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STRUCTURE OF THE MOUNT COPELAND AREA, MONASHEE
COMPLEX, SOUTHEASTERN BRITISH COLUMBIA

By

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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

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acceptance of the thesis "Structure of the Mount Copeland area, Monashee Complex,
Southeastern British Columbia"

Chairman, Department of Earth Sciences

(

Thesis supervisor

ABSTRACT

In southeastern British Columbia, high-grade metamorphic rocks of the Monashee complex are exposed in a tectonic window through the Selkirk Allochthon.

Paleoproterozoic paragneisses of North American affinity and granitic orthogneisses comprise the lowest exposed level of the complex. Basement gneisses are overlain by, and infolded with, a mantling metasedimentary sequence of clastic, carbonate, and volcanic rocks. The Monashee complex is bounded above by the Monashee décollement, a westerly rooted, northeasterly verging crustal scale zone of ductile shear. Various interpretations have been proposed for the development of the field area but recent mapping has shown a need for a reinterpretation of the stratigraphy and structure.

Detailed field mapping was performed at 1:20,000 scale focusing on minor structures and the sequence of lithologies. The regional lithology, representing a transposed stratigraphy, in ascending order is orthogneiss and migmatitic paragneiss of the crystalline basement, basal white quartzite, micaceous schist, a mixed unit of dominantly calc-silicate gneiss with minor quartzite and white marble, and micaceous schist. While this sequence was commonly seen throughout the area, repetition and inversion of units is attributed to early-stage isoclinal folding (F_1/F_2). A major structural feature, the Bews Creek Fault, has been reinterpreted as a zone of ductile strain involving attenuation of lithologic units without significant omissions or repetitions; there is no evidence of important brittle faulting. This new interpretation results in a simple lithological sequence and allows for correlation of units across the field area.

Two phases of pre- to syn-metamorphic intrafolial folds were observed based on the superposition of folds producing Type II and III interference patterns. Phase one folds are preserved as Type II interference patterns and have locally resulted in inverted sequences. Associated with phase two deformation are thrust nappes, sheath folds and a penetrative stretching lineation. These occurred at upper-amphibolite facies conditions and resulted in the dominant transposition foliation (S_2). Kilometre-scale, close to tight, northeasterly verging F_3 's are prominent in the area and fold the transposed stratigraphy. This is thought to have occurred during the waning stages of metamorphism in the complex. Late stage F_4 's are rarely seen as east-west trending, large, open warps.

Surface geology, relation of minor structures, and the refined lithologic sequence have allowed a reinterpretation of the three-dimensional geometry of the area using vertical cross-sections. These results show the subsurface geology is more complex than previously proposed.

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ORIGINAL CONTRIBUTION

The Mount Copeland area (Jordan River area) was originally mapped by Wheeler (1965), Fyles (1970), and Höy and Brown (1980). These early works provide the base map for the current project.

Fieldwork for this thesis commenced in the summer of 2000, with a follow-up season in 2001. Summer field mapping was done at 1:20,000 scale on parts of mapsheets 82M.008, 82M.009, 82M.018, and 82M.019 (North American Datum 1983). This involved a detailed investigation of lithologic units and the mapping of structures at all scales. Particular attention was paid to establishing the lithologic sequence and the correlation of minor structures with larger-scale structures for each generation of deformation. All rock samples were collected by myself, or by one of my field assistants.

I selected, cut and polished all thin section pucks, which were then mounted and ground by Mike Jackson, the Departmental thin section technician. I performed all thin section analyses and acknowledge discussions with Dan Gibson and Chris Taylor. Geologic maps were compiled in CorelDRAW by combining the above works with data collected for this study. The refined lithologic sequence is an important component in the structural analysis. Cross-sections were constructed at 2 km intervals across the area to allow correlation of structures and predict the three dimensional geometry of the Mount Copeland area.

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LIST OF ABBREVIATIONS AND PLACE NAMES

Mineral Abbreviations

Bio	Biotite
Musc	Muscovite
Qtz	Quartz
Plag	Plagioclase Feldspar
Kfs	Potassium (K) Feldspar
Chl	Chlorite
Gt	Garnet
Sill	Sillimanite (Fibrolite)
Ep	Epidote
Hbl	Hornblende

Geographic Abbreviations

MC	Monashee complex
MD	Monashee décollement
FCD	Frenchman Cap dome
TOD	Thor-Odin dome
MCA	Mount Copeland area
BCF	Bews Creek fault
FB	Foreland Fold and Thrust belt
OB	Omineca belt
IMB	Intermontane belt
CB	Coast belt
IB	Insular belt
SA	Selkirk allochthon
KT	Kootenay terrane
QT	Quesnel terrane

Other Abbreviations

PPL	Plane polarized light
XPL	Cross polarized light (crossed nicols)

Structural features used in this thesis and reference as necessary: Monashee Antiform (Journeay, 1986), Copeland synform (Fyles, 1970), Copeland antiform, Eagle Pass synform, Eagle Pass antiform, Hiren Creek antiform, Hiren Creek synform, Bews Creek Fault (Fyles, 1970), Grace Mountain Syncline (Wheeler, 1965; Höy, 1987).

Informal place names used for geographic reference in this thesis include (see Figure 1.4 for locations): Camp (Alpha) Lake, Beta Lake, Gamma Lake, Camp Ridge, and Grizzly Peak.

CHAPTER 1.

INTRODUCTION

Overview of the Canadian Cordillera

The western margin of North America has experienced a long and protracted tectonic history involving an interplay between sedimentation, rifting, orogenesis, and crustal extension. Geological evidence indicates the margin began in a passive platformal setting in the Paleoproterozoic with thick accumulations of shallow water, marine sediments (Scammell and Brown, 1990). Sediment eroded from the exposed North American craton (Laurentia) was deposited to the west in an extensive miogeocline. Rifting along the length of western Laurentia in the Meso- and Neoproterozoic removed the western margin of the continent (Hoffman, 1989; Gabrielse and Yorath, 1991).

For much of its history (1440 Ma) the miogeocline developed in a passive margin setting except for episodic volcanism and rifting in the outer part of the miogeocline in early Paleozoic time (Gabrielse and Yorath, 1991). A collage of island arcs and amalgamated terranes began to arrive from the west on the subducting Farallon Plate (proto-Pacific Ocean) beginning in the Jurassic (Monger et al., 1982; Brown et al., 1986). The miogeoclinal cover was telescoped and displaced eastward in front of the advancing orogen and resulted in a westerly derived source of sediment for the foreland basin. Shortening and progressive deformation of the pericratonic sediments continued into the Tertiary. Exhumation in the hinterland of the Cordillera began synchronously with the cessation of contractional tectonics, which was further facilitated by extension along crustal-scale detachment faults that persisted into the Eocene (Parrish et al., 1988;

Johnson and Brown, 1996).

Orogenic belts are the vestiges of past collisions between tectonic plates, and preserved within the rocks are the results of the progressive deformation associated with the evolving orogen. The information available depends on the level of exposure and as lower structural levels require a large amount of erosion, the deformation associated with tectonic processes that occurred at this level are not well known. Many questions about orogenesis remain unanswered and researchers have placed a greater emphasis on deeper crustal levels in their investigations. As a result, research efforts have focused on metamorphic core complexes (Coney, 1980; Journeay, 1986; Parrish et al., 1988), and more specifically, on once deeply buried areas such as the Monashee complex of the southern Omineca Belt, British Columbia.

The Monashee complex

Located in the hinterland of the Canadian Cordillera and comprising part of the Shuswap Metamorphic Core complex (Coney, 1980; Armstrong, 1982), the Monashee complex represents one of the western-most exposures of the North American continent (Crowley, 1999). Situated below the sole thrust to the Canadian Cordillera and brought to the surface through extension and exhumation, this area preserves the structures and thermal events previously associated with a ductile regime at mid-crustal levels. The exact origin of the Monashee complex remains uncertain, but it is thought the area was initially buried to mid-crustal levels during the Cretaceous (Johnson and Brown, 1996 and references therein). Although the area was not directly involved in Cordilleran deformation, the deep burial and proximity to the basal décollement allowed the rocks to preserve the effects of strain from the Cordilleran orogen that are now exposed at the

surface in the Monashee complex. Regional mapping (Wheeler, 1965; McMillan, 1969; Reesor and Moore, 1971; Psutka, 1978; Brown, 1980; Höy and Brown, 1980; Read, 1980; Read and Brown, 1981; Journeay, 1986; Scammell and Brown, 1990), metamorphic studies (Journeay, 1986; Scammell, 1986), geochronological dating (Wanless and Reesor, 1975; Parrish and Scammell, 1988; Parrish et al., 1988; Carr, 1992; Crowley, 1997a; Gibson, 1997) and seismic imaging studies (Ross et al., 1991; Cook et al., 1992; Johnson and Brown, 1996) have all provided new insights into the origin and evolution of the Monashee complex. See Figure 1.2 for location of previous studies in the area.

Purpose of Study

The Mount Copeland area (previously referred to as the Jordan River area; Fyles, 1970) is located on the southern flank of Frenchman Cap dome (Figure 1.1). Mapping has been performed at scales of 1:253,440 (Wheeler, 1965), 1:25,000 (Fyles, 1970), and 1:100,000 (Höy and Brown, 1980). Höy and Brown (1980) provided a different interpretation from the first detailed mapping done in the area by Fyles (1970). Since that time, there has been no further comprehensive study of the area to address the conflicting views.

The purpose of this study is to re-evaluate the work of Fyles (1970) and Höy and Brown (1980) based on data collected during two summers of detailed field mapping. The goals of this study are: to refine the lithological succession; to determine the contact relation between the basement and cover sequences; to determine the significance of the Bews Creek Fault; and to characterize the phases of deformation. This information will be used to redefine the geologic map of the Mount Copeland area, and predict the

subsurface geometry of the area by constructing vertical cross-sections that satisfy the relation of minor structures, refined stratigraphy, and observed geologic data. Finally, this study will conclude with a comparison with previous studies and known Cordilleran events to place the Mount Copeland area in a regional context.

Access into the field area is by a 10 minute helicopter flight from Revelstoke, BC. Tree line is located at an elevation of approximately 1800 m and above this rocks are well exposed. Maximum elevations in the area range up to 2560 m at Mt Copeland and 2700 m along parts of Derickson Ridge. Below tree line, especially in valleys, exposure is poor and outcrops where seen are highly weathered and covered in lichen. At higher elevations, snow persists late into the summer, and ice remains on some lakes year round. Ten weeks of detailed mapping have been carried out during the 2000 and 2001 summer field seasons. This was followed by comprehensive thin section analysis, and construction of geologic maps and cross sections in the laboratory.

Regional Geology

The northwest-southeast trending Canadian Cordillera is the result of Jurassic to Cretaceous arc collision and accretion to the paleo-North American craton with crustal shortening and thickening (Monger et al., 1982; Brown et al., 1986). From east to west, the Cordillera comprises five morphogeologic belts (Gabrielse et al., 1991; see inset, Figure 1.1): the Foreland thrust and fold belt is characterized by thin skinned deformation; the Omineca belt is the metamorphic and plutonic hinterland of the Foreland belt; the Intermontane and Insular belts represent amalgams of largely unmetamorphosed to low-grade metamorphic island arcs and oceanic terranes (Intermontane and Insular superterranes) accreted to the North American cratonic margin during Jurassic to

Cretaceous convergence; and a plutonic complex named the Coast belt that bisects the Intermontane and Insular belts.

In southern British Columbia, the Omineca belt is bounded in the east from the Foreland belt by the Rocky Mountain Trench and partly in the west by the Okanagan-Eagle River Fault (Figure 1.1). The eastern portion of the belt is composed of folded thrust sheets of variably metamorphosed miogeoclinal rocks, while the western portion contains high-grade rocks of the pericratonic Kootenay Terrane which includes thrust sheets and fold nappes of accreted terranes (Gabrielse et al., 1991; Johnston et al., 2000 and references therein). The central part of the belt exposes middle to upper amphibolite facies basement-cored culminations in the Monashee complex, which includes autochthonous basement of the North American craton. The Monashee complex is structurally overlain by the metasedimentary rocks of the Selkirk allochthon. These tectonic elements are separated from each other by the Monashee décollement, a westerly rooted crustal-scale shear zone that has been interpreted through seismic reflection data to correlate with the sole thrust of the Foreland thrust and fold belt (Brown et al., 1992; Cook et al., 1992). The Monashee complex together with the high-grade portion of the Selkirk allochthon constitute the Shuswap Metamorphic Core complex as defined by Coney (1980) and Armstrong (1982).

The Monashee complex is seen in a tectonic window through the Selkirk allochthon where two Eocene domal culminations, Frenchman Cap to the north and Thor-Odin to the south, expose rocks of Archean to Paleozoic age. The Monashee complex is bounded to the west by the Monashee décollement, and to the east by the Columbia River fault, an Eocene, crustal-scale, normal sense shear zone partly responsible for the tectonic

denudation of the complex. The exhumed upper-amphibolite facies rocks forming the core of the Monashee complex comprise Paleoproterozoic para- and orthogneisses (Armstrong et al., 1991; Crowley, 1999 and references therein). These are unconformably overlain by a polydeformed, Paleoproterozoic to Paleozoic, platformal metasedimentary cover sequence of quartzites, calc-silicate schists and gneisses, and marbles (Scammell and Brown, 1990) that were deposited along the rifted margin of ancestral North America beginning at ca. 1900 - 1852 Ma (Crowley, 1997b).

The study area is located in the Monashee Mountains approximately 20 km northwest of Revelstoke, British Columbia (Figure 1.1), on the southern flank of Frenchman Cap dome. The contact between Paleoproterozoic basement gneisses and the unconformably overlying metasedimentary cover sequence is located within the study area. Predominant structures in the area are northwest-southeast trending, and include kilometre-scale, open to close third generation folds that control map-scale geometry and cause repetition of the stratigraphy throughout the field area.

Previous work

The basement gneisses of the Monashee Mountains were first described by Dawson (1898) who referred to the rocks as Archean and noted the similarities between the Shuswap Series (Monashee complex) and the Grenville Series of Eastern Canada. Mapping and interpretations at various scales were done by Wheeler (1965), Fyles (1970), Reesor and Moore (1971), and McMillan (1973) that led to the interpretation of the basement gneiss domes as isolated exposures of the North American craton, involved in and uplifted during the Cordilleran orogeny (Journeay, 1986; Johnson and Brown, 1996). The Proterozoic tectonic history of continental rifting and sedimentation in the

northern Frenchman Cap dome has been documented by Scammell (1986), while Journeay (1986) investigated Mesozoic compression and Late Cretaceous to Early Tertiary extension. More recent work by Crowley (1997b and 1999) examined the age of deformation in crystalline basement rocks and provided refined age constraints.

The interpretation of the stratigraphy and structural hierarchy in the field area has evolved significantly over the past few decades. In the Mount Copeland area Fyles (1970) documented two phases of deformation; isoclinal folds and open to close folds were termed F_1 and F_2 , respectively. Fyles also described the stratigraphy in terms of 10 map units and reported that distinct units (eg. white quartzite) always occurred in contact with calc-silicates and that this relation appears to be repeated throughout the section. However, subsequent research (Höy and McMillan, 1979; Höy and Brown, 1980) has shown that some of the interpretations of Fyles may be incorrect due to a misinterpretation of the phases of deformation and the unknown character of the stratigraphy. Fyles (1970) made reference to several areas where he recognized unusual map patterns but was unable to interpret the patterns or correlate them with the phases of deformation. Patterns on the lithologic compilation map (Map 3) resemble Type II and Type III interference patterns, indicative of three phases of deformation. The interpretation of three phases of deformation has also been corroborated by other studies in the region (Figure 1.2; McMillan, 1973; Psutka, 1978; Brown, 1980; Höy and Brown, 1980; Journeay, 1986).

Research Procedure

This study builds upon the results from previous work to develop a geometric model of the subsurface geology through interpretation of stratigraphy and structures.

The initial analysis involved: 1) determining the sequence of lithologic units; 2) determining the structural relationship between the basement and cover sequence; 3) mapping and correlation of structures to determine the phases of deformation; and 4) determining the significance of the Bews Creek Fault to evaluate whether or not it exists, and if so, what affect it has on the local stratigraphy. Due to its proximity to the basement-cover contact (Figure 1.3), the fault could have important implications if it is determined that the fault cuts across, or cuts out stratigraphy. Protracted ductile deformation and poly-phase folding complicate structures in the field area, but by associating field observations and minor structures to map patterns, an interpretation of the three dimensional geometry emerged.

The map patterns of Fyles (1970) and Höy and Brown (1980) were then integrated and tested with the refined stratigraphy and structural data collected in the study by producing a subsurface model for the area. By also adding the findings of Scammell (1986), Journeay (1986), and Crowley (1997b, 1999), timing constraints and geologic history may be established and placed in a regional context. The important aspect of this project is to confirm the occurrence of basement gneisses in the southern part of the Mount Copeland area, and to correlate these in the subsurface using vertical cross-sections with known basement rocks to the north.

Figure 1.2: Simplified regional geology (modified after Wheeler and McFeely, 1991) and location of previous mapping in the northern Monashee Complex. Note distribution of basement in Mount Copeland area.

