillimanite in metasedimentary screens within undeformed granite along Sale Creek. In contrast to these M3 minerals, kyanite with a strongly preferred orientation that is interpreted to have grown during M2 in the southern part of the terrane has S2 perturbed around it.

Discussion

This review of the metamorphic assemblages shows that amphibolite facies regional metamorphic (M2) assemblages (kyanite-staurolite-muscovite-biotite $\pm$ fibrolitic sillimanite) in the base of the Clachnacudainn terrane progressively decrease upward to lower grade M2 assemblages (garnet-chlorite $\pm$ biotite) that continue across the Standfast Creek fault into the base of the Illecillewaet slice. I interpret the prismatic sillimanite-andalusite $\pm$ K-feldspar assemblages in the Clachnacudainn Range to have grown during contact metamorphism (M3) associated with intrusion of late-tectonic granitoids in the Clachnacudainn igneous complex. Chapter 3 presents the textural and compositional evidence which suggests that the late-tectonic granitoids in the Clachnacudainn Range are coeval with the 104 ± 1 Ma Albert stock. For this reason, M3 in the Clachnacudainn

Range is interpreted to be coeval with M3 in the Albert stock contact aureole.

This review establishes that once the distinction between regional and contact metamorphism is made, then it is evident that the rocks on either side of the Standfast Creek fault were at garnet grade during regional metamorphism. Hence, there is no evidence suggesting that a steep regional metamorphic gradient occurs across the Standfast Creek fault; the downward increase in regional metamorphic grade from garnet grade in the hanging wall to kyanite-staurolite grade at the base of the terrane over a vertical distance of over 1 km (isograds appear to be horizontal) suggests a normal regional metamorphic gradient (see below for estimates of the P-T differences between the hanging wall and base of the footwall. The apparently steep metamorphic gradient across the northernmost portion of the Standfast Creek fault can be explained by closely spaced M3 isograds in the Clachnacudainn igneous complex. In this locality, biotite grade rocks in the hanging wall (R.L. Brown, personal communication, 1991) occur about 5 km from sillimanite-bearing pelite in the footwall. This close spacing of M3 isograds is not unusual since the M3 biotite zone in Albert stock contact aureole occurs as close as 1.75 km to the M3 sillimanite - K-feldspar zone (Sears, 1979).

Parrish et al. (1988) used K-Ar mica cooling dates (reported in Read and Brown, 1981, Lane, 1984, Parrish et al., 1988) to infer a large temperature difference between the Clachnacudainn terrane and overlying Illecillewaet slice in the Eocene. I assert that since all of the mica cooling dates in the terrane were obtained from or near granitoids in the Clachnacudainn igneous complex, then the cooling dates signify that Late Cretaceous and younger intrusions heated the country rock above the mica closure temperature until the Eocene. A Late Cretaceous leucogranite from Mount Revelstoke (dated by the U-Pb zircon

method by Parrish, 1992) and crosscutting leucopegmatite are examples of such intrusives. The mica cooling data therefore cannot be used to determine whether or not the entire terrane was hot in the Eocene and should not be used as evidence that the entire terrane was rapidly tectonically denuded in the Eocene. An Ar-Ar mica cooling date traverse across the Standfast Creek fault beginning in the intrusive-free southern portion of the terrane would reveal whether the entire terrane was at high temperatures in the Eocene. I suggest that the southern, metasedimentary dominated part of the terrane was cool by the Late Cretaceous because the 104 ± 1 Ma Albert stock i) crosscuts the Standfast Creek faults mapped by Sears (1979) and inferred by Parrish et al., ii) has a contact aureole that crosscuts regional isograds, and iii) has 100 Ma hornblende and 83 Ma biotite K-Ar cooling dates (reported in Price et al., 1985).

The apparent mismatch in metamorphic grade and K-Ar mica cooling dates across the Standfast Creek fault therefore cannot be used to establish its tectonic significance. The fact that regional metamorphic isograds crosscut the fault does not disprove its existence because of the possibility of pre-metamorphic displacement or post-fault reequilibration; however, it is consistent with other data which suggest that the Standfast Creek is a minor tectonic feature.

Textural constraints on the timing of metamorphism versus deformation

Introduction

Aligned inclusion fabrics (Si) in porphyroblasts, chiefly garnet, are compared with matrix fabrics (Sm) in this section to aid in deciphering the tectonic history of the Clachnacudainn terrane. These porphyroblast-inclusion-

matrix textural relationships record the superposition of three phases of deformation and metamorphism and thereby allow for a relative assessment of timing between them. Figure 4-1 shows this relationship between deformation and metamorphism in graphical form.

Thin sections were cut from 107 samples in order to fully view the various deformation fabrics. Five "Areas" in the terrane are discussed separately because the timing of deformation relative to metamorphism varies according to structural level. The map in Figure 4-2 shows the location of the areas. Each Area is assigned a low, medium, or high structural level based on its position relative to the underlying Devonian orthogneiss. It must be emphasized that structural levels are not assigned based on the style or intensity of deformation in each level, as there is minor variation in these aspects across the terrane (see Chapter 5 on deformation). The textural descriptions and interpretations in the beginning of this section culminate in a discussion at the end which relates the diachroneity of the tectonothermal events to structural level.

Area 1 (kyanite, staurolite, garnet zones) - low to middle structural level, east slope of Columbia River valley, south of Illecillewaet River

Area 1 is assigned a low to middle structural level because it directly overlies the Devonian orthogneiss and continues 600-1000 m upward. I interpret the well developed inclusion fabrics (Si) in garnet from Area 1 to be coeval with the oldest recognizable schistosity (S1) in the matrix (S1m) because S1 and S1m have similar relationships to deformation and are defined by the same metamorphic minerals. Since S1 is interpreted to have formed during the oldest recognizable phase of deformation (D1), then garnet growth in Area 1 must have occurred after D1. Garnet growth occurred during the second, peak metamorphic

event (M2). The quartz-muscovite-chlorite assemblage that defines S1 suggests that D1 was accompanied by a low grade metamorphic event (M1). Since compositional layering (S0) is parallel with S1 throughout the study area, phase one folds (F1) may be large-scale isoclines responsible for the transposition.

The relationship between the second phase of deformation (D2) and garnet growth is more complex because inclusion fabrics are variably deformed. S1 was rarely preserved in the matrix because it was strongly transposed by second phase folds (F2) into the dominant, axial planar schistosity (S2). F2 folds are tight to isoclinal with gently inclined to recumbent, generally northeast to south-dipping axial surfaces and shallowly southeast-plunging hinge lines. One interpretation of the diverse S1i geometries is that garnet growth in Area 1 occurred at different times relative to D2; the degree of curvature of S1i is related to the amount of F2 folding that occurred prior to garnet growth. Straight S1i that are highly discordant to S2m are interpreted as a result of pre-D2 garnet growth, thereby preserving an unfolded S1 (Fig. 4-3). Variably curved S1i that are highly discordant to S2m are interpreted as a result of syn-D2 garnet growth; gently curved S1i resulted from garnet growth after slight F2 crenulation folding (Fig. 4-4a,b), whereas tightly folded S1i and the related incipient crenulation cleavage (S2i) resulted from garnet growth after moderate deformation of S1, during early stages in the transposition of S1 into the dominant S2 crenulation cleavage. The evidence which establishes that sigmoidal Si resulted from garnet overgrowth of a crenulated S1 rather than from syn-kinematic garnet rotation is relict F2 fold hinge zones in the matrix with similar geometries and orientations as Si (Fig. 4-4b). These relict matrix F2 fold hinge zones were preserved from further deformation because they occur in the low stress region of garnet pressure shadows. The straight and sigmoidal Si in Area 1 garnets record either garnet

growth during a hiatus in D2 or very rapid growth relative to D2 because Si geometries do not change progressively from core to rim.

The Area 1 garnets with core-to-rim variations in Si geometries are evidence for more than one period of growth. The typically crenulated Si in these garnet cores are discordant to rim Si and S2m, which are subparallel. Since the rim Si are slightly warped around the core and subparallel to S2m, they are interpreted as S2 that initially enveloped the garnet and were subsequently overgrown. These garnet with two distinct, discordant Si must have grown in two static stages; the curved core Si grew after an early stage of F2 crenulation folding whereas the rim Si that are warped around the core grew after formation of the penetrative matrix schistosity (S2m) formed. The smaller sized garnet with core Si which are subparallel to S2m probably represent slices through the porphyroblast rim. These inclusion fabrics prove that the rims of some Area 1 garnets grew after a more advanced stage of S2m development (destruction of crenulation folds and total transposition of S1 into S2) than previously described garnets with straight or weakly crenulated S1i throughout.

Other garnet Si in Area 1 unambiguously record syn-shearing growth. In these porphyroblasts, the core S1i are at moderate to high angles to S2m and gradually curve into parallelism with S2m within the rim (Fig. 4-4c). S1 is therefore interpreted to have been sheared and rotated around the porphyroblast by F2 folds into the S2 orientation during incremental garnet growth.

The preferred interpretation of the variation in Area 1 garnet Si is that garnet grew at different stages in D2; whereas some garnets ceased growing prior to crenulation folding, others ceased after formation of an incipient S2 crenulation cleavage, and yet others completed growth after total transposition of S1 into the S2 crenulation cleavage. Another viable interpretation is that the diverse Si

geometries resulted from garnet growth that occurred at the same time relative to deformation but in different strain regimes; whereas garnet with straight or slightly folded S1i that are highly discordant to S2m grew in the low stress region of F2 fold hinge zones, those with incipient S2i and tightly folded S1i grew along F2 fold limbs. Whatever the scenario, substantial D2 deformation must have occurred after garnet growth in Area 1 because all porphyroblasts have well developed pressure shadows which formed in response to F2 folding and they are enveloped by S2m (Figs. 4-3 and 4-4).

Two stages of biotite growth are observed in Area 1. M2 biotite is interpreted to have grown syn-D2 because it lies within the S2 that envelopes garnet (Figs. 4-3 and 4-4). M2 biotite may have grown after garnet because it is never included within the porphyroblast, but this relationship may be controlled by mineral kinetics rather than the timing of mineral growth. M3 biotite overprints S2. Most chlorite in Area 1 is the post-D2 retrograde (M3) product of garnet, biotite, and staurolite. However, rare syn-D2 chlorite is observed. Some Area 1 staurolite is post-D2 because it overprints S2m. Other staurolite and coarse kyanite are syn-D2 because they are aligned within and enveloped by S2 and contain sigmoidal S1i. Staurolite and kyanite are therefore interpreted to have grown during and after garnet.

Area 2 (garnet zone) - high structural level, west of Ghost Peak

Area 2 is assigned a high structural level because it directly underlies the Standfast Creek fault and is the part of the terrane farthest from the orthogneiss. Garnet growth in Area 2 is interpreted as late- to post-D2 because Si are parallel to and continuous with S2m (Fig. 4-5). Also, idioblastic garnet lacks pressure shadows and S2m is not perturbed at its margin. These relationships indicate that

garnet in Area 2 overgrew a completely developed S2.

Garnet growth in Area 2 is interpreted as pre-D3 because garnet Si are unfolded and highly discordant to the S3 crenulation cleavage that envelopes the porphyroblast (Fig. 4-5). Also, pressure shadows formed around garnet in response to F3 folding. F3 folds have steeply dipping axial planes and shallowly east-southeast-plunging hinge lines. This hinge line orientation is slightly more southerly than F2 fold hinge lines in most localities, but the two were also locally observed to be parallel.

Syn-D2 biotite in Area 2 is aligned within S2. Post-D2/pre-D3 biotite is evidenced by crystals that i) overgrew S2m and are enveloped by S3, ii) are kinked and folded by F3 folds, and iii) contain straight inclusion trails whereas S2m is crenulated by F3 folds. Syn-D2 chlorite is aligned within S2 and post-D3 chlorite is a retrograde product of biotite and garnet.

Area 3 (garnet and staurolite zones) - middle to high structural level, on ridge between Ghost and Albert peaks

Area 3 is assigned a middle to high structural level because it directly underlies the Standfast Creek fault and continues downward to within about 500 m of the Devonian orthogneiss. Garnet growth in Area 3 is interpreted to have begun late in D2 and continued after D2. Late-D2 garnet growth is shown by porphyroblasts that overgrew a moderately crenulated S1 and an associated S2 crenulation cleavage. Plagioclase exhibits this same relationship. These inclusion fabrics require that garnet grew after development of an incipient S2 but ceased prior to extreme tightening of F2 folds and total transposition of S1 into S2 that occurred in the matrix. Late- to post-D2 garnet growth is shown in other Area 3 samples by a lack of pressure shadows and by fabrics that pass undisturbed from

the matrix through entire idioblastic garnets (Fig. 4-5b).

The following relationships were also seen in other areas: i) garnet is pre-D3 because straight Si are at high to moderate angles to S3, which envelopes the porphyroblast (Fig. 4-5a); ii) chlorite, muscovite, and quartz define S1m in relict F2 fold hinge zones; and iii) post-D3 retrograde chlorite is ubiquitous. Unlike other areas, three stages of biotite growth are observed in Area 3: i) syn-D2 biotite is aligned within S2, ii) post-D2/pre-D3 biotite overgrows S2m and has straight inclusion fabrics whereas the matrix is crenulated, iii) post-D3 biotite is represented by large and undeformed crystals that overgrew F3 crenulation folds. Replacement of garnet by chlorite is interpreted to have occurred post-D3 because the opaque inclusion trails in the chlorite mats are unfolded whereas the matrix is pervasively crenulated by F3 folds.

Area 4 (garnet zone) - low structural level, Albert Canyon area

Area 4 is assigned a low structural level because it directly overlies the Devonian orthogneiss. Plagioclase and garnet in one sample from Area 4 have straight S1i (composed of muscovite and quartz) that are at about 55° to S2m, suggesting that M2 occurred between D1 and D2 (Fig. 4-3c). Garnet growth in Area 4 therefore occurred at a similar stage in the deformational history as some samples in Area 1 and at an earlier stage than all samples from Areas 2 and 3. Area 4 biotite grew syn- to post-D2. Chlorite grew syn-D1 (aligned within S1) through D3 (retrograde product of garnet).

Area 5 (garnet zone) - low to middle structural level, region south of Mount LaForme and north of West Woolsey Creek

Area 5 is assigned a low to middle structural level because it directly

overlies the Devonian orthogneiss and continues upward for about 500 m. Garnet in Area 5 is interpreted to have started growing prior to D2 and continued late in D2. The change in garnet inclusion trail geometry from core to rim (straight S_1 in the core are at 20-30° to S_{2m} and curve into parallelism with S_{2m} in the rim) signifies that the core grew pre-D2 and the rim grew during shearing and realignment of the foliation by F2 folds. The observation that S_{2m} envelope garnet indicates that D2 continued after garnet growth. Syn- to late-D2 growth is shown in other samples by garnets that overgrew a portion of their pressure shadows. All samples contain abundant syn-D2 biotite and rare biotite that is a retrograde product of garnet. Chlorite grew pre-D2 to post-D3.

Discussion

Gillett (1990) interpreted sigmoidal garnet inclusion fabrics in pelitic schist from the upper level of the Clachnacudainn terrane as a result of porphyroblast rotation due to noncoaxial shear associated with the Standfast Creek fault. On the contrary, I interpret the variety of inclusion fabrics as a result of garnet growth during pre-D2 through post-D2 conditions. Since fabrics associated with three phases of ductile deformation and metamorphic porphyroblasts are preserved throughout the Clachnacudainn terrane, then timing between deformation and metamorphism is compared between areas (Fig. 4-1).

M1 is a greenschist grade event that occurred pre- to syn-D1. This interpretation is made because S_1 , which is only observed in garnet inclusion trails and relict F2 fold hinge zones in the matrix, is invariably formed by quartz, muscovite, and chlorite. M2 is an upper greenschist to middle amphibolite grade event in the Clachnacudainn terrane and base of the Illecillewaet slice. M2 garnet grew pre- to late-D2 in the low structural level of the terrane (Areas 1, 4,

5) and late- to post-D2 in middle to high levels (Areas 2 and 3). Pre-D2 garnet growth in the low level is shown by straight (unfolded) S1i that are highly discordant to S2m (Fig. 4-3). Sigmoidal S1i are interpreted as a result of garnet growth after F2 crenulation folding and before formation of the S2 crenulation cleavage (Fig. 4-4a,b). S1i that vary continuously from highly (in core) to slightly (in rim) discordant to S2m record syn-D2 garnet growth because S1 must have been realigned by shearing related to F2 folding into parallelism with S2 during growth (Fig. 4-4c). Two stages of garnet growth in the low level are shown by core S1i that are discordant to subparallel rim S2i and S2m; the core grew pre-D2 and the rim grew late-D2. Well developed garnet pressure shadows and envelopment of the porphyroblast by S2m require that D2 continued after M2 in the low structural level. Garnet growth in middle to high structural levels (Areas 2 and 3) occurred during a more advanced stage of D2 (late- to post-D2) compared to the low level. Evidence for late-D2 garnet growth in these Areas is Si that are tightly crenulated but not totally transposed into S2. S2i which pass undisturbed through idioblastic garnet into S2m and the lack of garnet pressure shadows are evidence that growth occurred post-D2 (Fig. 4-5).

Post-D3 biotite, andalusite, and sillimanite grew during contact metamorphism (M3) associated with intrusion of late-tectonic granitoids in the Clachnacudainn igneous complex. This interpretation is made because these minerals are spatially related to the granitoids, and they overgrew all deformation fabrics and M2 minerals.

If it is assumed that folding (crustal thickening) associated with D2 was synchronous throughout the Clachnacudainn terrane, then times at which rocks passed through temperature conditions favourable for garnet growth must have varied. This diachroneity in garnet growth is most likely a function of depth of

burial because growth occurred at different times in rocks at different crustal levels in the Clachnacudainn terrane. The relatively early garnet growth (pre- to syn-D2) in the low structural level (Areas 1, 3, 5) compared to the later growth (late- to post-D2) in the high levels (Areas 2 and 3) readily supports this interpretation; since Area 1 is at least 1 km structurally lower than Areas 2 and 3, it would have reached temperatures necessary for garnet growth before the overlying levels simply because it received heat emanating from beneath sooner. Growth in the high structural level occurred post-D2 because D2 had ceased by the time these rocks were heated to temperatures necessary for garnet growth. Thermal models of Thompson and England (1984) support this interpretation by indicating that metamorphic reactions which are pre-deformational for more deeply buried rocks can be post-deformational for shallower rocks. In order for this hypothesis to be viable in the Clachnacudainn terrane, isotherms must have been upright and nearly horizontal. Such a thermal regime may be likely because the lowest structural level attained the highest metamorphic grade; the main source of heat must have been radioactive decay of thickened crust beneath. However, a static model of heat transfer during D2 may not be applicable because heat was probably advected by F2 folds.

Estimates of the peak P-T conditions of metamorphism

Introduction

The peak pressure and temperature (P-T) conditions of metamorphism in the Clachnacudainn terrane are estimated through i) comparison of mineral assemblages with experimentally and theoretically determined reaction equilibria and ii) use of compositional data in thermobarometric calculations.

Comparison of mineral assemblages with experimentally and theoretically determined reaction equilibria

AFM projections (Thompson, 1957) conveniently represent pelitic mineral assemblages and the reactions among them. AFM projections that are referred to below are shown in Figure 4-6 and the P-T positions of the metamorphic zones and reactions between them are shown in Figure 4-7. Map 2 shows the AFM assemblages in each sample and the inferred positions of isograds between the metamorphic zones.

The lowest grade AFM assemblage in the Clachnacudainn terrane is garnet-chlorite $\pm$ biotite, which is stable in the garnet zone. This assemblage requires that garnet zone pelite reached peak P-T conditions on the high temperature side of the discontinuous reaction

which breaks the chloritoid-biotite tie line and establishes the garnet-chlorite tie line. Staurolite becomes stable in schist with normal pelitic composition at a higher grade through the discontinuous reaction

which breaks the garnet-chlorite tie line and establishes the staurolite-biotite tie line. Figure 4-6 shows that this AFM topology is uniquely defined by the assemblage staurolite-chlorite-biotite (with garnet usually present as an extra phase that was probably stabilized by an additional component, such as Mn or Ca). The assemblage staurolite-garnet-biotite is stable in both the staurolite zone AFM topology and in the higher grade kyanite zone topology. The staurolite zone in the terrane contains the assemblage staurolite-garnet-biotite $\pm$ chlorite: since chlorite is locally in equilibrium with staurolite, the staurolite zone assemblage is chosen to be represented by the AFM topology shown in Figure 4-

6. However, staurolite zone rocks which lack equilibrium chlorite could also be represented by the kyanite zone AFM topology. Kyanite becomes stable in schist with normal pelitic composition through the discontinuous reaction

which breaks the staurolite-chlorite tie line and establishes the kyanite-biotite tie line. Kyanite zone pelitic schist in the terrane contains the diagnostic assemblage kyanite-staurolite-biotite. The kyanite zone AFM topology in Figure 4-6 shows that the assemblage staurolite-garnet-biotite is stable in relatively Fe-rich pelitic schist and kyanite-staurolite-biotite is stable in relatively Mg-rich pelitic schist. Thus, the difference between pelitic schist in the staurolite zone with the former assemblage and pelitic schist in the kyanite zone with the latter assemblage could be bulk composition rather than P-T conditions. However, I interpret the local presence of chlorite in equilibrium in the staurolite zone as indicating that Reaction (3) defines the isograd between the staurolite and kyanite zones. The presence of fibrolitic sillimanite in the kyanite zone and prismatic sillimanite in metasedimentary infolds within the Devonian orthogneiss (Thompson, 1972) suggests that an alumina-silicate phase boundary was encountered at this structural level. Since the sillimanite-bearing rocks documented by Thompson lack staurolite, then it is possible that they equilibrated on the high temperature side of the staurolite-out reaction

The sillimanite zone AFM topology is shown in Figure 4-6.

Pelitic schist in the garnet, staurolite, and kyanite zones attained peak P-T conditions in bathozone 4 (Carmichael, 1978). Pelitic schist in the garnet zone attained peak conditions in the P-T space bounded by Reaction (1) and Reaction (2) curves. On the pelitic petrogenetic grid (for the system $\text{K}_2\text{O}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-$

SiO₂-H₂O) in Figure 4-7 developed by Spear and Cheney (1989), at 5 kbar the temperature values on the Reaction (1) and (2) curves are 540°C and 560°C, respectively; at 7 kbar the values are 555°C and 575°C, respectively. Pelitic schist in the staurolite zone attained peak conditions in the P-T space that is bounded by Reaction (2) and Reaction (3) curves. Staurolite zone rocks therefore could have attained moderately higher temperature than garnet zone pelite (up 30°C higher at 5 kbar and 40°C at 7 kbar, Fig. 4-7). Pelitic schist in the kyanite zone attained peak conditions in the P-T space that is bounded by the Reaction (3) and staurolite breakdown curves. Minor fibrolitic sillimanite in kyanite-bearing pelite suggests that peak conditions occurred near the kyanite-sillimanite curve. The lowest temperature over this segment is 610°C at 6.4 kbar and the highest temperature is 715°C at 8.4 kbar (Fig. 4-7).

Andalusite within contact aureoles of undeformed granitoids in the Clachnacudainn igneous complex indicates that the younger, M3 Buchan-style metamorphism occurred at a significantly lower pressure than the regional, M2 Barrovian-style metamorphism. The minimum pressure difference between M2 and M3 on the southwest side of Albert stock is calculated with the following constraints: i) staurolite inclusions within andalusite require that peak M2 temperature conditions were above the Reaction (2) curve (it brings staurolite into schist with a normal pelitic bulk composition) and that M3 conditions were below the alumina-silicate triple point (about 3.8 kbar), and ii) peak M2 P-T conditions were in the kyanite field because kyanite is part of the M2 assemblage in the underlying metasedimentary rocks. Figure 4-7 shows that the minimum pressure attained during M2 is indicated by the intersection of the Reaction (2) and kyanite-sillimanite curves, which is 5.4 kbar at 570°C. The minimum pressure difference between M2 and M3 is therefore about 1.6 kbar (5.4 kbar - 3.8 kbar).

This suggests that a minimum of 5.3 km (assuming 3.3 km/kbar) of exhumation occurred between M2 and M3. The maximum pressure attained during M2 is indicated by the intersection of the staurolite breakdown and kyanite-sillimanite curves, which is 8.4 kbar at 715°C (Fig. 4-7). Assuming that the Albert stock was emplaced at a depth of at least 3.3 km (1 kbar), the maximum pressure difference between M2 and M3 is about 5.1 kbar. This suggests that a maximum of 16.8 km of exhumation occurred between M2 and M3. The sillimanite-K-feldspar-cordierite assemblage documented by Sears (1979) suggests that the peak temperature attained during M3 was higher than that attained during M2.

Thermobarometry

Thermobarometric estimates of the peak regional metamorphic P-T conditions generally agree with estimates made comparing mineral assemblages with the pelitic petrogenetic grid of Spear and Cheney (1989).

Description of analyzed samples

Thermobarometry was carried out on six pelitic rocks from the southern portion of the Clachnacudainn terrane, representing three structural levels, two regional metamorphic zones, and three lithologic units. Map 2 shows the location of the samples and the mineral assemblages in each sample are presented in Appendix 1. Samples 13b and 13d1 are quartz-rich pelitic schists that crop out within 100-200 m of each other and are correlated with the Broadview Formation. They are assigned a position in the low structural level because Devonian orthogneiss underlies them by a few hundred metres. Peak regional metamorphism (M2) at this level is pre- to syn-D2 (see above). Samples 13b and 13d1 contain the staurolite zone assemblage of staurolite-garnet-biotite-muscovite-

plagioclase-quartz. These minerals are considered to have been in equilibrium because all phases are in contact with sharp grain boundaries and no disequilibrium textures are observed other than minor retrograde chlorite on garnet and staurolite margins, and slightly chloritized biotite. The kyanite zone underlies these samples by a few hundred metres. Sample 135 is a pelitic schist that contains the characteristic garnet zone equilibrium assemblage garnet-biotite-muscovite-plagioclase-chlorite-quartz and is correlated with the Index Formation. Chlorite occurs both as an equilibrium phase and as a retrograde phase on garnet margins and in biotite intergrowths. Sample 135 is assigned a position in the middle structural level, in which M2 is syn- to late-D2, because it overlies 13b and 13d1 by a few hundred metres. Sample 63 is a graphitic pelitic schist that contains the characteristic garnet zone equilibrium assemblage and is correlated with the Index Formation. It is assigned a position in the high structural level, in which M2 is post-D2, because it is less than 50 m beneath the Standfast Creek fault. Two samples, 72a and 76, that are at the same structural level and metamorphic grade as 63 were analyzed only for a temperature estimate because they lack the phases required for a pressure estimate. 72a and 76 are schists from pelitic horizons within the quartzite-dominated Hamill Group. Unfortunately, no garnet-bearing samples with the proper mineral assemblages in the hanging wall of the Standfast Creek fault were available at the time of microprobe analysis.

Analytical methods

Mineral compositions were determined by analyzing carbon-coated, polished thin sections on the ARL-SEMQ electron microprobe at the University of Calgary. Accelerating voltage was 15 kV, beam current was 15 nA, and a focused beam diameter of about 1 micron was used for all analyses. Nine-

element analyses were carried out on garnet and plagioclase and eleven-element analyses were done on muscovite and biotite. The natural minerals and oxides used as standards are presented in Appendix 3. Data reduction was performed according to the procedure of Bence and Albee (1968).

Garnet

Selected garnet core and rim compositions generated by microprobe analysis are presented in Table 1 of Appendix 4. This table indicates that the $\text{Fe}/(\text{Fe} + \text{Mg})$ ratio in garnet rims is related to structural level and metamorphic grade. Rims in the staurolite zone samples (13b and 13d1) from the lowest structural level are richer in Mg than rims in the garnet zone sample (135) from the middle structural level, which are richer in Mg than rims in the garnet zone samples (63, 72a, 76) from the highest structural level. The grossular (Ca) and spessartine (Mn) components in garnet rims are also roughly related to metamorphic grade, as the highest grade rocks have the lowest values of each.