Figure 1.3: Simplified lithologic compilation map of the Mount Copeland area incorporating mapping by Fyles (1970), Høy and Brown (1980), Bosdachin (1989) and this study. Only features and units mapped by Fyles (1970) that have since been changed are highlighted; not all units are included.

CHAPTER 2.

LITHOLOGY OF THE MOUNT COPELAND AREA AND REGIONAL CORRELATIONS

Introduction

The Mount Copeland area (Map 2 and Figure 1.1) is located on the southern flank of Frenchman Cap dome of the Monashee complex; the field area largely encompasses the drainage basin of the Jordan River (Map 1). Rocks in the area are divided into three groups based on deformation history, lithology, and relative age. The three groups in order of decreasing relative age are basement gneisses, cover gneisses, and post-orthogneiss intrusions. These are further subdivided into lithologic map units (Map 3); see also Figure 2.1 for a generalized map of the area. The sequence of units was determined based on field observations, and the description of rock units from the analysis of over 120 hand samples and 62 thin sections.

Frenchman Cap dome is composed of crystalline basement rocks comprising paragneiss (unit 1) and orthogneiss (unit 2). Both units are well exposed in the north of the field area and in a finger of basement that is infolded with cover gneisses and wraps around the south end of the dome (Map 2). Basement gneisses are often migmatitic in appearance and can display a random, somewhat disorganized structure. Overlying the basement is a metamorphosed and intensely deformed sequence of platformal metasedimentary rocks that are laterally extensive and >2000 m thick (Scammell, 1986). In the study area only about 700 m of the succession is preserved. A white quartzite, commonly referred to as basal white quartzite (unit 3), is the lowest unit of the cover sequence and is seen in sharp contact with basement rocks. Overlying the basal quartzite

is a succession of biotite schist (unit 4), a mixed unit of calc-silicate gneiss and marble (unit 5), and biotite-sillimanite schist (unit 6); all units contain m-scale layers of quartzite. A thin Pb-Zn-rich layer is associated with the marble of unit 5, and together make a distinctive marker unit that is commonly referred to as the virgin marble (Brown, 1980; Brown pers. comm., 2000). There are three recognized episodes of later-stage intrusions distinguished by composition and relative timing.

Original bedding (S_0), and two transposition fabrics (S_1 and S_2) plus five phases of deformation complicate structural interpretations of the lithologic sequence; deformation has apparently destroyed all primary tops indicators within the units. Metamorphic rocks in the map area are of upper-amphibolite facies. The thickness of some units varies from metres to hundreds of metres so no inference about original sedimentary thickness can be made; where listed, measurements represent the structural thickness. Map units are presented in ascending order from structurally lowest to highest levels with intrusions described at the end of this chapter in order of inferred relative timing.

Through correlation of units with other studies from Frenchman Cap dome (McMillan, 1969; Psutka, 1978; Scammell, 1986; Journeay, 1986; Scammell and Brown, 1990) the lithologic sequence in the area was determined with reasonable certainty (Map 2). Supplementary to the information presented in this chapter, a summary of rock descriptions is provided in Figure 2.2, and correlation between the stratigraphy of Fyles (1970) and the lithologic sequence from the current study is listed in Figure 2.3.

Basement (Core) Gneisses

Basement rocks of the Monashee complex are present in the northern part of the

field area where excellent exposure facilitated detailed mapping of structures and lithology. This area, north of Copeland Creek (Map 3), is composed of known basement gneisses which were used to characterize the basement rocks to allow for comparison with other areas of suspected basement rocks. Various interpretations have been presented for the origin of the core gneisses (McMillan, 1973; Psutka, 1978), they are now interpreted as distinct units containing rocks of igneous and sedimentary origin for the orthogneiss and paragneiss, respectively, with the metre-scale layering of the paragneiss inferred as transposed compositional layering. Structurally, orthogneiss is the lowest unit in the area and was always seen separated from the cover sequence by an intervening layer of the paragneiss (Figure 2.4). The base of the basement gneisses is not observed in the study area but mapping indicates the exposed basement sequence to be a minimum of several hundred metres thick.

Unit 1: Plagioclase feldspar, biotite, migmatitic paragneiss

Unit one of the lithologic sequence is a heterogeneous mixture of amphibolite, psammite, and semi-pelite layers (Figure 2.2a) and will be referred to simply as basement paragneiss from its dominantly sedimentary origin. Leucosome is common and is especially evident in the amphibolite and semi-pelite units, which also display a well defined gneissic foliation of light (felsic) and dark (mafic) material varying from mm to cm in scale (Plate 2.1). Competent amphibolite layers with concordant boudins display pinch and swell structures with coarse grained quartzofeldspathic material infilling the 'pinch' (Plate 2.1a).

The fine to medium-grained semi-pelite is made up of >30% disseminated biotite defining the foliation, with approximately 20% each of quartz, potassium feldspar, and

plagioclase feldspar; minor muscovite and chlorite were seen replacing biotite. Hornblende, sillimanite, garnet, and muscovite were also observed in variable proportions, generally in close proximity to amphibolite layers. Fresh surfaces are medium grey coloured and display a prominent gneissosity. Weathered surfaces are more uniform medium to dark grey and commonly rust stained. Psammitic layers are similar in composition to the semi-pelite, but with a greater proportion of quartz at the expense of the other minerals. The well-defined alternations of semi-pelite and psammite most likely originate from transposed sedimentary layering. Semi-pelite and psammite define dm- to m-scale layering with gradational contacts. These resistant layers appear fairly homogeneous along strike.

Amphibolite layers are medium grained and rich in biotite and hornblende with discontinuous mm-scale plagioclase (An 30%) rich layers defining a strong foliation in outcrops. Rust stained weathered surfaces are common. The concordant nature and lateral extent of the amphibolite layers parallel with the host semi-pelite and psammite layers suggest the amphibolite layers were either intruded by mafic sills or the layers represent syn-depositional mafic volcanoclastics. Contacts are parallel to the foliation, and transposition and deformation preclude determination of the intrusive or depositional relation of the amphibolites to the host rocks. The composition of the layers does not appear to vary along strike although the amphibolite layers were observed pulled apart with leucocratic pegmatite infilling between the boudins (Plate 2.1a).

The appearance of paragneiss in outcrops is variable. In the northern end of the field area along Copeland Ridge, the paragneiss is migmatitic with alternating layers of leucocratic and melanocratic material (Plate 2.1c) defined by variably thick

quartzofeldspathic layers (leucosome) within a matrix of thin biotite-hornblende-rich layers (melanosome). In other areas that may have been at a higher structural level, the migmatites and gneisses are not as well developed and the basement paragneiss contains small elongate lenses of leucosome, within a dark, foliated aphanitic matrix (Plate 2.1d). This was referred to as grey gneiss by Fyles (units 6 and 10; Fyles, 1970) and distinguished from other units but composition and structural interpretations indicate the grey gneiss is correlative with basement rocks of unit 1. The grey gneisses are very fine grained as seen in hand sample with 10-20 cm thick quartz lenses, containing mm- to cm-scale quartz grains and smaller potassium feldspar crystals. In the South Hiren Valley (Map 1), large garnet crystals >1cm in diameter were seen in the amphibolite layers of the paragneiss near the basement-cover contact. In the field area, biotite makes up ~30% of the semi-pelite layer of paragneiss and in thin section defines the foliation.

Unit 2: Quartz, potassium feldspar augen orthogneiss

Orthogneiss, intruded as large plutons, comprises a large portion of the core of Frenchman Cap dome (McMillan, 1969) and was seen at several isolated localities in the field area (Map 3). In the study area orthogneisses are described as a poorly foliated, grey weathering, potassium feldspar augen orthogneiss (Plate 2.2a and b), and will herein be referred to as basement orthogneiss (unit 2).

Quartz grains are typically small (<1mm) and display evidence of dynamic recrystallization (undulose extinction) when viewed in cross polarized light (XPL) under the microscope. Some clusters of feldspar grains (augen) were seen up to 4 cm in length. Biotite grains, present as laths but more commonly disseminated, make up 10% of the rock and shows a poor alignment within the foliation (Plate 2.2c). Potassium feldspar is

dominant over plagioclase feldspar (An 25-30%) but both were partially altered to sericite. Hornblende has been reported in other studies (Fyles, 1970; MacMillan, 1973) but is only a minor component in this study. Relict augen potassium feldspar phenocrysts and the homogeneous nature support an igneous origin for the orthogneiss.

Porphyritic Granite Gneiss of Fyles (1970)

Porphyritic granite gneiss (Unit B of Fyles, 1970) was seen at several localities and forms a large exposure near Neil Lake and south of Hiren Creek (Figure 1.3). Originally classified as a late-stage granitic intrusion, this study supports the interpretation of the porphyritic granite gneiss as basement orthogneiss (unit 2). This is supported by the association of porphyritic granite gneiss with paragneiss, and structural evidence which indicates the porphyritic granite gneiss has experienced several phases of deformation. The rocks will be called porphyritic granite gneiss for the purposes of identification in the text where necessary, but are labeled as part of the suite of basement orthogneiss (unit 2) on geological maps. For a description and the mineralogy of the porphyritic granite gneiss see the discussion of the basement orthogneiss in the previous section.

Interpretation

Psammite and semi-pelite layers preserved in the paragneiss are inferred to represent original compositional layering indicating a sedimentary origin for the basement paragneiss. The relation of amphibolite to the other layers could not be determined, but the concordant appearance supports a volcanoclastic or extrusive mafic event.

The orthogneiss is interpreted as intruding into the paragneiss. Orthogneiss displays sharp margins while the host paragneiss near the contact appears more strongly

foliated with a greater abundance of biotite that made the host paragneiss appear more mafic than typical paragneiss. Orthogneiss is listed as unit 1 in other publications, presumably because it is structurally lower than paragneiss. However, geochronological evidence indicates the orthogneiss is younger than the paragneiss (see references in Figure 2.4) so the orthogneiss is referred to as unit 2 in this thesis. The base of the orthogneiss was not observed. Previous authors (Wheeler, 1965; McMillan, 1969; Höy and Brown, 1980; Armstrong et al., 1991; Crowley, 1997b, 1999) have mapped the orthogneiss as an independent unit so its distribution is well known around Frenchman Cap dome (Map 2).

Strain and upper-amphibolite facies metamorphism can preclude determination of the contact relationship of the core gneisses with the structurally overlying metasedimentary cover sequence. Transgression of orthogneiss plutons by unit 3 (quartzite) in the northern part of the field area indicates erosion of the basement prior to deposition of the cover sequence. Compositional layering, foliation, and contacts are all roughly parallel (Plate 2.3) which makes determination of primary relations difficult. In the northern Monashee complex where the deformation was less intense, Journeay (1986) investigated the contact relation of the basement and cover rocks and inferred the top of the basement to be an erosional unconformity. Field and geochronological evidence also suggest that there was a hiatus in deposition before deposition of the cover sequence. In this area, the contact between the basement gneisses and cover sequence is generally sharp (Plate 2.3c and d) and appears concordant. The regional foliation is parallel to sub-parallel to the contact although evidence does exist for an erosional unconformity. Primary sedimentary features such as erosional scars (Plate 2.3e), and a grit layer (Plate 2.3a) were identified although not confirmed. The erosional

unconformity is further supported by geochronological evidence (Crowley, 1997b, 1999) which shows a time gap between deformation and metamorphism of the basement (~2060 Ma) and the beginning of deposition of the cover sequence (1862 to 1852 Ma).

Geochronology and Correlation with the North American Craton

Correlation of the Monashee complex with the Precambrian rocks of eastern Canada is not a new idea (Dawson, 1898; Daly, 1915). The appearance and character of the rocks of the Monashee complex and Canadian Shield are similar, but separated by the intervening sedimentary strata of the Rocky Mountains and Prairies, and the idea only persisted as a hypothesis. Support was offered by applying geochronological (Crowley, 1999) and geophysical techniques (Ross et al., 1991) which provided a link across the thick Phanerozoic cover, and allowed probable correlation between the North American craton and Cordilleran exposures of suspected North American affinity. The understanding of the Phanerozoic evolution of the Monashee complex is still evolving, but there is enough information preserved in basement rocks to interpret some of the history back to the Paleoproterozoic (Armstrong et al., 1991; Crowley, 1997b, 1999). The effects of the Cordilleran overprint during Mesozoic to Tertiary orogenesis in deeper levels of the Monashee complex are weak such that Paleoproterozoic ages and events are preserved (Crowley and Shaubs, 1994).

Geochemical and geochronological analyses show the rocks that now form the paragneiss of the Monashee complex were formed from juvenile crust as old as 2300-2200 Ma (Armstrong et al., 1991). Crowley (1999) inferred the basement paragneisses were deposited between 2270-2080 Ma yielding a complex history of deposition, deformation, metamorphism and erosion occurred before deposition of the cover

sequence began between 1990 and 1852 Ma (Crowley, 1997b). No evidence of sedimentary structures were observed but the dm- to m-scale layering of psammite, semipelite, and amphibolite preserve the original compositional layering of the metasediments and metavolcanics. Deposition was followed by deformation and high-grade metamorphism producing a migmatitic gneissosity that apparently destroyed all sedimentary structures. Metamorphism with formation of migmatites in the basement paragneiss is reported as occurring between ~1900-1850 Ma (Parrish, 1995) and 2077-1848 Ma (Crowley, 1999). The migmatites observed in the basement gneisses in the Mount Copeland area are likely a composite of Paleoproterozoic and Cordilleran metamorphism and deformation.

Intrusions of orthogneiss were episodic with phases in Frenchman Cap dome reported at ~2270 Ma and ~2100-2080 Ma (Crowley, 1999), 2100-1850 Ma for an augen gneiss (Armstrong et al., 1991), and 2100-2060 Ma for the Malton Complex (McDonough and Parrish, 1991; Murphy et al., 1991) to the north of the current study. Most of these ages are older than the 2060 Ma age of deformation and metamorphism in the basement as reported by Crowley (1999) although the deformation is not well recorded in the orthogneiss.

Geochronology in the Alberta Basin and Canadian Shield are providing models for correlation of source materials and geophysical studies are extending the Precambrian basement beneath Alberta into B.C. (Hoffman, 1988, 1989; Armstrong et al., 1991; Ross et al., 1991). Based on earlier studies undertaken in the Alberta Basin (Hoffman, 1988; Ross et al., 1991), the crystalline rocks of the Monashee complex correlate with the Wopmay orogenic belt (Parkinson, 1991) and Paleoproterozoic rocks underlying northern

Alberta (Armstrong et al., 1991). Later refinements indicate that the basement rocks of the Monashee complex are older than 2060 Ma but do not contain Archean ages, supporting the correlation of the Monashee complex with rocks to the north of the Hearne Province. Westward extension of structural provinces along strike (Figure 2.5) points to a correlation of Frenchman Cap dome of the Monashee complex with the Paleoproterozoic Thorsby Terrane, which also matches the formation ages of Armstrong et al. (1991) with the 2300-2100 Ma juvenile crust ages of the Thorsby Terrane (Hoffman, 1988; Thériault and Ross, 1991). This is further supported by the similarity in intrusive episodes (Crowley, 1999) between the Monashee complex and intrusive suites of the Canadian Shield.

Cover (Metasedimentary) Gneisses

The cover gneisses comprise a thick succession of up to 2 km of autochthonous platformal metasedimentary rocks lying unconformably on basement gneisses (Scammell and Brown, 1990). The contact with basement rocks is well exposed (Plate 2.3). In the field area the cover sequence is exposed in a large embayment flanked by basement gneisses (Map 3 and Figure 2.1). The cover sequence has been divided into three parts (Journey, 1986) which represent upward-fining depositional cycles referred to by Journey as the Lower (Units 3, 4), Middle (Units 5, 6), and Upper (Units 7, 8, and 9) Assemblages. These assemblages are capped at the top by the Monashee décollement, a crustal-scale shear zone separating the Monashee complex from the overlying Selkirk allochthon. Only the lowest upward-fining sequence (Lower Assemblage) was observed in this study; the Monashee décollement is located in the southwest corner of the field area although in an area that was not mapped.

The original stratigraphic thickness of units is not possible to determine due to the infolding of units (Plate 2.3g), tectonic transposition, and attenuation of fold limbs. In other areas where metamorphism and strain are not as intense (Scammell, 1986), thicknesses were determined with more certainty but even the basal quartzite was found to vary from metre-scale to as much as 600 m (Journeay, 1986). Estimates of unit thickness have been made for the Mount Copeland area (Fyles, 1970; Figure 2.4). These values are highly variable and readers are referred to other publication for more precise figures (eg. Crowley, 1997b and references therein). In areas including the basement-cover contact, the sequence of white quartzite, biotite schist, calc-silicate gneiss plus marble, and biotite-sillimanite schist was commonly seen; see also Plate 2.6b. This is a similar sequence inferred from McMillan (1969), Psutka (1978), Scammell (1986), Journeay (1986), and Crowley (1997b) around Frenchman Cap dome (Map 2) and is shown graphically along with intrusive and depositional ages in Figure 2.4. As the cover sequence is dominated by rocks of sedimentary origin, they could be referred to as paragneisses, but to avoid confusion in this thesis the term paragneiss is reserved for the metamorphosed sedimentary layers of the basement sequence.

Unit 3: White quartzite; metaconglomerate

The base of the cover sequence is marked by a prominent white quartzite (Plate 2.4a) that appears flaggy in hillsides (Plate 2.4e to g) and blocky in hand sample. The rock contains >95% quartz with cm- to dm-scale layering defined by fine psammitic to micaceous partings (Plate 2.4c). Weathered surfaces are typically white but may also appear grey or rusty coloured depending on local compositional variations. The contact with the underlying basement paragneiss is generally sharp and in some locations the two

are infolded (Plate 2.3d and g); see also the description on the basement-cover contact in the interpretation section of the basement sequence.

Quartz grains are well sorted, anhedral and 1-2 mm in diameter. Centimetre-scale pure quartzite layers are interspersed with micaceous layers of fine grained muscovite, $\pm$ feldspar in layers that are <1 mm thick but variable and up to 0.5 cm thick, and create thin parting surfaces. In thin section, alternating layers of varying grain size give a banded appearance to the rocks. Other authors (eg. Journeay, 1986) have documented a greater assortment of layers within each unit than is reported for the study area; this will be discussed further in the interpretation section for the cover sequence.

A quartz pebble metaconglomerate (Plate 2.4b), <3 m thick, was observed at the base of the quartzite unit grading up to pure white quartzite. Elongate white quartzite pebbles are sub-parallel to the foliation, up to 5 cm long and 2 cm in diameter, but generally are more equidimensional, approximately 3 cm in diameter. A greater abundance of muscovite is present in the matrix in relation to the composition of the pure white quartzite.

This unit, composed of the basal quartzite and metaconglomerate, is generally 20-80 m thick although it was observed as thin as a few metres along fold limbs to hundreds of metres in the hinge zones of folds. Also present at the base, but considered part of the basement, is a pelitic grit layer which has been interpreted as a paleosol (Journeay, 1986). This suggests the top of the basement gneisses were an erosional surface as the cover sequence was deposited.

Unit 4: Mica schist; minor biotite-sillimanite schist

Overlying the white quartzite is a ~ 10 metre thick, lenticular layer of rusty

weathering mica schist. The lower contact with quartzite is sharp and considered stratigraphic. This unit appears to thin and pinch-out but it could not be determined if this is tectonic or a result of sedimentological processes. Where present, the unit forms distinct rusty layers easily identified in hillsides and outcrops (Plate 2.5a and b). The mineral assemblage is dominated by quartz, potassium feldspar, biotite, plagioclase feldspar, and sillimanite with garnet and sporadic muscovite present as accessory minerals.

Weathered surfaces are rust stained with a variable appearance. Rocks with a mottled texture due to white clusters of feldspar and sillimanite have a gneissic appearance and contain thin layers of quartz and feldspar. These rocks are blocky when fractured. Schistose appearing rocks have alternating grey and white layers, contain large biotite up to 3 mm in diameter, and randomly oriented sillimanite mats (fibrolite) when viewed in thin section.

Fyles (1970) interpreted a structural break at the top of unit 4, which he termed the Bews Creek Fault. Where observed the contact of unit 4 with the overlying calc-silicate gneisses of unit 5 is concordant and appears stratigraphic. The major stratigraphic break interpreted by Fyles (1970) was not identified and will be described more in Chapter 3.

Unit 5: Calc-silicate gneiss and marble; minor quartzite

Calc-silicate gneiss (unit 5) is widespread throughout the field area and can be further subdivided into four mappable subunits based on lithology (Höy and Brown, 1980; Journeay, 1986; Höy, 1987). The subunits are listed below in ascending order and will be described briefly:

- a) white quartzite; minor schist
- b) biotite hornblende gneiss; minor quartzite, marble, calc-silicate gneiss, mica schist
- c) white crystalline marble $\pm$ Pb-Zn sequence
- d) calc-silicate gneiss; sillimanite schist and biotite schist

The majority of the calc-silicate gneiss (unit 5) appears homogeneous (Plate 2.5c-e), and without marker horizons it was difficult to identify folds within the sequence. The calc-silicate gneisses were greatly affected by Cordilleran deformation due to intense ductile strain and isoclinal folding so the thickness of units could not be determined.

The lowest subunit (unit 5a) is a white quartzite with subordinate minor schist layers. This subunit is indistinguishable from the basal quartzite seen elsewhere and may represent periodic changes in sedimentary sources or depositional environment. Its location adjacent to calc-silicate in the sequence precludes it being a repetition of the basal quartzite and it is seen more commonly around Frisby Ridge (Map 3). The composition is largely quartz with minor muscovite and potassium feldspar defining cm- to dm-scale layering.

The grey gneiss, more appropriately called biotite, hornblende gneiss (unit 5b), is located mainly in the Copeland River valley and forms dark grey outcrops, with a dominant schistosity and lenses of quartz up to 10 cm thick. The quartz lenses preserve the hooks of isoclinal folds (Plate 3.3e). Composition of the matrix is quartz, potassium feldspar, biotite, hornblende, plagioclase feldspar, and minor calcite.

One of the most distinct units around Frenchman Cap dome is a buff to grey weathering, white crystalline marble, up to 16.3 m thick (Fyles, 1970), that serves as a regional marker horizon (Brown, 1980). This unit is prominently seen outlining the

Copeland Synform in Mount Copeland and is present with pelite at the eastern end of Derickson Ridge (Plate 2.5f). Fyles (1970) reported 3 marble layers in the calc-silicate gneiss (his unit 8) but only one pure marble layer was identified and correlated in this study. This unit is resistant to weathering and is located near the top of the sequence, commonly associated with a stratabound Pb-Zn sulphide layer (Fyles, 1970; Höy, 1987). The sulphide layer has long been of economic interest to geologists in the area (Fyles, 1970; Höy, 1987) although it has never been actively mined. The marble is composed almost entirely of calcite with minor proportions of biotite and muscovite identified in thin section. The sulphide layer, up to 3.3 m thick, consists of fine-grained galena, sphalerite, and pyrrhotite. Pb-isotope dates (Höy and Godwin, 1988) suggest an Early Cambrian age for the formation of the sulphide sequence, which has been interpreted as syn-depositional by Höy (1987).

The top subunit of the calc-silicate gneiss is a hornblende calc-silicate gneiss $\pm$ sillimanite schist (unit 5d). Unit 5d was not mapped during this study but the composition and description from Fyles (1970) appear similar to the calc-silicate gneiss unit (unit 5). The unit is fine to medium-grained, dark green-grey with an uneven weathered surface. Lenses of quartz were observed in the foliation with hooks at the end of the lenses preserving earlier folds.

Unit 6: Biotite-sillimanite schist

The upper unit identified in the cover sequence is a rusty weathering pelite. This unit was not mapped in this study by the author but it has been mapped and characterized in the field area by Fyles (1970) and Höy and Brown (1980); the description is based on Fyles (1970). This unit is exposed near Mt. Copeland and at the confluence of the

Copeland and Jordan Rivers (Map 3), which are the highest structural levels in the field area. Originally classified as part of unit 4, sillimanite schist (Fyles, 1970), recent mapping (Höy and Brown, 1980; Journeay, 1986; this study) indicate this rock type to be a distinct unit. Rocks in outcrop have a brown and white mottled appearance caused by clusters of sillimanite and feldspar. The Foliation is poorly defined and folds are outlined by lenses of sillimanite and feldspar. Samples are classified as quartz-biotite-sillimanite schist but also contain significant amounts of plagioclase and potassium feldspar.

Interpretation

The Paleoproterozoic history of the western North American craton is still only poorly understood. This is mainly due to the thick Phanerozoic cover that obscures the craton at its western extremity. The exposures of cratonic North America that do occur, such as the Monashee complex, have been overprinted by Cordilleran orogenesis such that earlier tectonic events have been obscured. Advances in geophysics, geochronology, and geochemistry are helping to clarify this early history.

The cover sequence is inferred to represent three upward-fining deposition cycles in a passive margin setting (Journeay, 1986; Scammell and Brown, 1990). Each assemblage (cycle) is characterized by (after Journeay, 1986): (a) a basal unit of pure quartzite with upward increasing proportions of semi-pelitic schist and calc-silicate gneiss, (b) a mixed unit of interlayered pelitic and semi-pelitic schist with layers of calc-silicate gneiss and minor quartzite, and (c) an upper unit of intercalated pelitic schist, quartzite and calc-silicate gneiss with pure and impure marble layers defining regionally continuous marker horizons. See Plate 2.6a and b for examples of the lithologic sequence at various scales in the Mount Copeland area. Journeay (1986) and Scammell (1986)

reported detailed lithologic sequences with a greater variety of sedimentary layers within the basal quartzite than is reported in this study. However, strain may have been more intense in the Mount Copeland area so the preservation of the individual layers was not as good. Since this study concentrated on detailed mapping of minor structures it was not always possible to follow lithologic units. There was some variation in the appearance of white quartzite at higher structural levels in the area, but the subunits of unit 5 were not prominent in outcrop. This may indicate a decrease in strain as one moves away from the core of Frenchman Cap dome, or variations in strain within the field area.

In the northern Monashee complex where deformation and metamorphism were less intense, several authors (Brown, 1980; Scammell, 1986; Journeay, 1986) observed preserved primary structures and reported that the stratigraphic sequence is generally right-way-up. Correlation of previous work with this study suggests that the sequence as reported is in normal stratigraphic order and is well known around Frenchman Cap dome (McMillan, 1969; Psutka, 1978; Brown, 1980; Journeay, 1986; Höy, 1987; Scammell, 1986; Bosdachin, 1989; Crowley, 1997b). Fyles (1970), McMillan (1973), Psutka (1978), and Höy (1987) reported stratigraphic facing directions as being right-way-up in metaconglomerates and quartzites of unit 3, which directly overlie the basement rocks. Where seen, contacts are generally sharp and parallel to the foliation (S_2). Possible evidence of primary structures was observed as a result of mapping during this study (Plate 2.3e and 2.4d), but it is not certain whether these features are of primary sedimentary origin or products of deformation.

Due to the persistence of marker horizons such as the basal quartzite and buff weathering marble around Frenchman Cap dome, the thinning of units is inferred to be

largely a result of attenuation of fold limbs during deformation rather than a result of a significant change in the sedimentary environment. Some thinning and disappearance of lithologic units may be attributed to primary facies changes, but the lateral extent of these units along strike would indicate similar processes persisted throughout the region, and through time. As these extensive units are invariably present in the same stratigraphic position (Höy, 1987) and persist throughout the Monashee complex (Map 2), the notion of a platformal sequence on a passive margin with only minor variation in the sedimentary environment would be an appropriate interpretation. Journeay (1986) inferred topographic basement highs during deposition of the lower assemblage of the cover sequence that may have produced local, although not regional, variations in the sedimentary environment. See Höy (1987) and Scammell and Brown (1990) for a more detailed description of the lithologic sequence and Journeay (1986) for an interpretation of the sedimentary environment.

Geochronology and regional correlations

Geochronological data indicate the base of the cover sequence was deposited between 1990 Ma and 1852 ± 4 Ma based on detrital zircons from the basal unit and a pegmatite that intruded into the lower part of the cover sequence (Crowley, 1997b). Constraints on the upper part of the lower sequence are provided by a syenite gneiss that intruded into the calc-silicate gneiss (Figure 2.4). The syenite gneiss has been dated in the vicinity of Mount Copeland with a crystallization age of 740 ± 36 Ma and a metamorphic age of 59 ± 1 Ma provided by zircons (Parrish and Scammell, 1988). Further up-section, Pb isotope dates (Höy and Godwin, 1988) yielded an Early Cambrian age for a stratabound Pb-Zn sulphide horizon that Höy (1987) interpreted as syn-

depositional with the host calc-silicate gneisses and marbles.

These data indicate the cover sequence began to be deposited before 1852 ± 4 Ma, older than any known rocks in the Canadian Cordillera (Crowley, 1997b), and also creates a controversy as to the source of the sediments. The cover sequence has previously been correlated with two known rift sequences in the Cordillera, the Mesoproterozoic (~1500-1200 Ma) Belt-Purcell Supergroup and the Neoproterozoic to Paleozoic Windermere Supergroup (Brown, 1980 and references therein), but the data from Crowley (1997b) indicate the basal unit of the cover sequence is older than the Belt-Purcell Supergroup.

Studies in the Alberta Basin (Ross et al., 1991; Villeneuve et al., 1993) and North American craton (Hoffman, 1989 and references therein) have provided data that may provide a correlation with the Monashee complex cover sequence. Detrital zircon ages from the basal quartzite indicate a source in the Alberta basement rather than the underlying basement. Possible sources include the Slave province for the Neoarchean zircon ages, 2950-2860 Ma, and the Rimbey magmatic arc for the Paleoproterozoic zircon ages 1850-1810, 1750 Ma (Figure 2.5; Crowley, 1997b). This provides evidence that during the Proterozoic, the western edge of North America extended beneath what was later to become the Cordillera at least as far west as the Monashee Complex (Armstrong et al., 1991; Crowley, 1999). The metasedimentary sequences of the Selkirk Allochthon were deposited 300 km to the west of their present position (Brown et al., 1993; Brown, in press) and thrust northeastward over the Monashee Complex during the Cenozoic. Therefore the margin of the North American craton in Early Jurassic time must have extended at least as far west as the current position of the Intermontane belt.

Late Intrusions

There are three major episodes of intrusions in the Mount Copeland area that occurred after the intrusion of the basement orthogneisses. These are distinguished on the basis of lithology and deformation history; basement orthogneiss intrusions were described in the section on basement gneisses so are not included in this section.

The first major phase is an extensive, polydeformed nepheline syenite gneiss that occurs in association with calc-silicate (unit 5; Figure 2.4). The next two phases are smaller isolated intrusions that cross-cut the foliation and are distinguished based on composition. A white pegmatite was intruded at the end of deformation in the basement gneisses whereas a rusty to grey weathering lamprophyre is interpreted to have been intruded in the Cenozoic after deformation and metamorphism had ceased. The white pegmatite is interpreted as having intruded after Cordilleran deformation in the Monashee complex. Lamprophyre dykes are thought to have been associated with Eocene extension; the dykes are north-south trending, vertical and cross-cut all structures.

Nepheline Syenite Gneiss; Syenite Gneiss

Nepheline syenite gneiss dated near Mt. Copeland at 740 ± 36 Ma (Parrish and Scammell, 1988), forms a large intrusion in the Hiren Creek valley and on Copeland Ridge. Field observations of a similar syenite gneiss on Derickson Ridge reveal it has undergone several phases of deformation. This is also evident on geological maps where nepheline syenite gneiss occurs in close proximity to calc-silicate gneiss units and fold patterns in the syenite gneiss closely follow the metamorphosed cover sequence. The unit is only present with calc-silicate gneiss and provides a minimum constraint on the deposition age of the upper part of the sequence.

The syenite gneiss is a fine- to medium-grained, fairly resistant unit (Plate 2.7a), which weathers white to light grey, and has a pitted surface giving it a sugary texture (Plate 2.7b). The composition of the syenite gneiss is dominated by potassium feldspar, partially altered to sericite, along with biotite, plagioclase feldspar, and calcite. Quartz comprises up to 5% of the rock. Nepheline was not identified in any samples but it may have weathered away accounting for the pitted surface of the syenite gneiss outcrops.

Pegmatite

A syn- to post-kinematic white pegmatite occurs in the low pressure area of pull-apart structures in the basement gneiss. The pegmatite is sub-parallel to the foliation, but can also be traced at a high angle to S_2 (Plate 2.7c; see also Plate 2.1a). It could not be determined if there was more than one episode of pegmatite intrusion. The pegmatite was observed only intruding basement paragneiss and locally contained rafts of the host paragneiss within the pegmatite (Plate 2.7d). Contacts with the host rock are sharp and lack a reaction margin. When viewed in thin-section biotite grains appear torn at the contact with a sharp reduction in grain size in the host rock.

Crystals of quartz and feldspar are 0.5-1.0 cm in diameter whereas randomly oriented biotite grains generally range from 1 to 3 mm in diameter, but reach up to 4 cm as biotite 'books.' Quartz grains display slight undulose extinction and euhedral to subhedral feldspar crystals are highly fractured and partially replaced by sericite in the cores.

Lamprophyre dykes

Vertical north-south trending lamprophyre dykes cross-cut all contacts and structures (Plate 2.7f). In outcrop the dykes are fractured, strongly weathered, and appear

dark grey to rust coloured. The light green-grey aphanitic matrix encloses large phenocrysts of quartz, 1-5 mm in diameter, and potassium feldspar, 2-7 mm in diameter (Plate 2.7e). In thin section, subhedral potassium feldspar crystals display concentric zoning with a core of sericite alteration that can sometimes also be seen in hand sample. Lamprophyre dykes contain a chilled margin and the host rock is partially fractured with a stockwork pattern invading up to 1 m into the host.

Interpretation

There are several phases of intrusions in the Mount Copeland area, and although geochronology in the area is limited, correlation with equivalent units from other studies constrain the timing of events. Work on crystalline basement rocks of Frenchman Cap by J.L. Crowley (Crowley, 1997b, 1999) provides constraints on deposition ages for paragneiss and cover rocks and crystallization ages for the orthogneiss.