Compositional profiles obtained by microprobe analysis across garnet from each of the six samples indicate significant zoning. The location of the profiles in garnets from 13b and 135 are shown in Figure 4-8, the profiles from these samples and 76 are shown in Figures 4-9, and a compositional map of Mn in garnet from 13b is shown in Figure 4-10. All analyzed garnets have overall prograde growth zoning profiles (Tracy, 1982); compared with the core, rims are enriched in almandine (Fe) and pyrope (Mg) and depleted in spessartine (Mn) and grossular (Ca). The preservation of this compositional zoning indicates that volume diffusion must have been slow in order for the core to be effectively isolated from the rest of the rock. However, at the edge of the garnet there are zoning perturbations and reversals that are probably due to retrograde reequilibration

and diffusion. Because these zoning perturbations and reversals greatly affect the thermobarometric estimation of peak P-T conditions, they are discussed in detail. Zoning profiles of garnets in 13b (Figure 4-9a,b,c,d) will be specifically used to outline these relationships because their zoning is best characterized.

Spessartine throughout most of the garnet has a smooth, classical "bell-shaped" zoning profile (decreasing from core to rim, Fig. 4-10). Within the outermost 150 microns, the curve flattens then gently reverses slope (slopes upward toward the edge of the grain) as spessartine slightly increases to the edge. The grossular profile is nearly flat across the core. At about 500 microns from the edge, the curve steepens greatly downward toward the edge as grossular decreases dramatically. The curve flattens again at about 200 microns from the edge as the rate of decrease in grossular moderates. Within the outermost 50 microns of the grain, the grossular sharply decreases again to the edge. Profiles such as this are referred to as "stair step" profiles because of their steep-flat-steep shape. The almandine in the core has a smooth, inverted "bell-shaped" zoning profile (increasing from core to rim). In the interval from 400 to 50 microns from the edge, the curve flattens and then slightly reverses slope (slopes downward toward the edge) as almandine decreases. Within the outermost 50 microns, the profile steepens dramatically upward again as the amount of almandine increases to the edge. The steeps and flats in the grossular and almandine stair step profiles occur at similar positions in every garnet; the grossular and almandine curves in the rims are nearly mirror images of each other. The pyrope profile is flat across the core. At about 500 microns from the rim (the same place in which grossular first decreases dramatically), the pyrope curve steepens moderately upward toward the edge. The pyrope curve levels at about 50 microns from the edge, then at 20 microns it reverses slope (slopes downward) decreases to the

edge. The resultant $\text{Fe}/(\text{Fe} + \text{Mg})$ curve is flat in the core, moderately downward sloping in the interval from 600 to 50 microns from the edge, and upward (reversed compared to the majority of the curve) sloping over about the outermost 50 microns. Detailed compositional profiles across garnet in 135 (Fig. 4-9e,f) show zoning curves similar to those in 13b. One difference is that the flats which occur in the interval from 400 to 50 microns from the edge of garnet 13b are compressed into a maximum of 100 microns in garnet 135. Another difference is that the grossular peak in 135 does not occur in the core but rather between the core and rim. Less detailed profiles across garnet in 63 and 76 also suggest similar compositional zoning. The grossular peak between the core and rim in garnet 76 is particularly well developed (Fig. 4-9g).

These compositional profiles can be used to infer the garnet-producing reactions and the changes in the reactant assemblage (Tracy, 1982). The decrease in the $\text{Fe}/(\text{Fe} + \text{Mg})$ ratio between the core and rim probably indicates that material supplied to the growing garnet edge over this interval is depleted in Fe. This trend is in accord with the many garnet-producing continuous reactions represented by a T-X(Fe-Mg) loop which moves to lower X_{Fe} upon increasing temperature (Thompson, 1976). Presumably, the fairly smooth and continuous curves from the core to about 300-400 microns from the edge in garnet 13b represent garnet growth during a single prograde reaction. One such typical garnet-producing reaction in chlorite-bearing pelite is the model reaction

The dramatic decrease in the amount of grossular between the core and rim could signal the growth of more calcic phases (such as plagioclase) or a change in the (Ca, Al) phase in the reaction assemblage. For instance, it could signal the onset of plagioclase reacting to form garnet rather than another (Ca, Al) phase, such as

zoisite, or vice-versa. Zoisite grains totally enclosed by chlorite in pelitic schist near 13b and 13d1 suggest that this reaction was operative.

The greater amount of shallowly sloping profiles in the rim of garnet 13b (in the interval from 400 to 50 microns from the edge) compared to garnet 135 (less than 100 microns of shallowly sloping profiles) could have resulted from the onset of diffusional homogenization in the former sample and the lack of diffusion in the latter. This relationship is expected because 13b attained a higher temperature than 135, allowing more time for diffusional homogenization to operate. The reversal in the slope of the spessartine, pyrope, and $\text{Fe}/(\text{Fe} + \text{Mg})$ curves and the steepening of the almandine and grossular curves in the outermost 50 microns is interpreted as a result of garnet-consuming reactions. These reactions could either be prograde garnet-consuming reactions or retrograde resorption of garnet during post-peak cooling and reequilibration. One prograde garnet-consuming reaction is the discontinuous Reaction (2) that allows staurolite growth in schists with normal pelitic compositions. Since zoning perturbations and reversals at the edge occur in staurolite-absent samples, then other garnet-consuming reactions must have been operative. The ubiquity of late chlorite on garnet margins suggests that retrograde resorption of garnet by chlorite and the resultant ion diffusion across the garnet-chlorite interface during reequilibration is responsible for the anomalous edge zoning. This inference is supported by the zoning profiles, as minor resorption of the garnet edge accompanied by production of a more magnesian phase such as chlorite drives garnet edge compositions toward a higher $\text{Fe}/(\text{Fe} + \text{Mg})$ ratio and higher spessartine content (Tracy, 1982). The accentuated decrease in grossular in the disturbed edge could be ascribed to production of some anorthite during the resorption reaction. Compositional evidence for this reaction in plagioclase is ambiguous; whereas

anorthite increases in some grains at the edge, it decreases in others. Although samples that were chosen for microprobe analysis have the least amount of retrograde chlorite possible (only a few chlorite laths occur on margins of idioblastic to subidioblastic garnet), enough retrogression must have occurred to alter the garnet edge compositions.

The thermobarometric estimates of peak P-T conditions are extremely dependent on which garnet composition is taken to have been in equilibrium with the matrix during peak metamorphism (Spear, 1989). Since all garnet is affected by retrograde reequilibration shown in the zoning reversals and perturbations outlined above, the highest P-T conditions are obtained from compositions that are closest to the edge yet unaffected by ion diffusion. For the thermobarometric calculations, I used compositions at the minima in the Fe/(Fe+Mg) curve (about 50 microns from the edge) because they yield the highest temperatures; these compositions presumably were not altered by post-peak reequilibration. The minima in the Fe/(Fe+Mg) curve occur a few tens of microns outside the minima in the spessartine curve, which is predictable because spessartine tends to decrease during prograde metamorphism and increase during cooling and/or retrograde metamorphism (Tracy, 1982). The compositions used in the thermobarometric calculations are termed "50 micron compositions" rather than "peak compositions" because I realize that the compositions stable during peak metamorphism were probably altered within the outermost 50 microns during post-peak cooling and are hence unrecoverable. The 50 micron compositions therefore yield lower P-T estimates than true peak P-T conditions. However, peak compositions may not have differed greatly from the 50 micron compositions because the altered zones in garnet edges are narrow and the porphyroblast is not noticeably embayed by chlorite (garnet is idioblastic to subidioblastic). The

disturbed "edge compositions" could represent post-peak conditions, but diffusive alteration during retrogression may have changed the edge to a composition that does not accurately reflect the temperature of cooling. The thickness of the rim affected by retrogression (50 microns or less) reveals the necessity for obtaining numerous detailed compositional profiles across garnet rims. For instance, without these detailed profiles this study would have been oblivious to the zoning perturbations and reversals, and edge compositions would have been used to yield erroneous estimates of peak P-T conditions.

Biotite

Selected biotite core and rim compositions generated by microprobe analysis are presented in Table 2 of Appendix 4. This table indicates that biotite is virtually unzoned and that Fe/(Fe + Mg) ratios are roughly related to metamorphic grade, as highest grade samples are richest in Mg.

Peak P-T conditions will only be obtained through thermobarometric calculations if compositions that reflect peak conditions are used. It is therefore necessary to obtain biotite compositions that were in equilibrium with garnet prior to post-peak reequilibration (Spear, 1989). If biotite that was adjacent to garnet continued exchanging ions after the metamorphic peak, it is possible that the composition of biotite in the matrix away from garnet may more accurately reflect the peak metamorphic composition. This possibility was checked by assessing the consistency of the biotite with the microprobe; there is no discernible compositional variation between matrix grains and those adjacent to garnet. Since Fe-Mg diffusion in biotite is rapid compared to garnet, biotite compositions could be consistent because analyzed samples generally contain a large modal amount of biotite, which, due to mass balance constraints, may have assured that biotite

compositions changed little after peak temperature (Spear, 1989).

Muscovite

Selected muscovite core and rim compositions generated by microprobe analysis are presented in Table 3 of Appendix 4. This table indicates that muscovite compositions do not vary systematically with structural position and zoning was not discerned.

Plagioclase

Selected plagioclase core and rim compositions from matrix and inclusion (in garnet) grains generated by microprobe analysis are presented in Table 4 of Appendix 4. Compositional profiles obtained by microprobe traverses across plagioclase indicate strong to weak zoning in the anorthite (Ca) and albite (Na) components. These profiles are shown in Figure 4-11. The compositional profiles will be described in detail for each sample from the lowest to highest structural position because zoning has profound effects on the thermobarometric estimation of peak P-T conditions.

Some plagioclases in 13b have flat zoning profiles (Fig. 4-11a), others have cores slightly enriched in anorthite compared to rims (An22 versus An16, Fig. 4-11b), whereas others have cores slightly depleted in anorthite (An17 versus An20, Fig. 4-11c). At the edges of some grains there are abrupt zoning reversals and anomalous variations in composition (the slope of curve remains the same but steepens dramatically). These edge features are interpreted as results of post-peak reequilibration. Since there is no consistent variation at the edges in 13b (i.e., similar variations were not measured at all edges in the same grain), a reequilibration mechanism cannot be proposed. In all samples, the plagioclase

composition that is used in the thermobarometric calculation of peak P-T conditions is from just inside the reversals and anomalous variations at the edge. Cores of matrix plagioclase in 13d1 are slightly depleted in anorthite relative to the rims and slightly enriched in anorthite compared to plagioclase 13b cores (Fig. 4-11d). Most plagioclases in 13d1 that are included within garnet are unzoned and moderately richer in anorthite compared to matrix grains (roughly An35 versus An30, Fig. 4-11e). Two plagioclase inclusions have discrete zones of anomalously high anorthite (up to An51, Fig. 4-11f) and one grain is extremely anorthite-rich throughout (from An57 to An71, Fig. 4-11g). The origin of these anomalously anorthite-rich compositions is unknown. Cores of two matrix plagioclase grains in 63 are significantly depleted in anorthite relative to the anorthite peaks in the rims (An29 versus An48, and An26 versus An43, Fig. 4-11h). The edges of the matrix plagioclase in 63 are reversely zoned, as anorthite decreases by An08 from the anorthite peaks over a distance of 15 microns. This variation in anorthite at the edges significantly affects thermobarometric calculations. Because plagioclase either becomes more calcic or sodic during regional metamorphism, it is assumed that peak conditions are recorded (or nearly so) by the anorthite peaks in the rims, and the edge reversals are interpreted as resulting from post-peak reequilibration. The cores of plagioclase inclusions within garnet in 63 (Figs. 4-11i,j) have anorthite contents equal to the cores of matrix grains. However, the anorthite peaks in the rims of inclusion grains (between An35 and An40) are lower than that in matrix grains. Also, edge reversals in the inclusion plagioclase are less significant, as some edges show no reverse zoning and the maximum decrease is An02 over 37 microns. The lack of edge zoning reversals in inclusion plagioclase is evidence that the edge reversals in matrix plagioclase are due to post-peak reequilibration with other matrix

phases. Cores of the two analyzed plagioclase grains 135 are enriched in anorthite relative to the rims (An₃₄ versus An₂₉ (Fig. 4-11k) and An₃₈ versus An₂₈ (Fig. 4-11l)).

Spear (1989) noted that the effective bulk composition of pelitic rocks changes through time and is continuously being depleted in the components that are enriched in garnet, mostly Ca and Mn. If garnet and plagioclase are the only Ca-bearing phases in a closed system, then mass balance constraints suggest that as the matrix source becomes depleted in Ca garnet should zone toward lower grossular in the rim and plagioclase inclusions within garnet and plagioclase cores should be richer in anorthite than rims of matrix plagioclase. The compositional profiles described above indicate that these expected Ca zoning trends in garnet and plagioclase are rare; grossular peaks in most garnets occur between the core and rim and plagioclase cores are either richer or poorer in anorthite than the rims. In some cases, these zoning variations can be interpreted as a result of the breakdown of other Ca-rich phases, such as epidote and zoisite, which may have released sufficient Ca to permit both garnet and coexisting plagioclase to become increasingly Ca-rich (Spear, 1989).

Results of GE0-CALC

Estimates of temperature and pressure conditions during peak metamorphism (M2) were made using the PTAXSS computer program, which is part of the GE0-CALC software package (Berman et al., 1987; Brown et al., 1988). The calculations were made using the thermodynamic data base of Berman (1988, 1990) and non-ideal mixing is accounted for using the calibrations of Berman (1988, 1990) for garnet, McMullin et al. (1991) for biotite, Fuhrman and Lindsley (1988) for plagioclase, and Chatterjee and Froese (1975) for

muscovite. Only three equilibria are generated by garnet, biotite, muscovite, and plagioclase in the system K_2O - CaO - FeO - MgO - SiO_2 - Al_2O_3 . The thermometer is the following equilibrium that exchanges Fe and Mg between biotite and garnet:

The barometer is the net transfer equilibrium that exchanges Ca between the anorthite component of plagioclase and the grossular component of garnet and is balanced with muscovite and biotite. The Fe and Mg end members are

and

These three equilibria should intersect at a single point in P-T space. The resultant P-T point represents peak metamorphic conditions only if the mineral compositions upon which the equilibria are based were in equilibrium during the metamorphic peak. Determining the compositions which were in equilibrium during the metamorphic peak, even with detailed compositional profiles of compositionally zoned minerals, is a difficult endeavour (Spear, 1989). For reasons given above, the inferred peak compositions of garnet and plagioclase are generally not at the edge, but rather inside the edge by 10-70 microns.

The compositions used in the thermobarometric calculations are taken from minerals within the same domain. Domains are regions that are generally less than 200 microns in diameter and are centred on garnet rims that were profiled with detailed microprobe traverses. Table 4-1 presents all reasonable and permissible combinations of mineral compositions within the same domain which were used in the thermobarometric calculations. The compositions in Table 4-1 and Appendix 4 represent single analyses, averages of analyses from the same grain, and averages of analyses from multiple grains. Most calculations are based

on closely spaced and often mutually adjacent garnet, biotite, and muscovite that are grouped with the nearest plagioclase, which is usually not in mutual contact. Other calculations are based on minerals that are not in mutual contact but rather may be separated by up to hundreds of microns. Since compositions of minerals in mutual contact do not vary in any systematic manner from compositions of minerals that are separated, calculations based on each are similar. As outlined above, the most crucial aspect of the thermobarometric calculations is choosing compositions that were in equilibrium during peak metamorphism, a task that is difficult at best in complexly zoned garnet and plagioclase.

Samples 13b and 13d1 both yield a cluster of equilibria intersections on the GE0-CALC P-T diagram in the vicinity of 600°C and 6.8 kbar (Table 4-1). Boxes outlining the scatter of equilibria intersections are shown on the petrogenetic grid of Spear and Cheney (1989) in Figure 4-12. It is important to note that this petrogenetic grid and the GE0-CALC software use the same thermodynamic data base. The similar P-T conditions for 13b and 13d1 are expected because they crop out within 100-200 m of each other. The amount of scatter in the intersections from the centre of the boxes is about $\pm 25\text{-}30^\circ\text{C}$ and 0.5-0.6 kbar, which is a realistic amount of uncertainty. Figure 4-12 shows that the scatter boxes lie on the high pressure side of the kyanite-sillimanite reaction curve, mostly on the high temperature side of the staurolite-in Reaction (2) curve, and mostly on the low temperature side of the alumina-silicate-in Reaction (3) curve. These peak P-T conditions are expected for 13b and 13d1 because these staurolite zone schists overlie the kyanite zone by a few hundred metres.

Samples 63 and 135 yield equilibria intersections that cluster around 550°C and 5.2 kbar (Table 4-1). Boxes outlining the scatter in the P-T intersections are shown on the P-T grid in Figure 4-12. Figure 4-12 shows that the scatter boxes

overlap Reactions (1) and (2) and straddle the kyanite-sillimanite boundary. Such peak P-T conditions in these garnet zone pelitic schists are expected from comparison with the petrogenetic grid (Figure 4-7).

Since 72a and 76 occur at a similar structural level as 63 and 135 and their pressure at peak metamorphism cannot be estimated because they lack plagioclase, their peak temperature was estimated at 5.0 kbar. 72a yields temperature estimates between 560 and 610°C and 76 yields estimates between 560 and 580°C (Table 4-1). This temperature range overlaps the Reaction (2) curve (Figure 4-12), which is expected since these staurolite-absent schists are near staurolite-bearing schist mapped by Sears (1979).

Estimation of a P-T path using techniques based on garnet and plagioclase zoning (Spear and Selverstone, 1983) was not attempted because of the lack of biotite inclusions in garnet and because of the ambiguities imposed by not knowing which minerals were present during early stages of garnet growth (Frost and Tracy, 1991). For instance, it is uncertain when plagioclase, staurolite, and biotite began to grow because of the lack of inclusions in garnet.

Discussion of thermobarometric P-T estimates

As one would expect in equilibrated samples, thermobarometric P-T estimates (Figure 4-12) agree closely with estimates made by comparing mineral assemblages with the pelitic petrogenetic grid of Spear and Cheney (1989, Figure 4-7). These estimates support my assertion that maximum regional metamorphic grade is middle amphibolite facies in the base of the Clachnacudainn terrane and lower amphibolite facies at the uppermost level.

Figure 4-12 shows that the boxes outlining the scatter of equilibria intersections from samples in the lower structural level, 13b and 13d1, do not

intersect the scatter boxes from samples in the middle to upper structural levels, 135 and 63. Although the temperature estimates slightly overlap, the pressure estimates of the former are considerably higher than the latter; taking the centre of the scatter boxes as the average pressure, lower level pressure estimates are greater by about 1.6 kbar than the middle and upper levels, a pressure difference corresponding to a 5.3 km difference in depth (at 3.3 km/kbar). This pressure difference is greater than that expected if the isobars were nearly horizontal because 13b and 13d1 underlie 135 by about 800 m in elevation and 63 by about 1.4 km. One explanation for the larger than expected pressure difference is that the pressure calculations are erroneous because of problems discussed above concerning the choice of peak compositions in complexly zoned garnet and plagioclase. Another possibility is that the pressure difference is real, a result of extensional displacement between the localities of the apparently higher and lower pressure samples. Although I consider a discrete fault to be unlikely because mapping shows that lithologic units are neither attenuated nor omitted between the localities (Thompson, 1972; Read and Thompson, 1980; this study), it is possible that the entire Clachnacudainn terrane is in a zone of distributed extensional shear. These shears have perhaps not been documented because they parallel the main foliation and exposure on the heavily timbered slopes is poor. The possibility of faulting between the lower and upper levels of the terrane clearly needs to be explored by additional fieldwork before it is ruled out or considered further.

Table 4-1. P-T estimates based on thermobarometric calculations outlined in the text from all reasonable combinations of mineral compositions within the same domain. The letters in the columns below garnet (gt), biotite (bt), muscovite (mu), and plagioclase (pl) refer to compositions presented in Tables 1-4, respectively, of Appendix 4. These P-T estimates are shown on the P-T grid in Figure 4-12.

Sample 13b

Gt	Bt	Mu	Pl	T(C)	P(kbar)
A	A	A	A	615	7.25
A	A	A	B	614	6.95
B	A	A	A	607	6.85
B	A	A	B	605	6.55
C	A	A	A	614	7.57
C	A	A	B	613	7.27
D	A	A	A	611	7.26
D	A	A	B	610	6.97
E	A	A	A	612	7.01
E	A	A	B	610	6.71
F	B	A	D	623	7.40
F	B	A	C	623	7.32
G	B	A	D	607	6.58
G	B	A	C	607	6.52
H	C	B	E	572	6.27
H	C	B	F	577	7.31

Sample 13d1

Gt	Bt	Mu	Pl	T(C)	P(kbar)
A	A	A	A	589	6.51
A	A	A	B	580	6.82
A	A	A	D	587	6.42
A	A	A	C	588	6.59
B	B	A	E	572	6.70
B	B	A	G	572	6.60
B	B	A	F	573	6.76
C	C	A	A	609	6.75
C	C	A	D	609	6.55
D	B	A	E	585	6.19
D	B	A	G	585	6.08
D	B	A	F	585	6.23
A	D	A	A	628	7.14
A	D	A	B	627	7.39
A	D	A	D	625	6.95
A	D	A	C	628	7.14
E	D	A	A	633	7.82
E	D	A	D	632	7.55
E	D	A	C	634	7.82

Sample 63

Gt	Bt	Mu	Pl	T(C)	P(kbar)
A	A	A	A	519	5.02
B	A	A	A	547	5.18
C	A	A	A	554	5.12
C	A	A	B	552	4.96
C	A	A	C	553	5.04
C	A	A	D	554	5.27
C	A	A	E	554	5.21
C	A	A	F	553	5.15
C	B	A	G	553	5.16
C	B	A	B	565	5.14
C	B	A	C	566	5.21
C	B	A	D	567	5.45
C	B	A	E	567	5.38
C	B	A	F	566	5.32
C	B	A	G	566	5.33
C	C	A	B	544	4.86
C	C	A	C	544	4.92
C	C	A	D	545	5.15
C	C	A	E	545	5.09
C	C	A	F	545	5.03
C	C	A	G	545	5.03

Sample 135

Gt	Bt	Mu	Pl	T(C)	P(kbar)
A	A	A	A	538	4.67
A	A	A	B	539	4.73
A	B	A	A	570	5.12
A	B	A	B	571	5.17
B	C	B	A	526	4.90
B	C	B	C	526	4.96
C	C	B	A	521	4.95
C	C	B	C	521	5.00
D	C	B	A	516	4.63
D	C	B	C	516	4.68
B	B	B	A	576	5.67
B	B	B	C	578	5.73
B	D	B	A	553	5.35
B	D	B	C	553	5.37
C	B	B	A	572	5.70
C	B	B	C	572	5.77
C	D	B	A	548	5.35
C	D	B	C	548	5.40
D	B	B	A	566	5.37
D	B	B	C	567	5.43
D	D	B	A	542	5.02
D	D	B	C	543	5.08

Sample 72a

Gt	Bt	T(C) at 5 kbar
A	A	585
B	A	556
A	B	589
C	C	578
C	D	609

Sample 76

Gt	Bt	T(C) at 5 kbar
A	A	577
B	B	563
B	C	564
B	D	580
C	E	574
C	F	564
C	G	576
C	H	567

Figure 4-1. Relative timing of metamorphism versus deformation in each of the five Areas that are outlined in Figure 4-2. See sketches in Figures 4-3 through 4-5 for the evidence upon which these interpretations are based.

Figure 4-2. Location of the five textural areas referred to in text and Figure 4-1. Akolkolex anticline (AA), Albert Canyon (AC), Albert Canyon anticline (ACA), Albert Peak (AP), Columbia River fault (CRF), Drimmie Creek syncline (DCS), East Twin Creek (ETC), Ghost Peak (GP), Ghost Peak anticline (GPA), lower Standfast Creek fault (LSCF), Mount Cartier syncline (MCS), Mount Klotz (MK), Mount LaForme (MLF), Mount Llewelyn (ML), Mount Revelstoke (MR), North Albert Peak (NAP), Sale Creek (SC), Twin Creek fault system (TCFS), upper Standfast Creek fault (USCF), West Twin Creek (WTC), West Woolsey Creek (WWC).

Figure 4-3. Sketches from photomicrographs of garnet (gt, sketches a and b) and plagioclase (pl, sketch c) that grew entirely before the second phase of deformation (D2) in the Clachnacudainn terrane. This interpretation is made because the oldest schistosity is preserved within the porphyroblasts as inclusion fabrics (S1i) that are nearly undeformed (they are straight) by second phase of folds (F2). F2 folding transposed S1 into the dominant fabric in the matrix (S2m), which envelopes the porphyroblasts. Note the quartz (qz)-rich pressure shadows around the porphyroblasts. S1i is defined by flattened quartz whereas S2m is defined by flattened quartz, muscovite (mu) and biotite (bt).

pre-D2
(Area 1) pre-D2
(Area 1) pre-D2
(Area 4)

Figure 4-4. Sketches from photomicrographs of garnet (gt) that grew during the second phase of deformation (D2) in the Clachnacudainn terrane. This interpretation is made because the S1 inclusion fabrics (S1i) are variably deformed by second phase folds (F2). F2 folding must have continued in the matrix after garnet grew because it transposed S1 into the dominant fabric in the matrix (S2m), which envelopes the garnets. Note the quartz (qz)-rich pressure shadows around the garnets. S1i is defined by flattened quartz and opaques (black) whereas S2m is defined by these minerals plus muscovite (mu), biotite (bt), and staurolite (st). The relict F2 fold hinge zone in a pressure shadow in sketch b has a similar (but tighter) orientation as the folded S1i. This relationship proves that garnet overgrew a crenulated S1 rather than having undergone syn-tectonic rotation. Garnets in sketches a and b are interpreted to have grown either during a punctuation in D2 or very rapidly relative to F2 folding because there is no core-to-rim change in the geometry of S1i. In contrast, the straight S1i in the core of the garnet in sketch c curves into parallelism with S2 in the rim. S1 is therefore interpreted to have been sheared and rotated around the garnet by F2 folds into the S2 orientation during incremental garnet growth.

syn-D2
(Area 1) syn-D2
(Area 1) syn-D2
(Area 1)

Figure 4-5. Sketches from photomicrographs of garnet (gt) that grew entirely after the second phase of deformation (D2) and prior to the third phase of deformation (D3) in the Clachnacudainn terrane. This interpretation is made because i) inclusion fabrics in garnet are the second schistosity (S2i) because they are continuous with and parallel to the dominant matrix fabric (S2m) and ii) garnet did not overgrow the third phase crenulation folds (F3) or the third schistosity (S3) that are well developed in the matrix. Note that in response to F3 folding, quartz (qz)-rich pressure shadows formed around garnet and S3 was perturbed around garnet. S2i is formed by flattened quartz and opaques (black) whereas S2m is formed by these minerals plus muscovite (mu).

post-D2/pre-D3 (Area 2)

post-D2/pre-D3 (Area 3)

Figure 4-6. AFM projections (Thompson, 1957) for the metamorphic zones in the Clachnacudainn terrane. Numbers beneath the reactions between topologies refer to those discussed in the text. Abbreviations are those shown in Appendix 2. Letters within the topologies represent assemblages in samples that I studied in thin section, which are presented in Appendix 1.

Figure 4-8. Location of microprobe profiles across garnets 13b-1, 13b-2, and 135. Profiles are shown in Figure 4-9. Letters A and B at end of profiles refer to rim locations in Figure 4-9 and Table 1 of Appendix 4. Mineral abbreviations are defined in Appendix 2.