Geochronology on the Mount Copeland syenite gneiss (Parrish and Scammell, 1988) revealed an igneous crystallization age of 740 ± 36 Ma, and a metamorphic age of 59 ± 1 Ma that is interpreted as corresponding to tectonic events within the Monashee complex. The upper intercept (740 Ma) provides a minimum age constraint for deposition of the upper part of the cover sequence and for deformation in the Monashee complex. The lower intercept may provide an age for peak metamorphic conditions before the onset of extension.

The white pegmatite occurs in the 'pinch' between pull-apart structures and is also seen as large veins discordant to the paragneiss foliation. The boudins are present in the more competent amphibolite layers in association with basement paragneiss. Leucocratic intrusions are common in the Omineca Belt making correlations with other intrusive

suites difficult. As the pegmatites only intruded the basement gneisses they may correlate with a leucogranite crystallization age of 1862 Ma by Crowley (1997b) but this is unlikely as the intrusions do not appear to be deformed. Some leucocratic intrusions appear coeval with the third phase of deformation, however, another intrusion appears to post-date deformation indicating there may be more than one episode of leucocratic intrusions in the area.

There are no intrusive ages available for the pegmatites in the Mount Copeland area, but the lack of a foliation would presumably indicate they are younger than Cordilleran deformation and may be related to lithospheric extension and uplift of the complex. The timing could be constrained by dating the leucocratic pegmatite cross-cutting the paragneiss layers in the basement. The pegmatite could also be related to rifting in the Neoproterozoic. However, the most likely scenario is the pegmatite is a result of an anatectic melt produced during unroofing and decompression of the complex. The lack of a foliation is the main evidence for the pegmatite being younger than the Cretaceous.

Lamprophyre dykes are interpreted as intruding during extension in the Eocene; the dykes cut across all structures and contacts in hillsides. This would fit with tapping an upper mantle source during extension of the lithosphere approximately 45 Ma (Parrish et al., 1988), allowing intrusion of the lamprophyre. The dykes do not appear to have been deformed and if dated they may provide a minimum age for extension in the Mount Copeland area.

Figure 2.1: Simplified lithologic compilation map of the Mount Copeland area incorporating detailed mapping by Fyles (1970), Höy and Brown (1980), Bosdachin (1989) and this study.

Figure 2.2a: Lithology and description of basement gneisses and metasedimentary cover map units, see Map 2 for distribution.

	MAP UNIT	LITHOLOGY	UNIT DESCRIPTION	MODAL MINERALOGY
1.	Migmatitic Paragneiss	Plagioclase feldspar, potassium feldspar, biotite migmatitic paragneiss	<ul style="list-style-type: none"> - fine to medium grained - foliation defined by alternating 0.5-1.0 cm thick quartz-feldspathic layers, with melanosome dominated by biotite, and fine grained amphibole in amphibolites - transposed compositional layering defined by dm- to m-scale psammite and semipelite, and dm-scale 	15% plagioclase (An 32-37%) 50% quartz <5% potassium-feldspar 20-60% biotite ± garnet ± muscovite (retrograde)
2.	Orthogneiss	Potassium feldspar augen orthogneiss	<ul style="list-style-type: none"> - poorly foliated, medium to large grained, grey weathering resistant unit - 0.5-1.0cm potassium feldspar augen 	10% plagioclase (An 25-30%) 35% quartz 40% potassium feldspar 15% biotite ± chlorite (after biotite)
	Unconformity			
3.	Basal Quartzite	White quartzite (local quartz pebble metaconglomerate and pelite/grit)	<ul style="list-style-type: none"> - fine to medium grained, white to grey coloured blocky resistive unit - thin (<1cm) psammitic and micaceous layers of quartz, biotite K-feldspar and muscovite, partings - metaconglomerate contains quartzite clasts 2-3cm long 	95% quartz 5% muscovite ± biotite ± potassium feldspar
4.	Biotite Schist	Biotite schist	<ul style="list-style-type: none"> - fine to medium grained - white-brown mottled appearance in close-up view - forms distinct rusty weathering cliffs 	40% potassium feldspar 25% biotite 20% quartz 5-10% plagioclase feldspar <5% sillimanite (fibrolite) ± muscovite

	MAP UNIT	LITHOLOGY	UNIT DESCRIPTION	MODAL MINERALOGY
5.	Calc-Silicate Gneiss	Hornblende calc-silicate gneiss	<ul style="list-style-type: none"> - fine to medium grained, medium to dark green-grey coloured with uneven weathering on surface - rare hornblende crystals up to 1 cm in length within the foliation - weak foliation seen in thin section, possibly from dominance of calcite, some muscovite folded 	up to 50-60% calcite (dolomite?) 5-20% diopside 5-10% tremolite <10% biotite 5% muscovite ± sillimanite, epidote
5a.	Quartzite	White quartzite	<ul style="list-style-type: none"> - thinly layered, blocky, fine to medium grained - large lenticular lenses with m-scale thickness 	95% quartz 5% muscovite
5b.	Grey Gneiss	Biotite hornblende gneiss	<ul style="list-style-type: none"> - dark green-grey coloured in outcrop - lenses of quartz up to 10cm thick outline tight folds 	30-35% potassium feldspar 20-25% biotite 20% quartz 15% hornblende 0-10% Plagioclase up to 5% calcite
5c.	White Crystalline Marble	White Crystalline Marble	<ul style="list-style-type: none"> - buff to grey weathering, pure to slightly siliceous marble - associated with sulphide layer 	95% calcite ± biotite ± muscovite
*5d.	Calc-Sil Gneiss	Hornblende calc-silicate gneiss ±sillimanite schist	Not seen in study area - similar to calc-silicate above	calcite diopside biotite muscovite ± sillimanite, epidote, tremolite
*6.	Biotite Sillimanite Schist	Biotite sillimanite schist	Not seen in study area, occurs on side of Mount Copeland - medium grained, brown-rusty weathering - brown-white mottled appearance - similar to unit 4, biotite schist	potassium feldspar biotite quartz plagioclase feldspar sillimanite ± muscovite

* Due to mapping method, unit was not sampled; description based on interpretation of equivalent units from Fyles (1970), Psutka (1978), and Scammell (1986).

Figure 2.2b: Lithology and description of intrusive map units

	MAP UNIT	LITHOLOGY	UNIT DESCRIPTION	MODAL MINERALOGY
N	Nepheline Syenite Gneiss	Nepheline Syenite Gneiss, Syenite Gneiss	<ul style="list-style-type: none"> - associated with calc-silicate - sugary weathering appearance in outcrop 	75% potassium feldspar 7% biotite 5% plagioclase feldspar 3% calcite <5% quartz
	Pegmatite	K-feldspar, quartz pegmatite dyke	<ul style="list-style-type: none"> - variable locations, sub-parallel to foliation - medium to coarse grained. - sporadic biotite 'books' up to 5cm in diameter and 1cm thick - only seen within basement paragneiss - unit 10a of Fyles (1970) - majority of intrusion is ~2 m wide but varies from veinlets to 4 m thick 	45% quartz 45% potassium feldspar 5% biotite 5% plagioclase feldspar ± tourmaline
	Lamprophyre	Lamprophyre dyke	<ul style="list-style-type: none"> - north-south trending, highly weathered and fractured - stockwork present in host - phenocrysts of K-feldspar (1cm) and quartz (1cm) - feldspar phenocrysts centre partially altered to sericite - 1 to 4 m wide 	phenocrysts 15% potassium feldspar 10% quartz 75% aphanitic matrix

Figure 2.3: Correlation of lithologic sequences in the Jordan River area.

	Fyles (1970)	Unit	Unit	Bjornson (2003)	
	Lithology (from Table 1 of Fyles)	Unit	Unit	Lithology	
	Lamprophyre dykes			Lamprophyre dykes	
	Pegmatite dykes			Pegmatite dykes	
	Porphyritic granite gneiss	B	2	Basement Orthogneiss	
	Nepheline syenite gneiss	A	N	Nepheline syenite gneiss	
	Amphibolite			Amphibolite layers in Paragneiss unit	
	Brown mica schist, grey gneiss	10	1	Basement paragneiss	
	White quartzite	9	2	Basal Quartzite; semipelite, quartz pebble metaconglomerate	
	Biotite sillimanite schist	4	5d	Calc-silicate gneiss, sillimanite kyanite schist, biotite gneiss	
	Lead-zinc sequence	5	5c	White crystalline marble	
	Greyish and greenish-grey gneiss	6	5b	Biotite hornblende gneiss; quartzite, marble	
	White quartzite	7	5a	White quartzite; minor schist, semipelite	
	Calc-silicate gneiss and marble	8	5	Calc-silicate gneiss and marble; minor schist, quartzite	
	Bews Creek Fault			Conformable contact	
	Mica schist and quartzite	3	4	Micaceous schist; biotite sillimanite schist	
	Quartzite and conglomerate	2	3	Basal quartzite; semipelite, quartz pebble metaconglomerate	
	Mixed gneiss	1	1, 2	Basement gneisses; Paragneiss, Orthogneiss

Thicknesses are approximate due to variability of units and are presented as maximums based on field observations and interpretation from maps. Some units thin considerably.

Figure 2.4: Lithologic sequence for the Mount Copeland area of southern Frenchman Cap dome and correlated geochronological data from previous studies in the Monashee complex. References as indicated above.

Figure 2.5: Tectonic subdivisions of the western North America craton, modified after Hoffman, 1989. Major tectonic elements include: 1, Archean craton, 3.3-2.6 Ga.; 2, 2.0-1.8 Ga. orogenic belts; 3, 2.4-2.1 Ga. crust; 4, 2.0-1.8 Ga. magmatic belts; 5, crustal wedges along the Snowbird Tectonic Zone, locally 2.3 Ga.; 6, juvenile Proterozoic terranes of the Trans-Hudson Orogen. Structural elements: GSL, Great Slave Lake shear zone; MC, Monashee complex; T, Thorsby Terrane; R, Rimbey magmatic arc. Wopmay orogenic belt; F, Fort Simpson arc (1.85 Ga.); G, Great Bear arc (1.88 to 1.84 Ga.); H, Hottah (1.95-1.90 Ga.) arc built on 2.3-2.1 Ga. crust; N, Nahanni (possibly 2.3-2.1 Ga.).

Plate 2.1: Representative field photographs of basement paragneiss. (a) Pulled-apart and folded amphibolite layer of basement paragneiss with leucocratic material filling in the pull apart. (b) Layered amphibolite and semi-pelitic paragneiss with disharmonic folding, station 04. (c) Close-up of pelitic paragneiss in an area of high strain, station 262. (d) Migmatitic paragneiss in an area of high strain, station 262.

Plate 2.2: Representative field photographs of basement orthogneiss; photos are of the unit Fyles (1970) termed Porphyritic granite gneiss. (a) Relict phenocrysts in foliated orthogneiss, station 112. (b) D3 shear bands folding S2 and flattened phenocrysts of basement orthogneiss. (c) Rusty weathering orthogneiss with shear bands, station 118.

Plate 2.3: Contact of basement gneisses and cover sequence. (a) Close-up of quartzite and basement paragneiss, station 230. (b) Inverted sequence with basement paragneiss (right) on top of basal quartzite, station 124. (c) Steeply dipping contact in the South Hiren Valley at station 230; basement gneisses are on the right (south) side of the photo.

Plate 2.3: Contact of basement gneisses and cover sequence (continued).
 (d) Layer of paragneiss overlying basal quartzite, station 247. Also seen at this station was a metaconglomerate next to the basement gneiss. (e) Possible scour marks on top surface of basement paragneiss with basal quartzite on top. Sequence is right way up, station 228. (f) Close-up view on pavement surface of wavy contact dipping to the left. Contact is between quartzite (right) and semi-pelitic paragneiss in the South Hiren Valley.

Plate 2.3g: Oblique view to the northwest of interfingering quartzite and paragneiss on an uneven surface. Contact continues with northeastern vergence off to the right of the photo. Rocks dip to the left (south).

Plate 2.4: Representative photographs of basal quartzite. (a) Close-up of typical white quartzite in outcrop, station 58. (b) Polished hand sample of pure quartzite with schistose layer at bottom. (c) Metaconglomerate at base of quartzite, station 165. Quartzite pebbles are flattened and aligned within the foliation. (d) Polished hand sample of pure quartzite with fine schistose laminations near base that may represent primary cross-bedding. Rock samples shown actual size.

Plate 2.4: Representative photographs of basal quartzite (cont.). (e) Steeply dipping, flaggy quartzite in hillside. This example contains semi-pelitic layers and displays the typical strong foliation. View is to the north. (f) Close-up of flaggy quartzite interlayered with thin layers of psammite. (g) Steeply dipping quartzite in hillside, view is to the north.

Plate 2.5: Representative field photographs of rock units of the metasedimentary cover sequence. (a) View of pelite layer in hillside on north side of Derickson Ridge. Structurally below the pelite is the basal quartzite (unit 3) and paragneiss (unit 1).

Plate 2.5: Representative field photographs of rock units of the metasedimentary cover sequence (continued). (b) Rusty weathering pelite at base of photo seen in sharp contact with overlying calc-silicate. Photo taken near top left of Plate 2.5a. (c) Typical banding and weathered appearance in calc-silicate, near Camp Lake. (d) Close-up of F1/F2 in calc-silicate with strong, recessive weathering. (e) Tight folding in impure marble, station 256.

Plate 2.5: Representative field photographs of rock units of the metasedimentary cover sequence. (f) View to the southeast of the contact of pelite and impure marble, station 262, about 50m SE of Plate 2.5a.

Plate 2.6a: Lithological succession including the basement-cover contact. The basement-cover contact is located at the bottom of the gully, just to the right of the lake, and at the top of the gully. The established lithologic sequence was followed up-section from basement next to the lake in the background, up the gully, with an inverted sequence seen at the top. No folds were apparent. Foliation is vertical, view is to the north.

Plate 2.6b: View to west of lithologic sequence in hillside. The contact seen in Plate 2.5c is located approximately 200 m south of this photo.

Plate 2.7: Intrusions in the Mount Copeland Area. (a) View to north of strongly foliated syenite gneiss along Derickson Ridge, station 267. (b) Surface of syenite gneiss along Hiren Rd, station 133. (c) Leucocratic intrusion of granodiorite composition invading the host paragneiss, station 239. The intrusion has no apparent fabric and is discordant to the foliation (S2) and layering in the paragneiss.

Plate 2.7: Intrusions in the Mount Copeland area (continued). (d) Late stage leucocratic intrusions into paragneiss, cutting across S2 and a fold in lower right of the photo. Note pod of paragneiss within the intrusion. (e) Polished hand sample of lamprophyre intrusion with phenocrysts of quartz and feldspar; sample shown actual size. (f) Approximately 1 m wide intrusion of N-S trending lamprophyre dyke, cutting across the foliation

CHAPTER 3.

STRUCTURAL GEOLOGY OF THE MOUNT COPELAND AREA

Introduction

Field mapping was done at 1:20,000 scale in selected areas, labeled Camps I to V. See Map 1 for station and camp locations and Map 4 for the geologic map of the region mapped by Fyles (1970). Structural data will be interpreted and described based on these locations. Map 4 includes structures that were identified in the field and in cross-sections (Figure 3.1).

Sedimentary rocks of the Mount Copeland area were affected by two metamorphic events and five phases of deformation including two episodes of transposition. The most prominent fabric is a penetrative regional transposition foliation (S_2). Poles to foliation show there is little difference, if any, between data for the basement gneisses and cover sequence. In some areas the basement and cover rocks are so highly infolded it is hard to distinguish the units. Therefore, structures in the basement and cover are treated together.

The deformation hierarchy has been complicated by at least one episode of transposition but the phases of deformation were interpreted based on the overprinting relation of fabric elements. Minor structures associated with all phases of deformation are preserved in the area to varying degrees. Minor structures are used to characterize the style of folding associated with larger-scale structures of the same generation of deformation. By also incorporating the lithologic sequence, large-scale structures were identified on geologic maps of the field area. The results of this chapter will be used in Chapter 5 to interpret the three-dimensional geometry and the tectonic evolution of the Mount Copeland area.

The principal fabric and structural elements identified in the Mount Copeland area in the course of this study include:

- S₀ - original bedding
- S_p - inferred Paleoproterozoic transposition foliation in basement gneisses that pre-dates deposition of the cover sequence. Largely overprinted by deformation during Cordilleran orogenesis so may not be preserved
- deposition of cover sequence
- S₁, F₁ - folding of S₀ about F₁ fold axes, with S₁ axial surface and associated axial planar cleavage
- S₂, F₂, L₂ - folding of S₀₊₁ about F₂ fold axes with S₂ axial surfaces. L₂ mineral lineation parallel with F₂ fold axes and development of composite transposition foliation S₂. Development of a shear band foliation (S₂) in orthogneiss. The Eagle Pass synform-antiform pair, with fold limbs close to 10 km long, are the largest features associated with phase two deformation
- Ls₂ - mineral stretching lineation normal to the trend of F₂ hinge lines, composed of trains of, or elongate grains of, quartz, feldspar or biotite
- S₃, F₃, L₃ - folding about F₃ axes and development of crenulation cleavage (L₃) in pelitic units. Km-scale synforms and antiforms, including the Copeland Synform
- F₄ - folding and warping about F₄ axes
- D₅ - post-metamorphic brittle faulting

Proterozoic History

The correlation of the Monashee Mountains (Dawson, 1898; Daly, 1915) with the Precambrian rocks of the Canadian Shield has been supported by geochronological (Crowley, 1999) and geochemical (Armstrong et al., 1991) data. Evidence also indicates a Paleoproterozoic history for the basement gneisses, which involved: (a) intrusion of

granitic plutons, (b) deformation including formation of migmatites, (c) exhumation, and (d) the beginning of deposition of the cover sequence. As was discussed in the lithology chapter, erosion of the basement gneisses occurred before the base of the cover sequence began to be deposited in the Paleoproterozoic. While there is some evidence for a Paleoproterozoic history in the Monashee complex, the Paleoproterozoic deformation is best preserved in the deepest exposed structural level of the Monashee complex (Crowley and Schaub, 1994), while at higher levels such as the Mount Copeland area, deformation associated with Mesozoic to Tertiary convergence may have completely overprinted any Paleoproterozoic structures. Subsequent to the Paleoproterozoic history and deposition of the sedimentary cover sequence, both sequences of rocks experienced a common deformation history. The Paleoproterozoic history will not be considered in the structural hierarchy but the foliation in the basement rocks may be considered a composite of Paleoproterozoic ($S_0 + S_P$) and Cordilleran ($S_1 + S_2$) deformation.

Primary origin of basement gneisses

The layering in the basement gneisses (S_2) is a composite of compositionally distinct units (S_0) and metamorphic layering (S_1) that were folded and rotated into parallelism with the axial planes of isoclinal F_2 folds. The preserved dm- to m-scale layering of amphibolite, psammite, and semi-pelite represent transposed primary sedimentary bedding (S_0); see also Chapter 2, lithology. Segregation of layers with contrasting mineralogy can be produced by in situ melting during metamorphism. The thickness and scale of repetition of amphibolite, psammite and semi-pelite in the study area are large and could only be the result of original sedimentary bedding. Thinning of the layers is largely attributed to attenuation of fold limbs during tectonic processes. The

well-defined interlayering of psammite and semi-pelite suggests protoliths of impure sandstone and clay, respectively, with sporadic occurrences of 5-30 cm thick amphibolite layers interpreted to be metamorphosed mafic volcanoclastic rocks or mafic sills.

Basement orthogneiss, forming a large portion of the core of Frenchman Cap dome (Map 2), were intruded as granitic plutons prior to deformation of the basement paragneisses and are generally separated from the cover sequence by an intervening layer of paragneiss. Structural fabrics are not prominent in the orthogneiss but in the Mount Copeland area it appears as though these intrusions experienced all phases of deformation associated with Cordilleran orogenesis. The appearance of the felsic orthogneiss is variable but at several localities relict igneous phenocrysts were observed; no other igneous textures or features were identified. A comparison of orthogneiss outcrops shows the primary phenocrysts are better preserved at higher levels (Plate 3.1g and h); to the south in the field area (Map 3). Field mapping did not extend further north into Frenchman Cap dome to see if this variation persisted into the core of the dome. The more foliated orthogneiss (Plate 3.1h) was observed close to the cover sequence in the northern part of the field area, where all units occur as thin layers, which may indicate an area of high strain. The orthogneiss is interpreted to have intruded before Mesozoic metamorphism and deformation as it displays evidence for several phases of deformation and may be correlated based on mineralogy and structures with a 2077 ± 2 Ma augen orthogneiss (Crowley, 1999) further north.

Structures in the orthogneiss are not as prominent as in the paragneiss, and in some cases it appears as though the orthogneiss may be largely undeformed (Plate 3.1e). This may have led Fyles (1970) to interpret a large body of orthogneiss near Neil Lake

(see Figure 1.3 and Map 3) as a separate late stage granitic intrusion he called porphyritic granite gneiss (his Unit B). As will be shown in the following sections, fabric elements and field relationships indicate the porphyritic granite gneiss of Fyles (1970) has experienced several phases of deformation, and due to its close stratigraphic association with basement paragneiss, it is herein reclassified as part of the suite of granitic orthogneiss to augen orthogneiss of the basement. There is a noticeable progression of increasing deformation closer to the margins of orthogneiss bodies. The orthogneiss in the southern part of the field area (Plate 3.1e and g) is not as strongly foliated as the orthogneiss further to the north (Plate 3.1h). This may represent different intrusive episodes or a variation in strain during Cordilleran deformation but there is not enough data available in the area to substantiate either claim. Geochronology could be used to confirm the age of the orthogneiss and the proposed correlation.

Paleoproterozoic deformation of the Basement Gneisses

Phase one structures in the basement do not necessarily correspond to phase one structures in the cover. Upper-amphibolite facies metamorphism is interpreted to have occurred at 2060 ± 1 Ma based on monazite growth in pelitic schists (Crowley, 1999). Deformation with formation of migmatites in the basement rocks occurred between 2077 ± 2 and 1848 ± 3 Ma based on dated intrusions that contain and post-date the fabric, respectively (Crowley, 1999). The early deformation that affected the basement rocks would have occurred before deposition of the cover sequence (approximately 1852 ± 4 Ma; Crowley, 1997b). However, it cannot be determined if the migmatitic gneissosity in the Mount Copeland area is a result of Paleoproterozoic and/or Cordilleran deformation; this could be resolved by trying to date the age of deformation in the orthogneiss.

Primary origin of the Cover Sequence

Gneisses of the cover sequence exhibit a large-scale compositional layering interpreted to be transposed bedding. Suggested protoliths for the sequence are quartz arenites, shales, impure sandstones, and impure limestones, representing a typical shallow shelf or platformal assemblage (see Chapter 2, lithology). Transposition has locally resulted in an inverted (Figure 3.1) and repeated sequences, but where present with basement gneisses the stratigraphic facing direction could be inferred based on regional correlations (Scammell and Brown, 1990 and references therein).

The main intrusive body in the cover sequence is the Mount Copeland syenite gneiss that has been described by Fyles (1970) and Currie (1976) and more recently, including radiometric dating, by Parrish and Scammell (1988). Structural observations show the syenite gneiss has experienced several phases of deformation indicating it was intruded before deformation and metamorphism occurred in the cover sequence. The 740 ± 36 Ma (Parrish and Scammell, 1988) intrusive age for the syenite indicates the immediate host calc-silicate gneisses must be older and all deformation and metamorphism in the cover sequence is younger than this age. As the syenite gneiss experienced the same history as the rest of the cover sequence, the cover rocks can be treated as a single package.

The laterally extensive carbonatites and alkaline magmatic rocks (Scammell and Brown, 1990) have been interpreted as a result of Late Proterozoic extension along the margin of pre-Cordilleran western North America (Hoffman, 1989), and may be associated with the initiation of Windermere rifting about 760 Ma (Anderson and Davis, 1995). The nepheline syenite gneiss is highly folded but the disappearance of the unit in

cross sections (Figure 3.1) is a feature of primary igneous origin. Marker units in the cover sequence are observed throughout the map area (Map 3) and regionally (Map 2) so the disappearance of an igneous suite on maps and cross sections is likely a result of an originally discontinuous intrusion.

Cordilleran Deformation in Basement and Cover Rocks

Five generations of structures are recognized in rocks of the Mount Copeland area largely based on overprinting relations (Plate 3.2). Structures are well preserved in the basement paragneiss so this unit will be used to characterize deformation in the basement, however, the orthogneiss will be described where correlation of structures with known phases is possible. The regional foliation (S_2) is the most prominent structural feature in the study area and defines the antiformal nature of Frenchman Cap dome. This prominent foliation in the basement and cover gneisses, defined by the preferred orientation of biotite and migmatitic layering parallel to compositional layering, is a typical transposition foliation (Hobbs et al., 1976). Figure 3.3 is a graphical interpretation of transposition and Figure 3.3d shows how transposed bedding may be incorrectly identified as original bedding. Phase two structures are commonly seen at the outcrop scale as rootless, isoclinal folds (Plate 3.3d). The foliation (S_2) is interpreted to be regional in extent (Psutka, 1978; Scammell, 1986; Journeay, 1986; Johnson, 1994; Crowley, 1997b, 1999). Figure 3.4 is a graphical interpretation of the phases of deformation and may be a convenient reference for the discussion in the following sections in this chapter.

Phase 1 (D_1) Structures

The first phase of deformation, producing isoclinal folds, was recognized due to

overprinting by later deformation which produced Type II and III interference patterns (Plate 3.2a). Type II interference patterns are interpreted as a product of phase one deformation and later non-coaxial deformation. Hinge lines recorded from Type II interference patterns (Plate 3.2a) show a large scatter in data (Figure 3.5) but with only 15 measurements no reliable inference regarding the geometry of F_1 folds can be made. Orientation of phase one fold hinge lines (L_1) trace a north-northwest to south-southeast striking great circle on stereonet. On maps the trace of the axial surface appears more north-south trending, and while generally inferred to be east verging folds, due to later folding the orientations are not reliable. Folds are intrafolial with respect to the foliation and axial surfaces are parallel to compositional layering ($S_{0+1} = S_1$) (Figure 3.5b).

Attenuated limbs and thickened hinge zones are seen on geologic maps and in outcrops. This includes a thinned layer of paragneiss inferred from cross sections (Figure 3.1b) enclosing a large exposure of quartzite in the hinge zone in the proximity of Camp II (Map 4 and Figure 3.2). F_1 folds are preserved as remnant, pre-peak metamorphic, isoclinal folds with limb lengths of approximately 1-2 kilometres. Figure 3.1a also shows that the basement rocks were involved with this generation of deformation.

Interference patterns identified in outcrops (Plate 3.2) and on geologic maps (Map 3) confirm the existence of the first phase of deformation, but it is often difficult to reliably identify phase one folds, especially on cross sections as the hinge lines may be in the plane of view. This difficulty is also largely a result of overprinting from later deformation obscuring phase one structures. For the most part, macro-scale phase one folds were identified on maps and cross sections where inverted or repeated stratigraphy cannot be explained by phase two or three folding. These appear as northerly trending,

easterly verging km-scale folds.

The definitive assignment of phase one folds is not always possible and hence some isoclinal folds thought to be phase one may actually be phase two, and vice versa (Plate 3.2e and f). As phase one (Plate 3.2a) and two (Plate 3.2b) folds are both intrafolial with respect to the foliation they can only be reliably assigned to a generation of deformation where overprinting relations exist (Plate 3.2c and d). Plate 3.3 shows several examples of isoclinal folds in outcrop and hand samples that can only be assigned as F_1/F_2 due to lack of overprinting evidence.

Phase 2 (D_2) Structures

Cordilleran deformation, associated with upper-amphibolite facies metamorphism during the second generation of deformation, produced large-scale isoclinal folds and the penetrative regional transposition foliation (S_2) (Plate 3.4). Minor structures include preserved asymmetric, isoclinal folds, with attenuated limbs and thickened hinge zones.

A comparison of the trend of hinge lines recorded from Type II and III interference patterns, representing F_1 and F_2 folds respectively, overprinted by later deformation, show there is little to no variation in the scatter and trend of data (Figures 3.5 and 3.6). While approximately aligned along a northwest-southeast striking great circle, there is a large amount of scatter in the plunge of hinge lines. Fold limbs of first and second generation structures are both intrafolial with respect to the regional foliation and are difficult to differentiate. Compositional layering (S_0) is approximately parallel to S_2 ($S_{0+1+2} = S_2$); S_{0+1} is folded around the hinges of F_2 folds but is axial planar to F_1 (Plate 3.2a and b). Little information remains about phase one and two structures in the rocks and in many examples it cannot be determined if folds are F_1 or F_2 so mesoscopic folds

will generally be listed as F_1/F_2 . Phase 2 folds have stretched and pinched limbs so sense of vergence could not always be ascertained (Plate 3.3a and g). While there is good evidence for two phases of deformation, attenuated limbs limit the usefulness of vergence reversals in structural interpretations and repeated stratigraphy must often be assumed.

Hinge lines of phase two folds scatter about a west-northwest and southeast orientation (Figure 3.6) with a southwesterly plunging Ls_2 stretching lineation measured from elongate quartz grains (Figure 3.7) on foliation surfaces. Poles to S_2 (Figure 3.8) show some scatter but rotate about a northwest-southeast axis, from the overprinting effects of phase three folds. Foliations (S_2) generally dip $50-60^\circ$ to the southwest, but also $60-70^\circ$ to the northeast on the short limb of F_3 's (Figure 3.8). Foliations at Camp I (Figure 3.8a) consistently dip 70° to the south. The trace of hinge lines (L_2) about two great circles (Figure 3.5 and 3.6) can be a result of rotation of initially straight hinge lines by drag folds and refolding. The orientation of the great circle aligned with the S_2 foliation (Figure 3.8) could indicate refolding by phase three deformation after formation of drag folds during phase two. The presence of sheath folds in the northern part of the field area support this interpretation. Hinge lines predominantly plunge $\sim 30^\circ$ to the northwest, consistent with the dip of the foliation, and also along a great circle that strikes north-northwest to south-southeast showing rotation of the L_2 hinge lines through 180° . This pattern is best illustrated in Figure 3.6b.

Associated with the foliation, and contributing to its development is a small-scale gneissosity of alternating light coloured quartzofeldspathic bands and dark coloured biotite bands in pelitic schists. Migmatitic basement paragneiss displays a strong cm-scale foliation of alternating bands and lenses of quartz and feldspar (leucosome) in a fine

to medium grained micaceous matrix (Plate 3.4a and e). Calc-silicate gneisses have thick bands of quartz lenses and appear migmatitic. Some quartzite samples contained thin bands of finer grained quartz apparent in thin sections and hand samples. This alternation in quartzite is associated with areas of high strain and in some examples may be classed as protomylonite; see description on quartzite in Chapter 4 (Petrography and Microstructures). This may have been where Fyles (1970) identified the Bews Creek Fault. Quartz grains show a preferred lattice orientation and evidence of static recrystallization when viewed in thin-sections under the microscope. Clusters of rodded quartz with feldspar and sillimanite (faserkiesel) are apparent in pelitic schists. See metamorphic chapter for further discussion.

Phase two structures are preserved throughout the study area as Type III interference patterns from the overprinting of phase two by later deformation (Plate 3.2b). This produced a swirled pattern (Höy, 1987) and rootless isoclinal folds (Plate 3.3e and g). These structures along with sheath folds (Plate 3.5), which show up well on the pavement surfaces, occur in all units but predominately in calc-silicate gneisses. Both sheath folds and F₂ fold vergence (where determined) indicate a northeast sense of shear (Plate 3.5b). Lithological units are predominately discontinuous and could not be traced around large-scale fold closures. This lack of continuity may be the result of complete transposition by F₁/F₂ folds or disruption of large folds by later stage extensional deformation (D₅). Pull-apart structures (Plate 3.6) indicate a high competency contrast between layers and units. Amphibolite within paragneiss commonly forms boudins (Plate 3.6b and d), which when rotated give a top to the northeast sense of shear.

The margin of orthogneiss units near paragneiss are highly foliated (Plate 3.1h)

but the effects do not penetrate far into the orthogneiss, leaving the core relatively unaltered (Plate 3.1g). This may represent a strain gradient due to the competency contrast between paragneiss and orthogneiss. The foliation in orthogneiss is poorly- to moderately-well developed, with a shear band foliation (S_2) on the limbs of F_2 folds, at a high angle to the dominant foliation, S_1 (Plate 3.1f). This becomes progressively rotated into parallelism with the foliation near the basement-cover contact. Seen in thin section is a poorly defined biotite foliation ($S_{1/2}$) along with annealed quartz displaying undulose extinction and subgrains (Plate 3.1d). Unlike the paragneiss, the orthogneiss does not have a strong gneissosity and is not migmatitic (Plate 3.1c), but where present the orientation of the foliation in the orthogneiss appears similar to the foliation in the adjacent paragneiss. The poorly defined biotite foliation ($S_{1/2}$) is interpreted as occurring contemporaneously with Cordilleran deformation (D_2) and metamorphism (M_2) but due to the afore mentioned competency contrast, it is not well developed.

The Eagle Pass synform and antiform are the largest structural features associated with phase two deformation. Although refolded by Phase three deformation lengths of long limbs are ~8 km and the overturned limb of the antiform seen in cross section lines A-A' and B-B' (Figure 3.1a and 3.1b) may contain an inferred attenuated limb. Folds are recumbent and slightly plunging. Vergence is generally northeast but as some areas may be structurally inverted and fold limbs can be attenuated, the true sense of vergence could not always be determined.

Phase 3 (D_3) Structures

Structures associated with the third phase of deformation are post-metamorphic, northeasterly verging, open to tight folds. Interlimb angles are generally 58-80° but did

range up to 100° (Plates 3.7 and 3.8). Phase three deformation affected the regional transposition foliation (Plate 3.7) and lacks an axial planar cleavage. These structures are upright to overturned (Figure 3.1) and control the structural geometry over much of the field area (Map 4). Minor structures associated with F_3 folds show reversals in asymmetry across the area indicating repeated northwest-southeast trending synforms and antiforms, but at a larger scale all structures verge to the north/northeast (Plate 3.7c).