Figure 4-9. Garnet compositional profiles obtained with the microprobe. See text for discussion of the strong zoning in the almandine (alm), grossular (grs), pyrope (prp), and spessartine (sps) components. The sample number and profile number are separated by a dash above each profile. Locations of the profiles in 13b and 135 are shown in Figure 4-8. Rim A and Rim B refer to specific garnet rims in Figure 4-8 and Table 1 of Appendix 4.

a. Garnet profile 13b-1(2)

b. Garnet profile 13b-1(3)

c. Garnet profile 13b-1(1)

d. Garnet profile 13b-2

e. Garnet profile 135-1(1)

f. Garnet profile 135-1(2)

g. Garnet profile 76-1

Figure 4-10. Mn (spessartine) zoning in garnet 13b-2. This qualitative picture of Mn zoning was obtained by counting for 4 seconds about every 2 microns with the microprobe. Sketch of this garnet in Figure 4-8 shows the scale and location of the quantitative compositional profile across this garnet (Figure 4-9d). Normal prograde growth zoning is indicated by the greatest Mn concentration in the core (pink coloured areas) and minor reverse zoning at the edge is indicated by a slight increase in Mn within the outer 50 microns.

Figure 4-11. Plagioclase compositional profiles obtained with the microprobe. See text for discussion of the zoning in the anorthite (an) and albite (ab) components. The sample number and profile number are separated by a dash above each profile. Rim A and Rim B refer to specific plagioclase rims in Table 4 of Appendix 4.

a. Plagioclase profile 13b-4

b. Plagioclase profile 13b-3-4

c. Plagioclase profile 13b-2-2

d. Plagioclase profile 13d1-2-2(1)
matrix

e. Plagioclase profile 13d1-2l(a)
inclusion

f. Plagioclase profile 13d1-2l(e1)
inclusion

g. Plagioclase profile 13d1-2l(g)
inclusion

h. Plagioclase profile 63-1a
matrix

i. Plagioclase profile 63-P2
inclusion

j. Plagioclase profile 63-P3
inclusion

k. Plagioclase profile 135-1-2

l. Plagioclase profile 135-2

Figure 4-12. Scatter of P-T estimates through thermobarometry on the P-T grid of Spear and Cheney (1989). See text for the thermobarometry method used to make these P-T estimates. Boxes for 13b, 13d1, 63, and 135 outline the P-T limits shown in Table 4-1 and the lines for 72a and 76 show the temperature limits at 5 kbar. Numbers on reactions refer to those discussed in the text. See Figure 4-7 for a comparison with P-T estimates based on mineral assemblages. Mineral abbreviations are defined in Appendix 2.

CHAPTER 5. Deformation

Introduction

Three structural fabrics and associated folds in the Clachnacudainn terrane and base of the overlying Illecillewaet slice are evidence for three phases of penetrative ductile deformation. An upper age limit of 104 ± 1 Ma is placed on this deformation by intrusion of the late-tectonic Albert stock (U-Pb zircon age, Chapter 3) and a lower age limit of Late Devonian is provided by a penetratively deformed orthogneiss (U-Pb zircon age from Parrish, 1992). Brittle faulting related to the youngest deformational phases post-dated intrusion of the Albert stock. The main conclusion in this chapter is that deformational phases on both sides of the Standfast Creek fault are correlatable, which is consistent with stratigraphic evidence that suggests the Standfast Creek fault is not a boundary between different tectonic levels.

Following the descriptions of the deformational phases, evidence for ductile shearing along the Standfast Creek fault is presented. The final section of the chapter discusses the deformation in relation to the fault. The relative timing between deformation and metamorphism is outlined in Chapter 4.

First phase of deformation (D1)

Evidence for the first phase of deformation (D1) is major map scale F1 folds (such as the Akolkolex anticline and Ghost Peak anticline, see below), possibly accompanied by the rarely observed oldest schistosity (S1). S1 is poorly

preserved because it is strongly folded and transposed by phase 2 folds (F2) into the dominant, axial planar second phase schistosity (S2). S1 is therefore only seen distinct from S2 in F2 fold hinge zones and as inclusion trails in porphyroblasts. S1 was not recognized by previous fieldworkers. The presence of S1 as inclusion trails within peak metamorphic minerals and the fact that it is formed exclusively of quartz, muscovite, and chlorite indicate that S1 developed prior to peak metamorphism. Since S1 always parallels compositional layering (S0), then F1 folds must be large isoclinal responsible for the transposition. I interpret the major isoclinal in the study area, such as Akolkolex and Ghost Peak anticlines, as F1 folds (see below). No definite refolded F1 minor folds were observed.

Second phase of deformation (D2)

The second phase of deformation (D2) was penetrative in the entire Clachnacudainn terrane and base of the Illecillewaet slice, as second phase folds (F2) are well developed throughout. F2 minor folds are characterized by attenuated, often rootless, tight to isoclinal, overturned to recumbent folds with thickened hinge zones and axial surfaces that form the penetrative S2 (Plate 5-1). S0 and S1 are transposed into the S2 orientation along F2 fold limbs and are only discernible from S2 in F2 fold hinge zones. S2 orientations in eight structural domains (which are located on the map in Figure 5-1) are shown on stereonet in Figure 5-2, and F2 hinge lines are shown on stereonet in Figure 5-3. North of the Illecillewaet River, S2 generally dips moderately (35-50°) northeast. Large-scale third phase folds (F3) locally warp S2 in this region. South of the Illecillewaet River, the orientation of S2 varies widely; S2 in the Albert Peak area dips moderately northeast, along the Illecillewaet-Akolkolex divide S2 dips moderately south, in the Akolkolex River and Standfast Creek valleys it dips

moderately to shallowly northeast through east, and it has no consistent orientation in the Ghost Peak area and on the east slope of the Columbia River valley. Large-scale F3 folds may be responsible for some of this dispersion. The shallowly plunging F2 fold hinge lines generally trend southeast, which is up to 20° more southerly trend than F3 fold hinge lines. Chapter 4 provides evidence that proves peak metamorphism occurred syn- to post-D2.

Thompson (1972) named the folds with F2 characteristics as "type 1 folds" in the Clachnacudainn terrane and base of the Illecillewaet slice. He interpreted type 1 folds as the oldest structures because their axial planar schistosity was the oldest fabric he observed. Thompson concluded that structural style remains consistent across the Standfast Creek fault because type 1 folds occur in its footwall and hanging wall (he placed structural domains for type 1 folds across the fault). Zwanzig (1973) noted that moderately inclined, nappe-like folds, such as the Albert Canyon anticline, dominate the base of the Illecillewaet slice. Sears (1979) noted that the axial planar schistosity of tight overturned folds is the dominant fabric in the terrane and base of the Illecillewaet slice; Sears' F1 folds are equivalent to my F2.

Major folds

Thompson (1972) mapped two major type 1 folds, the Akolkolex anticline and Drimmie Creek syncline, in the base of the Illecillewaet slice. These recumbent isoclines have amplitudes approaching 6.5 km and a relatively constant interlimb thickness of 800 m or less (Thompson, 1972). The Akolkolex anticline was ascribed a northeast vergence (I define vergence as the direction of the upper member of the rotational couple implied by the asymmetry of the fold) based on a projection of its apparent closure down the plunge of minor folds which

Thompson assumed are related to the major structure. The Drimmie Creek syncline was ascribed a northeast vergence based on stratigraphy. Structural and stratigraphic evidence gained during my fieldwork indicate that the major folds may instead be northwest-verging structures that pre-date the minor folds measured by Thompson. Each of the major folds are described below from the highest structural level in the base of the Illecillewaet slice to the lowest level in the base of the Clachnacudainn terrane. Cross-sections A-D (in back pocket) were constructed across these folds.

Ghost Peak anticline

Thompson (1972) mapped the Drimmie Creek syncline as a northeast-verging, west-closing antiform that is everywhere truncated along its lower limb by the Standfast Creek fault. According to Thompson's cross-sections, the hinge zone of the structure is composed of multiple minor antiforms. The antiform in the lowest structural position is cored by a thin (15-35 m), east-dipping clastic layer that is sandwiched between east-dipping limbs composed of Badshot marble that close westward around it (Plate 4-7 in Thompson; Plates 2-5, 2-6, and 2-7 in this study). The limbs close within Ghost Peak, a klippe above the Standfast Creek fault composed of Badshot marble in the thickened hinge zone of the antiform (Plate 2-8). Thompson placed the axial surface trace of the Drimmie Creek syncline east of Ghost Peak through this band of clastic rock.

Whether this antiform is interpreted as part of an east-verging syncline (Drimmie Creek syncline of Thompson, 1972) or a west-verging anticline depends solely on the correlation of the clastic core. Thompson mapped the antiform as a syncline because he correlated the core with the Index Formation, which regionally overlies Badshot marble. I, however, correlate the core with the Hamill

Group, which regionally underlies Badshot marble. My correlation requires the antiform to be a west-verging anticline, which I name the Ghost Peak anticline. I retain the name Drimmie Creek syncline for the structure that directly overlies the Ghost Peak anticline because the Index Formation in the core also forms the main core of Thompson's Drimmie Creek syncline. The basis for my correlation, which is given in detail in Chapter 2 on stratigraphy, is the predominance of medium- to thick-bedded (generally more than 1 m) quartzite and micaceous quartzite (Plates 2-6 and 2-7) that bear strong resemblance to Hamill Group in the core of the Akolkolex anticline (Plate 2-1). Quartzite exposed in the cliffs east of Ghost Peak is also readily correlated with the Hamill Group and is interpreted to be the core of the Ghost Peak anticline. Cross-section A (in back pocket), which is drawn through Ghost Peak, is based on this interpretation.

I interpret the axial surface trace of the Ghost Peak anticline as continuing northeast of Ghost Peak along the Illecillewaet-Akolkolex divide into the Albert Peak area. My interpretation is based on stratigraphic evidence which suggests that numerous layers of marble and siliceous clastics (between three and seven layers each) in this interval are isoclinal fold limbs (Plates 2-2 and 2-3). Read and Thompson (1980) correlated the Badshot Formation with the marble and the Hamill Group with the siliceous clastics. My interpretation differs slightly because I correlate the Hamill Group with the lighter coloured siliceous clastics and the Index Formation with the darker, graphitic siliceous clastics (see the Chapter 2 for the details). Thompson (1972) suggested that at the head of Standfast Creek an appressed west-verging syncline is cored by the rocks which I correlate with Index Formation and a west-verging anticline is cored by the rocks which I correlate with the Hamill Group. These cores and limbs composed of Badshot marble are discernible in Plates 2-2 and 2-3. However, Thompson noted that the remaining

clastic layers do not exhibit the consistent alternation that is expected to result from folding, possibly because of the high variability in clastic lithology both along and across strike. Another possible explanation for the lack of consistent alternation is that shearing along the limbs of the isoclinal folds attenuated the section. Sears (1979) correlated the Hamill Group with the quartzite-dominated unit directly below the Standfast Creek fault and the Horsethief Creek Group with underlying marble and siliceous clastic layers. Sears' alternative hypothesis, that the sequence is largely inverted, isoclinally folded repetitions of Hamill, Badshot, and Lardeau strata, is similar to my hypothesis but was considered unlikely by Sears because the symmetrical repetition of units and major fold hinges were not detected.

Structure contour lines drawn across the layers in the limbs of the Ghost Peak anticline indicate that they dip shallowly (about 16°) to the southeast in the headwaters of Greeley Creek and dip moderately ($30-40^{\circ}$) south-southeast in the headwaters of Standfast Creek. The orientation of these layers is similar to the orientation of the limbs of the overlying Akolkolex anticline (see below for description of this fold). Cross-sections B and C (in back pocket), which are drawn through the portion of the Ghost Peak anticline shown in Plates 2-2 and 2-3, are based on these limb orientations.

Sears (1979) mapped a distinctive quartzite-dominated unit, which he correlated with the Hamill Group, as well as marble layers from Standfast Creek northeastward for over 10 km, through the heavily intruded Albert Peak area, to South Albert Creek. These lithologies abut the Standfast Creek fault at South Albert Creek (Sears, 1979). Hamill quartzite and Badshot marble in the core of the west-verging Albert Canyon anticline (Wheeler, 1963; Zwanzig, 1973) crop out in the immediate hanging wall and are in apparent continuity with footwall

stratigraphy mapped by Sears. I therefore suggest that stratigraphy in the upper part of Clachnacudainn terrane crosses the Standfast Creek fault into the base of the Illecillewaet slice with little or no disruption. My hypothesis is corroborated by Gilman's (1972) mapping of the unit underlying the Hamill quartzite (Sears, 1979) as being continuous across South Albert Creek around a fold closure. My hypothesis can only be disproved by mapping which reveals that either a shear zone in South Albert Creek disrupts the apparent continuity or Sears' correlation of the Hamill Group with the quartzite in the footwall is erroneous. The corollary of my hypothesis is that the Ghost Peak anticline may continue across the Standfast Creek fault as the Albert Canyon anticline, thereby forming structure with an amplitude of over 25 km. Before such a correlation is considered further, one must confirm the continuity of the Hamill quartzite through the heavily intruded Albert Peak area and across South Albert Creek.

The Ghost Peak anticline is widest at the head of East Twin Creek and thins southwestward to the Akolkolex River valley, where, according to Thompson (1972), it terminates either in a closure or is truncated by the underlying Standfast Creek fault (uncertainty is due to the lack of exposure in the valley).

Drimmie Creek syncline

I map the Drimmie Creek syncline as being cored by the thick section of Index Formation that overlies Badshot marble in the upper limb of the Ghost Peak anticline and underlies Badshot marble in the lower limb of the overlying Akolkolex anticline. This section of Index Formation composes the cores of the structurally highest antiforms in Thompson's (1972) Drimmie Creek syncline. In my interpretation shown in cross-sections A-C, the Drimmie Creek syncline is a west-verging synform. Like the Ghost Peak anticline, this structure is a few

hundred metres wide at the head of Drimmie Creek and is attenuated southward to where it is truncated by the Standfast Creek fault on the south side of the Akolkolex River valley (Thompson, 1972).

Akolkolex anticline

Hamill Group in the core of the Akolkolex anticline (Thompson, 1972) overlies Badshot marble in the upper limb of the Drimmie Creek syncline. Thompson interpreted the Akolkolex anticline as a synformal northeast-verging structure based on a down-plunge projection of its apparent closure in the southeast-trending cliff east of lower Standfast Creek. Thompson assumed that the Akolkolex anticline closed in this cliff because thin layers of Badshot Formation in the limbs "converge northeastward [in the cliff] as a thickened hinge zone of limestone" (Thompson, 1972, p. 85, Plate 4-5). It is important to note that Thompson mapped the closure by inferring that thickened Badshot marble in the cliff represents a hinge zone; he did not follow the marble around the closure on foot, but rather viewed it from a distance across Standfast Creek valley and traversed the slope north of the thickened marble. Thompson used southeast-plunging minor fold hinge lines measured north of the cliff to project the apparent closure down to the southeast. This projection requires that the anticline is northeast-verging in this location. Since minor fold hinge lines are consistently southeast-plunging over the entire anticline, Thompson inferred that the structure is northeast-verging throughout. Thompson (Plate 4-6) asserted that the hinge zone of the Akolkolex anticline is also exposed in a peak at the head of East Twin Creek. This assertion was made because the peak contains appressed folds of Hamill Group and Badshot marble. Since this exposure is up plunge of the minor folds (to the northwest) from the apparent closure along lower Standfast

Creek, it was used by Thompson to confirm that the hinge line is southeast-plunging.

In order for Thompson's (1972) down-plunge projection to be valid as evidence supporting the northeast vergence of the Akolkolex anticline over its entire length, the minor folds, from whose orientation the projection was made, must have formed during and been related to folding of the larger structure. I contend that Thompson's assumption of cogenational folding may be incorrect because evidence suggests that the minor folds are unrelated to and superposed upon the larger scale Akolkolex anticline. This evidence is presented below. If the folds are not cogenational, then the hinge line of the anticline may not be southeast-plunging, and the structure is therefore not necessarily northeast-verging throughout.

Thompson (1972) probably contended that the minor folds and Akolkolex anticline were the oldest structures and cogenational because the axial planar schistosity of the minor folds was the oldest fabric observed and no interference patterns were observed. The first evidence gained during my fieldwork that contests Thompson's assumption is the presence of an older schistosity (S1) which is deformed by the minor folds. S1 is observed in the hinge zones of minor folds and as porphyroblast inclusion trails. The axial planar schistosity of the minor folds (S2) crosscuts S1 in F2 fold hinge zones and is transposed into parallelism with S1 in F2 fold limbs. The presence of S1 requires that the minor folds are not related to the older S1-forming structures. For this reason, I associate S1 with F1 folds, whereas folds that deform S1, such as minor folds measured by Thompson, are interpreted as F2 folds.

The second evidence which suggests that F2 minor folds are superposed upon the large-scale Akolkolex anticline is based on a comparison of fold

orientation data. If the minor folds and Akolkolex anticline are cogenational, then they should have similar orientations. However, the data show that the orientation of the minor folds differs from the orientation of the Akolkolex anticline, thereby implying that they are not cogenational. In Thompson's (1972) work and this study, minor fold hinge lines are shown to be shallowly to moderately ($10-25^\circ$) southeast-plunging over the entire structure (Fig. 5-3, domains 2 and 4). The orientations of the axial surfaces of the minor folds outline three subdomains: i) in the cliff along lower Standfast Creek within which the fold apparently closes, Thompson documented shallowly southeast- to west-dipping axial surfaces (plotted within domain 4 in Figure 5-2); ii) in the well exposed cirque at the head of the southernmost tributary of Standfast Creek, Thompson and I both documented shallowly to moderately ($15-50^\circ$) northeast-dipping axial surfaces (plotted within domain 4); and iii) on the slope south of the Akolkolex River, Thompson documented mostly northeast- to south-dipping axial surfaces (plotted within domain 2). Since the Akolkolex anticline is tight to isoclinal (Thompson, 1972), then the orientation of its axial surface is approximated by the orientation of its upper and lower limbs. Structure contours drawn along its limbs indicate that they consistently dip shallowly ($12-20^\circ$) southeast throughout the structure. The axial surface of the Akolkolex anticline is therefore not subparallel to the minor fold axial surfaces (S2) within each domain; the orientation of the Akolkolex anticline axial surface differs significantly from the moderately northeast-dipping minor fold axial surfaces in subdomain ii, and minor fold axial surfaces in subdomains i and iii vary considerably, whereas the limbs of the anticline are consistently southeast-dipping. These data suggest that F2 minor folds with generally northeast- to southeast-dipping axial surfaces are superposed upon the southeast-dipping limbs of the F1 Akolkolex anticline. The

F2 minor folds should therefore only change vergence across the axial surface of the Akolkolex anticline in the case where the axial surfaces of the Akolkolex anticline and F2 folds coincide. These data also indicate that Thompson projected the closure of the Akolkolex anticline down the dip of its limbs. This geometry requires that the anticline is reclined, which conflicts with the observation that F2 minor folds are not reclined. Since there is no evidence suggesting that Thompson projected the Akolkolex anticline down its hinge line, I consider the projection to be invalid.

Additional evidence noted by Thompson (1972) for a southeast-plunging hinge line of the Akolkolex anticline is that the apparent closure in the cliff along lower Standfast Creek is southeast of the apparent closure in the peak at the head of East Twin Creek. As noted above, Thompson, without actually having followed the marble around the closure on the ground, inferred that thickened Badshot marble in the cliff along lower Standfast Creek represents a closure; he mapped the closure by observing it from across Standfast Creek valley and traversing the timbered slope north of the apparent closure. My observations and photographs of the cliff from a helicopter and across Standfast Creek valley indicate that layers of Hamill Group are continuous through thickened Badshot marble; tan Hamill Group layers are easily differentiated from grey Badshot marble layers. My inference requires that, although Hamill Group in the anticline core is severely appressed and Badshot marble is substantially thickened in this cliff, the structure does not close within it. The closure instead may occur north of this cliff, implying that the hinge line is east-plunging rather than southeast-plunging.

Northeast vergence of the Akolkolex anticline is required if Z-folds (viewed southeastward) in the upper portion of the peak at the head of West Twin Creek occur in the lower limb of the anticline and if the hinge line is

southeast-plunging. Aside from the evidence presented above which suggests that the Akolkolex anticline hinge line is not southeast-plunging, I consider these Z-folds to be equivocal concerning vergence of the larger structure because Thompson did not provide any data which suggest they are in the lower limb of the anticline. If the fold axial surface instead lies within the thick section of Mohican Formation at the base of the peak at the head of West Twin Creek, then the Z-folds are in the upper limb and a southwest vergence is required. Badshot marble is not observed at the base of the peak in the inferred lower limb, but this is a common attribute of the lower limb of the Akolkolex anticline (Thompson, 1972).

Based on this evidence, I suggest that the Akolkolex anticline is an F1 fold that pre-dates the southeast-plunging F2 minor folds. Since the orientation of Akolkolex anticline hinge line must be known in order to accurately ascertain its vergence, further fieldwork is required. However, I hypothesize that the Akolkolex anticline is a northwest-verging antiformal F1 anticline because the limbs of the anticline (S0, which is transposed into S1 and refolded by F2) are consistently southeast-dipping and the oldest rocks occur in the core (Hamill quartzite is enveloped by Badshot marble). If the anticline is northwest-verging and does indeed close in the cliff along lower Standfast Creek and in the peak at the head of East Twin Creek, then the hinge line must rotate in a sheath-fold-like manner (canoe-shaped and doubly-plunging). My hypothesis that the Akolkolex anticline is a northwest-verging F1 sheath fold will remain tentative pending further fieldwork. Cross-sections A to C are based on this hypothesis. These cross-sections were constructed as southeast-striking in order to show the true dip of the limbs of major folds. This orientation is nearly orthogonal to the cross-sections constructed by Thompson (1972) and nearly parallel with the trend of the

minor F2 folds.

Thompson documented that the thickness of the Hamill Group in the core of the Akolkolex anticline is reduced from about 500 m at the head of the southernmost tributary of Standfast Creek to less than 50 m at the southern edge of his map area, on the south slope of the Akolkolex River. This drastic southward reduction in thickness may indicate that the anticline closes in the poorly exposed ground to the south. In addition, the structural distance between the Akolkolex anticline and Standfast Creek fault decreases rapidly between the Akolkolex River and the south slope of the valley, where it directly overlies the fault. More fieldwork in this heavily timbered area is obviously required to determine whether a closure exists and if the anticline is truncated by the fault to the south. If the fold closes here, then it would be evidence supporting my hypothesis of a northwest-verging sheath fold because this geometry reconciles the closure in the south with the closure along lower Standfast Creek. Additional fieldwork could determine whether the hinge line of the sheath fold rotates between the closures. However, minor folds associated with the Akolkolex anticline (F1) may be difficult to differentiate from the well developed F2 minor folds.

The change of vergence of minor folds across the Akolkolex anticline in the well exposed cirque at the head of the southernmost tributary of Standfast Creek was used by Crowley et al. (1992) to suggest that the anticline is southwest-verging in this location. These minor folds were assumed to be cogenational with the anticline because they have the same shape as the larger structure. Southwest vergence is indicated by the vergence of the minor folds; when observed to the southeast (down the shallow plunge of the hinge line) minor folds in the upper limb are Z-shaped, those in the core are M-shaped, and those in the

lower limb are S-shaped. Bedding-cleavage relationships in the upper limb also imply southwest vergence. In light of the data presented above that indicate the minor folds are unrelated to the Akolkolex anticline, the asymmetric minor folds and bedding-cleavage relationships probably outline a west-verging F2 antiform which is superposed on the larger scale F1 Akolkolex anticline.

Mount Cartier syncline

Stratigraphy outlines an overturned syncline in the Clachnacudainn terrane west of Ghost Peak, directly beneath the Ghost Peak anticline (see cross-section A). This syncline is similar to the structure shown by Read and Thompson (1980), with some differences resulting from further mapping by Read (unpublished data, 1982) and this study. Read and Thompson correlated the Jowett Formation with the amphibolite layer mapped by Thompson (1972) on the upper east slope of the Columbia River valley and with a similar-looking amphibolite layer mapped by Read (1980) that occurs approximately 300 m lower down the slope. They interpreted these two amphibolite layers as a single folded layer rather than faulted repetitions or as two separate stratigraphic horizons. Read (unpublished data, 1982) mapped the northeast-dipping lower limb around a fold closure (1.5 km south of Mount Mackenzie) to the upper limb, thereby corroborating Read and Thompson's interpretation of the two amphibolite layers as a single folded unit. The structure must be a syncline because the Broadview Formation is correlated with the core (Read and Thompson, 1980; this study). I name it the Mount Cartier syncline, after the peak southwest of Ghost Peak that dominates the core. Read and Thompson (1980) and Brown and Lane (1988) inferred that the Mount Cartier syncline is west-verging because its limbs are east-dipping. However, its vergence is speculative because the plunge of its hinge line

is unknown and no minor folds are documented. I also interpret the Mount Cartier syncline as west-verging because its upper limb coincides with the lower limb of the west-verging Ghost Peak anticline. Accordingly, these two structures form an overturned anticline-syncline pair that crosses the Standfast Creek fault. If these folds are indeed F1, then D1 structural style must be continuous from the base of the Illecillewaet slice down to the base of the Clachnacudainn terrane. This is further evidence that the Standfast Creek fault is not a boundary between tectonic levels.

Beneath the Mount Cartier syncline, in the lower slopes of the Columbia River valley, alternating amphibolite and schist layers that I correlate with the Jowett and Broadview formations, respectively, are interpreted as isoclinal fold limbs. The geometry and orientation of these folds remain speculative pending further fieldwork in this heavily timbered area.

Discussion of major folds

The vergence of each major fold has profound implications for interpretations of surrounding structures and lithologic correlations. For instance, the correlation of the Index Formation with dark grey graphitic phyllite in the immediate footwall of the Standfast Creek fault in the Ghost Peak area (Read and Thompson (1980); this study) is highly compatible with a west-verging Ghost Peak anticline; Index in the hanging wall of the Standfast Creek fault must have been continuous with Index in the footwall over Badshot marble in the hinge zone of the anticline. Inverted Lardeau Group directly beneath Ghost Peak is therefore in the lower limb of the Ghost Peak anticline. Footwall Index Formation beneath the Ghost Peak anticline requires minimal displacement on the Standfast Creek fault, in agreement with structural and metamorphic evidence.

On the contrary, footwall Index Formation beneath an east-verging Drimmie Creek syncline (Thompson, 1972) requires significant displacement on the Standfast Creek fault. A west-verging Ghost Peak anticline is also compatible with overlying structures (west-verging Akolkolex anticline and Drimmie Creek syncline), underlying structures (west-verging Mount Cartier syncline), and nearby structures (west-verging Albert Canyon anticline) as I interpret them. On a regional scale, westerly vergence of the major folds in my map area is compatible with westerly vergence of the major folds in the Selkirk allochthon to the north, such as the Carnes nappe (Brown and Tippet, 1978; Brown et al., 1983; Brown and Lane, 1988; Brown, 1991). The tectonic model in Chapter 7 discusses the implications of this regional compatibility in more detail.

Ductile shearing along the Standfast Creek fault

No evidence of ductile shearing along the Standfast Creek fault was reported by previous fieldworkers. The only physical description of the fault was given by Sears (1979); a boulder that was supposedly derived from the fault zone contains brecciated schist in a calcareous matrix. The only inference I draw from this description is that the fault has at least a partly brittle displacement history. Price et al. (1985) inferred that the Devonian orthogneiss is an annealed mylonite because of its strong fabric. Whether or not this rock is mylonitic, there is evidence suggesting that its fabric is not related to motion on the Standfast Creek fault (see Chapter 3 on the Clachnacudainn igneous complex for the evidence backing this contention).