Hinge lines of F_3 folds plunge an average 20° towards the northwest and southeast (Figure 3.9). Data from Camps I, II and III have good clusters of points, but even data from Camps IV and V show the same general trend. Hinge lines from Camps II, III, and IV also plot along a northwest to southeast striking great circle, which would indicate rotation of F_3 fold hinge lines during or after deformation. There may be some undulation of the crest lines of folds that could account for some scatter which was observed in the field and can be inferred on cross sections. However, the rotation of the hinge line of F_3 folds through 180° defined by great circles on stereonet, would not be produced by the simple undulations. Poles to axial planes of F_3 folds indicate steeply dipping axial planes with surfaces predominantly dipping about 70° to the southwest but also $\sim 60^\circ$ to the northeast (Figure 3.10). A comparison of axial plane data (Figure 3.9) with poles to S_2 's (Figures 3.8) show similar trends indicating D_2 and D_3 may have been affected by a progressive deformation event. Phase three structures are penetrative and were observed in thin section folding biotite in the S_2 foliation (Plate 3.7e-h).

Structural data (L_{S2} and S_2) was recorded from the exposed foliation surface of a large-scale, F_3 fold at Camp V in paragneiss (station 265; see Map 1) near the basement-cover contact. Analysis of the data on stereographic plots for L_{S2} mineral (elongate

quartz) stretching lineations (Figure 3.11d) show the trace of a great circle, indicating the mechanism of deformation was flexural flow folding for phase three deformation.

Foliation measurements (S_2) from the same fold (Figure 3.11c) indicate, at least for this area, phase three deformation resulted in cylindrical folds. The fold axis is approximately 05/180 (Figure 3.11c) close to the average (Figure 3.9e) which can also be inferred from poles to S_2 for camps II and V (Figure 3.8). The disharmonic fold patterns can result in a random appearance of structures in outcrop (Plate 3.4h) and in hand sample (Plate 3.8d).

Interpretation of F_3 folds using dip isogon patterns (Plate 3.8) indicate Class 1b and 1c folds (Ramsey, 1967) for basement paragneiss. Other units had more disharmonic folds and did not prove useful for classification based on dip isogon patterns. Alternating Class 1b/1c and Class 3 folds are due to interlayered competent and incompetent units (Twiss and Moores, 1992), and as the classification used the more competent layers, it is likely the calc-silicates would yield Class 3 folds. This conclusion is similar to that of Journeay (1986) who described Class 1c folds alternating with Class 3 folds, in competent and incompetent layers, respectively.

Phase three structures are well preserved in rocks, probably due to this phase occurring after the metamorphic peak. The region was still in the ductile regime as the biotite foliation and layering are folded, although lower than the peak-metamorphic conditions associated with Cordilleran deformation as there is no axial planar cleavage nor metamorphic minerals associated with this phase. Many major structures are part of this phase such as the Copeland Synform, which can be seen prominently in the hillside of Mount Copeland (Plate 3.9), and the Hiren Creek antiform and synform. These structures were recognized previously (Fyles, 1970; Höy and Brown, 1980; Journeay,

1986) although not all were named.

The Hiren Creek antiform-synform pair and the Copeland antiform are interpreted on cross sections to extend the length of the field area. These structures produced the northwest to southeast trend of structures as seen on maps (Map 4). Other phase three folds were identified in the field area (Map 4), however, most of these folds are not as significant and die out along strike of the axial surface.

Phase 4 (D₄) Structures

The final phase of folding in the Mount Copeland area is rarely observed as approximately east-west trending, large, open warps and may have contributed to some of the scatter in S₂ and other data. These folds were not observed overprinting F₁/F₂ folds, however, at Camp V, an F₄ antiform clearly overprints a third generation synform (Plate 3.10). The limb lengths observed at this locality are on the order of tens of metres. The wavelength of F₄ folds in this locality may be smaller accounting for their preservation. The interlimb angle is 170° or more and the axial surface is generally upright with a horizontal hinge line (L₄). Rocks were still in a ductile regime as the foliation (S₂) is folded, but this was presumably during the waning stages of metamorphism as no metamorphic minerals are associated with this phase and may represent a progressive cooling from peak metamorphic conditions. Several folds seen at Camp V are the only evidence for the existence of the fourth generation of folding.

The tectonic significance of phase four structures is unknown, and due to its sporadic development in the field, regional correlations may not be possible. The folding may represent the final arching of the dome related to extension in the Monashee complex. Based on the orientations of F₄ folds, the association with later extension

seems reasonable. The size of the folds and limited extent of mapping do not allow for the possible correlation of F₄'s with larger scale structures.

Phase 5 (D₅) Structures

The final phase of deformation is expressed as late brittle faults of minor displacement. It may relate to crustal-scale extension that brought the once mid-crustal rocks of the Monashee complex near to the surface along a ductile to brittle transition. Brittle faults were observed from the offset of lithologic contacts on maps, but displacements are generally small, on the order of 1 to 10's of metres. Faults are typically vertical to steeply dipping and strike roughly north-south; Fyles (1970) also recognized another dominant strike of about 30°. The faults are not well exposed but generally inferred as normal as interpreted from map patterns. Rocks were not greatly affected by phase five deformation and phase 4 ductile deformation may have progressed into phase 5 brittle deformation similar to the observed ductile to brittle transition in the Columbia River Fault zone (Psutka, 1978; Lane, 1984). Crustal-scale extension during the Eocene is well documented (Lane, 1984; Parrish et al., 1988; Johnson, 1994), however, in the field area it resulted in only minor affects.

Late stage intrusive lamprophyre dykes are commonly contained within or associated with the brittle faults (D₅). The dykes have chilled margins and display a stockwork pattern in the host rocks that makes them appear fractured at the contact. The intrusion of lamprophyre dykes may have released of a volatile phase during emplacement as another lamprophyre dyke is associated with fault breccia in calc-silicate (Plate 3.12c and d). The dykes are north-south trending and cut across all units and most structures (Plate 2.7f). The lamprophyre dykes are not foliated but petrographic analysis

of quartz phenocrysts observed in thin section display minor undulose extinction. Quartz with undulose extinction provides evidence that the dykes were affected by some deformation but the strain associated with deformation is not significant.

Interpretation

A high competency contrast between paragneiss and orthogneiss resulted in the paragneiss absorbing the bulk of the strain and may have allowed the orthogneiss to preserve relict, igneous phenocrysts. The cover sequence appears to have been weak relative to the basement rocks, and absorbed a significant amount of the strain from the Cordilleran orogen.

Preserved fabrics in the rocks allow interpretation of the structural geology of the Mount Copeland area and indicate the area has experienced a long and complex deformation history. Minor evidence exists for Paleoproterozoic (pre-cover) deformation but this could not be confirmed so only Mesozoic deformation is considered. Pre- to syn-metamorphic intrafolial phase one folds are rarely preserved but have locally resulted in inverted sequences. Phase two deformation occurred under conditions of upper-amphibolite facies metamorphism and produced the large-scale antiform-synform pair (Eagle Pass Synform and Antiform) that resulted in an inverted sequence of basement gneisses on top of the cover sequence. These units are seen infolded at the contact. Map scale folds are largely a result of phase three deformation. Several third generation folds, such as the Hiren Creek antiform-synform, are inferred to span the length of the field area. F₃ folds are common in outcrop folding the S₂ foliation, which can also be seen in hillsides. The geometry of the Mount Copeland area will be discussed in more detail in Chapter 5.

Stereographic plots show most of the data sets for each camp have similar orientations and degree of scatter. The only exception could be the foliation (S_2) at Camp I, where the tightly grouped average foliation (089/53) differs from the average for the area (136/48). This is represented in F_2 fold hinges, F_3 axial plane and fold hinges, and poles to S_2 . The tight grouping of data points for Camp I may be due to a smaller area of coverage in the field rather than more consistent data or structures; Camp II had a lot of scatter but also covers a large area. Most of the scatter is attributed to phase 4 deformation warping the area, but the large, open phase 4 folds are hardly recognizable when mapping so no corrections can be made. Rotation of fabrics and isoclinal folding during phase 1 and 2 deformation, producing a complicated geometry, may have been a minor contributing factor. Poles to foliations at Camp I and IV (Figure 3.8a and d) trace northeast-southwest great circles on stereonets, which may indicate the presence of a larger unrecorded structure. The plot of poles to the foliation (S_2) are similar to F_3 axial planes and indicate a northwest to southeast trending hinge line, similar to the hinge lines of F_3 folds. Similarities in F_2 and F_3 style of deformation suggest a progressive deformation event. Shear and drag folds with a northeastern sense of shear were observed on a small scale, and may also be present on a larger scale. Sheath folds were also identified in the field.

The interpretation by Fyles (1970) of the Bews Creek Fault as a major stratigraphic break is not supported in this study. Rock units are commonly highly folded and attenuated, and contain evidence for high strain, but a continuous sequence of basement paragneiss, basal quartzite, pelite, and calc-silicate gneiss $\pm$ marble is invariably present. This indicates the units are generally right-way-up with no stratigraphic breaks.

Investigations in the proximity of the Bews Creek Fault with observed protomylonite and fault breccia, indicate locally there are discontinuous zones of concentrated shear (Plates 3.12) but there is no evidence for a major fault. There does not appear to be a break in the lithologic sequence and continuous units could be traced throughout the area. These zones of concentrated shear are present in different lithological units but were observed close to the basement-cover contact.

Figure 3.2: Simplified geologic map of the Mount Copeland area incorporating detailed mapping by Fyles (1970), Höy and Brown (1980), Bosdachin (1989) and this study. See Figure 3.1 for a more detailed map. Included are large-scale structural features and cross-section locations. Also shown are camp locations referred to in the chapter.

Figure 3.3: Schematic diagram illustrating the sequences in the development of a transposition foliation, modified after Turner and Wiess (1967). Attenuation of alternate fold limbs during penetrative asymmetric folding of bedding (S_1) into the new foliation (S_2). Note that a stratigraphic section defined along traverse line A-A' in d) would be at a high angle to the true stratigraphic succession.

a) Primary Bedding

Compositional layering preserves primary bedding or layering, S_0 . Primary sedimentary features such as ripple marks (a), scours (b), and cross-bedding (c), indicate sedimentary origin and stratigraphic facing direction.

b) Generation I

S

Folding preserved as an inverted sequence of units above 'right-way-up' units. Tight to isoclinal folding results in compositional layering appear parallel to the foliation along the limbs of folds resulting in $S_{0+1} = S_1$ but S_0 and S_1 occur at a high angle in the hinge area. Deformation and high strain can result in development of metamorphic layering (gneissosity or migmatites) parallel to S_1 and segregation of minerals. The extent of development of an axial planar foliation depends on the grade of metamorphism and intensity of deformation.

Figure 3.4: Diagrammatic representation illustrating the progression of the phases of deformation recorded by folding in the Mount Copeland area and the stages in the development of the regional transposition foliation, S_2 . Geographic reference frame listed at top of diagrams to show the orientation of view.

d) Generation III

Although not shown, a repetition of the lithologic sequence, and Type II interference patterns preserve the second generation of folding.

e) Generation IV

Phase 4 deformation produced large-scale warping of the foliation (S_2) and axial surfaces of F_3 folds.

Figure 3.4: Diagrammatic representation illustrating the progression of the phases of deformation in the Mount Copeland area (continued).

(a) Camp I

N = 0

(b) Camp II

N = 7

(c) Camp III

N = 7

(d) Camp IV

N = 1

(e) Camp V

N = 0

(f) Camps I to V combined

N = 15

Figure 3.5: Equal area, lower hemisphere projections of hinge lines (L_1) for phase one folds (F_1).

Figure 3.6: Equal area, lower hemisphere projections of fold hinge lines (L_2) for phase two folds (F_2).

Figure 3.7: Equal area, lower hemisphere projections of L2 mineral (elongate quartz) stretching lineations.

Figure 3.8: Equal area, lower hemisphere projections of poles to transposition foliation (S_2) for camps I to V. Solid square = basement data, open cross = cover data, with numbers listed as basement/cover.

(a) Camp I

(b) Camp II

(c) Camp III

(d) Camp IV

(e) Camp V

(f) Camps I to V combined

Figure 3.9: Equal area, lower hemisphere projection of fold hinge lines (L_3) of third generation folds (F_3) for camps I to V. Solid square = basement data, open cross = cover data, numbers listed as basement/cover.

(a) Camp I

N = 10/7

(b) Camp II

N = 110/100

(c) Camp III

N = 26/9

(d) Camp IV

N = 5/22

(e) Camp V

N = 3/8

(f) Camps I to V combined

N = 154/146

Figure 3.10: Equal area, lower hemisphere projections of poles to axial planes of third generation folds (F_3) for camps I to V. Solid square = basement data, open cross = cover data, numbers listed as basement/cover.

(a) Quartzite Pebbles

N = 19

(b) Crenulation lineations

N = 15

(c) Station 265
Foliation (S₂)

N = 11

(d) Station 265
Qtz stretching lineation (Ls₂)

N = 11

Figure 3.11: Equal area, lower hemisphere projections of: a) quartzite pebble (L₂) lineations; b) L₃ crenulation lineations; and c) foliations (S₂) and d) quartz stretching lineations (L₂) for fold style characterization at station 265.

Plate 3.1: Representative photographs of basement orthogneiss. (a) Foliated and folded augen orthogneiss with a northeast vergence. (b) Recessive weathering and protruding quartz-feldspar augen within foliation of orthogneiss. (c) Flattened phenocrysts in augen orthogneiss. (d) Thin section photo of poorly foliated orthogneiss (Sample LB118) showing quartz with subgrains and undulose extinction and plagioclase. Width of view 5.0 mm. XPL. Foliation (S₂) seen in outcrop photos is a second phase fabric.

Plate 3.2: Overprinting relations of fold generations seen in outcrops. (a) Type II interference pattern in outcrop of basal quartzite (unit 3). View is approximately to the southwest. (b) Type III interference pattern between F2 and F3 in basement paragneiss near Camp Lake with preserved F1 fold on bottom limb of F2 fold.

Plate 3.2: Overprinting relations of fold generations seen in outcrops (continued). (c) Type II and III Interference patterns in outcrop of basement paragneiss (station 03). Patterns formed by phase three folds (F3) refolding first (F1) and second (F2) generation folds. (d) Three generations of folding in basement paragneiss.

Plate 3.2: Overprinting relations of fold generations seen in outcrops (continued). (e) Type III interference pattern. Tight isoclinal folds of phase two deformation are refolded by more open F3 folds. (f) Diopside-rich layer in core of an F1/F2 isoclinal fold that has been refolded by F3 deformation. Axial planes of both F1/F2 and F3 are perpendicular to the plane of the page.

Plate 3.3: Representative field photographs of pre- to syn-metamorphic intrafolial phase one and two folds. (a) Rootless isoclinal fold in basal quartzite. (b) Close-up view of hinge area of isoclinal fold with the foliation (S2) passing through the layering in the hinge zone of the fold. (c) Isoclinal folding in basement paragneiss outlined by quartz-rich layer. (d) Rootless isoclinal phase two fold outlined by quartz rich layer in migmatite.

Plate 3.3: Representative field photographs of pre- to syn-metamorphic intrafolial phase one and two folds (continued). (e) Isoclinal fold in basement paragneiss. (f) Intrafolial F1/F2 fold in hand sample (possibly a preserved Paleoproterozoic structure). (g) Rootless isoclinal fold of quartz-rich layer surrounded by psammite; the northerly trending fold hinge could indicate this is a phase one fold. Hand samples shown actual size.

Plate 3.4: Representative photographs of foliation in basement and cover rocks. (a) Leucosome lenses in basement gneiss, station 262. (b) Pull-apart structures and gneissic layering (S2) in basement gneiss. Quartzofeldspathic lenses at bottom of photograph. (c) Strong foliation in rusty weathering basal quartzite. (d) Foliation in white quartzite dipping to southwest.

Plate 3.4: Representative photographs of foliation in basement and cover rocks (continued). (e) Well foliated quartzite with the typical flaggy appearance seen in outcrop and hillsides. (f) Close-up view of flaggy psammite in outcrop. (g) Migmatitic basement paragneiss with well defined foliation. (h) Folded gneissic foliation with shear bands.

Plate 3.5: Second generation sheath folds. (a) Oblique down-plunge view of sheath folds in migmatitic basement paragneiss. (b) Sheath folds in calc-silicate exposed on pavement surface.

Plate 3.5: Second generation sheath folds (continued). (c) Down-plunge view of a sheath fold in quartzite refolded by third generation fold. View is to the southwest, fold closes to the northeast. (d) Down plunge view of sheath fold preserved in quartzite with concentric zoning in centre of fold.

Plate 3.6: Representative field photographs of pull-apart structures. (a) Rotated boudins in basement paragneiss. (b) Amphibolite boudins concordant with the foliation. (c) pinch and swell of psammitic layer. (d) Pulled apart amphibolite layer surrounded by quartzite.

Plate 3.7: Representative photographs of phase three folds; field photos. (a) Upright phase three fold in basement paragneiss with similar appearance to the fold in Mount Copeland (Plate 3.8). (b) Looking west at northeasterly verging F3 fold in basement paragneiss; note stair stepping appearance. (c) Open F3 fold viewed to the west. (d) Open F3 fold with northerly trending hinge line in northern part of the field area. Steep foliation (S2) near snow patch on left becomes gently dipping in centre of photo.

Plate 3.7: Representative photographs of phase three folds; thin-sections. (e) Biotite and sillimanite are folded by phase three deformation in pelitic schist. Width of view 5.0 mm. PPL. (f) View of e) under XPL. (g) Folded biotite foliation in calc-silicate. Width of view 5.0 mm. XPL. (h) Folded biotite and hornblende-rich layer alternating with quartz-rich layers in amphibolite. Width of view 5.0 mm. PPL.

Plate 3.8: Representative hand samples illustrating the characteristics of phase three deformation in the Mount Copeland area. Hand samples shown at actual size. (a) Basement paragneiss (Sample LB11b) from the northern part of the field area. (b) Basement paragneiss (Sample LB03a) with folded gneissic layering. Dip isogon patterns also shown.

Plate 3.8: Representative hand samples illustrating the characteristics of phase three deformation in the Mount Copeland area (continued). Hand samples shown actual size. (c) Folded amphibolite from northern part of the field area (Sample LB15a). (d) Small phase three folds preserved in calc-silicate gneiss (Sample LB254), from east end of Copeland Ridge.

Plate 3.8: Representative hand samples illustrating the characteristics of phase three deformation in the Mount Copeland area (continued). Hand sample shown actual size. (e) Large, open fold in basement paragneiss (Sample LB03a) from the northern part of the field area, with prominent gneissic layering and crenulations in the centre of the fold. Dip isogon patterns for fold class determination are also shown.

Plate 3.9: View to the east of Mount Copeland with marble layer of unit 5c outlined. Note the prominence of the folded white marble layers even when seen from a distance. This view, photographed from Station 04, is at approximately the location of cross-section line C-C' where it intersects Copeland Ridge. In the valley below the F3 fold, the structure appears simple but this may also be due to a lack of noticeable marker horizons to trace structures.

Plate 3.10: Field photograph of phase four deformation at the eastern end of Derrickson Ridge. The foliation is folded by an open, upright F3 synform, that is then folded by an F4 warp. View is approximately to the north. Similar to Plate 3.7d.

Plate 3.11: (a) Trace of late stage brittle fault (D5), viewed to the south/southeast. The vertical dipping, north-south trending fault is located in the valley bottom and can be followed onto the far hillside. The distant peaks are part of Thor-Odin dome. (b) View to south at north end of the same valley as in a). See * in Plate 3.11 (d) for photo location.

(c)

(d)

(c)

Plate 3.11: (c) Located in the bottom of a side valley, a late stage brittle fault, a late stage brittle fault, D5 runs the length of the valley. View to NNW. (d) A continuation of the view from (c), the trace of the fault is seen running through the valley and across the saddle to the Copeland valley. * indicates location of Plate 3.11 (b). Water drainage patterns in the area generally run east-west, however this valley is north-south.

Plate 3.12: The 'Bews Creek Fault' (a) Outcrop of mylonitic quartzite near location of Bews Creek Fault, station 09. (b) Close-up of mylonite with cross-cutting late-stage quartz vein. (c) Fault breccia in calc-silicate near location of Bews Creek Fault, station 263. (d) Close-up of fault breccia in calc-silicate, station 263.

CHAPTER 4.

PETROGRAPHY AND MICROSTRUCTURES OF THE MOUNT COPELAND AREA

Introduction

The Mount Copeland area of the Monashee complex was subjected to upper-amphibolite facies metamorphism as a result of tectonic accretion and crustal thickening in Cretaceous to Tertiary time. Metamorphic grade was interpreted from the mineral assemblage of pelitic, semi-pelitic and calc-silicate rock units; other rock types yielded inconclusive results. There was no noticeable variation in pressure-temperature conditions across the area based on the mineralogy of pelitic schists; detailed pressure-temperature work was not performed in this study. However, as the occurrence of pelites with diagnostic assemblages was sporadic, this may be due to the mapping and sampling methods rather than the lack of significant changes in pressure-temperature conditions. Pelites are typically composed of quartz, biotite, sillimanite, potassium feldspar and plagioclase feldspar, with minor amounts of garnet, muscovite, and chlorite. Calc-silicate gneisses contain calcite, diopside, tremolite and biotite. Late-stage granitic and mafic intrusions were not strongly affected by deformation but some minerals contain evidence for retrogression.

Metamorphic mineral assemblages

In the study area most rock types were subjected to high-grade regional metamorphism. The dominant mineral assemblage of pelitic schists is quartz, potassium feldspar, biotite, plagioclase feldspar and sillimanite indicating peak metamorphic conditions of upper-amphibolite facies (Plates 4.1 and 4.2). Garnet, chlorite, and

muscovite are present as accessory and retrograde minerals depending on local variations in lithology and/or chemical composition.

Calc-silicates gneisses (unit 5) comprise a large proportion of the rocks in the field area (Map 3). Mineral assemblages of calc-silicate gneisses identified in thin sections are composed of calcite ($\pm$ dolomite), diopside, quartz, tremolite and biotite $\pm$ potassium feldspar. A fluid phase with unknown composition complicates the interpretation of calc-silicates but mineral assemblages generally yield, high-grade metamorphic conditions for most samples (Yardley, 1989).

Microfabrics and Mineral Descriptions

Quartz

Quartz grains in most rocks types are typically fine grained, flattened, subhedral to anhedral crystals. The long dimension of the crystals are aligned parallel to the foliation plane and the grains display evidence for dynamic recrystallization (Plate 4.3). In psammitic and semi-pelitic rocks quartz grains are more variable in size and can occur in millimetre to centimetre thick, quartz-rich bands alternating with micaceous layers. Quartz displays subgrains especially in the larger crystals.

Quartzite is typically composed of quartz grains with a uniform size, which display evidence for dynamic recrystallization and have a preferred lattice orientation. In some samples quartz crystals are slightly flattened with the long axis aligned with mica grains (Plate 4.3c). Samples with bands containing larger quartz grains that are flattened and elongated in the plane of the foliation and display weak undulose extinction alternate with thin bands composed of fine grained quartz. These samples of quartzite may be classified as a protomylonite (Plate 4.3b and d), which would indicate these rocks have

experienced high strain. Quartz grains in fine grained bands have a preferred lattice orientation and display evidence of some dynamic recrystallization. Partial annealing of quartz grains is apparent in more pure samples (Plate 4.3).

Feldspar

Potassium and plagioclase feldspar crystals are fine grained and were observed in association with quartz in most rock types. Feldspar crystals are typically small and anhedral.

Euhedral potassium feldspar porphyroblasts in the leucocratic pegmatite and some crystals in the basement orthogneiss preserve varying degrees of sericite alteration (Plate 4.4a and b). The appearance of a microbreccia texture in some feldspar grains in the pegmatite may be the result of fluid interaction near the end of deformation. Potassium feldspar commonly shows a perthitic texture (Plate 4.5a-c) whereas plagioclase feldspar is myrmekitic (Plate 4.5d). These exsolution textures in feldspar were identified in the pegmatite and some samples of augen orthogneiss but were not observed in any other rock type.

Phyllosilicates

Biotite is ubiquitous in all rock types. Crystals are fine to medium grained, euhedral to subhedral laths aligned along and defining the foliation (Plate 4.2). In thin section biotite laths are folded around F_3 folds (Plate 3.7) and are rarely seen as folded crystals.

The habit of muscovite is variable. Muscovite was observed replacing biotite (Plate 4.2b) in some samples but this was a rare occurrence. The textural relation of biotite and muscovite indicates retrogression of biotite, as opposed to relicts of muscovite

from a lower grade (Plate 4.2). In some samples, rare chlorite may also replace biotite (Plate 4.4c).

Muscovite is abundant in one locality in the southern part of the field area and comprises 50% of some samples, although with no aluminosilicate. The muscovite in these samples define the foliation. The location may represent a local change in the bulk chemical composition. Other pelite samples collected nearby have the characteristic mineral assemblage of upper-amphibolite facies (Kfs + sill + melt) and there was no noticeable transition in mineralogy between the rock types. The high occurrence of muscovite was only observed at one locality. There was no further data available in the area.

Aluminosilicates

Sillimanite typically forms fibrolite mats in pelitic schists (Plate 4.6 and 4.7). In outcrop and hand sample, sillimanite occurs with quartz and feldspar in clusters and knots 2-3 cm long, termed *faserkiesel* (Spry, 1969; Brown, pers. comm. 2000), aligned parallel to hinge lines of second phase folds. In thin section fibrolite occurs mainly with biotite, but sillimanite also occurs independent of biotite as sporadic fine grained, acicular crystals (Plate 4.7 c and d). Fibrolite mats in thin section are generally elongated parallel to the foliation and appear somewhat contorted although Spry (1969) noted this as being due to rapid crystal growth, not deformation.

Later stage microfractures or shear fractures typically cut across or pull apart the fibrolite mats (Plate 4.6a-b and Plate 4.8). These fractures are orthogonal to the foliation, do not penetrate into the surrounding micaceous layers, and contain fine quartz, muscovite, and feldspar within the plane of fracture (Plate 4.8c). The lack of a noticeable

break or shear in the micaceous layers likely means the rocks were still in a ductile regime but below the stability field of sillimanite. These microfractures are sutured as the infilling crystals are interlocked and the rocks break across the fractures. There was no evidence for kyanite or andalusite crystals.

Garnet

Garnet is not a common mineral in the field area and was only observed in 3 samples of basement paragneiss. Where present, garnet porphyroblasts are 2-4 mm in diameter although one outcrop has garnet up to 1 cm in diameter. In thin section, garnet crystals are rounded with a poorly defined internal foliation. In Plate 4.9, a garnet crystal with an internal foliation (Si) is at a high angle to the external foliation (Se). Most rocks show no evidence of this earlier foliation. Inclusions in garnet are mainly quartz although some small biotite crystals are also observed.

Garnet crystals have reaction rims composed of fine to medium quartz, plagioclase feldspar, and biotite with minor potassium feldspar (Plate 4.9b). In thin section, it appears as though 50-100% of the garnet has been replaced but the biotite foliation still wraps around the garnet pseudomorph.

Interpretation

Metamorphism in Frenchman Cap dome consists of two recorded events; an early poorly preserved event, and an upper-amphibolite facies metamorphic event that occurred during Cretaceous to Tertiary collision and deformation. Later stage retrogression has also been identified in biotite and garnet. As garnets were only observed in basement gneisses it is uncertain whether this earlier fabric (Si) is a product of the Cordilleran orogeny or is a preserved Paleoproterozoic fabric. As there is no strong evidence for

preserved Paleoproterozoic fabrics in the Mount Copeland area these fabrics are considered to be a product of early Cordilleran orogenesis. Paleoproterozoic metamorphism will not be discussed.

Metamorphism associated with Cordilleran orogenesis is largely responsible for the current mineralogy and the gneissic and migmatitic textures of rock units.

Metamorphism and reactions identified in the Mount Copeland area and relation to phases of deformation (see Chapter 3) include:

D ₀		- Original bedding and compositional layering (S ₀)
D ₁	M ₁	- Preservation of quartz and biotite in planar inclusion trails in garnet (S ₁). Synchronous with deformation.
D ₂	M ₂	- Development of composite transposition foliation (S ₂). Rotation or recrystallization of biotite parallel to the axial planes (S ₂) of F ₂ folds. Upper-amphibolite facies metamorphism synchronous with deformation. Formation of gneisses and migmatites
D ₃		- Persistence of high-grade conditions but no associated metamorphism
D ₄		- No associated metamorphism
D ₅		- Retrogression. Minor replacement of biotite and garnet. Sericitic alteration of feldspars in granitic rocks.

An internal planar foliation defined by quartz is present in some garnet porphyroblasts. Where preserved, the internal foliation is at a high angle to the external biotite foliation (S_i ≠ S_e) indicating that garnet growth is pre-tectonic to the second phase of deformation. Since the foliation in rocks is interpreted as a second generation fabric (S₂), this internal foliation represents a preserved earlier generation fabric. No other evidence for this earlier metamorphic episode was preserved but regional metamorphism tends to destroy original textures resulting in a new preferred orientation of minerals

(Spry, 1969). The occurrence of garnet could also represent a relict from prograde metamorphism. Based on interpreted phase one folds (Chapter 3) quartz, biotite and feldspar were folded during phase one deformation.

The main foliation in the Mount Copeland area was produced during upper-amphibolite facies regional metamorphism as recorded in the mineral assemblage of pelitic schists. This second metamorphic event corresponds to rotation or recrystallization of biotite and sillimanite into the axial plane of second generation folds. Pelites are typically composed of quartz, potassium feldspar, biotite, plagioclase feldspar and sillimanite. McMillan (1969) reported a northerly trending muscovite-out isograd ($\text{musc} + \text{qtz} \rightleftharpoons \text{Kfs} + \text{sill} + \text{H}_2\text{O}$) to the west of the study area near the Perry River (Map 2). The rocks to the east of the Perry River, adjacent to this study, are above the sillimanite-potassium feldspar isograd similar to the assemblages seen in this study. Variations in the mineral assemblage of pelitic schists is likely the result of a change in the bulk composition of the rocks rather than variations in the pressure and temperature conditions. The mineral assemblages associated with the peak metamorphic conditions appear frozen into the rocks as reflected in the lack of significant retrogression of minerals which may be the result of quick exhumation and cooling of the complex.

The metamorphic grade of upper-amphibolite facies was interpreted from the apparent equilibrium mineral assemblage ($\text{Kfs} + \text{sill} + \text{melt}$) of pelitic schists. This corresponds to temperatures of 700-800 °C and pressures of 6-8 kilobars (Yardley, 1989; Spear, 1993), which suggest a burial depth of 22-29 km at the time of peak metamorphism. This depth is in agreement with metamorphic studies in northern Frenchman Cap dome (Journeay, 1986; see Figure 1.2) which have shown burial depths of

21-24 km (6.4-7.25 kb). Even after D₂ the high-grade conditions of M₂ must have persisted as there is no retrogression associated with D₃ although the rocks are strongly folded.

Muscovite in one locality comprises up to 50% of the rock but there was no observable transition in mineralogy from the assemblage of pelitic schists (Kfs + qtz + sill). While muscovite typically indicates low and medium grade metamorphism, the samples contained little to no aluminosilicate. The simple mineralogy of these rocks may not have been conducive for the complete break-down of muscovite at higher grades.

Late stage alteration is represented by minor retrogression of biotite and garnet. Muscovite and chlorite locally replace biotite although this is rare and only observed in granitic intrusions (pegmatite and some basement orthogneiss). Garnet crystals are partially to totally replaced by reaction rims of randomly oriented biotite, quartz, plagioclase feldspar, ± muscovite that preserve the original form of the garnet crystals. The biotite foliation wraps around the garnet. The lack of displacement of the surrounding minerals (bio + qtz + plag), could indicate the area was not affected by deformation during or after the retrogressive phase.

The majority of the evidence for retrogression is preserved in the pegmatite, which is a later stage event. The timing or spatial relationship of the retrogression to Cordilleran deformation episodes is not entirely known. While likely associated with extension, the retrogression may also be the result of fluid interaction any time after the second metamorphic event.

The microfractures formed in sillimanite after phase two deformation (peak-metamorphic conditions) but the exact time is not known. There does not appear to be

any offset across the fracture plane so the time of formation is likely during or after phase four deformation and before the onset of phase five brittle deformation. The sillimanite mats are pulled apart and the fracture plane is filled in with quartz, biotite, muscovite, and plagioclase feldspar. The assemblage of minerals associated with the fractures could indicate the area was at a lower grade than peak metamorphic conditions but still in a ductile regime.

Although in most cases biotite laths define the S₂ foliation, Plate 4.10a shows the weak development of a cross-cutting foliation (S₃) in the hinge zone of an F₃ fold. In this case, two sets of biotite laths are seen at a high angle to each other and may represent the beginning of mineral rotation into the axial plane of F₃. This sample was collected from the northern part of the field area at Camp I (Map 1) which may be the lowest structural level seen in the study. Similarly, Plate 4.10b shows high strain as seen in sheared micaceous bands aligned with the foliation.

The results of this study are in agreement with other studies in Frenchman Cap dome (McMillan, 1969; Scammell, 1986; Journeay, 1986). The common occurrence of gneisses and migmatites indicate the area experienced high grade regional metamorphism, which is also interpreted from the mineral assemblage (Kfs + sill + melt) of pelitic schists.

Plate 4.1: Representative samples of pelitic schists. (a) Polished hand sample LB13b. (b) Polished hand sample of sillimanite, biotite schist with clusters of sillimanite and feldspar, sample 69b. (c) Sillimanite mat in sillimanite biotite schist, from sample LB276. XPL (d) Well foliated sillimanite biotite schist, sample LB229c. PPL. Hand samples shown actual size. Thin section width of view 5.0 mm.

Plate 4.2: Well foliated pelitic schist, sample 26. (a) Plane polarized view of biotite, muscovite and bladed sillimanite lying within the foliation. PPL (b) The micas display prominent birds eye extinction in this cross-polarized view of (a). Near the top of the photo muscovite is replacing biotite. Quartz displays strong undulose extinction resulting from dynamic recrystallization. Width of view 1.2 mm.