The only evidence I found for ductile displacement along the Standfast Creek fault is in the lower limb of the Ghost Peak anticline in the Ghost Peak area; where the fault is not overprinted by a low-angle brittle fault (Plate 5-2),

dark grey graphitic phyllite that is correlated with the Index Formation (Read and Thompson, 1980; this study) is thinly (2-20 cm) interlayered with Badshot Formation marble over a 1 to 3 metre-wide zone (Plate 5-3). I interpret that this zone resulted from ductile shearing and transposition of isoclinal folds during development of the Ghost Peak anticline. My interpretation that this interlayering as a result of ductile shearing rather than depositional interbedding is corroborated by other evidence, such as boudins of Index Formation phyllite that are enclosed within Badshot marble (Plate 5-4). Although the amount of displacement due to this ductile shearing is unconstrained, it is probably minor (metres to tens of metres) because regionally conformable lithologies are juxtaposed across the fault according to correlations by Read and Thompson (1980) and this study (Plates 5-2, 5-3, and 5-4). This shearing is not unique to the lower limb of the Ghost Peak anticline, as local attenuation of the entire Badshot marble occurred along the lower limbs of the Akolkolex (Thompson, 1972) and Albert Canyon (Zwanzig, 1973) anticlines.

Highly strained, dark grey garnet-bearing micaceous quartzite crops out along the Trans-Canada Highway in Albert Canyon near the Standfast Creek fault mapped by Wheeler (1963, 1965). This lithology was correlated with the Lardeau Group by Zwanzig (1973). The fine quartz grain size and fractures in garnet probably resulted from cataclastic deformation. Evidence for ductile flow, such as quartz ribbons, is not observed. No other anomalously high strain zones were observed along the Standfast Creek fault, as fabrics in rocks at the fault are identical to and have the same orientations as fabrics in rocks tens of metres from the fault. This suggests that any ductile high strain zones which resulted from a large magnitude of displacement on the fault were recrystallized or there were no ductile high strain zones.

Third phase of deformation (D3)

The third phase of deformation (D3) is evinced by upright and open folds (F3) that warp S0, S1, and S2 (Plate 5-1). F3 folds penetratively deform middle to upper levels of the Clachnacudainn terrane and base of the Illecillewaet slice. F3 folds are subordinate to F1 and F2 folds in terms of intensity of deformation in the map area, yet they dominate the eastern Illecillewaet slice (Zwanzig, 1973; Sears, 1979). They vary in scale from crenulations to major folds with a few kilometres of amplitude. The vertical to steeply dipping, east-southeast-striking axial planar schistosity (S3, Fig. 5-4) is well developed in graphitic phyllite and locally in schist. F3 fold hinge lines generally plunge shallowly to the east-southeast (Fig. 5-5), which is a slightly (up to 20°) more easterly trend than F2 fold hinge lines in the Akolkolex River area (Thompson, 1972). F3 fold hinge lines are also locally shallowly west-northwest-plunging. I interpret D3 to have post-dated the metamorphic peak because S3 wraps around garnet with pressure shadows that formed during F3 folding (see Chapter 4 on metamorphism).

Thompson (1972) termed the larger folds with F3 characteristics "type 2 folds", and the smaller crenulations, which differed only in scale, were termed "type 3 folds". Zwanzig (1973) noted that northwest-trending, upright folds at structurally high levels in the Illecillewaet slice give way to more highly attenuated, moderately inclined folds at the base of the slice. Sears (1979) mapped the Akolkolex River fault as the boundary between tectonic levels because "younger upright folds [F3] appear to die out rather abruptly downwards across a zone of detachment in the Lardeau phyllites above an older tectonic slide, the Akolkolex River fault" (p. 105). Mapping by Thompson (1972) and this study shows that even though upright folds (F3) are apparently less pervasive in the footwall of the Akolkolex River fault, they penetratively deformed the

footwall and therefore do not "die out rather abruptly" beneath the fault. F3 folds most likely post-dated all motion on the Standfast Creek fault because fabrics in the fault zone are deformed by F3 folds; presumably F3 folds would easily be destroyed upon ductile shearing across the fault.

Twin Creek antiform

The Twin Creek antiform may be a large-scale F3 fold because it is an upright structure that warps S2 and has 3 km of amplitude (Sears, 1979). It has the regional F3 fold orientation and geometry where I observed it on the west ridge of North Albert Peak. The hinge line of the Twin Creek antiform coincides with a major transition in the orientation of S2; to the southwest of the hinge line S2 is moderately south-dipping and to the northeast S2 is moderately to steeply northeast-dipping. Sears noted that the curvilinear nature of the axial surface trace around the Albert stock indicates that the Twin Creek antiform may be only partly related to southeast-plunging, regional F3 folds. The Albert stock lies in a structural depression created by steeply dipping limbs of the Twin Creek antiform which face into it (Sears, 1979). Sears suggested that the antiform may "represent structures that were formed or modified during emplacement of the Albert stock and were subsequently cut by the stock during the final stages of its emplacement" (p. 106). Final emplacement of the Albert stock post-dated F3 folding because it shows no evidence of ductile deformation and porphyroblasts in its contact metamorphic aureole overgrew S3.

Maunder Creek antiform

The large-scale antiform on the ridge north of West Woolsey Creek is interpreted as an F3 fold because it warps S2 and has regional F3 fold orientation

and geometry. I name this fold the Maunder Creek antiform after the creek to the north of its exposure. The Maunder Creek antiform is intruded on its west limb by late-tectonic granodiorite (Plate 3-9, cross-section D in back pocket). Other large-scale F3 folds that gently warp S2 occur in the Ghost Peak area.

Youngest phases of deformation

Brittle faults are the youngest phases of deformation in and above the Clachnacudainn terrane. They are divided into low-angle faults, high-angle normal faults, and strike-slip faults. Orientations of the brittle faults are shown on the stereonet in Figure 5-6.

Low-angle faults

Brittle low-angle faults occur along the Standfast Creek fault mapped by Thompson (1972) beneath Ghost Peak (Plate 5-2), south of Albert Peak in the footwall of the Standfast Creek fault, and north of Albert Peak in the hanging wall of the Standfast Creek fault. Because the latter set of faults are continuous from metasedimentary rocks through dykes extending from the Albert stock, this phase of brittle deformation is most likely younger than mid-Cretaceous. These faults are characterized by 1-10 m thick, shallowly east-dipping (Fig. 5-6), highly fractured, commonly silicified zones with numerous slickenside surfaces. Most slickenside surfaces are randomly oriented, but the most prominent surfaces and the zones themselves generally dip shallowly ($10-25^\circ$) to the east and contain striae that plunge down dip. The sense and amount of displacement on these low-angle faults are unknown because no offset marker layers were observed, yet similar-looking lithologies across the faults suggests that displacement is minor. The brittle fault that partly coincides with the Standfast Creek fault in the Ghost

Peak area has variable lithologies across it (Plate 5-2); directly beneath Ghost Peak there is Badshot marble in the hanging wall and Index phyllite in the footwall, but to the south the fault cuts down into the Index phyllite. The low-angle brittle fault south of Albert Peak is distinct because schistosity in the hanging wall was rotated relative to the footwall.

High-angle normal faults

High-angle normal faults occur in Albert Canyon, along the Illecillewaet-Akolkolex divide, and in numerous localities in the hanging wall and footwall of the Standfast Creek fault. These faults are characterized by 0.2-5 m thick fault zones that contain gouge, highly fractured rock, discrete brittle shears, slickenside surfaces, and rare breccia. The faults generally dip between 50-70° to the east. The down-dip direction of displacement is inferred by slickenside lineations and the extensional sense of displacement is inferred by drag folds and offset markers.

Three thin (less than 1 m), moderate to steeply (50-70°) northeast-dipping brittle faults are exposed along the Trans-Canada Highway in Albert Canyon near the upper Standfast Creek fault of Price (1986). A similar-looking, moderately steeply (55-60°) northeast-dipping brittle fault occurs in the hanging wall of the Standfast Creek fault at the head of the Akolkolex River. A northeast-side-down sense of displacement on these faults is inferred from drag folds and foliation in the gouge zones that is oblique to the zone boundaries. The amount of displacement on each is most likely minor because identical-looking rock types occur in the hanging wall and footwall.

The high-angle, mostly east-dipping normal faults mapped by Thompson (1972) and Sears (1979) along the Illecillewaet-Akolkolex divide compose the Twin Creek fault system. I name the fault system so because at least seven faults

are prominently displayed in cliffs at the headwaters of East Twin and West Twin creeks (Plates 5-5 and 5-6). Offset marker layers indicate that each fault has between 5 and 150 m of extensional displacement. The fault with the greatest amount of displacement is exposed south of the lake at the head of East Twin Creek (Plate 5-6) as a fault breccia with a remobilized marble matrix. The steeply east-dipping normal fault mapped by Sears (1979) on the west ridge of North Albert Peak juxtaposes metasedimentary rocks against Devonian orthogneiss (Plate 5-7). This fault was inferred by Parrish et al. (1988) as lying along the ductile extensional branch of the Standfast Creek fault that skirts the Albert stock. I contend that the fault is part of the Twin Creek system because my mapping corroborates Sears' interpretation of the fault as a high-angle brittle fault (Plates 5-7 and 5-8). It is unlikely that the brittle fault overprinted an east-dipping ductile fault because the foliation in the gneiss and metasedimentary rocks outside the brittle zone is south-dipping and there is no ductile strain gradient away from the fault.

Strike-slip faults

Zones with vertical to steeply dipping, chloritized slickenside surfaces which contain shallowly plunging striations are interpreted as strike-slip faults. The amount and sense of displacement on these faults are unknown because no offset marker layers were observed. However, the fact that identical-looking rock types occur on both sides of each fault suggest that displacement is minor. The orientations of each of the faults discussed below are shown in Figure 5-6.

Brittle strike-slip faults occur within the upper portion of the Devonian orthogneiss in the col south of Mount LaForme, on the west ridge of North Albert Peak (Plate 5-9), and in Albert Canyon. The fault zones in the former two

localities are between 1 and 4 m wide and are marked by anastomosing, vertical discrete brittle shears. Since the vertical shears are parallel to the gneissosity, which dips moderately to the northeast outside of the fault zone, then the fabric in the fault zone must have rotated into the shear plane during faulting. In the col south of Mount LaForme, it is uncertain whether the strike-slip fault is the contact between the orthogneiss and metasedimentary rocks or whether it occurs entirely within the orthogneiss because the contact is not exposed. The fault on the west ridge of North Albert Peak is separated from metasedimentary rocks by 15-20 m of unfaulted gneiss (Plate 5-9). The fault in Albert Canyon is about 15-20 m wide, lies at least 500 m from the metasedimentary rocks, and is marked by highly fractured, hydrothermally altered gneiss with pseudotachylyte (Price et al., 1985) along the fractures.

In all localities the strike of the fault zone parallels the nearby orthogneiss-metasediment contact, suggesting that the faults accommodated motion between the two units. The strike-slip faulting in the upper portion of the orthogneiss may be related to the insignificant strike-slip displacement documented by Lane (1984) in the Columbia River valley. He suggested that strike-slip motion occurred after dip-slip motion in a local zone of reduced shear strength.

A 100 m wide zone of highly fractured marble occurs at the upper Standfast Creek fault on the ridge north of West Woolsey Creek (Plate 3-9). The most prominent of three fracture sets strikes north-northeast and is vertical. Shallowly plunging slickenside striae in vertical slickenside surfaces that form the west boundary of the fracture zone indicate strike-slip motion. The amount of displacement is probably minor because similar-looking metasedimentary rocks that are correlated with the Index Formation occurs on both sides of the fault.

Thin (less than 1 m wide) strike-slip zones in quartz diorite at the head of

Siegfried Creek post-dated intrusion of the Albert stock. I therefore suggest that all brittle strike-slip faults are younger than mid-Cretaceous.

Deformation in relation to the Standfast Creek fault

Wheeler (1965) first noted that the intensity of deformation increases toward a structural culmination in the Clachnacudainn terrane. Thompson (1972) contended that the Standfast Creek fault was the boundary between "two distinct structural levels" which have major differences in "structural fabric". However, he mapped the same types of folds in the hanging wall and footwall and thereby noted that the two levels "appear to have the same structural style". He further contradicted his interpretation of "distinct structural levels" by placing structural domains across the Standfast Creek fault. Zwanzig (1973) noted that folds become more highly attenuated, inclined, and nappe-like and faults less important as the Clachnacudainn terrane is approached from the east. He positioned the boundary between these structural levels east of the Standfast Creek fault; it was in the Illecillewaet synclinorium "for an extended period of time during deformation but finally the boundary migrated toward the west [to the Albert Canyon anticline]" (p. 79). Sears (1979) positioned the boundary between structural levels at the Akolkolex River fault because a greater contrast in structural style occurs across it compared to the Standfast Creek fault, which has upright (F3) and inclined folds (F2) in its hanging wall and footwall.

I therefore agree with previous fieldworkers' observations that overturned, tight folds (F1 and F2) and upright, open folds (F3) deformed the entire Clachnacudainn terrane and base of the overlying Illecillewaet slice. This leads to the conclusion that the Standfast Creek fault does not separate distinct tectonic levels.

Figure 5-1. Location of eight structural domains. Fabric and fold data from domains are plotted on stereonets in Figures 5-2 to 5-5. Akolkolex anticline (AA), Albert Canyon (AC), Albert Canyon anticline (ACA), Albert Peak (AP), Columbia River fault (CRF), Drimmie Creek syncline (DCS), East Twin Creek (ETC), Ghost Peak (GP), Ghost Peak anticline (GPA), lower Standfast Creek fault (LSCF), Mount Cartier syncline (MCS), Mount Klotz (MK), Mount LaForme (MLF), Mount Llewelyn (ML), Mount Revelstoke (MR), North Albert Peak (NAP), Sale Creek (SC), Twin Creek fault system (TCFS), upper Standfast Creek fault (USCF), West Twin Creek (WTC), West Woolsey Creek (WWC).

Figures 5-2 to 5-5. Stereonets (equal area) representing fabric and fold data from the eight domains located on the map in Figure 5-1. Structural data are from Thompson (1972), Zwanzig (1973), Sears (1979), and this study.

Figure 5-4. S3
Equal Area
C.I. = 2.0%/1% area

Figure 5-6. Brittle faults
Equal Area Stereonets

Plate 5-1. Overturned tight F2 refolded by upright open F3 in Badshot marble, upper limb of Ghost Peak anticline, east of Ghost Peak. F2 axial surfaces in pelitic schist form the dominant S2.

Plate 5-2. Low-angle brittle fault within the Index Formation (In) at the Standfast Creek fault beneath Ghost Peak. The Index Formation is highly fractured, silicified, and cut by randomly oriented slickenside surfaces. Grey Badshot marble (Bd) lies in the hanging wall of the Standfast Creek fault and the low-angle brittle fault. The low-angle brittle fault cuts downward into the Index Formation to the south, where the Standfast Creek fault appears as it does in Plates 5-3 and 5-4.

Plate 5-3. Evidence for ductile shearing on the Standfast Creek fault is a 1-3 m wide zone of interlayered grey Badshot marble (Bd) and dark grey graphitic Index Formation phyllite (In), southeast of Ghost Peak. Since this shearing occurs in the lower limb of the Ghost Peak anticline, I interpret it as a result of transposition of isoclinal folds during development of the Ghost Peak anticline. The amount of displacement is probably minor since the fault juxtaposes units that are in conformable contact in the hanging wall.

Plate 5-4. Additional evidence of ductile shearing on the Standfast Creek fault is a boudin of dark grey graphitic Index Formation phyllite (In) enclosed within Badshot marble (Bd), southeast of Ghost Peak.

Plate 5-5. Southward view of a high-angle east-directed brittle normal fault in the Twin Creek fault system (my terminology, mapped by Sears, 1979) that cuts through lithologies that Read and Thompson (1980) and this study correlate with Badshot Formation (Bd) and Hamill Group (Ha) in the upper level of the Clachnacudainn terrane, palisades south of the lake at the head of East Twin Creek. Arrows show corresponding layers across the fault. Plate 5-6 shows this fault at a larger scale.

Plate 5-6. Southward view of high-angle east-directed brittle normal faults in the Twin Creek fault system that cut the section above the lake (lk) at the head of East Twin Creek. The fault to the east has the greatest known displacement (a few hundred metres) of any fault that I observed in the Twin Creek fault system. Arrows show corresponding layers across the fault. The Twin Creek fault system also consists of west-directed normal faults. Amphibolite dykes (am) crosscut stratigraphy and are cut by the normal faults. Possible Index Formation (In?) underlies the layers of Badshot marble (Bd) and Hamill Group (Ha).

Plate 5-7. Southwestward view of the high-angle east-directed brittle normal fault that juxtaposes semipelitic schist in the hanging wall that I correlate with the Broadview Formation (Bw) against Devonian orthogneiss (Dogn) in the footwall, west ridge of North Albert Peak. Backpacks on top of the ridge (beneath arrow) are the scale. Parrish et al. (1988) inferred that this boundary is a ductile Eocene normal fault branch of the Standfast Creek fault. Since the highly fractured appearance of the fault zone in Plate 5-8 indicates that latest displacement on the fault is brittle and there is no evidence of older ductile motion, then I interpret the fault as part of the Twin Creek fault system (see Plates 5-5 and 5-6 for other examples).

Plate 5-8. Brittle fractures in Devonian orthogneiss within the high-angle east-directed normal fault shown in Plate 5-7, west ridge of North Albert Peak.

Plate 5-9. Northward view of vertical brittle shears within a strike-slip fault through Devonian orthogneiss that is exposed from just left of the lens cap to the ice ax in the background, west ridge of North Albert Peak, nearby brittle normal fault in Plates 5-7 and 5-8. Sense and amount of displacement are unknown. See text for descriptions of similar brittle strike-slip faults through Devonian orthogneiss in the Clachnacudainn terrane.

CHAPTER 6. Standfast Creek fault

Introduction

The conclusion that the Standfast Creek fault is a minor tectonic feature is the basis for my tectonic model for the Clachnacudainn terrane and Standfast Creek fault. This model is outlined in the Chapter 7. This chapter presents the evidence for this conclusion by synthesizing my observations with those made by previous fieldworkers (Thompson, 1972; Zwanzig, 1973; Sears, 1979). This chapter reveals that although this conclusion is highly incompatible with previous fieldworkers' interpretations and recent tectonic models proposed by Read and Brown (1981), Price (1986), and Parrish et al. (1988), it is compatible with many of the fieldworkers' observations.

The conclusion that the Standfast Creek fault is a minor tectonic feature is based mostly on two premises. The first is that faults mapped by individual fieldworkers in their map areas should not have been correlated and joined together as one composite fault because the sense, amount, and timing of displacement on each was poorly known or thought to differ. This premise, which was mostly reached by reviewing previous fieldworkers' manuscripts, is outlined in the first section of the chapter. The second premise, based mostly on my fieldwork, is that the Clachnacudainn terrane is an integral part of the Selkirk allochthon in stratigraphic, structural, and metamorphic aspects; the terrane therefore could not have been displaced greatly from the overlying Illecillewaet slice. The final section of the chapter compares my conclusions with previous

fieldworkers' observations and interpretations.

First premise - the Standfast Creek fault is not a composite fault

Sears (1979) was the first to propose a composite Standfast Creek fault by linking together the fault that he mapped in the Albert Peak area with faults mapped i) by Wheeler (1963, 1965), Zwanzig (1973), and Hakkinen (unpublished data, 1976, 1977) in the lower limb of the Albert Canyon anticline, ii) by Thompson (1972) in the metasedimentary rocks south of the Illecillewaet River. Sears correlated and joined these fault segments into the composite Standfast Creek fault with minimal supporting evidence; the only reason given by Sears is that each segment apparently attenuates stratigraphy and the metamorphic gradient. The sense, direction, amount, timing, and style of displacement on each segment were therefore not considered. The following review shows that these important aspects were thought to differ or were poorly constrained by previous fieldworkers. I assert that the lack of evidence supporting Sears' hypothesis of a composite Standfast Creek fault makes it highly untenable.

Wheeler (1963) noted that the lower limb of the Albert Canyon anticline is sheared off and suggested that it "may lie along a fault that separates northwest-trending Selkirk structures from the divergent structures in and around the [Clachnacudainn salient]" (p. 26). Other than the apparent juxtaposition of structural levels, the only physical evidence given by Wheeler (1963) for the fault is tightly drag-folded Hamill quartzite in the core of the anticline. Wheeler (1965) contended that "highly sheared rocks and a fault zone near Albert Canyon" (p. 20) suggest that in Albert Canyon there is a zone similar to if not the same as the mylonitic fault zone that underlies much of the Columbia River valley north of LaForme Creek. This correlation was made on the basis that both fault zones

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apparently juxtapose high grade metasedimentary and igneous rocks to the west against low grade rocks to the east; this relationship requires that the fault zones are post-metamorphic.

Thompson (1972) mapped the Standfast Creek fault as a shallowly southwest-dipping, southwest-directed normal fault through metasedimentary rocks south of the Illecillewaet River. Thompson interpreted the fault as southwest-dipping because "truncation of the footwall sequence occurs both to the west and south" (p. 83). Between the two localities where truncation is evident, the head of Standfast Creek and south of the Akolkolex River, the fault was extrapolated along the lower limb of the Drimmie Creek syncline. A moderate amount of truncation on the Standfast Creek fault is evident south of the Akolkolex River because Thompson's map shows that i) the Jowett Formation (Read and Thompson, 1980; this study) is juxtaposed against Badshot marble across the fault, as the Index Formation (correlated by Read and Thompson (1980); this study) in the footwall beneath Ghost Peak is entirely omitted at this locality; ii) the structural distance between the Akolkolex anticline and the Standfast Creek fault is drastically decreased from a few hundred metres at the Akolkolex River to directly overlying the fault in this locality; and iii) the Ghost Peak anticline either closes or is truncated by the fault at the Akolkolex River. It is important to note that the locality of apparent truncation in the heavily timbered slope south of the Akolkolex River is relatively poor compared to the excellent alpine exposures to the north. Assuming that the relationships mapped by Thompson are correct and that all of the apparent attenuation in this region is attributable to the Standfast Creek fault, then there is a maximum of about 750 m of additional attenuation of section on the segment south of the Akolkolex River compared to the segment in the Ghost Peak area. At the head of Standfast Creek, the fault cuts down to the

west into marble and siliceous clastic layers in the footwall. The fault in this interval juxtaposes the Index Formation against a quartzite-dominated unit that is correlated with the Hamill Group (Sears, 1979; Read and Thompson, 1980; this study).

Thompson inferred a southwesterly extensional sense of displacement on the Standfast Creek fault largely because the hanging wall was interpreted as an upper tectonic level, the footwall as a lower level, and the fault cuts down section to the southwest. It is important to note that this sense of displacement is not based on kinematic indicators or offset marker layers because Thompson stated that "no obvious mesoscopic features suggestive of differential translation were observed along the fault trace" (p. 81). The Hamill Group, Badshot Formation, and Lardeau Group were correlated with metasedimentary rocks in the upper tectonic level. Thompson suggested that the lower tectonic level could contain part of the Hamill Group and possibly part of the Upper Proterozoic Horsethief Creek Group, but concluded that the lack of marker units made correlation largely indeterminable. Although lower level metasedimentary rocks were not firmly correlated and structural style was documented to be consistent across the Standfast Creek fault, Thompson still asserted that it separated rocks from "different areas and tectonic levels" (p. 85). It is not known whether the fault was ductile or brittle because none of its physical attributes were described. Other than contending that the fault juxtaposes different tectonic levels, Thompson could not estimate the amount of displacement because the metasedimentary rocks in the footwall were not correlated. The apparent omission of metamorphic isograds across the fault allowed him to make the only estimate of its timing; it post-dated peak metamorphism.

Zwanzig (1973) correlated the Hamill Group, Badshot Formation, and

Lardeau Group with metasedimentary rocks in the hanging wall and footwall of the fault that was first noted by Wheeler (1963, 1965) in the lower limb of the Albert Canyon anticline. Zwanzig termed this an "unsubstantiated fault" (p. 84) because sheared quartzite in one creek bed and the apparent truncation of stratigraphy with few marker units were the only evidence. He instead suggested that the lower limb of the Albert Canyon anticline is a zone of "intense ductile deformation" (p. 130) which contains a series of unfaulted, appressed isoclinal folds. Zwanzig documented that all early linear structures and fold axes in this zone trend northeast. This observation allowed for his conclusion that i) "the maximum extension was northeast parallel with the direction of displacement" (p. 143) and ii) the Albert Canyon anticline formed because the "gneissic rocks at Albert Canyon were emplaced laterally [and] carried the miogeocline succession northeast as they flowed underneath it" (p. 142). Hakkinen (unpublished data, 1976, 1977) mapped the fault northwestward from Albert Canyon into the eastern Clachnacudainn Range between rocks that he correlated with the Hamill Group (footwall) and Lade Peak Formation.

Sears (1979) interpreted the Standfast Creek fault in the Albert Peak area as having two episodes of motion. An early history of east-directed extension was inferred because the fault is "characterized by an omission of stratigraphic units" (p. 108). Sears considered flattening due to vertical compression to be responsible for this extension as well as the flat-lying schistosity and highly attenuated recumbent isoclines, all of which were synchronous with regional metamorphism. A younger episode of west-directed compression on the Standfast Creek fault was inferred by Sears because "structural relationships across the [fault] zone indicate that some contraction faulting occurred" (p. 108). The Albert Canyon anticline provided the "structural relationships" that indicate young

thrusting; since the anticline is west-verging and the lower limb is truncated along the fault mapped by Wheeler (1963, 1965), Zwanzig (1973), and Hakkinen (unpublished data, 1976, 1977), then Sears assumed "it was formed by a relative displacement of the rocks of the lower tectonic level [Clachnacudainn terrane] toward the northeast beneath the anticline" (p. 109). Upright, open folds of foliation with southeast-trending hinge lines (F3 folds in my terminology) also provided Sears with evidence that thrusting resulted from southwest-northeast horizontal shortening. This model is identical to that proposed by Zwanzig (1973) for the Albert Canyon area. Like Zwanzig, Sears did not provide any kinematic evidence from the gneiss or overlying metasedimentary rocks which indicates that the terrane flowed northeast under the Standfast Creek fault. They also did not identify a fault in Albert Canyon on which the displacement occurred because the fault that Wheeler (1963, 1965) mapped was considered "unsubstantiated" by Zwanzig (1973, p. 84).

The Standfast Creek fault mapped by Sears (1979) in the Albert Peak area juxtaposes lithologies that he correlated with the Hamill Group (footwall) and Lade Peak Formation. According to his assignment of the Lade Peak Formation with the base of the Lardeau Group, the Standfast Creek fault must have omitted the Badshot marble; there could be as little as a few metres of throw on this fault because Badshot marble thins to 3 m along the limbs of the Albert Canyon anticline (Zwanzig, 1973) and is discontinuous along the limbs of the Akolkolex anticline (Thompson, 1972). Minor displacement is also required by Sears' and Gilman's (1972) mapping which show that lithologies are continuous across the fault mapped by Sears in South Albert Creek valley. Sears estimated that the fault attenuated 2-3 km of stratigraphy by correlating lithologies juxtaposed across the fault outside of his map area; since he correlated footwall metasedimentary

rocks in Thompson's map area with the Horsethief Creek Group, then most of the Hamill Group and upper part of the Horsethief Creek Group must have been omitted because Badshot marble lies in the immediate hanging wall. Sears suggested that reactivation of the younger thrust fault on the older normal fault surface explains why the thrust coincides with an omission of stratigraphy. All ductile displacement on the Standfast Creek fault must have pre-dated intrusion of the Albert stock because it shows no evidence of ductile deformation where it crosscuts the fault and porphyroblasts in its contact aureole overgrew the youngest ductile fabric (Sears, 1979). Sears' observations about the nature of the fault are conflicting as to whether it is brittle or ductile. It can be gleaned from Sears' only description of the fault's physical attributes that it had a partly brittle history. The inference by Sears that the fault was active during formation of the dominant schistosity suggests that he thought it also has a ductile history.

Sears (1979) linked the Standfast Creek fault in the Albert Peak area together with the faults mapped by Wheeler (1963, 1965), Zwanzig (1973), and Hakkinen (unpublished data, 1976, 1977) to the north and Thompson (1972) to the south. Sears therefore broadened Wheeler's (1963, 1965) definition of the Clachnacudainn salient to also include the lower grade metasedimentary rocks mapped by Thompson south of the Illecillewaet River. Sears joined these fault segments into the composite Standfast Creek fault because each is apparently marked by attenuation of stratigraphy and the metamorphic gradient. In doing so, he disregarded these pertinent points: i) Thompson concluded that the normal fault in his map area is southwest-directed instead of east-directed, ii) attenuation of stratigraphy on the fault mapped by Thompson is mostly conjecture because footwall metasedimentary rocks were not correlated, iii) regional metamorphic isograds cross the fault in the Albert Peak and Albert Canyon areas (Zwanzig,

1973; Sears, 1979), iv) only minor omission of stratigraphy on the "unsubstantiated fault" in Albert Canyon can be surmised from tentative correlations by Zwanzig, and v) only minor omission of stratigraphy can be surmised from correlations by Sears and Hakkinen in their map areas. I consider Sears' theory of a composite Standfast Creek fault that is a major tectonic feature to be invalid because of these inconsistencies.

Price (1986) modified Sears' Standfast Creek fault by bifurcating it north of the Albert stock into two subparallel branches. The eastern branch (which I name the upper Standfast Creek fault) is the fault mapped by Wheeler (1963, 1965), Zwanzig (1973), and Hakkinen (unpublished data, 1976, 1977). The western branch (the lower Standfast Creek fault) is the contact mapped by Ross (1968) and Hakkinen (unpublished data, 1976, 1977) between Devonian orthogneiss (thought to be Precambrian basement by Ross) and lower Paleozoic metasedimentary rocks. Ross suggested that a slice of basement was wedged upward into cover metasedimentary rocks along this contact during formation of a large east-verging, basement-cored nappe. The west-dipping, mylonitic fault that formed at the base of the wedge supposedly resulted from a ductility contrast between basement and cover. Although not stated explicitly by Ross, he appears to be claiming that the fault is a ductile east-directed thrust. Although not revealed by Price (1986), he most likely inferred the lower Standfast Creek fault because orthogneiss in Albert Canyon is highly fractured, is cut by pseudotachylyte, and is mylonitic (Price et al., 1985). I contend that the lack of evidence supporting the hypothesis that these features are related to motion on the Standfast Creek fault (see Chapter 3) makes it highly untenable.

Second premise - Clachnacudainn terrane is an integral part of the

Selkirk allochthon

The stratigraphic, structural, and metamorphic continuity that I mapped across the Standfast Creek fault indicate that the Clachnacudainn terrane is an integral part of the Selkirk allochthon. This continuity requires that the Standfast Creek fault is a minor tectonic feature.

Recognition of the Hamill Group, Badshot Formation, and Lardeau Group in the Clachnacudainn terrane (Read and Thompson, 1980; this study) suggests that it is not a "suspect terrane", as was proposed by Price (1986), because these lithologies also occur in the immediate hanging wall of the Standfast Creek fault (Thompson, 1972; Zwanzig, 1973; Sears, 1979). Since the Badshot and Index formations were mapped by all previous workers as conformable, then the occurrence of Index in the immediate footwall of the Standfast Creek fault (Read and Thompson, 1980; this study) and Badshot in the hanging wall in the Ghost Peak area is inconsistent with Sears' estimate of 2-3 km of attenuation. Attenuation south of the Akolkolex River is apparently greater than at Ghost Peak because the entire Index Formation is omitted and the Akolkolex anticline directly overlies the fault (Ghost Peak anticline and Drimmie Creek syncline are omitted). Along the Illecillewaet-Akolkolex divide and in the Albert Peak area, the fault entirely omits Badshot marble in the upper limb of the Ghost Peak anticline (Badshot marble is also locally absent along the limbs of the Akolkolex and Albert Canyon anticlines (Thompson, 1972; Zwanzig, 1973)). Therefore, I suggest that displacement on the segment of the Standfast Creek fault north of the Akolkolex River is attributable to fold limb attenuation on the scale of tens to a few hundred metres and that attenuation on the segment south of the Akolkolex River is considerably greater, perhaps up to 1 km. The greater amount of attenuation on the segment south of the Akolkolex River could be due to younger

southwest-directed extension, as suggested by Thompson. However, if this overprinted extension is younger than mid-Cretaceous, then it must be localized on this segment and not continue northeastward toward the Albert stock for the reasons given above.

Any motion along the Standfast Creek faults mapped by Sears (1979) and inferred by Parrish et al. (1988) pre-dated intrusion of the 104 ± 1 Ma Albert stock (U-Pb zircon age). The amount of displacement across the upper Standfast Creek fault north of the Trans-Canada Highway is constrained by identical lithologies, which I correlate with the Index Formation, on both sides of the fault. Any motion along the lower Standfast Creek fault must have pre-dated intrusion of granodiorite in the Clachnacudainn Range. This granodiorite is most likely coeval with the Albert stock.

The documentation of the same three phases of deformation on both sides of the Standfast Creek fault suggest that it is not a boundary between structural levels. Instead, I interpret the Clachnacudainn terrane to have the same deformational history as the base of the overlying Illecillewaet slice; two phases of tight to isoclinal, overturned folding (the first of which are large-scale and west-verging) are followed by a final phase of open, upright folding. This contention alone does not prove that the Standfast Creek fault is a minor tectonic feature, but it is consistent with stratigraphic correlations which suggest that the terrane is not the footwall of any major faults. The evidence put forth in Chapter 3 suggest that the fabric in the Devonian orthogneiss is not related to displacement between the orthogneiss and overlying metasedimentary rocks. Instead, I interpret this contact as a transposed intrusive contact and the gneissic fabric as the result of regional deformation. The ductile displacement along the Standfast Creek fault beneath Ghost Peak is attributed to shearing along the lower limb of the Ghost

Peak anticline.

Mapping of regional metamorphic isograds across the Standfast Creek fault requires that it is not associated with a steep or attenuated metamorphic gradient and suggests that the Clachnacudainn terrane has the same metamorphic history as the base of the Illecillewaet slice (see Chapter 4 on metamorphism). I contend that high grade (sillimanite-andalusite $\pm$ muscovite $\pm$ K-feldspar) assemblages in the terrane grew during contact metamorphism related to intrusion of post-tectonic granitoids in the Clachnacudainn igneous complex. Parrish et al. (1988) contended that K-Ar mica cooling dates are evidence for a large temperature difference between the Clachnacudainn terrane and the overlying Illecillewaet slice in the Eocene. They used these mineral cooling data to infer that the terrane was rapidly tectonically denuded in the Eocene by a ductile Standfast Creek fault that skirts the late-tectonic Albert stock. I assert that since all the cooling dates in the terrane were obtained from or near granitoids in the Clachnacudainn igneous complex, they merely signify that young intrusions heated the surrounding granitoids above the mica cooling temperature in the Eocene. A Late Cretaceous leucogranite (dated by the U-Pb zircon method by Parrish, 1992) and cross-cutting leucopegmatite are examples of young intrusives. These mineral cooling data therefore do not indicate whether the entire terrane was at high temperatures in the Eocene and should not be used to contend that the entire terrane was rapidly tectonically denuded in the Eocene. A K-Ar mica cooling date traverse across the Standfast Creek fault beginning in the intrusive-free southern portion of the terrane would reveal whether the entire terrane was at high temperatures in the Eocene. I suggest that the southern portion of the terrane was cool by the mid-Cretaceous because the Albert stock (with its 83 Ma K-Ar biotite cooling date) and its contact aureole crosscut the Standfast Creek

faults mapped by Sears (1979) and inferred by Parrish et al. (1988). Although the lack of a steep regional metamorphic gradient across the Standfast Creek fault and the probability that cooling dates are identical across its southern extent do not disprove its existence (there is the possibility of pre-metamorphic displacement or post-fault reequilibration), they are consistent with other data which suggest the Standfast Creek is a minor tectonic feature.

Comparison with previous fieldworkers' observations and interpretations

My review of previous fieldworkers' manuscripts reveals that although their interpretations differ widely from my conclusion that the Standfast Creek fault is a minor tectonic feature, their observations often support it. For instance, Thompson (1972) interpreted the fault as a boundary between tectonic levels even though he observed structural style to be consistent across it; he even placed his structural domains across the fault. Thompson's observation that "no obvious mesoscopic features suggestive of differential translation were observed along the fault trace" (p. 81) is consistent with my observation that the fault is not marked by mylonite or any high ductile strain zone. Zwanzig's (1973) mapping of Hamill, Badshot, and Lardeau stratigraphy on both sides of the fault that he considered "unsubstantiated" agrees with my conclusion. Sears (1979) estimated that the fault had 2-3 km of attenuation by correlating footwall lithologies in Thompson's map area rather than in the Albert Peak area. If Sears had based the attenuation estimate on lithologies mapped by himself, then his estimate would have been drastically lower; perhaps only a few tens of metres of attenuation is required to omit Badshot marble in the Albert Peak area, and Hamill quartzite is apparently continuous across the fault in South Albert Creek valley.

CHAPTER 7. Tectonic model

Introduction

I interpret the Clachnacudainn terrane as an integral part of the overlying Selkirk allochthon based on stratigraphic, structural, and metamorphic links between the terrane and the base of the overlying Illecillewaet slice that have been outlined in previous chapters. This final chapter incorporates my interpretation into a new tectonic model for the Clachnacudainn terrane and base of the overlying Illecillewaet slice that is similar to those proposed for the northern Selkirk Mountains, compares the model with those previously proposed for the Clachnacudainn terrane and base of the overlying Illecillewaet slice, and suggests future tests of the model.

Age of deformation and metamorphism

Since I interpret the Standfast Creek fault as a minor tectonic feature, it follows that the age and style of deformation and metamorphism in the Clachnacudainn terrane are similar to those in the overlying Selkirk allochthon. The issue then becomes which deformational and metamorphic event, as numerous events in the allochthon have been documented or inferred. Southeast of the Clachnacudainn terrane, in the Kootenay Arc, there is evidence of middle Paleozoic tectonic events (Read and Wheeler, 1976) with superposed events in the Middle Jurassic and Late Cretaceous (Archibald et al., 1983, 1984). In the northern Selkirk Mountains, major deformation and associated regional metamorphism have been ascribed to Middle Jurassic (Brown et al., 1986) and

Cretaceous (Price, 1986) tectonic events. In the northern Monashee Mountains, mid-Cretaceous upper amphibolite facies metamorphism outlasted second phase folding (Sevigny et al., 1990). In the Shuswap metamorphic complex, sillimanite grade metamorphism is Late Cretaceous to Paleocene (Parrish et al., 1988).

The three phases of regional ductile deformation in the Clachnacudainn terrane and base of the Illecillewaet slice must pre-date 104 ± 1 Ma (U-Pb zircon age from the late-tectonic Albert stock) and post-date 356 ± 6 Ma (U-Pb zircon age obtained by Parrish (1992) from the orthogneiss which appears to have been involved in all deformation). Minor post-Late Cretaceous (U-Pb zircon age from lineated leucogranite obtained by Parrish, 1992) deformation is localized in areas such as Mount Revelstoke. Read and Wheeler (1976) suggested that the Akolkolex anticline and Drimmie Creek syncline are middle Paleozoic, possibly Devonian D1 structures. They contended that D1 is older than the latest early Mississippian to Pennsylvanian or Permian Milford Group because randomly oriented and foliated clasts of the Broadview Formation occur in the basal conglomerate of the Milford Group, which truncates underlying Broadview phyllite. The possibility that the pre-peak metamorphic S1 and associated F1 folds in the Clachnacudainn terrane and base of the overlying Illecillewaet slice are middle Paleozoic cannot be ruled out with available geochronological data.

Since I interpret the Akolkolex and Ghost Peak anticlines as northwest-verging F1 folds, then it is likely that they are coeval with major pre-peak metamorphic west-verging folds in the northern Selkirk and Monashee mountains, such as the Scrip nappe (Simony et al., 1980; Raeside and Simony, 1983) and its southern extension, the Carnes nappe (Brown et al., 1983; Brown and Lane, 1988; Brown, 1991). Brown and Lane contended that these west-verging folds originated during accretion of western allochthonous oceanic and island arc

terrane (Superterrane I, see Monger et al., 1982) in the Early to Middle Jurassic and prior to eastward displacement of the Selkirk allochthon on the Monashee decollement. Brown and Lane (1988) and Brown (1991) interpreted the Scrip and Carnes nappes as pre-dating peak metamorphic F1 folds and the map-scale west-verging folds which were superposed on them as synchronous with peak metamorphic F2 folds. Brown et al. (1992) established that west-verging folding and regional metamorphism in the northern Selkirk Mountains are pre-Late Jurassic, thereby noting that a lower age limit on these events is unconstrained. Since the age and style of deformation in the Clachnacudainn terrane and base of the overlying Illecillewaet slice are consistent with those in the northern Selkirk Mountains, then my tectonic model correlates the major deformational events in both areas.

In Chapter 4 on metamorphism, I present evidence which suggests that S1 and associated F1 folds in the Clachnacudainn terrane and base of the overlying Illecillewaet slice pre-date peak regional metamorphism (M2, upper greenschist to middle amphibolite facies). The only age constraint on M2 is that it must be older than mid-Cretaceous because contact metamorphism (M3) around the 104 ± 1 Ma Albert stock post-dated M2. M3 must have post-dated M2 by at least a few million years because M3 assemblages equilibrated at a depth of at least 5.3 km less than that at which M2 assemblages equilibrated (see Chapter 4 for details of these depth estimates). M2 is diachronous with respect to F2 folds across the Clachnacudainn terrane; F2 folds and M2 are coeval at the lowest structural level of the terrane, whereas M2 post-dates D2 in the highest levels. All contact metamorphism (M3) in the Clachnacudainn Range is interpreted as mid-Cretaceous because it occurs in contact aureoles of intrusives that are most likely coeval with the Albert stock.

The tectonic model that I propose also holds that pre-mid-Cretaceous peak regional metamorphism (M2) in the Clachnacudainn terrane and base of the overlying Illecillewaet slice is correlative with pre-Late Jurassic peak regional metamorphism in the northern Selkirk Mountains. If further geochronological and metamorphic studies reveal that this proposal is true, then it would be further evidence supporting my contention that i) major, pre-M2, northwest-verging F1 folds in my study area, such as the Akolkolex and Ghost Peak anticlines, are correlative with major, pre-peak metamorphism, west-verging F1 folds in the northern Selkirk Mountains, such as the Carnes nappe and ii) syn-peak metamorphism F2 folds in my map area are correlative with map-scale, syn-peak metamorphism F2 folds in the northern Selkirk Mountains.

**Clachnacudainn igneous complex - the aspect that differentiates
the terrane from the rest of the eastern Selkirk allochthon**

Chapter 3 on the Clachnacudainn igneous complex describes the wide variety of intrusives that differentiate the terrane from the rest of the Selkirk allochthon east of the Monashee complex; although other regions in the eastern Selkirk allochthon have a greater abundance of intrusives (such as the Battle Range batholith to the east and Kuskanax batholith to the south), none contain as many phases of diverse age and deformational state as the Clachnacudainn igneous complex. Intrusives in the terrane that are unique to the eastern Selkirk allochthon include Devonian orthogneiss, leucogranite (some of which is Late Cretaceous), and leucopegmatite. Since these intrusives also crop out in the Seymour Range (Parrish, 1992), which is in the Selkirk allochthon west of the Monashee complex, Parrish et al. (1988) suggested that the Seymour Range and Clachnacudainn terrane contain the same structural and lithological assemblages

which were transported eastward by the Monashee decollement and then separated during Eocene arching and normal faulting. The terrane is also differentiated from the remainder of the eastern Selkirk allochthon by abundant sillimanite and andalusite grade contact metamorphism in the aureoles of the late-tectonic granitoids in the Clachnacudainn igneous complex.

Wheeler (1963, 1965) used these differentiating igneous and metamorphic aspects as the basis for grouping most of the Clachnacudainn igneous complex into the Clachnacudainn salient of the Shuswap metamorphic complex. The salient only subsequently came to include the intrusive-free, lower grade metasediment south of the Illecillewaet River when Sears (1979) joined fault segments into the composite Standfast Creek fault.

Comparison with recent tectonic models

The stratigraphic, structural, and metamorphic continuity that I mapped across the Standfast Creek fault invalidates tectonic models for the Clachnacudainn terrane and Standfast Creek fault proposed by Read and Brown (1981), Price (1986), and Parrish et al. (1988). The terrane is therefore not a structural horse that was accreted to the hanging wall of the Monashee decollement, not the leading edge of a "suspect terrane" that was wedged into the Selkirk allochthon, and not a high grade salient of the Shuswap metamorphic complex that was rapidly tectonically denuded in the Eocene by a ductile splay from the Columbia River fault. It is important to note that these models were based largely on inference and previous fieldworkers' interpretations.

A regional implication of this tectonic model is that the Standfast Creek fault did not accommodate the large amount of Eocene ductile extensional displacement proposed by Parrish et al. (1988) that is apparently absent on the

segment of the Columbia River fault beneath the Clachnacudainn terrane (Lane (1984) estimated 1-10 km of brittle motion on this fault segment). There are two possible explanations for the apparent lack of Eocene extensional shear zones beneath the Clachnacudainn terrane: i) there was negligible Eocene extensional displacement between the Monashee complex and Clachnacudainn terrane or ii) some of the post-metamorphic mylonite in the Columbia River valley is related to Eocene extension rather than Mesozoic to Paleocene compression on the Monashee decollement. Lane et al. (1989) preferred to interpret the post-metamorphic mylonite as having a compressional origin, but they could not rule out the possibility of an Eocene extensional origin for the mylonite because its age is poorly constrained.

Tests of the tectonic model

The tectonic model proposed here could be tested primarily by further geochronological investigations. The first investigation should be an Ar-Ar mica cooling age traverse across the Standfast Creek fault, beginning in the intrusive-free southern portion of the Clachnacudainn terrane, which would reveal whether the fault juxtaposes levels with different thermal histories. According to the proposed model, rocks in the immediate hanging wall and footwall have identical thermal histories; there should be no discrepancy in mica cooling dates across the Standfast Creek fault. The mica cooling dates could be as young as Late Cretaceous if the sampled rocks were affected by the contact metamorphism (M3) due to the Albert stock, or the dates could be substantially older than mid-Cretaceous if the rocks were unaffected by M3. The second investigation should be to date the deformed and undeformed leucogranite dykes in Albert Canyon. The age of these dykes would effectively test the age of deformation proposed by

this tectonic model. Other future geochronologic investigations in the Clachnacudainn terrane should include dating emplacement of the following late-tectonic intrusives in order to determine whether they truly are coeval with the 104 ± 1 Ma Albert stock: i) the dyke that crosscuts the Devonian orthogneiss-metasediment contact on the ridge north of North Albert Peak, ii) quartz diorite in the Siegfried Creek body, iii) granodiorite that intrudes the Jowett Formation at the lower Standfast Creek fault on the ridge north of West Woolsey Creek, and iv) granite along Sale Creek and the headwaters of West Woolsey Creek.

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Appendix 1. Mineral assemblages in thin section samples. Sample localities are shown in Map 2. Abbreviations are Hamill Group (Ha), Mohican Formation (Mo), Badshot Formation (Bd), Lade Peak Formation (LP), Index Formation (In), Jowett Formation (Jt), Broadview Formation (Bw), Devonian orthogneiss (Dg), Clachnacudainn igneous complex (Cc), uncorrelated metasedimentary rock (?). Letters in the assemblage column refer to AFM topologies shown in Figure 4-6 (those in capital letters are contact metamorphic assemblages).

Sample	Formation	Assemblage	Quartz	Plagioclase	K-feldspar	Muscovite	Biotite	Chlorite	Garnet	Staurolite	Kyanite	Sillimanite	Calcite	Hornblende	Actinolite	Zircon	Tourmaline	Rutile	Titanite	Talc	Zoisite
JC5A	Bw	c	x	x		x	x	x		x	x	x									
JC6A	Jt	j	x	x			x	x							x	x					
JC10A	Bw	i	x	x		x	x	x	x							x	x				
JC11	Jt		x										x	x					x		
JC11B	Jt	i	x	x		x	x	x	x						x						
JC13A	Bw	d	x			x	x	x	x	x						x	x				
JC13B	Bw	d	x	x		x	x	x	x	x						x	x				
JC13C	Bw	d	x	x		x	x	x	x	x						x	x				
JC13D	Bw	d	x	x		x	x	x	x	x						x	x				
JC14B	Bw	f	x	x		x	x	x	x				x			x					
JC14D	Bw	d	x	x		x		x	x	x						x	x		x		x
JC15	In		x			x							x								
JC19	In	i	x	x		x	x	x	x							x	x		x		
JC21A	In	e	x	x		x		x	x				x			x	x			x	x
JC25	Dg		x	x	x	x	x	x					x			x	x				
JC26B	Bw		x			x										x					
JC27	In	l	x			x	x	x								x	x				
JC32B	In	e	x			x		x	x							x					
JC36B	In	e	x			x		x	x							x	x	x			
JC39A	In		x			x															
JC40	In		x			x		x					x			x	x	x			
JC41B	In	l	x			x	x	x					x			x					
JC43B	In	l	x			x	x	x								x					
JC44A	In	e	x			x		x	x												
JC44B	In		x			x															
JC47	Bw		x			x															
JC48A	Bw	f	x	x		x	x	x	x							x					
JC54A	Ha	e	x	x		x		x	x							x	x				
JC54B	Ha	f	x	x		x	x	x	x				x			x	x				x
JC56	Ha	i	x	x		x	x	x	x							x	x				x

Sample	Formation	Assemblage	Quartz	Plagioclase	K-feldspar	Muscovite	Biotite	Chlorite	Garnet	Sillimanite	Calcite	Hornblende	Actin/Trem	Zircon	Tourmaline	Rutile	Titanite	Talc	Zoisite
JC57	Ha	f	x			x	x	x	x		x			x	x				
JC58B	Ha	f	x			x	x	x	x					x	x		x		
JC60	In	e	x	x		x		x	x					x	x				
JC63	Ha	f	x	x		x	x		x					x	x				
JC65	In	e	x	x		x		x	x					x	x				x
JC72A	Ha	i	x			x	x	x	x		x			x	x				
JC72B	Ha	i	x			x	x	x	x					x	x				
JC73B	Ha	f	x			x	x	x	x					x	x				
JC73C	Ha	l	x			x	x	x						x	x			x	
JC75	In	i	x	x		x	x	x	x		x			x	x			x	
JC76	Ha	f	x			x	x	x	x					x	x		x	x	
JC81	In	k	x			x					x			x	x			x	
JC84A	Ha	i	x			x	x	x	x						x	x			
JC89A	LP		x			x		x			x		x	x					x
JC93A	In	A	x	x	x	x	x	x	x	x				x					
JC101A	In	f	x	x		x	x		x		x			x					
JC101C	In	m	x	x		x	x	x	x					x				x	x
JC101D	In	f	x	x		x	x	x	x					x			x	x	x
JC101E	Dg		x	x			x	x				x	x	x			x		
JC105	In		x	x			x				x	x		x		x			
JC107B	Dg		x	x		x	x	x					x	x					
JC109A	In	f	x	x		x	x	x	x										x
JC114	Cc		x	x	x	x	x	x						x					
JC126B	In	i	x	x		x	x	x	x					x	x				
JC126C	In	i	x	x		x	x		x					x	x				
JC132	Jt	j	x	x			x				x	x		x			x		
JC133	Bw	i	x	x		x	x		x					x	x	x	x		
JC134	Jt		x	x			x	x				x					x		x
JC135	Bw	f	x	x		x	x	x	x					x	x				
JC138B	Mo	l	x	x			x	x			x			x	x				

Sample	Formation	Assemblage	Quartz	Plagioclase	K-feldspar	Muscovite	Biotite	Chlorite	Garnet	Staurolite	Andalusite	Sillimanite	Calcite	Hornblende	Actin/Trem	Zircon	Tourmaline	Rutile	Titanite
JC140	In		x			x										x	x		
JC141	Dg		x	x			x							x	x	x			x
JC157	Cc		x	x			x	x						x	x	x	x		x
JC160	Bw	l	x	x	x	x	x	x								x			x
JC162	Cc		x	x	x	x	x	x								x			x
JC174B	Bw	d	x			x	x									x	x		
JC174C	Bw	d	x	x		x	x	x	x	x						x	x		
JC190	Ha	j	x	x		x	x	x	x	x						x	x		
JC196	Ha	B	x	x		x	x					x					x		
JC196B	Cc	i	x	x		x	x	x	x										
JC198	Ha	f	x	x		x	x	x	x				x		x	x			
JC199D	Ha	i	x	x		x	x	x	x								x		
JC200	Ha	A	x	x		x	x	x	x	x	x					x	x		
JC200B	Ha	A	x	x		x	x		x	x	x	x					x		
JC203	Ha	l	x			x	x	x					x			x			
JC204A	LP		x																
JC204B	LP		x													x			x
JC205	LP		x										x						
JC208B	In	i	x	x		x	x	x	x							x			
JC211	Dg		x	x			x	x					x	x		x			
JC215A	In	f	x			x	x	x	x							x	x		
JC215C	In	j	x			x	x	x									x		
JC216C	In	l	x			x	x						x						x
JC219	Cc			x	x			x								x			x
JC219E	In	l	x			x	x	x											x
JC223	Cc		x	x	x		x							x					x
JC229	In		x	x		x							x						
JC233D	Jt		x		x		x	x										x	
JC233E	Bw	d	x	x		x	x	x	x							x	x		
JC234	In	l	x	x		x	x	x					x						

Sample	Formation	Assemblage	Quartz	Plagioclase	K-feldspar	Muscovite	Biotite	Chlorite	Garnet	Sillimanite	Calcite	Hornblende	Actin/Trem	Zircon	Tourmaline	Rutile	Titanite
JC240	?	B	x	x		x	x			x							
JC241	Cc			x		x	x					x	x	x			x
JC244B	Bw	f	x			x	x	x	x								
JC246	Bw	i	x	x		x	x	x	x								
AS1	Cc		x	x	x		x	x				x		x		x	x
AS2	Cc		x	x	x	x	x	x			x			x	x		
SST	Cc		x	x	x		x	x				x		x			x

APPENDIX 2. Mineral abbreviations

ab - albite
alm - almandine
an - anorthite
ann - annite
as - alumino-silicate
bt - biotite
cd - cordierite
ch - chlorite
ct - chloritoid
grs - grossular
gt - garnet
kf - K-feldspar
ky - kyanite
mu - muscovite
or - orthoclase
pl - plagioclase
prp - pyrope
qz - quartz
si - sillimanite
sps - spessartine
st - staurolite

APPENDIX 3. Microprobe standards

garnet: Si, Al, Fe - almandine
 Ca, Mg - Kakanui garnet
 Ti - aenigmatite
 Mn - rhodonite
 Na - albite
 F - phlogopite

biotite: Si, Al, Fe, Mg - biotite
 Ca - Kakanui garnet
 Ti - aenigmatite
 Mn - rhodonite
 Na - albite
 F - phlogopite

muscovite: Si, Al - muscovite
 Fe, Mg - biotite
 Ca - Kakanui garnet
 Ti - aenigmatite
 Mn - rhodonite
 Na - albite
 F - phlogopite

plagioclase: Na, Al, Si - albite
 Fe, Mg - biotite
 Ca - An50 glass
 K - orthoclase
 Ba - hyalophane
 F - phlogopite

Appendix 4. Selected mineral compositions generated by microprobe analysis. Sample number and mineral analyzed are given at the top of the tables. Abbreviations in the row headed by "mineral" or "min" indicate whether the analysis is from the rim (R) or core (C) of the mineral, which is abbreviated (see Appendix 3). Numbers and letters in the row headed by "location" or "loc" refer to the analysis' location in the thin section. In the garnet tables (Table 1) these are read as follows: garnet crystal (profile number)-subdomain. In the biotite and muscovite tables (Tables 2 and 3) these are read as follows: domain-subdomain. In the plagioclase table (Table 4) these are read as follows: domain-subdomain (profile number). Numbers in the row headed by "analysis #" or "an#" are analysis numbers assigned by the microprobe. Compositional data that are an average of multiple analyses are denoted in this row by AVG-number of analyses. Letters in the row headed by "P-T" refer to compositions used in thermobarometry; the other minerals used in the calculations and the P-T results are shown in Table 4-1.

MINI FIVE 1111

Notes includes mapping by
 completed in part by Hong
 Hsiao to completed data
 Thompson, Chant, and comp.

STRATIGRAPHIC CH

CRETACEDUS

LATE CRETACEOUS

[Klg] Muscovite biotite leuco

MIDDLE CRETACEOUS

[K7d] Biotite hornblende unit

[K7gd] Biotite granodiorite

[K7gr] Muscovite biotite (gr 91)

[Kmg] Megacrystic K-feldspar

[K7qd] Biotite hornblende 19.2

Kqm Biotite hornblende gneiss

MOUNT RICHMOND MAP

CHIAIACUDAINN TERRANE &

BAY OF THE LYPPLING ILLECILLEWAET SLICE

PART III LAMBUENE (82 K/13), REVELSTOKE (82 L/16),

MOUNT REVELSTOKE (82 M/1), & ILLECILLEWAET (82 N/4)

MAP AREAS

BRITISH COLUMBIA

Notes includes mapping by J.L. Crowley (1990, 1991) and a compilation of mapping modified in part by Thompson (1972), Zwanzig (1973), Read and Wheeler (1976), Bakken (unpublished data, 1976, 1977), Sears (1979), Read (1980), Read and Thompson (1980), Read (unpublished data, 1982), and Price (1986).

STRATIGRAPHIC COLUMN

CRETACEOUS

LATE CRETACEOUS

[Klg] Muscovite biotite leucogranite and pegmatite

MIDDLE CRETACEOUS

[K7d] Biotite hornblende diorite

[K7gd] Biotite granodiorite

[K7gr] Muscovite biotite granite

[Kmg] Megacrystic K-feldspar granite

[K7qd] Biotite hornblende quartz diorite

[Kqn] Biotite hornblende quartz monzonite

MIOCENE TO PALEOGENE (MOUNT RICHMOND)

Table 1 of Appendix 4. Selected garnet compositions based on 12 oxygen atoms

Sample 13b - garnet											
mineral	RG1	RG2	RG3	RG4	RG5	RG6	RG7	RG8	CG1	CG2	CG3
location	1(1)-A	1(2)-A	1(2)-B	1(3)-A	1(3)-B	2-A	2-B	3-2	1(2)	1(3)	2
analysis #	344	375	454	462	528	283	321	21	424	502	300
P-T	A	B	C	D	E	F	G	H			
SiO ₂	37.87	36.68	37.06	37.56	37.62	37.63	37.29	37.36	36.96	37.78	37.49
TiO ₂	0.00	0.06	0.02	0.02	0.01	0.17	0.01	0.01	0.13	0.14	0.11
Al ₂ O ₃	20.49	20.68	20.79	21.25	20.81	20.29	20.05	21.34	20.86	20.95	20.34
FeO	38.31	38.16	37.93	38.44	39.12	36.97	37.13	38.75	27.56	27.67	27.48
MnO	0.48	0.50	0.60	0.50	0.49	0.64	0.57	0.73	7.11	7.24	7.32
MgO	2.81	2.76	2.73	2.77	2.86	2.78	2.74	2.25	0.72	0.73	0.74
CaO	1.98	1.84	2.24	2.06	1.90	2.33	1.98	1.2	6.84	7.06	7.65
Total	101.94	100.68	101.37	102.63	102.81	100.81	99.77	102.38	100.18	101.57	101.33
Si	3.015	2.965	2.973	2.974	2.980	3.021	3.028	3.004	2.987	3.008	3.003
Al(M)	0.000	0.005	0.027	0.026	0.020	0.000	0.000	0.000	0.013	0.000	0.000
	3.015	3.000	3.000	3.000	3.000	3.021	3.028	3.004	3.000	3.008	3.003
Al(V)	1.923	1.936	1.938	1.955	1.923	1.920	1.919	1.953	1.975	1.966	1.920
Ti	0.000	0.004	0.001	0.001	0.001	0.010	0.001	0.001	0.008	0.008	0.007
Fe ²⁺	2.551	2.580	2.544	2.543	2.592	2.482	2.521	2.530	1.863	1.842	1.841
Mn	0.032	0.034	0.041	0.034	0.033	0.044	0.039	0.037	0.487	0.488	0.497
Mg	0.333	0.332	0.326	0.327	0.338	0.333	0.332	0.331	0.087	0.087	0.088
Ca	0.169	0.159	0.3	0.175	0.161	0.200	0.172	0.165	0.592	0.602	0.674
	5.008	5.046	5.043	5.034	5.048	4.988	4.984	5.016	5.011	4.993	5.027
alm	0.827	0.831	0.820	0.826	0.830	0.811	0.823	0.826	0.615	0.610	0.594
spe	0.010	0.011	0.013	0.011	0.011	0.014	0.013	0.012	0.161	0.162	0.160
prp	0.108	0.107	0.105	0.106	0.108	0.108	0.108	0.108	0.029	0.029	0.029
grs	0.055	0.051	0.062	0.057	0.052	0.066	0.056	0.054	0.196	0.199	0.217
Fe/(Fe+Mg)	0.884	0.886	0.886	0.886	0.885	0.882	0.884	0.884	0.956	0.955	0.954
Sample 13d1 - garnet											
mineral	RG1	RG2	RG3	RG4	RG5	CG1	CG2				
location	1-1	1-1	1-4	2-3	2-4	2	1				
analysis #	144	147	166	163	162	172	149				
P-T	B	D	E	A	C						
SiO ₂	37.13	37.56	37.80	38.01	37.59	37.99	37.96				
TiO ₂	0.01	0.01	0.08	0.02	0.00	0.26	0.09				
Al ₂ O ₃	20.12	20.80	21.09	21.04	20.37	20.35	20.81				
FeO	33.88	35.06	35.10	35.25	34.73	26.20	28.13				
MnO	1.20	1.38	0.79	1.22	1.26	7.59	6.18				
MgO	2.64	2.75	2.90	2.97	2.86	0.89	0.89				
CaO	4.30	3.84	4.69	3.90	3.70	8.20	8.06				
Total	99.28	101.40	102.45	102.41	100.51	101.48	102.12				
Si	3.016	2.993	2.977	2.994	3.017	3.023	3.005				
Al(M)	0.000	0.007	0.023	0.006	0.000	0.000	0.000				
	3.016	3.000	3.000	3.000	3.017	3.023	3.005				
Al(V)	1.926	1.946	1.934	1.947	1.927	1.908	1.942				
Ti	0.001	0.001	0.005	0.001	0.000	0.016	0.005				
Fe ²⁺	2.301	2.336	2.312	2.322	2.331	1.744	1.863				
Mn	0.083	0.083	0.053	0.081	0.086	0.512	0.414				
Mg	0.320	0.327	0.340	0.349	0.342	0.108	0.105				
Ca	0.374	0.328	0.386	0.329	0.318	0.689	0.684				
	5.005	5.030	5.040	5.029	5.004	4.984	5.013				
alm	0.748	0.758	0.746	0.754	0.758	0.570	0.608				
spe	0.027	0.030	0.017	0.026	0.028	0.167	0.135				
prp	0.104	0.108	0.110	0.113	0.111	0.035	0.034				
grs	0.122	0.106	0.128	0.107	0.103	0.228	0.223				
Fe/(Fe+Mg)	0.878	0.877	0.872	0.869	0.872	0.943	0.947				

Table 1 of Appendix 4. Selected garnet compositions based on 12 oxygen atoms

Sample 63 - garnet

mineral location	RG1 1	RG1 1	RG1 1	RG1 1	CGT 1	CGT 1
analysis #	AVG-3	AVG-3	AVG-2	344	350	349
P-T	A	B	C			
SiO ₂	37.66	37.26	37.23	37.41	37.37	37.62
TiO ₂	0.05	0.02	0.02	0.04	0.11	0.10
Al ₂ O ₃	21.02	20.71	20.41	20.28	20.35	20.74
FeO	35.00	35.77	35.99	35.06	31.11	30.17
MnO	1.24	0.79	0.71	1.05	4.39	4.21
MgO	1.57	1.65	1.92	1.82	1.19	1.19
CaO	5.64	4.76	4.43	4.89	6.53	6.90
Total	102.19	101.17	100.71	100.55	101.05	100.90
Si	2.987	2.988	3.001	3.015	3.002	3.009
Al(IV)	0.013	0.012	0.000	0.000	0.000	0.000
	3.000	3.000	3.001	3.015	3.002	3.009
Al(VI)	1.953	1.946	1.939	1.926	1.927	1.955
Ti	0.003	0.001	0.001	0.002	0.007	0.006
Fe ²⁺	2.322	2.399	2.426	2.363	2.090	2.018
Mn	0.084	0.054	0.048	0.072	0.299	0.285
Mg	0.186	0.221	0.231	0.219	0.142	0.142
Ca	0.480	0.409	0.382	0.422	0.562	0.581
	5.027	5.031	5.027	5.005	5.026	4.998
alm	0.756	0.778	0.786	0.768	0.676	0.665
sps	0.027	0.017	0.016	0.023	0.097	0.094
prp	0.061	0.072	0.075	0.071	0.046	0.047
grs	0.156	0.133	0.124	0.137	0.182	0.195
Fe/(Fe+Mg)	0.926	0.916	0.913	0.915	0.936	0.934

Sample 72a - garnet

mineral location	RG1 1	RG1 1	RG1 1	RG1 1	RG1 1	RG1 2	RG1 2	RG1 2	CGT 1	CGT 1
analysis #	109	111	116	110	121	137	138	139	118	119
P-T	A	B				C				
SiO ₂	38.47	37.91	38.34	38.94	37.43	38.12	38.12	37.87	37.27	37.13
TiO ₂	0.07	0.06	0.10	0.05	0.06	0.06	0.10	0.10	0.16	0.09
Al ₂ O ₃	21.04	20.81	20.34	21.37	20.53	21.12	20.64	20.48	19.98	20.37
FeO	32.82	34.66	34.35	33.22	33.99	34.11	33.89	34.62	32.81	33.87
MnO	2.39	2.72	2.62	2.45	2.29	2.51	2.48	2.24	2.86	2.16
MgO	1.47	1.39	1.36	1.43	1.34	1.39	1.34	1.29	1.15	1.28
CaO	4.56	5.17	5.63	4.57	5.66	5.12	5.86	6.07	5.80	5.78
Total	100.82	102.72	102.74	102.03	101.32	102.43	102.43	102.67	100.03	100.68
Si	3.061	3.001	3.031	3.062	3.000	3.012	3.017	3.002	3.024	2.997
Al(IV)	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.003
	3.061	3.001	3.031	3.062	3.000	3.012	3.017	3.002	3.024	3.000
Al(VI)	1.974	1.942	1.895	1.980	1.940	1.967	1.926	1.914	1.911	1.935
Ti	0.004	0.004	0.006	0.003	0.005	0.004	0.006	0.006	0.010	0.005
Fe ²⁺	2.184	2.294	2.271	2.184	2.279	2.254	2.243	2.295	2.226	2.288
Mn	0.161	0.182	0.175	0.163	0.155	0.168	0.166	0.150	0.197	0.148
Mg	0.174	0.164	0.160	0.168	0.160	0.164	0.158	0.152	0.139	0.154
Ca	0.389	0.438	0.477	0.385	0.486	0.433	0.497	0.516	0.504	0.500
	4.886	5.024	4.985	4.884	5.025	4.989	4.997	5.033	4.987	5.029
alm	0.751	0.745	0.736	0.753	0.740	0.747	0.732	0.737	0.726	0.740
sps	0.055	0.059	0.057	0.056	0.050	0.056	0.054	0.048	0.064	0.048
prp	0.060	0.053	0.052	0.058	0.052	0.054	0.052	0.049	0.045	0.050
grs	0.134	0.142	0.155	0.133	0.158	0.144	0.162	0.166	0.164	0.162
Fe/(Fe+Mg)	0.926	0.933	0.934	0.929	0.934	0.932	0.934	0.938	0.941	0.937

Table 1 of Appendix 4 Selected garnet compositions based on 12 oxygen atoms

Sample 76 - garnet

mineral location analysis #	RG1 1-1 AVG-2 A	RG1 1-1 AVG-3 B	RG1 1-1 AVG-7 C	RG1 1-A 181	RG1 1-B 217	RG1 1-B 218	RG1 1-B 221	RG1 2 258	RG1 2 259	RG1 2 260	CGT 1 199	CGT 1 200	CGT 1 201	CGT 1 180
P-T	A	B	C				D		E					
SiO ₂	38.06	38.12	38.22	37.94	38.02	37.81	37.70	38.07	38.07	37.99	37.31	37.25	36.95	38.04
TiO ₂	0.01	0.05	0.03	0.02	0.05	0.05	0.02	0.06	0.05	0.05	0.08	0.08	0.13	0.10
Al ₂ O ₃	21.04	21.27	21.15	21.47	21.07	21.79	21.31	20.99	21.20	20.95	21.11	21.04	20.72	21.35
FeO	36.62	35.89	36.59	36.67	32.85	33.19	36.13	33.74	33.60	34.02	32.11	32.07	31.79	32.72
MnO	1.83	2.53	1.99	1.89	2.93	2.91	1.70	2.02	2.44	2.28	4.93	4.87	4.96	4.97
MgO	2.03	1.86	1.95	1.93	1.74	1.91	2.02	1.70	1.74	1.72	1.30	1.32	1.34	1.35
CaO	3.92	4.19	3.96	4.11	5.81	4.51	3.55	6.02	5.83	5.93	4.89	4.78	4.80	5.07
Total	103.49	103.94	103.90	104.03	102.47	102.17	102.43	102.59	102.93	102.94	101.73	101.42	100.69	103.60
Si	2.991	2.983	2.991	2.968	2.999	2.983	2.985	3.001	2.992	2.991	2.979	2.983	2.983	2.984
Al(IV)	0.009	0.017	0.009	0.032	0.001	0.017	0.015	0.000	0.008	0.009	0.021	0.017	0.017	0.016
	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.001	3.000	3.000	3.000	3.000	3.000	3.000
Al(vi)	1.339	1.944	1.942	1.947	1.958	2.009	1.973	1.950	1.955	1.938	1.968	1.968	1.954	1.957
Ti	0.000	0.003	0.002	0.001	0.003	0.003	0.001	0.004	0.003	0.003	0.005	0.005	0.008	0.006
Fe 2+	2.407	2.348	2.395	2.399	2.167	2.190	2.392	2.224	2.208	2.240	2.144	2.148	2.146	2.146
Mn	0.122	0.168	0.132	0.125	0.196	0.194	0.114	0.135	0.162	0.152	0.333	0.330	0.339	0.330
Mg	0.237	0.220	0.228	0.225	0.205	0.225	0.238	0.200	0.204	0.202	0.155	0.158	0.161	0.158
Ca	0.330	0.351	0.332	0.344	0.491	0.381	0.301	0.508	0.491	0.500	0.418	0.410	0.415	0.426
	5.035	5.034	5.032	5.042	5.019	5.001	5.020	5.020	5.024	5.033	5.022	5.019	5.024	5.024
alm	0.777	0.781	0.778	0.775	0.709	0.732	0.785	0.725	0.720	0.724	0.703	0.705	0.701	0.701
spe	0.039	0.054	0.043	0.040	0.064	0.065	0.037	0.044	0.053	0.049	0.109	0.108	0.111	0.108
prp	0.077	0.071	0.074	0.073	0.067	0.075	0.078	0.065	0.068	0.065	0.051	0.052	0.053	0.052
grs	0.107	0.114	0.108	0.111	0.161	0.127	0.099	0.166	0.160	0.162	0.137	0.135	0.136	0.139
Fe/(Fe+Mg)	0.910	0.914	0.913	0.914	0.914	0.907	0.909	0.917	0.915	0.917	0.933	0.932	0.930	0.931

Sample 135 - garnet

mineral location analysis #	RG1 1-1 211	RG1 1-1 217	RG1 1-A 227	RG1 2-A 537	CGT 2-B 571	CGT 1 245	RCGT 1 231	CGT 2 555	RCGT 2 546
P-T	A	B	C	D					
SiO ₂	37.89	38.55	37.78	37.72	37.41	37.35	37.95	38.98	37.49
TiO ₂	0.01	0.01	0.02	0.03	0.03	0.08	0.04	0.06	0.08
Al ₂ O ₃	20.85	21.16	20.90	20.63	21.04	20.33	20.55	20.95	21.00
FeO	37.22	36.43	37.24	36.24	36.17	29.40	33.68	30.10	32.59
MnO	1.83	1.85	1.54	1.90	1.59	6.29	2.77	6.43	3.04
MgO	2.26	2.24	2.28	2.17	2.15	1.16	1.65	1.16	1.46
CaO	2.76	3.06	3.23	3.37	3.02	6.02	6.03	5.23	6.01
Total	102.82	103.30	102.99	102.08	101.41	100.83	102.67	100.91	101.65
Si	2.988	3.020	2.985	3.003	2.981	3.011	3.001	2.977	2.986
Al(IV)	0.002	0.000	0.014	0.000	0.009	0.000	0.000	0.023	0.014
	3.000	3.020	3.000	3.003	3.000	3.011	3.001	3.000	3.000
Al(vi)	1.942	1.954	1.932	1.938	1.973	1.932	1.916	1.986	1.958
Ti	0.001	0.001	0.001	0.002	0.002	0.005	0.002	0.004	0.004
Fe 2+	2.463	2.387	2.461	2.413	2.418	1.982	2.228	2.027	2.171
Mn	0.123	0.123	0.103	0.128	0.108	0.430	0.186	0.439	0.205
Mg	0.267	0.262	0.269	0.258	0.256	0.139	0.195	0.139	0.173
Ca	0.234	0.257	0.274	0.287	0.259	0.520	0.511	0.451	0.513
	5.029	4.982	5.040	5.024	5.016	5.068	5.037	5.025	5.024
alm	0.798	0.798	0.792	0.782	0.795	0.645	0.714	0.663	0.709
spe	0.040	0.041	0.033	0.042	0.035	0.140	0.059	0.144	0.067
prp	0.088	0.086	0.086	0.083	0.084	0.045	0.062	0.046	0.057
grs	0.076	0.085	0.088	0.093	0.085	0.169	0.164	0.148	0.167
Fe/(Fe+Mg)	0.902	0.901	0.902	0.904	0.904	0.934	0.920	0.938	0.926

Table 2 of Appendix 4. Selected biotite compositions based on 22 oxygen atoms

Sample 13b - biotite

min	RBT	RBT	RBT	RBT	RBT	RBT	RBT	CBT	CBT	CBT	CBT	CBT	CBT	CBT
loc	2-1	2-2	2	1	3-2	3	4	2-1	2-2	2	1	3-2	3-3	3
anal	AVG-3	AVG-2	AVG-6	AVG-3	24	AVG-5	AVG-4	AVG-2	AVG-2	AVG-5	AVG-4	25	27	AVG-4
P-T	B		B	A	D							E	C	
SiO ₂	35.99	35.38	35.76	35.13	35.69	35.88	35.96	35.86	35.56	35.67	35.72	35.51	35.34	35.69
TiO ₂	1.42	1.29	1.37	1.56	1.63	1.90	2.03	1.70	1.51	1.68	2.15	1.67	1.58	1.77
Al ₂ O ₃	20.26	20.19	20.24	19.87	19.87	19.97	19.33	19.93	19.68	19.86	19.91	20.01	20.16	19.93
FeO	21.63	21.60	21.64	21.50	20.87	20.51	21.32	21.38	21.93	21.69	21.31	20.85	20.11	20.81
MnO	0.02	0.01	0.02	0.02	0.02	0.01	0.01	0.02	0.02	0.02	0.02	0.02	0.02	0.04
MgO	9.31	9.24	9.24	8.98	9.45	9.22	9.33	9.27	9.40	9.18	8.86	9.35	9.48	9.32
CaO	0.00	0.06	0.02	0.03	0.13	0.03	0.01	0.02	0.06	0.02	0.04	0.08	0.06	0.03
Na ₂ O	0.27	0.24	0.26	0.26	0.29	0.33	0.30	0.24	0.30	0.28	0.31	0.30	0.33	0.31
K ₂ O	8.86	8.68	8.79	8.81	8.12	8.72	8.51	8.90	8.51	8.73	8.76	8.47	8.46	8.66
F	0.29	0.26	0.28	0.28	0.31	0.25	0.28	0.23	0.28	0.25	0.28	0.28	0.30	0.30
H ₂ O	3.89	3.86	3.88	3.83	3.83	3.88	3.86	3.90	3.85	3.88	3.87	3.85	3.82	3.85
Total	101.69	100.54	101.24	100.02	99.94	100.47	100.67	101.21	100.82	101.03	100.97	100.12	99.36	100.43
Si	5.356	5.330	5.347	5.325	5.374	5.379	5.306	5.362	5.347	5.350	5.351	5.349	5.347	5.362
Al(M)	2.644	2.670	2.653	2.671	2.626	2.621	2.604	2.638	2.653	2.650	2.649	2.651	2.653	2.638
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Ti	0.159	0.146	0.155	0.177	0.186	0.214	0.229	0.191	0.170	0.189	0.242	0.189	0.178	0.200
Al(V)	0.910	0.914	0.914	0.883	0.900	0.908	0.815	0.874	0.835	0.890	0.868	0.901	0.942	0.890
Fe	2.692	2.722	2.706	2.728	2.628	2.572	2.676	2.672	2.758	2.722	2.670	2.627	2.546	2.615
Mn	0.003	0.001	0.002	0.003	0.003	0.002	0.001	0.003	0.003	0.003	0.003	0.003	0.003	0.004
Mg	2.064	2.074	2.068	2.031	2.120	2.080	2.096	2.065	2.107	2.053	1.977	2.099	2.139	2.087
	5.828	5.857	5.835	5.822	5.835	5.755	5.807	5.804	5.872	5.827	5.780	5.818	5.806	5.798
Ca	0.001	0.009	0.003	0.006	0.021	0.005	0.001	0.002	0.007	0.004	0.008	0.010	0.008	0.004
Na	0.078	0.070	0.076	0.076	0.085	0.097	0.088	0.089	0.088	0.080	0.089	0.088	0.097	0.090
K	1.681	1.663	1.677	1.704	1.580	1.667	1.630	1.698	1.633	1.671	1.675	1.628	1.633	1.680
	1.760	1.742	1.757	1.786	1.665	1.769	1.716	1.770	1.726	1.755	1.770	1.725	1.738	1.754
F	0.136	0.121	0.132	0.125	0.148	0.119	0.134	0.109	0.133	0.120	0.130	0.133	0.144	0.141
H	3.864	3.879	3.868	3.875	3.852	3.881	3.866	3.891	3.867	3.880	3.870	3.867	3.856	3.869
Fe/Mg	1.304	1.312	1.315	1.343	1.239	1.248	1.283	1.264	1.309	1.326	1.350	1.251	1.190	1.253

Sample 13d1 - biotite

min	RBT	RBT	RBT	RBT	RBT	CBT	CBT	CBT	CBT
loc	1	2-1	2-4	3-2	1	2-1	2-3	3-2	
anal	AVG-5	10	AVG-2	17	19	AVG-4	11	16	20
P-T	B		C	D			A		
SiO ₂	36.83	36.35	35.89	35.65	35.88	35.73	35.84	36.67	36.10
TiO ₂	1.88	1.78	1.16	1.01	1.74	1.91	1.86	1.54	1.76
Al ₂ O ₃	19.88	19.56	20.01	20.26	19.83	19.72	19.19	20.72	19.82
FeO	19.17	19.57	20.15	20.19	19.02	18.96	20.82	18.99	19.07
MnO	0.04	0.02	0.04	0.04	0.03	0.03	0.04	0.03	0.02
MgO	10.61	10.32	10.20	10.03	10.52	10.54	10.32	10.40	10.62
CaO	0.02	0.01	0.07	0.09	0.02	0.01	0.00	0.00	0.00
Na ₂ O	0.21	0.22	0.18	0.17	0.23	0.21	0.19	0.19	0.24
K ₂ O	8.77	8.81	8.43	8.20	8.68	8.81	8.73	8.98	8.92
F	0.35	0.32	0.33	0.29	0.36	0.33	0.34	0.39	0.29
H ₂ O	3.85	3.87	3.84	3.84	3.83	3.84	3.84	3.89	3.88
Total	100.29	100.54	99.98	99.51	99.82	99.77	100.67	101.45	100.47
Si	5.350	5.419	5.387	5.372	5.378	5.381	5.370	5.392	5.381
Al(M)	2.650	2.581	2.613	2.628	2.624	2.639	2.630	2.608	2.619
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Ti	0.211	0.200	0.130	0.114	0.196	0.216	0.210	0.170	0.197
Al(V)	0.849	0.866	0.928	0.970	0.878	0.850	0.780	0.984	0.883
Fe	2.394	2.440	2.529	2.544	2.383	2.378	2.584	2.335	2.377
Mn	0.006	0.003	0.006	0.006	0.004	0.003	0.006	0.004	0.003
Mg	2.361	2.293	2.280	2.252	2.349	2.368	2.305	2.279	2.369
	5.821	5.791	5.871	5.887	5.811	5.804	5.883	5.772	5.798
Ca	0.003	0.002	0.010	0.015	0.003	0.002	0.000	0.000	0.000
Na	0.061	0.064	0.061	0.060	0.067	0.080	0.055	0.064	0.089
K	1.671	1.676	1.613	1.576	1.669	1.686	1.689	1.685	1.688
	1.735	1.741	1.674	1.641	1.729	1.748	1.724	1.739	1.768
F	0.164	0.151	0.154	0.138	0.171	0.158	0.161	0.181	0.137
H	3.836	3.849	3.846	3.882	3.829	3.842	3.839	3.819	3.883
Fe/Mg	1.014	1.064	1.106	1.130	1.015	1.009	1.121	1.025	1.008

Table 2 of Appendix 4 Selected biotite compositions based on 22 oxygen atoms
Sample 63 - biotite

min loc anal	RBT 1 AVG-2	RBT 1 AVG-3	RBT 1 AVG-3	RBT 1 AVG-2	RBT 1 AVG-17	CBT 1 12	CBT 1 21	CBT 1 24	CBT 1 AVG-7
P-T	B			C	A				
SiO ₂	35.13	34.74	35.13	35.14	34.92	35.11	34.73	35.17	34.84
TiO ₂	1.82	1.62	1.63	1.65	1.75	1.57	1.54	1.70	1.65
Al ₂ O ₃	18.13	18.85	18.61	18.83	18.57	18.96	18.68	18.46	18.60
FeO	22.37	21.26	21.60	21.25	21.52	20.97	22.72	21.57	21.40
MnO	0.08	0.07	0.08	0.08	0.07	0.07	0.08	0.05	0.08
MgO	8.74	8.89	8.72	8.88	8.75	8.98	8.65	9.02	8.94
CaO	0.04	0.01	0.04	0.03	0.06	0.00	0.03	0.01	0.02
Na ₂ O	0.06	0.09	0.08	0.07	0.08	0.09	0.12	0.11	0.11
K ₂ O	8.77	9.47	9.08	9.29	8.95	9.43	9.34	9.35	9.30
F	0.87	0.70	0.86	0.85	0.84	0.73	0.69	0.75	0.72
H ₂ O	3.57	3.55	3.57	3.59	3.58	3.58	3.57	3.54	3.54
Total	98.77	98.64	98.58	98.68	98.32	98.82	99.53	99.07	98.52
Si	5.422	5.384	5.419	5.400	5.400	5.394	5.348	5.405	5.384
Al(IV)	2.578	2.636	2.581	2.600	2.600	2.606	2.652	2.595	2.616
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Ti	0.211	0.188	0.189	0.192	0.203	0.181	0.178	0.197	0.191
Al(IV)	0.721	0.795	0.802	0.811	0.785	0.828	0.739	0.750	0.771
Fe	2.887	2.746	2.787	2.731	2.784	2.695	2.925	2.773	2.768
Mn	0.010	0.009	0.008	0.008	0.009	0.009	0.008	0.007	0.008
Mg	2.009	2.048	2.004	2.034	2.018	2.056	1.985	2.085	2.058
	5.836	5.785	5.789	5.775	5.800	5.770	5.835	5.791	5.795
Ca	0.008	0.001	0.007	0.005	0.010	0.000	0.005	0.002	0.003
Na	0.018	0.026	0.023	0.022	0.023	0.027	0.038	0.033	0.032
K	1.727	1.866	1.786	1.821	1.768	1.848	1.839	1.833	1.833
	1.751	1.893	1.816	1.849	1.799	1.875	1.880	1.888	1.868
F	0.325	0.340	0.324	0.316	0.311	0.355	0.336	0.365	0.354
H	3.675	3.680	3.676	3.684	3.689	3.645	3.684	3.635	3.646
Fe/Mg	1.437	1.342	1.391	1.343	1.380	1.310	1.474	1.342	1.344

Sample 72a - biotite

min loc anal	RBT 1 A	RBT 1 B	RBT 1 5	RBT 1 23	RBT 1 AVG-8	RBT 2 25	RBT 2 32	RBT 2 AVG-8	CBT 1 2	CBT 1 7	CBT 1 20	CBT 1 AVG-9	CBT 2 26	CBT 2 31	CBT 2 AVG-5
P-T	A	B				C	D								
SiO ₂	35.35	35.75	34.24	34.88	35.05	34.89	34.91	34.89	35.24	35.17	35.05	35.07	35.46	34.67	34.83
TiO ₂	1.20	0.76	1.04	1.84	1.47	1.77	1.88	1.78	1.27	1.93	2.10	1.81	1.78	1.71	1.72
Al ₂ O ₃	19.35	18.58	18.63	18.75	18.39	18.11	18.07	18.26	18.30	17.58	18.13	18.20	17.94	18.03	18.08
FeO	23.38	23.82	24.39	25.97	24.47	25.24	25.82	25.26	24.51	24.82	24.86	24.72	25.63	24.95	25.26
MnO	0.09	0.11	0.10	0.08	0.09	0.09	0.07	0.08	0.09	0.09	0.10	0.08	0.08	0.08	0.09
MgO	7.09	7.25	7.39	5.89	7.16	6.92	6.67	6.57	7.31	7.25	6.95	7.07	6.86	6.88	6.83
CaO	0.03	0.01	0.43	0.03	0.12	0.07	0.05	0.05	0.38	0.01	0.03	0.10	0.02	0.01	0.03
Na ₂ O	0.02	0.00	0.00	0.01	0.01	0.02	0.03	0.02	0.01	0.03	0.02	0.04	0.04	0.03	0.03
K ₂ O	9.39	9.40	8.29	9.34	9.17	9.45	9.07	9.34	8.26	9.54	9.84	9.21	9.25	9.56	9.32
F	0.21	0.28	0.20	0.14	0.23	0.19	0.25	0.19	0.18	0.22	0.23	0.21	0.21	0.18	0.20
H ₂ O	3.80	3.75	3.73	3.81	3.76	3.78	3.75	3.77	3.78	3.77	3.78	3.78	3.79	3.78	3.76
Total	99.73	99.47	98.28	100.59	99.72	100.38	100.33	100.04	99.17	100.21	100.89	100.10	100.79	99.70	99.98
Si	5.441	5.528	5.372	5.595	5.429	5.402	5.407	5.415	5.484	5.445	5.402	5.421	5.480	5.405	5.412
Al(IV)	2.559	2.474	2.628	2.805	2.571	2.598	2.593	2.585	2.538	2.555	2.598	2.579	2.540	2.595	2.588
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Ti		0.088	0.123	0.214	0.171	0.208	0.217	0.208	0.148	0.225	0.243	0.210	0.208	0.201	0.201
Al(IV)	0.951	0.911	0.817	0.814	0.788	0.708	0.737	0.755	0.808	0.854	0.895	0.737	0.717	0.718	0.721
Fe	3.009	3.079	3.200	3.380	3.171	3.288	3.345	3.278	3.178	3.214	3.204	3.195	3.301	3.253	3.283
Mn	0.012	0.014	0.013	0.008	0.012	0.009	0.009	0.011	0.012	0.012	0.013	0.011	0.010	0.011	0.011
Mg	1.826	1.872	1.728	1.358	1.853	1.597	1.540	1.521	1.889	1.873	1.598	1.828	1.554	1.599	1.582
	5.736	5.785	5.881	5.753	5.793	5.791	5.817	5.772	5.835	5.777	5.752	5.781	5.788	5.781	5.789
Ca	0.005	0.002	0.072	0.005	0.020	0.012	0.008	0.008	0.083	0.002	0.005	0.017	0.003	0.002	0.005
Na	0.005	0.000	0.000	0.003	0.003	0.006	0.009	0.005	0.003	0.009	0.006	0.011	0.012	0.009	0.008
K	1.843	1.854	1.659	1.843	1.812	1.867	1.792	1.850	1.834	1.884	1.895	1.816	1.817	1.901	1.847
	1.854	1.855	1.732	1.851	1.835	1.884	1.810	1.883	1.700	1.895	1.908	1.843	1.832	1.912	1.851
F	0.102	0.137	0.099	0.088	0.114	0.093	0.122	0.095	0.088	0.108	0.112	0.105	0.102	0.089	0.095
H	3.898	3.863	3.901	3.932	3.888	3.907	3.878	3.905	3.912	3.892	3.888	3.895	3.898	3.911	3.904
Fe/Mg	1.851	1.841	1.852	2.474	1.918	2.047	2.172	2.156	1.882	1.921	2.007	1.983	2.124	2.036	2.075

Table 2 of Appendix 4. Selected biotite compositions based on 22 oxygen atoms

Sample 76 - biotite

min	FBT	FBT	FBT	FBT	FBT	FBT	FBT	FBT	FBT	CBT	CBT	CBT	CBT	CBT	CBT	CBT
loc	1-1	1-1	1-1	1-1	1	2-1	2-1	2-1	2	1-1	1-1	1-1	1	2-1	2-1	2
anal	7	13	15	16	AVG-9	21	22	27	AVG-7	5	11	17	AVG-6	24	28	AVG-4
P-T	B	A	C	E	G	I	J	K	L	S	D	F	H	M	N	O
SiO ₂	35.54	35.54	35.73	35.81	35.10	35.07	35.58	35.29	35.20	35.43	35.68	35.14	35.70	35.83	36.55	36.24
TiO ₂	1.54	2.00	1.88	1.80	1.82	1.32	1.21	1.70	1.42	2.07	1.75	1.83	1.85	1.82	1.83	1.81
Al ₂ O ₃	19.49	19.04	18.92	19.22	19.01	18.80	18.75	18.93	18.98	18.38	18.98	19.51	19.02	19.53	18.98	19.24
FeO	22.75	22.57	22.55	22.68	23.07	23.57	21.63	23.28	22.70	22.45	22.44	22.18	22.53	22.09	22.81	22.79
MnO	0.10	0.04	0.05	0.03	0.05	0.07	0.08	0.05	0.05	0.05	0.04	0.05	0.04	0.02	0.05	0.04
MgO	8.53	8.39	8.33	8.25	8.40	8.64	8.35	8.30	8.46	8.32	8.47	8.34	8.42	8.43	8.29	8.50
CaO	0.00	0.00	0.01	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
Na ₂ O	0.02	0.01	0.00	0.05	0.01	0.00	0.30	0.03	0.02	0.02	0.04	0.09	0.05	0.07	0.05	0.05
K ₂ O	9.71	9.67	8.92	9.61	9.47	9.62	9.72	9.14	9.49	9.55	9.86	9.54	9.53	9.52	9.44	9.32
F	0.28	0.31	0.26	0.24	0.28	0.34	0.25	0.31	0.30	0.31	0.26	0.31	0.29	0.28	0.28	0.29
H ₂ O	3.84	3.81	3.82	3.86	3.80	3.76	3.83	3.79	3.78	3.77	3.83	3.85	3.82	3.85	3.81	3.81
Total	101.55	101.11	100.25	101.33	100.78	100.68	100.18	100.54	100.14	100.05	100.91	101.58	101.01	101.19	100.82	100.82
Si	5.359	5.380	5.430	5.404	5.345	5.386	5.553	5.379	5.384	5.422	5.408	5.419	5.402	5.394	5.408	5.358
Al(v)	2.641	2.620	2.570	2.598	2.655	2.634	2.447	2.821	2.616	2.578	2.592	2.581	2.598	2.608	2.592	2.642
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Ti	0.175	0.228	0.215	0.204	0.208	0.152	0.138	0.195	0.164	0.238	0.201	0.208	0.210	0.208	0.187	0.183
Al(vi)	0.823	0.777	0.819	0.823	0.757	0.720	0.907	0.779	0.808	0.733	0.795	0.867	0.795	0.880	0.808	0.807
Fe	2.869	2.857	2.888	2.863	2.939	3.016	2.746	2.957	2.905	2.873	2.844	2.781	2.852	2.781	2.902	2.900
Mn	0.013	0.005	0.008	0.004	0.007	0.009	0.010	0.008	0.008	0.008	0.005	0.008	0.005	0.003	0.008	0.005
Mg	1.917	1.893	1.887	1.858	1.907	1.970	1.889	1.885	1.929	1.897	1.913	1.884	1.899	1.892	1.879	1.928
	5.797	5.759	5.784	5.749	5.819	5.867	5.891	5.833	5.811	5.749	5.758	5.725	5.782	5.742	5.782	5.823
Ca	0.000	0.000	0.002	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000	0.000
Na	0.008	0.003	0.000	0.015	0.003	0.000	0.000	0.009	0.008	0.008	0.012	0.028	0.018	0.020	0.015	0.015
K	1.858	1.857	1.729	1.850	1.840	1.878	1.882	1.777	1.852	1.884	1.888	1.825	1.841	1.829	1.832	1.808
	1.874	1.870	1.731	1.865	1.843	1.878	1.882	1.788	1.858	1.870	1.880	1.851	1.857	1.849	1.847	1.823
F	0.134	0.148	0.125	0.115	0.134	0.165	0.120	0.149	0.144	0.150	0.125	0.147	0.138	0.133	0.135	0.138
H	3.866	3.852	3.875	3.885	3.886	3.835	3.880	3.851	3.858	3.850	3.875	3.853	3.882	3.867	3.865	3.882
Fe/Mg	1.497	1.510	1.519	1.543	1.542	1.531	1.454	1.574	1.508	1.514	1.487	1.492	1.502	1.470	1.544	1.504

Sample 135 - biotite

min	FBT	FBT	FBT	FBT	FBT	FBT	CBT	CBT	CBT
loc	1-1	1-1	1-2	1-3	1-3	2	1-1	1-3	2
anal	1	6	10	11	13	19	5	12	16
P-T	A	B			D			C	
SiO ₂	36.49	36.42	36.55	36.81	36.80	36.38	35.95	36.25	36.32
TiO ₂	1.68	1.82	1.82	1.29	1.37	2.10	1.63	1.59	1.74
Al ₂ O ₃	20.30	19.28	20.19	19.17	19.57	19.92	19.70	19.71	19.92
FeO	19.82	21.78	19.95	22.49	21.47	19.93	20.64	20.72	19.88
MnO	0.17	0.13	0.07	0.20	0.07	0.05	0.16	0.04	0.03
MgO	8.68	8.75	8.55	9.08	9.45	9.42	9.43	9.93	9.10
CaO	0.19	0.10	0.26	0.07	0.22	0.08	0.03	0.08	0.14
Na ₂ O	0.07	0.04	0.03	0.02	0.14	0.08	0.09	0.11	0.08
K ₂ O	9.01	8.98	7.85	8.92	8.61	9.17	9.29	8.91	7.89
F	0.42	0.48	0.32	0.45	0.46	0.35	0.38	0.53	0.51
H ₂ O	3.81	3.78	3.84	3.76	3.82	3.88	3.83	3.77	3.73
Total	100.26	101.13	98.14	100.86	101.57	101.01	100.79	101.15	98.67
Si	5.458	5.454	5.485	5.405	5.482	5.412	5.392	5.397	5.482
Al(v)	2.542	2.546	2.505	2.595	2.538	2.598	2.608	2.603	2.518
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Ti	0.189	0.205	0.208	0.148	0.153	0.235	0.184	0.178	0.198
Al(vi)	1.038	0.857	1.073	0.816	0.888	0.905	0.875	0.857	1.025
Fe	2.480	2.728	2.508	2.839	2.685	2.480	2.589	2.580	2.482
Mn	0.022	0.016	0.009	0.028	0.009	0.008	0.020	0.005	0.004
Mg	1.935	1.953	1.916	2.043	2.080	2.089	2.108	2.203	2.047
	5.883	5.780	5.711	5.870	5.803	5.715	5.778	5.823	5.755
Ca	0.030	0.018	0.042	0.011	0.035	0.010	0.005	0.010	0.023
Na	0.020	0.012	0.009	0.008	0.040	0.023	0.026	0.032	0.023
K	1.719	1.716	1.335	1.718	1.630	1.740	1.778	1.693	1.519
	1.770	1.743	1.558	1.735	1.708	1.773	1.809	1.734	1.585
F	0.199	0.227	0.152	0.215	0.216	0.185	0.171	0.250	0.243
H	3.801	3.773	3.848	3.785	3.784	3.835	3.829	3.750	3.757
Fe/Mg	1.281	1.397	1.309	1.390	1.275	1.187	1.228	1.171	1.212

- [Kgd] Biotite granodiorite
- [Kgr] Muscovite biotite granite
- [Kmgr] Megacrystic K-feldspar granite
- [Kiqd] Biotite hornblende quartz diorite
- [Kqm] Biotite hornblende quartz monzonite

MISSISSIPPIAN TO PENNSYLVANIAN OR PERMIAN

UPPER MISSISSIPPIAN TO PENNSYLVANIAN OR PERMIAN

MILFORD GROUP

- [uMMC] Grey and white limestone
- [uMMcg] Conglomerate

DEVONIAN TO CRETACEOUS

- [?d] Hornblende diorite, leucogranite, and leucopegmatite
- [Inu] Undifferentiated intrusives

DEVONIAN

LATE DEVONIAN

- [Dgn] Hornblende biotite orthogneiss

CAMBRIAN AND ORDOVICIAN

LOWER CAMBRIAN TO LOWER ORDOVICIAN

LARDEAU GROUP

- [LPLu] UNDIFFERENTIATED LARDEAU GROUP

BROADVIEW FORMATION

- [PBw] Sempelitic and pelitic schist, micaceous quartzite, quartzite

- [PBwc] Limestone, grey phyllitic limestone, and grey phyllite

JOWETT FORMATION

- [PJt] Amphibolite, marble, calc-silicate schist, and phyllite

INDEX FORMATION

- [PIN] Graphitic phyllite, marble, semipelitic schist, amphibolite, minor quartzite and micaceous quartzite

- [PIng] Quartz grit, minor gritty phyllite

LADE PEAK FORMATION

- [IPLP] Graphitic marble and graphitic siliceous siltstone

CAMBRIAN

LOWER CAMBRIAN

BADSHOT FORMATION

- [CBd] Grey and white marble

PROTEROZOIC AND/OR CAMBRIAN

UPPER PROTEROZOIC (WINDERMERE) AND/OR LOWER CAMBRIAN

NANTU GROUP

minor quartzite and micaceous quartzite

[PIng] Quartz grit, minor gritty phyllite

LADE PEAK FORMATION

Table 3 of Appendix 4 Selected muscovite compositions based on 22 oxygen atoms

Sample 13b - muscovite

min	RMU	RMU	RMU	RMU	RMU	CMU	CMU	CMU	CMU	CMU
loc	1-3	2-3	3-2	4-2		1-3	2-3	3-2	4-2	
anal	5	3	7	11	AVG-6	6	4	8	10	AVG-6
P-T	B					A				
SiO ₂	47.07	46.28	46.62	46.81	46.38	46.78	46.40	46.34	46.29	46.51
TiO ₂	0.40	0.33	0.42	0.46	0.37	0.40	0.39	0.37	0.30	0.36
Al ₂ O ₃	37.07	36.93	37.45	36.50	36.89	36.97	36.59	36.83	36.37	36.63
FeO	1.43	1.21	1.26	1.19	1.23	1.12	1.03	1.09	1.04	1.06
MnO	0.01	0.00	0.02	0.01	0.01	0.01	0.00	0.01	0.02	0.01
MgO	0.46	0.53	0.48	0.54	0.49	0.47	0.52	0.46	0.55	0.50
CaO	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.00
Na ₂ O	1.53	1.46	1.53	1.48	1.50	1.51	1.54	1.57	1.52	1.56
K ₂ O	8.60	8.44	8.87	8.32	8.49	8.85	8.55	8.99	8.66	8.66
F	0.06	0.14	0.06	0.04	0.08	0.06	0.06	0.04	0.09	0.07
H ₂ O	4.58	4.48	4.58	4.54	4.52	4.56	4.51	4.54	4.48	4.52
Total	101.16	99.70	101.23	99.88	99.89	100.68	99.54	100.20	99.24	99.81
Si	6.122	6.098	6.068	6.151	6.104	6.116	6.125	6.097	6.133	6.127
Al(IV)	1.878	1.902	1.932	1.849	1.896	1.884	1.875	1.903	1.867	1.873
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al(IV)	3.808	3.833	3.814	3.805	3.825	3.813	3.818	3.808	3.813	3.815
Ti	0.039	0.033	0.041	0.045	0.037	0.039	0.039	0.037	0.030	0.036
Fe	0.156	0.133	0.137	0.131	0.135	0.122	0.114	0.120	0.115	0.116
Mn	0.001	0.000	0.002	0.001	0.001	0.001	0.000	0.001	0.002	0.001
Mg	0.089	0.104	0.093	0.108	0.098	0.092	0.102	0.089	0.109	0.099
	4.080	4.103	4.088	4.088	4.084	4.087	4.073	4.066	4.089	4.067
Ca	0.000	0.000	0.000	0.003	0.001	0.000	0.000	0.000	0.000	0.000
Na	0.388	0.378	0.388	0.377	0.384	0.383	0.394	0.401	0.391	0.388
K	1.427	1.419	1.473	1.395	1.425	1.476	1.440	1.509	1.454	1.456
	1.813	1.797	1.859	1.775	1.809	1.859	1.834	1.910	1.854	1.854
F	0.025	0.058	0.025	0.017	0.032	0.021	0.025	0.017	0.038	0.029
H	3.975	3.942	3.975	3.983	3.968	3.979	3.975	3.983	3.962	3.971
X_K	0.787	0.790	0.792	0.787	0.788	0.794	0.785	0.790	0.789	0.785

Sample 13d1 - muscovite

min	RMU	RMU
loc	2-1	2-1
anal	1	2
P-T	A	
SiO ₂	47.24	46.80
TiO ₂	0.54	0.43
Al ₂ O ₃	36.28	35.16
FeO	1.10	1.05
MnO	0.02	0.04
MgO	0.58	0.74
CaO	0.02	0.02
Na ₂ O	1.07	1.11
K ₂ O	9.01	9.10
F	0.09	0.40
H ₂ O	4.26	4.31
Total	100.12	98.80
Si	6.188	6.231
Al(IV)	1.812	1.769
	8.000	8.000
Al(IV)	3.788	3.749
Ti	0.053	0.043
Fe	0.120	0.117
Mn	0.002	0.005
Mg	0.113	0.147
	4.077	4.060
Ca	0.003	0.003
Na	0.272	0.287
K	1.487	1.520
	1.782	1.810
F	0.037	0.168
H	3.983	3.832
X_K	0.945	0.841

Table 3 of Appendix 4 Selected muscovite compositions based on 22

oxygen atoms

Sample 63 - muscovite

min	RMU	RMU	RMU	RMU	RMU	CMU	CMU	CMU	CMU
loc	1	1	1	1	1	1	1	1	1
anal	3	4	7	13	AVG-9	8	12	15	AVG-5
P-T	A								
SiO ₂	45.97	46.50	46.63	46.49	46.24	46.06	46.56	45.18	46.00
TiO ₂	0.35	0.25	0.56	0.36	0.37	0.59	0.29	0.42	0.40
Al ₂ O ₃	34.65	36.08	33.95	35.23	34.95	34.80	36.40	33.31	35.12
FeO	2.64	1.27	2.03	1.45	2.09	1.63	1.24	4.52	2.32
MnO	0.00	0.01	0.01	0.00	0.01	0.02	0.01	0.00	0.01
MgO	0.79	0.69	1.08	0.83	0.82	0.90	0.61	0.87	0.81
CaO	0.02	0.03	0.00	0.00	0.02	0.00	0.00	0.01	0.01
Na ₂ O	0.59	0.56	0.46	0.73	0.60	0.74	0.71	0.68	0.68
K ₂ O	9.72	10.34	10.73	10.51	10.24	10.44	10.38	10.00	10.38
F	0.14	0.24	0.16	0.20	0.17	0.12	0.09	0.19	0.12
H ₂ O	4.40	4.42	4.41	4.42	4.41	4.43	4.52	4.32	4.44
Total	99.15	100.28	99.88	100.04	99.76	99.62	100.81	99.34	100.15
Si	6.170	6.143	6.227	6.174	6.167	6.155	6.122	6.136	6.129
Al(m)	1.830	1.867	1.773	1.828	1.833	1.845	1.878	1.865	1.871
	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al(v)	3.652	3.781	3.572	3.688	3.682	3.638	3.778	3.467	3.644
Ti	0.035	0.025	0.056	0.036	0.037	0.059	0.029	0.043	0.040
Fe	0.296	0.140	0.227	0.161	0.234	0.182	0.136	0.513	0.260
Mn	0.000	0.001	0.001	0.000	0.001	0.002	0.001	0.000	0.001
Mg	0.158	0.136	0.215	0.164	0.163	0.179	0.120	0.178	0.180
	4.142	4.063	4.071	4.049	4.067	4.059	4.081	4.199	4.106
Ca	0.003	0.004	0.000	0.000	0.002	0.000	0.000	0.001	0.001
Na	0.154	0.169	0.119	0.188	0.155	0.192	0.181	0.179	0.171
K	1.664	1.743	1.528	1.781	1.742	1.780	1.741	1.732	1.764
	1.821	1.916	1.947	1.989	1.899	1.972	1.922	1.913	1.936
F	0.059	0.100	0.068	0.084	0.072	0.051	0.037	0.082	0.051
H	3.941	3.900	3.932	3.916	3.928	3.949	3.963	3.918	3.949
X_K	0.916	0.912	0.939	0.905	0.918	0.903	0.906	0.906	0.912

Sample 135 - muscovite

min	RMU	RMU	CMU	CMU	CMU
loc	1-1	2	1-3	1-3	2
anal	6	3	4	5	2
P-T	A		8		
SiO ₂	47.70	47.43	46.68	46.75	48.12
TiO ₂	0.35	0.30	0.29	0.21	0.18
Al ₂ O ₃	36.63	36.46	35.64	36.96	36.18
FeO	1.51	1.17	1.20	1.16	1.20
MnO	0.00	0.01	0.00	0.01	0.01
MgO	0.76	0.67	0.76	0.54	0.83
CaO	0.00	0.03	0.00	0.01	0.01
Na ₂ O	0.92	1.12	1.07	1.27	1.01
K ₂ O	9.49	9.04	9.37	9.05	9.74
F	0.06	0.07	0.14	0.04	0.06
H ₂ O	4.60	4.56	4.46	4.56	4.60
Total	101.97	100.80	99.48	100.52	101.89
Si	6.172	6.187	6.188	6.123	6.230
Al(m)	1.828	1.813	1.812	1.877	1.770
	8.000	8.000	8.000	8.000	8.000
Al(v)	3.758	3.794	3.757	3.830	3.751
Ti	0.034	0.029	0.029	0.021	0.018
Fe	0.163	0.128	0.133	0.127	0.130
Mn	0.000	0.001	0.000	0.001	0.001
Mg	0.147	0.130	0.150	0.106	0.180
	4.102	4.082	4.089	4.084	4.080
Ca	0.000	0.004	0.000	0.001	0.001
Na	0.231	0.283	0.275	0.323	0.254
K	1.567	1.505	1.585	1.512	1.609
	1.797	1.792	1.880	1.836	1.864
F	0.025	0.029	0.069	0.017	0.025
H	3.975	3.971	3.941	3.983	3.975
X_K	0.872	0.842	0.862	0.824	0.864

Table 4 of Appendix 4 Selected plagioclase compositions based on 8 oxygen atoms

Sample 13b - plagioclase rim

min	RPL	RPL	RPL	RPL	RPL	RPL	RPL	RPL	RPL	RPL	CPL	CPL	CPL	CPL	CPL
loc	2-2-A	2-2-B	1-4-A	1-4-B	3-4(1A)	3-4(1B)	3-4(2A)	3-4(2B)	4-A	4-B	2-2	1-4	3-4(1)	3-4(2)	4
anal	86	101	125	134	146	173	174	202	206	220	99	128	156	180	214
P-T	C	D	A	B		F	E								
SiO ₂	64.33	64.78	65.13	64.82	63.99	65.77	66.07	65.73	63.93	63.88	64.67	65.40	63.31	63.86	64.22
Al ₂ O ₃	23.04	23.27	22.67	22.88	22.85	22.42	22.43	21.24	21.96	22.07	22.78	22.91	23.34	23.39	23.06
FeO	0.10	0.13	0.17	0.16	0.07	0.43	0.08	0.14	0.12	0.17	0.06	0.12	0.02	0.03	0.02
CaO	3.99	4.00	3.43	3.71	3.79	2.90	2.92	2.38	3.41	3.31	3.43	3.44	4.57	4.56	3.42
Na ₂ O	9.12	9.32	9.39	9.43	9.25	9.84	7.81	10.11	9.62	9.73	9.53	9.53	8.99	9.05	9.49
K ₂ O	0.10	0.07	0.07	0.08	0.06	0.07	0.71	0.06	0.06	0.08	0.09	0.07	0.08	0.08	0.08
BaO	0.03	0.00	0.00	0.00	0.01	0.00	0.21	0.03	0.02	0.00	0.02	0.00	0.00	0.00	0.00
Total	100.71	101.57	100.88	101.09	100.01	101.43	100.23	99.88	99.11	99.24	100.58	101.47	100.29	100.75	100.28
Si	2.817	2.814	2.842	2.827	2.820	2.858	2.887	2.897	2.844	2.840	2.833	2.838	2.790	2.792	2.821
Al	1.189	1.191	1.166	1.176	1.187	1.148	1.155	1.103	1.152	1.158	1.175	1.172	1.212	1.209	1.193
Fe	0.004	0.005	0.006	0.006	0.003	0.016	0.003	0.005	0.004	0.006	0.002	0.004	0.001	0.001	0.001
Ca	0.167	0.168	0.160	0.173	0.179	0.136	0.137	0.112	0.163	0.158	0.161	0.160	0.216	0.214	0.161
Na	0.775	0.785	0.795	0.798	0.791	0.829	0.682	0.884	0.830	0.836	0.810	0.802	0.768	0.770	0.808
K	0.006	0.004	0.004	0.006	0.003	0.004	0.040	0.003	0.003	0.005	0.005	0.004	0.003	0.003	0.004
Ba	0.001	0.000	0.000	0.000	0.000	0.000	0.004	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000
	4.978	4.985	4.974	4.986	4.983	4.987	4.886	4.985	4.998	5.004	4.987	4.979	4.990	4.990	4.989
an	0.194	0.191	0.167	0.178	0.184	0.139	0.163	0.115	0.163	0.158	0.165	0.166	0.219	0.217	0.165
ab	0.801	0.806	0.829	0.817	0.813	0.857	0.760	0.882	0.834	0.838	0.830	0.830	0.778	0.780	0.830
or	0.006	0.004	0.004	0.006	0.003	0.004	0.047	0.003	0.003	0.005	0.005	0.004	0.003	0.003	0.005
an/ab	0.242	0.237	0.202	0.000	0.226	1.175	0.207	0.000	0.196	0.188	0.199	0.199	0.281	0.278	0.199

Sample 13d1 - matrix plagioclase

min	RPL	RPL	RPL	RPL	RPL	RPL	RPL	CPL	CPL	CPL
loc	1-2-A	1-2-B	2-2(1A)	2-2(1A)	2-2(1B)	2-2(2B)	2-2(2B)	1-2	2-2(1)	2-2(2)
anal	232	241	254	255	263	266	271	235	258	277
P-T	E	G	A		C	D		F	B	
SiO ₂	60.50	60.34	60.25	61.20	60.99	60.76	59.88	61.27	60.75	59.32
Al ₂ O ₃	24.87	24.78	25.15	24.62	24.87	25.56	26.12	24.72	24.71	24.45
FeO	0.07	0.09	0.81	0.37	0.13	0.37	0.22	0.07	0.22	0.30
CaO	6.39	6.56	6.03	6.03	6.40	6.80	7.11	6.24	6.07	6.34
Na ₂ O	7.92	7.77	7.17	7.82	7.88	7.53	7.52	7.93	7.99	7.72
K ₂ O	0.04	0.07	0.25	0.08	0.06	0.06	0.06	0.07	0.07	0.08
BaO	0.00	0.02	0.03	0.00	0.01	0.00	0.01	0.00	0.00	0.02
Total	99.79	99.63	99.69	100.10	100.13	101.19	100.92	100.30	99.81	98.23
Si	2.696	2.694	2.689	2.715	2.706	2.673	2.645	2.712	2.706	2.689
Al	1.306	1.304	1.323	1.287	1.300	1.332	1.380	1.290	1.297	1.307
Fe	0.003	0.003	0.030	0.014	0.005	0.014	0.008	0.003	0.008	0.011
Ca	0.306	0.314	0.288	0.287	0.304	0.321	0.337	0.296	0.290	0.308
Na	0.684	0.673	0.621	0.673	0.680	0.642	0.644	0.681	0.680	0.679
K	0.002	0.004	0.014	0.003	0.003	0.003	0.003	0.004	0.004	0.005
Ba	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000
	4.995	4.982	4.987	4.979	4.977	4.984	4.998	4.985	4.994	4.989
an	0.308	0.317	0.312	0.298	0.314	0.332	0.342	0.302	0.294	0.311
ab	0.690	0.679	0.672	0.699	0.683	0.665	0.655	0.694	0.702	0.685
or	0.002	0.004	0.016	0.004	0.003	0.003	0.003	0.004	0.004	0.005
an/ab	0.446	0.468	0.465	0.426	0.460	0.489	0.522	0.435	0.420	0.454

Table 4 of Appendix 4. Selected plagioclase compositions based on 8 oxygen atoms

Sample 13d1 - inclusion plagioclase

min	RPL	RPL	RPL	RPL	RPL	RPL	RPL	CPL	CPL	CPL	CPL	CPL
loc	2A-A	2C-A	2E-1A	2F-A	2F-B	2G-A	2G-B	2A	2C	2E-1	2F	2G
en#	276	301	325	352	362	364	370	281	300	324	357	366
P-T												
SiO ₂	59.80	59.83	58.47	61.11	61.45	54.47	50.83	59.30	60.68	55.43	60.75	50.30
Al ₂ O ₃	25.74	25.98	25.90	25.25	25.06	28.54	31.71	25.68	25.37	28.67	24.72	31.72
FeO	0.46	0.56	2.10	0.47	0.51	0.64	0.53	0.55	0.43	0.57	0.38	0.50
CaO	7.38	7.34	7.42	6.28	6.69	11.62	13.56	7.26	6.58	10.54	6.41	14.45
Na ₂ O	7.41	7.25	6.63	7.85	7.71	4.83	3.68	7.50	7.83	5.52	7.93	3.28
K ₂ O	0.05	0.05	0.04	0.05	0.04	0.03	0.01	0.04	0.06	0.02	0.06	0.02
BaO	0.00	0.00	0.03	0.00	0.01	0.02	0.02	0.00	0.00	0.00	0.00	0.00
Total	100.84	101.01	100.59	101.01	101.46	100.25	100.44	100.33	100.96	100.75	100.22	100.27
Si	2.649	2.645	2.615	2.692	2.697	2.459	2.309	2.642	2.678	2.482	2.699	2.289
Al	1.344	1.354	1.365	1.311	1.296	1.519	1.694	1.349	1.320	1.513	1.294	1.702
Fe	0.017	0.021	0.079	0.017	0.019	0.024	0.020	0.020	0.016	0.021	0.013	0.019
Ca	0.350	0.348	0.356	0.298	0.315	0.562	0.659	0.347	0.311	0.508	0.306	0.706
Na	0.636	0.621	0.575	0.671	0.656	0.432	0.329	0.648	0.670	0.479	0.683	0.290
K	0.003	0.003	0.002	0.003	0.002	0.002	0.001	0.002	0.003	0.001	0.003	0.001
Ba	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
	4.999	4.991	4.991	4.990	4.985	4.988	5.006	5.009	4.999	5.002	4.997	5.006
an	0.354	0.358	0.381	0.308	0.323	0.565	0.670	0.348	0.316	0.513	0.308	0.708
ab	0.643	0.639	0.616	0.691	0.674	0.434	0.329	0.650	0.681	0.486	0.689	0.291
or	0.003	0.003	0.002	0.003	0.002	0.002	0.001	0.002	0.003	0.001	0.003	0.001
an/ab	0.550	0.559	0.618	0.442	0.479	1.302	2.036	0.535	0.464	1.055	0.447	2.434

Sample 63 - matrix plagioclase

min	RPL	RPL	RPL	RPL	RPL	RPL	RPL	CPL	CPL
loc	1(1)	1(1)-A	1(1)-B	1(2)-A	1(2)-B	1(2)-B	1(2)-B	1(1)	1(2)
en#	4	213	236	241	264	266	267	221	251
P-T	A	E	F	G	B	C	D		
SiO ₂	55.79	56.58	55.82	57.18	54.84	55.40	57.24	61.74	60.25
Al ₂ O ₃	26.90	26.70	26.48	26.14	27.56	27.32	26.17	23.21	23.79
FeO	0.20	0.04	0.12	0.06	0.09	0.08	0.10	0.02	0.01
CaO	8.97	8.58	8.75	8.79	9.93	9.46	8.22	5.36	6.06
Na ₂ O	6.42	6.70	6.49	6.54	5.91	6.23	6.80	8.46	8.08
K ₂ O	0.05	0.05	0.04	0.04	0.04	0.06	0.08	0.07	0.07
BaO	0.03	0.00	0.00	0.00	0.00	0.01	0.04	0.00	0.00
Total	98.36	98.65	97.50	98.72	98.36	98.55	98.65	98.86	98.26
Si	2.548	2.570	2.559	2.593	2.509	2.527	2.597	2.767	2.724
Al	1.448	1.430	1.436	1.398	1.486	1.469	1.400	1.226	1.268
Fe	0.008	0.002	0.005	0.002	0.003	0.003	0.004	0.001	0.000
Ca	0.439	0.418	0.431	0.427	0.487	0.462	0.400	0.257	0.294
Na	0.569	0.590	0.579	0.575	0.524	0.551	0.598	0.736	0.708
K	0.003	0.003	0.002	0.002	0.002	0.003	0.006	0.004	0.004
Ba	0.001	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.000
sum cat	5.014	5.012	5.013	4.987	5.012	5.016	5.004	4.990	4.998
an	0.434	0.413	0.426	0.425	0.480	0.455	0.399	0.258	0.292
ab	0.563	0.584	0.572	0.573	0.517	0.542	0.597	0.738	0.704
or	0.003	0.003	0.002	0.002	0.002	0.003	0.006	0.004	0.004
an/ab	0.772	0.708	0.745	0.743	0.928	0.839	0.668	0.360	0.414

Table 4 of Appendix 4 Selected plagioclase compositions based on 8 oxygen atoms

Sample 63 - inclusion plagioclase

min loc an# P-T	RPL 1B-A 289	RPL 1B-B 305	RPL 1C-A 309	RPL 1C-B 332	RPL 1D-A 334	RPL 1D-B 359	CPL 1B 295	CPL 1C 317	CPL 1D 342
SiO ₂	57.83	57.61	58.26	57.48	56.79	56.75	60.51	60.24	60.24
Al ₂ O ₃	25.15	25.16	25.07	25.17	25.80	25.48	23.13	24.36	23.88
FeO	0.21	0.18	0.30	0.52	0.33	0.25	0.03	0.06	0.04
CaO	7.20	7.62	7.41	7.95	8.08	7.89	5.11	5.82	5.80
Na ₂ O	7.48	7.24	7.41	7.09	6.93	7.07	8.69	8.27	8.48
K ₂ O	0.04	0.06	0.06	0.04	0.03	0.03	0.06	0.06	0.08
BaO	0.01	0.00	0.00	0.06	0.00	0.00	0.00	0.01	0.00
Total	97.72	97.86	98.50	98.30	97.98	97.47	97.53	98.81	98.32
Si	2.638	2.633	2.644	2.623	2.598	2.608	2.752	2.709	2.723
Al	1.358	1.355	1.341	1.354	1.391	1.380	1.240	1.291	1.272
Fe	0.008	0.007	0.011	0.020	0.013	0.010	0.001	0.002	0.002
Ca	0.363	0.373	0.380	0.389	0.396	0.389	0.249	0.280	0.271
Na	0.664	0.642	0.662	0.627	0.615	0.630	0.767	0.721	0.743
K	0.002	0.003	0.003	0.002	0.002	0.002	0.003	0.003	0.006
Ba	0.000	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000
sum cat	5.019	5.012	5.013	5.015	5.015	5.018	5.013	5.008	5.015
an	0.348	0.387	0.366	0.382	0.391	0.381	0.244	0.279	0.286
ab	0.651	0.630	0.642	0.618	0.607	0.618	0.752	0.718	0.729
or	0.002	0.003	0.003	0.002	0.002	0.002	0.003	0.003	0.006
an/ab	0.532	0.582	0.563	0.620	0.644	0.617	0.325	0.389	0.386

Sample 135 - plagioclase

min loc an# P-T	RPL 1-2-A 422	RPL 1-2-A 423	RPL 1-2-B 433	RPL 1-2-B 435	RPL 2 437	RPL 2 438	RPL 2-A 443	RPL 2-A 444	RPL 2-D 470	CPL 1-2 428	CPL 2 462	CPL 2 464
SiO ₂	60.62	60.02	60.66	59.48	61.40	60.98	60.86	61.39	60.51	59.15	58.46	59.18
Al ₂ O ₃	24.73	24.57	25.20	25.98	24.39	24.47	24.51	24.94	24.71	25.50	26.01	25.97
FeO	0.39	0.29	0.17	0.25	0.26	0.23	0.24	0.18	0.20	0.08	0.06	0.06
CaO	6.40	5.94	6.21	7.37	5.85	5.93	6.01	5.84	6.16	7.04	7.79	7.74
Na ₂ O	7.73	8.17	7.84	7.35	8.26	8.22	8.15	8.25	7.92	7.52	6.97	7.08
K ₂ O	0.08	0.07	0.07	0.08	0.09	0.09	0.07	0.07	0.07	0.06	0.06	0.04
BaO	0.01	0.00	0.04	0.00	0.00	0.00	0.03	0.04	0.01	0.00	0.00	0.04
Total	99.98	99.06	100.28	100.51	100.25	99.92	99.86	100.71	99.58	99.34	99.33	100.11
Si	2.699	2.697	2.689	2.641	2.722	2.713	2.710	2.708	2.701	2.864	2.826	2.837
Al	1.298	1.301	1.317	1.380	1.274	1.283	1.287	1.297	1.300	1.348	1.377	1.364
Fe	0.015	0.011	0.008	0.009	0.010	0.009	0.009	0.007	0.007	0.003	0.002	0.002
Ca	0.306	0.288	0.295	0.351	0.278	0.283	0.287	0.278	0.295	0.338	0.375	0.370
Na	0.667	0.712	0.683	0.633	0.710	0.709	0.704	0.706	0.686	0.854	0.807	0.812
K	0.005	0.004	0.004	0.005	0.005	0.005	0.004	0.004	0.004	0.003	0.003	0.002
Ba	0.000	0.000	0.001	0.000	0.000	0.000	0.001	0.001	0.000	0.000	0.000	0.001
sum cat	4.988	5.011	4.995	4.998	4.999	5.002	5.001	4.998	4.993	5.001	4.980	4.988
an	0.312	0.285	0.301	0.356	0.280	0.284	0.288	0.280	0.299	0.340	0.381	0.378
ab	0.683	0.711	0.696	0.641	0.715	0.711	0.708	0.718	0.697	0.657	0.616	0.622
or	0.006	0.004	0.004	0.006	0.006	0.006	0.004	0.004	0.004	0.003	0.003	0.002
an/ab	0.457	0.402	0.432	0.554	0.391	0.399	0.407	0.391	0.430	0.517	0.618	0.604

normal fau

minor quartzite and micaceous quartzite

[Png] Quartz grit, minor gritty phyllite

LADY PEAK FORMATION

[LPL] Graphitic marble and graphitic siliceous siltstone

CAMBRIAN

LOWER CAMBRIAN

BADSHOT FORMATION

[18Ba] Grey and white marble

PROTEROZOIC AND/OR CAMBRIAN

UPPER PROTEROZOIC (WINDERMERE) AND/OR LOWER CAMBRIAN

HAMILL GROUP

[18Ha] Quartzite, micaceous quartzite, semipelitic and pelitic schist, phyllite, and minor limestone

AGE UNCERTAIN

[msu] Undifferentiated metasediment

[msa] Amphibolite

[mscs] Calc-silicate schist

[msm] Marble

[mops] Pelitic schist

[msq] Quartzite

GEOLOGICAL SYMBOLS

stratigraphic and intrusive contacts (exposed, inferred, uncertain)

⊙ fossil locality (archaeocyathids in Badshot Formation)

AL JA' line of cross-section A to A'

MAJOR STRUCTURES

normal fault (exposed, inferred) peg on hanging wall

fault (exposed, inferred) sense of displacement unknown

axial surface trace of first phase anticline

axial surface trace of first phase syncline

normal fault (exposed, inferred) peg on hanging wall

Cross-sections A,B,C,D (see Map 1 for locations)

C?gd granodiorite (likely Cretaceous)
 BW Broadview Formation
 Dogn Devonian orthogneiss
 Jt Jowett Formation
 In Index Formation
 Inh heterogeneous Index formation
 Ins siliceous Index Formation
 LP Lade Peak Formation
 Bd Badshot Formation
 Ha Hamill Group

AA Akolkolex antiform
 ACA Albert Canyon antiform
 ARF Akolkolex River fault
 DCS Drimmie Creek syncline
 GPA Ghost Peak antiform
 LSCF lower Standfast Creek fault
 MCA Maunder Creek antiform
 MCS Mount Cartier syncline
 SCF Standfast Creek fault
 USCF upper Standfast Creek fault

— stratigraphic contact
 — fault
 — orientation of S2
 - - - erosional surface

- AA Akalkolex anticline
- ACA Albert Canyon
- ARF Akalkolex River
- CRF Columbia River
- DCS Drimlie Creek
- GPA Ghost Peak ant.
- MCA Maunder Creek
- MCS Mount Cartier
- SCF Standfast Creek
- TCA Twin Creek anticline
- TCFS Twin Creek fault

- AA
- ACA
- ARF
- CRF
- DCS
- GPA
- MCA
- MCS
- SCF
- TCA
- TCFS

MINDR STRUCTURE

- 15-17 strike and dip, trend and plunge
- 15-15 trend and plunge, second phase
- 15-17 strike and dip, trend and plunge
- 15- trend and plunge

- 15-17
- 15-15
- 15-15
- 15-17
- 15-17

E A N

- AA Akolkolex anticline
- ACA Albert Canyon anticline
- ARF Akolkolex River fault
- CRF Columbia River fault
- DCS Drimlie Creek syncline
- GPA Ghost Peak anticline
- MCA Maunder Creek antiform
- MCS Mount Cartier syncline
- SCF Standfast Creek fault
- TCA Twin Creek antiform
- TCFS Twin Creek fault system

MINDR STRUCTURES

- 15 strike and dip of second phase axial planar cleavage or foliation
- 15 trend and plunge of second phase fold axis
- 15 trend and plunge of S-, Z-, and M-shaped (looking down plunge) second phase folds
- 15 strike and dip of third phase axial planar cleavage or foliation
- 15 trend and plunge of third phase fold axis
- 15 trend and plunge of mineral lineation

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1. The map is a reproduction of the original map and is not a new map.

2. The map is a reproduction of the original map and is not a new map.

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4. The map is a reproduction of the original map and is not a new map.

5. The map is a reproduction of the original map and is not a new map.

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7. The map is a reproduction of the original map and is not a new map.

8. The map is a reproduction of the original map and is not a new map.

9. The map is a reproduction of the original map and is not a new map.

10. The map is a reproduction of the original map and is not a new map.

THE UNITED STATES OF AMERICA

1960

QUANTITATIVE OF SOIL MATERIAL

1960

4 of/de 4

PM-1 3½" x 4" PHOTOGRAPHIC MICROCOPY TARGET
NBS 1010a ANSI/ISO #2 EQUIVALENT

PRECISIONSM RESOLUTION TARGETS

PIONEERS IN METHYLENE BLUE TESTING SINCE 1974

MICRO

1000 COUNTY ROAD 8, BURNING WOOD, WV 26037 USA
TEL: 812 486 7827 FAX: 812 486 7887 TLA: 3100000000

MAP 2. METAMORPHIC

CLAUDE

BASE OF THE

PARTS OF THE

MOUNT REVELSTOCK

Note: includes mapping by
(modified in part) by Thompson
Hakkinen (unpublished data)
Thompson (1980), Reed (unpubl.)

STRATIGRAPHIC

CRETACEOUS

LATE CRETACEOUS

[Klg] Muscovite biotite leu

MIDDLE CRETACEOUS

[K7d] Biotite hornblende di

[K7gd] Biotite granodiorite

[K7gr] Muscovite biotite gr

[Kng] Megacrystic K-felds

[K7gd] Biotite hornblende q

[Kgn] Biotite hornblende q

MISSISSIPPIAN TO PENN

MAP 2. METAMORPHIC AND SAMPLE LOCATION MAP

CLACHNACUBAINN TERRANE &

BASE OF THE OVERLYING ILLECILLEWAET SLICE

PARTS OF CAMBORNE (82 K/13), REVELSTOKE (82 L/16),

MOUNT REVELSTOKE (82 M/1), & ILLECILLEWAET (82 N/4)

MAP AREAS

BRITISH COLUMBIA

Note: includes mapping by J.L. Crowley (1990, 1991) and a compilation of mapping (modified in part) by Thompson (1972), Zwanzig (1973), Read and Wheeler (1976), Hakkinen (unpublished data, 1976, 1977), Sears (1979), Read (1980), Read and Thompson (1980), Read (unpublished data, 1982), and Price (1986).

STRATIGRAPHIC COLUMN

CRETACEOUS

LATE CRETACEOUS

[Klg] Muscovite biotite leucogranite and pegmatite

MIDDLE CRETACEOUS

[Ktd] Biotite hornblende diorite

[Ktd] Biotite granodiorite

[Ktg] Muscovite biotite granite

[Kng] Megacrystic K-feldspar granite

[Ktd] Biotite hornblende quartz diorite

[Kqn] Biotite hornblende quartz monzonite

MISSISSIPPIAN TO PENNSYLVANIAN OR PERMIAN

- [Klg] Muscovite biotite granite
- [Kng] Megacrystic K-feldspar granite
- [K7qd] Biotite hornblende quartz diorite
- [Kqn] Biotite hornblende quartz monzonite

MISSISSIPPIAN TO PENNSYLVANIAN OR PERMIAN

UPPER MISSISSIPPIAN TO PENNSYLVANIAN OR PERMIAN

MILFORD GROUP

- [UMMc] Grey and white limestone
- [UMMcg] Conglomerate

DEVONIAN TO CRETACEOUS

- [?d] Hornblende diorite, leucogranite, and leucopegmatite
- [inu] Undifferentiated intrusives

DEVONIAN

LATE DEVONIAN

- [Dgn] Hornblende biotite orthogneiss

CAMBRIAN AND ORDOVICIAN

LOWER CAMBRIAN TO LOWER ORDOVICIAN

LARDEAU GROUP

- [LPLu] UNDIFFERENTIATED LARDEAU GROUP

BROADVIEW FORMATION

- [LPLw] Sempelitic and pelitic schist, micaceous quartzite, quartzite
- [LPLwc] Limestone, grey phyllitic limestone, and grey phyllite

JOWETT FORMATION

- [LPLt] Amphibolite, marble, calc-silicate schist, and phyllite

INDEX FORMATION

- [LPLn] Graphitic phyllite, marble, semipelitic schist, amphibolite, minor quartzite and micaceous quartzite
- [LPLng] Quartz grit, minor gritty phyllite

LADE PEAK FORMATION

- [LPLP] Graphitic marble and graphitic siliceous siltstone

CAMBRIAN

LOWER CAMBRIAN

BADSHOT FORMATION

- [LBd] Grey and white marble

PROTEROZOIC AND/OR CAMBRIAN

UPPER PROTEROZOIC (WINDERMERE) AND/OR LOWER CAMBRIAN

HAMILL GROUP

- [Hm] Quartzite, micaceous quartzite, semipelitic and pelitic

- CAMBRIAN
- LOWER CAMBRIAN
- BADSHOT FORMATION
- [16Ba] Grey and white marble
- PROTEROZOIC AND/OR CA
- UPPER PROTEROZOIC (W)
- HAMILL GROUP
- [16Ba] Quartzite, micaceous schist, phyllite, and
- AGE UNCERTAIN
- [NSU] Undifferentiated metas
- [NSa] Amphibolite
- [NSCS] Calc-silicate schist
- [NSM] Marble
- [NSPS] Pelitic schist
- [NSQ] Quartzite
- GEOLOGICAL SYMBOLS
- stratigraphic
- normal fault
- fault (exposed)
- ARF Akalkolex River
- CRF Columbia River
- SCF Standfast Cr
- TCFS Twin Creek fa
- GEOCHRONOLOGY
- ① 104 ± 2.5 Ma (U-Pb z)
- ② 100 Ma hornblende, 83
- ③ 359 ± 6 Ma (U-Pb zir
- ④ 71 ± 1 Ma (U-Pb zirc
- ⑤ mid-Cretaceous (U-Pb
- METAMORPHIC SYMBOLS
- a=andalusite
- k=kyanite
- s=sillimanite

CAMBRIAN

LOWER CAMBRIAN

BADSHOT FORMATION

[18Bd] Grey and white marble

PROTEROZOIC AND/OR CAMBRIAN

UPPER PROTEROZOIC (WINDERMERE) AND/OR LOWER CAMBRIAN

HAMILL GROUP

[18Ha] Quartzite, micaceous quartzite, semipelitic and pelitic schist, phyllite, and minor limestone

AGE UNCERTAIN

[nsu] Undifferentiated metasediment

[nsa] Amphibolite

[nscs] Calc-silicate schist

[nsn] Marble

[nspis] Pelitic schist

[nsq] Quartzite

GEOLOGICAL SYMBOLS

ARF Akolkolex River fault
CRF Columbia River fault
SCF Standfast Creek fault
TCFS Twin Creek fault system

GEOCHRONOLOGY SAMPLE LOCATIONS

- (A) 104 ± 2.5 Ma (U-Pb zircon; this study)
- (1) 100 Ma hornblende, 83 Ma biotite (K-Ar; reported in Price et al, 1985)
- (11) 358 ± 6 Ma (U-Pb zircon; Parrish, 1992)
- (11) 71 ± 1 Ma (U-Pb zircon; Parrish, 1992)
- (12) mid-Cretaceous (U-Pb zircon; R.L. Armstrong, unpublished data, 1989)

1000 1000 1000 1000 1000 1000 1000 1000 1000 1000

$$P = \begin{bmatrix} 1 & 1 & 1 & 1 \\ 1 & 1 & 1 & 1 \\ 1 & 1 & 1 & 1 \\ 1 & 1 & 1 & 1 \end{bmatrix}$$

- (A) 104 + 2.5 Ma (U-Pb zircon)
- (B) 100 Ma hornblende, 83 Ma zircon
- (C) 358 + 6 Ma (U-Pb zircon)
- (D) 71 + 1 Ma (U-Pb zircon)
- (E) mid-Cretaceous (U-Pb zircon)

a=andalusite
k=kyanite
s=silimanite

regional metand

○ contact metand

10 this study's th

13 sample analyzed

inferred position
box on high ground

contact (M3) me
tick on high gro

Use the graph only to obtain answers to questions 10 through 14.

Utilisez le diagramme que vous obtenez les années
1961-1962 ou les années 1962-1963
ou les années 1963-1964.

[illegible]

GEOCHRONOLOGY SAMPLE LOCATIONS

- ④ 104 ± 2.5 Ma (U-Pb zircon; this study)
- ② 100 Ma hornblende, 83 Ma biotite (K-Ar) reported in Price et al., 1985
- ③ 358 ± 6 Ma (U-Pb zircon; Parrish, 1992)
- ⑩ 71 ± 1 Ma (U-Pb zircon; Parrish, 1992)
- ⑬ mid-Cretaceous (U-Pb zircon; R.L. Armstrong, unpublished data, 1989)

METAMORPHIC SYMBOLS

a=andalusite
k=kyanite
s=sillimanite
st=staurolite
g=garnet
chl=chloritoid
ch=chlorite
bt=biotite

Lines represent tie lines in AFM topologies (Thompson, 1957) in pelitic schist.
All assemblages also contain quartz and muscovite.
Numbers next to symbols are this study's thin section numbers.
Symbols without numbers are assemblages reported by Thompson (1972), Zwanig (1973), and Sears (1979).
Circles without attached lines are non-pelitic schist samples.

● regional metamorphic (M2) assemblage

○ contact metamorphic (M3) assemblage

10 this study's thin section number

13 sample analyzed with microprobe by this study

inferred position of regional metamorphic (M2) isograds
box on high grade side

contact (M3) metamorphic isograd from Sears (1979)
tick on high grade side

TABLÉAU D'ENSEMBLE DU SYSTÈME NATIONAL
DE RÉFÉRENCE CARTOGRAPHIQUE

51 30	51 30	51 30
02 M 7	02 M 8	02 M 5
02 M 7	02 M 1	02 M 6
02 M 15	02 M 16	02 M 13
50 45	50 45	50 45
51 30	51 30	51 30

INDEX TO ADJOINING MAPS OF
THE NATIONAL TOPOGRAPHIC SYSTEM

The diagram is a schematic representation of the
geological map of the area shown in the
figure. It is not a map of the area shown in the
figure. It is a schematic representation of the
geological map of the area shown in the figure.

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PROF. 100-11

100-11

[illegible]

THE THOUSAND METRE
CONCRETE TOWER, 1000 METRES (3280 FT)
ZONE II
QUADRILLAGE DE MILLE METRES
SCALE 1:1000

T₁ INVERSION SCALE FOR ELEVATIONS

ECHELLE DE CONVERSION DES ALTITUDES

[illegible]

1992 年 12 月 15 日 星期一

$$e = \frac{1}{2} \left(\frac{1}{2} + \frac{1}{2} \right) = \frac{1}{2} \quad \text{and} \quad e = \frac{1}{2} \left(\frac{1}{2} + \frac{1}{2} \right) = \frac{1}{2}$$
[illegible]
$$\mathcal{H}^1(\mathbb{R}^n) \subset \mathcal{H}^1(\mathbb{R}^n) \subset \mathcal{H}^1(\mathbb{R}^n)$$

Sistema de información: Q. 00124100 - 01/10/2011

$$1 \leq p \leq \frac{1}{\alpha} \leq p^* \leq p \leq 1$$

Stress is a result of the interaction of the individual and the environment.

END

2/5 0/5 93

FIN