Plate 4.3: Foliation in Basal quartzite. (a) Polished hand sample of typical quartzite. (b) Sample LB09 displaying horizontal alteration of course and mylonitic bands. (c) Thin section view of foliated quartzite. Majority of the rock is quartz although some small biotite grains are also present. From sample 96. (d) Thin section showing the development of mylonite layers in quartzite, foliation is oriented vertically. Quartz grains display undulose extinction and annealed textures. From sample LB09 seen in (a). Hand samples shown actual size. Width of view for thin sections is 5.0 mm. XPL

Plate 4.5: Microstructures and textures. (a) Quartz with undulose extinction and perthitic K-feldspar. Sample 236A, width of view 5.0 mm. (b) and (c) Perthitic texture in K-feldspar. Sample 108b and width of view 5.0 mm for both photos. (d) Myrmekitic texture of quartz and plagioclase. Sample 118, width of view 1.2 mm. All photos XPL

Plate 4.6: Sillimanite, biotite pelitic schist (a) Hand sample of sillimanite, biotite pelitic schist with foliation and fractures outlined. White clusters composed of fine- to medium-grained sillimanite mats (fibrolite) and K-feldspar grains overprinting quartz and biotite (b) Thin-section of sillimanite mat with vertical, quartz-filled, cross-cutting fractures. Biotite is being replaced by muscovite in the fractures. PPL (c) View of (b) under XPL. Thin section width of view 5.0 mm. Hand samples shown actual size.

Plate 4.7: Representative examples of sillimanite in thin section. (f) Fine sillimanite mats (fibrolite) aligned along foliation with small cross-cutting microfaults. (g) View of (f) under XPL. (h) Bladed sillimanite seen at an angle to the biotite foliation. (i) View of (h) under XPL. Width of view 1.2 mm.

Plate 4.8: Microfractures in well foliated pelitic schist, sample 266.
 (a) Plane polarized thin section of sillimanite and muscovite aligned with biotite foliation. Elongate garnet grain is seen to the right of centre. Width of view 5.0 mm. (b) View of (a) under XPL. (c) Close-up view to left of centre in (b). View shows pulled-apart sillimanite and cross-cutting vertical microfracture with infilled muscovite and quartz. Width of view 1.2 mm.

Plate 4.9: Representative example of garnet. (a) Polished hand sample of garnet, biotite gneiss from basement shown actual size. (b) Close-up view of garnet from (a) in thin section. Quartz inclusions in the garnet preserve an internal foliation at a high angle to horizontal external foliation with a surrounding reaction rim of quartz and plagioclase. XPL (c) Close-up view of garnet in thin section with sillimanite, biotite and muscovite. PPL (d) View of (c) under XPL. Thin section width of view 5.0 mm.

Plate 4.10: Indications of high strain seen in thin section. (a) Beginning of S3 axial planar cleavage. Thin section view of hinge zone of an F3 fold shows the biotite foliation at a high angle and cross-cutting the transposition foliation (S2). Matrix is largely quartz with small amounts of potassium and plagioclase feldspar. Width of view 1.2 mm. XPL (b) Development of shear bands along the micaceous foliation in basement paragneiss. Majority of the rock is quartz. Width of view 5.0 mm. XPL

CHAPTER 5.

STRUCTURAL GEOMETRY OF THE MOUNT COPELAND AREA

Introduction

The results from the preceding chapters have revealed a complicated structure for the Mount Copeland area. This chapter integrates the findings of the previous chapters to propose a model of the three-dimensional geometry for the area. The geometry and orientation of folds related to each phase of deformation without overprinting effects are shown in Figure 5.1. The phases of deformation are combined in the block diagram in Figure 5.2 to show the geometry of the area. Also discussed is the tentative assignment of the phases of deformation to tectonic events during formation of the Canadian Cordillera.

Geometry of the Mount Copeland area

While inferred on geologic maps and cross-sections, phase one deformation is only considered to locally contribute to the structural geometry of the area. Several phase one folds have been recognized and are shown in Figure 5.1a but there are probably more that were missed at the present scale of mapping. The geometry of the area may be more complicated than can be determined with the available data as shown by comparing Figure 3.1c and Plate 5.1. Phase one folds have limb lengths of 1-2 kilometres, and on a local scale produce inverted and repeated stratigraphy. Several phase one folds are recognized and shown in Figure 5.2 on the southwest surface of the block diagram and are represented on the top surface by Type II interference patterns. These folds, inferred as easterly verging, can be traced on cross sections through the subsurface by following inverted units.

The structural geometry of the Mount Copeland area is largely defined by the

second generation Eagle Pass synform and antiform (Figure 3.1 and 5.1b). In cross section these folds are large-scale, isoclinal, northeasterly verging, recumbent z-folds with limb lengths up to tens of kilometres. The overturned limb of the antiform appears to be attenuated near the western end of the field area, which may indicate the early stages of nappe formation (Figure 3.1a). The Eagle Pass antiform is cored by basement paragneiss and shows the basement rocks were highly involved in Cordilleran deformation. The majority of the basement gneiss is paragneiss but also includes structurally inverted orthogneiss near Neil Lake (Figure 3.1a). The Eagle Pass synform is cored by rocks of the cover sequence with the Mount Copeland syenite gneiss present in the centre of the fold and enclosed in an embayment surrounded by basement gneisses (Map 3). Most of the Eagle Pass antiform has been eroded so the length of fold limbs cannot be determined. The antiform must close before the Copeland synform due to the lack of basement and basal quartzite in this area, which provides maximum length for the long limbs of folds. Second generation minor structures are common but the Eagle Pass synform and antiform were the only map-scale folds identified as part of this phase.

The distribution of lithologic units on geologic maps is largely controlled by phase three deformation. Northeasterly verging F₃ synforms and antiforms with amplitudes of ~3 km and limb lengths of kilometres are seen in cross sections across the area (Figure 3.1 and 5.1c). Disharmonic folds in the Hiren Valley die out along strike of the axial surface. Folds also become smaller but more abundant in the eastern part of the field area. Minor structures are overturned, northeasterly verging, open to close folds, which are parasitic to the km-scale antiforms and synforms. Map-scale folds are shown in Figure 3.1 and Map 4.

There are no minor structures nor large-scale features associated with the east-west trending fourth phase of deformation (Figure 5.1d). Doming of the complex occurred after the majority of the deformation had ceased and produced what Journeay (1986) termed the Monashee antiform. The uplift and arching of Frenchman Cap and Thor-Odin domes produced a regional-scale Type I interference pattern (Basin and Dome pattern; Ramsey, 1967) along the north-south trend of the complex. There is a variation in fold style across the dome resulting from the rotation of asymmetric F_3 minor folds. Long limbs of folds in the west dip to the west, but have been rotated as a result of the arching such that in the east the long limbs are horizontal or dip shallowly to the east (Plate 5.2). The late timing of phase four deformation in the history of the region results in poorly developed structures.

An important component of this study was to determine the distribution of basement gneisses across the field area (Map 3) that is highlighted by the basal quartzite (unit 3) in cross sections (Figure 3.1) which bounds the upper contact of basement gneisses. There are two important conclusions interpreted from the maps and cross-sections. First, maps (Map 3) and cross sections (Figure 3.1) show the basement gneisses and cover sequence were affected by all phases of deformation. Beginning with phase one deformation, infolding of the basement gneiss and cover sequence were observed in outcrops and interpreted on maps. This indicates the stress associated with each phase of deformation were transmitted through the sequence of rocks and show the basement rocks were involved in all phases of deformation. Second, the cross sections and maps (Map 3) indicate the rocks seen around Kirkup Creek are basement gneisses as has been previously reported by Höy and McMillan (1979) and Höy and Brown (1980). This

supports the correlation through the subsurface of basement gneisses at the south end of the field area with known basement rocks to the north (Figure 3.1).

Evolution of the Mount Copeland area

Geologic evidence indicates the history of the Mount Copeland area dates back to the Paleoproterozoic (Armstrong et al., 1991; Crowley, 1997b). There are recorded rift and orogenic events but for most of the time, 1440 Ma (Gabrielse and Yorath, 1991), the area remained buried under thick accumulations of miogeoclinal strata. The cover sequence was deposited along a passive margin that existed into the Paleozoic. Two rift events occurred along the western edge of ancestral North America interpreted from the sedimentary sequences in the Canadian Cordillera. The intrusion of nepheline syenite gneiss in the Mount Copeland area (Currie, 1976; Parrish and Scammell, 1988; Scammell and Brown, 1990) provides further evidence supporting the latter rift event. The phases of deformation in the Mount Copeland area can be tentatively correlated with tectonic events in the Canadian Cordillera (eg. Johnson, 1994; Johnston, 1998; Crowley, 1999), however, without more geochronological studies in the Mount Copeland area, it is not possible to conclusively assign events.

The arrival of exotic terranes in the Jurassic is largely responsible for the current geological architecture of the Canadian Cordillera although the Monashee complex was not affected until the Cretaceous (Brown et al., 1993). The relative timing of D₁ is inferred because of the strong association of second generation structures (D₂) and metamorphism (M₂) with obduction and accretion of the Intermontane Superterrane (Terrane I of Monger et al., 1982). There is no data to correlate D₁ with events in the Cordillera. The Monashee complex was buried as the miogeoclinal rocks (Kootenay

Terrane) were telescoped and displaced eastward in front of the advancing accreted terrane (QT in Plate 5.2c) that resulted in D₂ structures. This produced the peak metamorphic conditions of upper-amphibolite facies as the Selkirk Allochthon (part of the Kootenay Terrane) was thrust over the Monashee complex. The high strain associated with D₂ is preserved in the rocks as: a transposed fabric, sheath and drag folds, and large-scale isoclinal folds with attenuated and pinched limbs, all indicating a northeasterly directed sense of shear. The third phase of deformation is interpreted to be a result of the accretion of the amalgamated terranes of the Insular belt (Terrane II of Monger et al., 1982). Asymmetry of preserved minor structures still indicate a northeast sense of shear (Plate 5.2a), similar to phase two, although the effects in the Mount Copeland area are not as pervasive.

Lithospheric attenuation with crustal-scale extension resulted in exhumation of the Monashee complex. This occurred along the Columbia River Fault zone and Okanagan Valley-Eagle River Fault, which indicate ductile to brittle shearing during the Eocene (Lane, 1984; Johnson, 1994). Geochronological studies in the Monashee complex indicate rapid uplift, unroofing and cooling in the Eocene (Parrish et al., 1988) that produced the regional arching of the dome (Plate 5.2c).

Regional correlations

The regional-scale Eagle Pass synform and antiform are the only structures in the Mount Copeland area large enough to allow possible correlation. Journeay (1986) proposed two possible correlations with the Grace Mountain Syncline. The Jordan Range synform-antiform pair are located to the north of the field area so will not be addressed, although Scammell (1986) also regarded this correlation as not being very plausible. The

Grace Mountain Syncline has not been traced to the south, but the correlation with the Eagle Pass synform is possible. The argument (Journeay, 1986) was partly based on the Ratchford Creek fault correlating with the Bews Creek fault, although the inferred attenuation on the overturned limb on the Eagle Pass synform-antiform may represent a similar feature. The similar structural style and stratigraphic position of the Grace Mountain Syncline and Eagle Pass synform in relation to the basement gneisses allow a plausible correlation, and could indicate the phase one folds in the Mount Copeland area are parasitic to a larger, unrecognized, structure.

Interpretation

The results of this study have shown the preservation to varying degrees of at least four phases of folding. The Eagle Pass synform and antiform are the largest structural features identified, which were overprinted by smaller antiforms and synforms as a result of phase three deformation. Phase two and three deformation both indicate a northeast vergence. Several small phase one folds were identified where the distribution of units could not be explained by phase two or three deformation. Phase four deformation is not well preserved in the area but is associated with the uplift and large-scale arching of Frenchman Cap dome. Faults were identified in the area but are interpreted as being insignificant. Other studies in the northern Monashee complex (Psutka, 1978; Scammell, 1986; Journeay, 1986; Crowley, 1997b, 1999; Gibson, 1997) have reported similar deformation as presented in this study.

Figure 5.1: Geometry of inferred phase one (a) and two (b) folds with overlay of simplified lithologic compilation map of the Mount Copeland area.

Figure 5.1: Geometry of inferred phase three (c) and four (d) folds with overlay of simplified lithologic compilation map of the Mount Copeland area. Legend as per (a).

Figure 5.2: Generalized block diagram of basement-cover contact surface (base of the basal quartzite, unit 3) in the Mount Copeland area. The diagram combines the overprinting relations from Figure 3.12a-d as viewed to the north; the top surface of the block diagram represents the field area as seen in Map 3 and Figure 3.1. Folds are approximately located based on vertical cross-sections and Map 4.

Plate 5.1: Panoramic view of Grizzly Peak and hillside above Camp V with structures highlighted. View for this composite photograph is north-northwest. For close-up views, see plates as indicated by box outlines or arrows. Beginning at approximately the location indicated by Plate 2.5a no more folds are apparent to the northeast along the ridge although inverted sequences were identified. M, Marble layer; P, pelitic layer; Sy, Syenite gneiss layer; A, antiform; S, synform. Fold generations same as Map 3.

Plate 5.2: Variation of F3 fold style across Frenchman Cap dome (FCD). (a) Foliations on the west side of the dome dip predominantly to the west and structures verge to the northeast. (b) On the east side of the dome, structures indicate northeast displacement but due to rotation during uplift and arching of FCD foliations dip to the east. (c) Simplified crustal-scale cross-section of the Monashee complex between latitudes 51° and 52°N (modified after Brown and Journeay, 1987; Johnson and Brown, 1996) with location (a) and (b) indicating the variation in fold style across the dome. View is to the north. CRF, Columbia River fault; ERF, Eagle River fault; MD, Monashee Decollement; SA, Selkirk Allochthon.

CHAPTER 6.

SUMMARY OF CONCLUSIONS

The Mount Copeland area has experienced complex deformation and regional metamorphism. Preserved fabric and structural elements have allowed a detailed investigation of the three dimensional geometry. The results of this study have shown the subsurface geometry is more complicated than previously proposed. The location of the area for much of its history at a low structural level preserved structures related to a long tectonic history. The findings in the Mount Copeland area as a result of this study include:

- (1) The lithologic sequence, although transposed, is described in terms of 6 mappable units (2 Basement and 4 Cover), plus 4 subunits identified within unit 5.
- (2) The contact of basement gneisses and the overlying metasedimentary cover sequence may represent a regional unconformity indicating a time gap between events in the basement gneisses and deposition of the cover sequence. The cover sequence does not appear to have any stratigraphic breaks.
- (3) The interpretation of the Bews Creek Fault as a major stratigraphic break (Fyles, 1970) was found to be incorrect. Displacement parallel to S_2 may have occurred but was insignificant in terms of structural geometry and produced only local discontinuous shear zones.
- (4) Five generations of deformation have been recognized based on the overprinting of structural fabrics and relation of fold and fabrics elements to the foliation:
 - i) Phase one deformation produced km-scale, north-south trending, easterly verging folds that are inferred on a local basis.

- ii) Phase two deformation resulted in the large-scale Eagle Pass synform and antiform, formed at peak metamorphic conditions. The trace of fold hinge lines strikes along a northwest-southeast axis, and folds verge to the northeast.
 - iii) Phase three deformation is responsible for the map-scale distribution of units and produced northwest-southeast trending, northerly verging, synforms and antiforms repeated across the field area. Axial surfaces dip predominantly to the southwest.
 - iv) Phase four deformation is rarely preserved as east-west trending folds with upright axial planes and interlimb angles of 160-180°.
- (5) Peak-metamorphism as recorded in the mineral assemblage of pelitic schists (Qtz + Kfs + Plag + Bio + Sill $\pm$ Gt) indicate upper-amphibolite facies conditions.
- (6) Evidence for post-peak metamorphic reactions but minor component of mineralogy. Biotite was replaced by $\pm$ Chl $\pm$ Musc; garnet replaced by Qtz + Plag + Bio $\pm$ Kfs $\pm$ Hbl.

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Map 1: Station and Camp Location Map of Southern Frenchman Cap Dome

This map is a compilation of detailed mapping by Wheeler (1965) Fyles (1970), Höy and Brown (1980), Journeay, 1986, Bosdachin, 1989, and this study.

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Contours at 100 m interval

Scale 1:40 000

Universal Transverse Mercator
Projection

Topography from maps 82M.008,
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Note: Due to space limitations at Camp III, not all stations are numbered.

Representative stations are numbered along traverse lines so station numbers can be determined.

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Map 2: Correlation Map of Lithology around Frenchman Cap Dome, Monashee complex.

(Modified after Wheeler, 1965; Fyles, 1970; Höy and Brown, 1980; Journeay, 1986; Scammell, 1986; Höy, 1987; and Bosdachin, 1989).
Scale Approximately 1:200,000
Contour interval = 325 metres (1000 feet)

Legend

- Amphibolite
- Allochthonous Rocks
- Undifferentiated rocks of the Selkirk Allochthon
- Kootenay Terrane
- Autochthonous Cover Rocks
- Marble (+/- sulphides)
- Undifferentiated biotite schists, calc-silicate gneisses, quartzites, and biotite sillimanite schists
- Basal Quartzite

- Autochthonous Basement Gneisses
- Granitic Augen Orthogneiss (Porphyratic Granite Gneiss of Fyles, 1970)
- Mixed Paragneiss and Orthogneiss
- Granitic Augen Orthogneiss

Symbols

- Contacts: defined, approx. assumed
- Fault: defined, approx. assumed
- Limit of mapping, overburden
- Ductile shear zone and fault
- Thrust sense, normal sense
- Trans-Canada Highway

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Map 3: Lithologic Compilation Map of Southern Frenchman Cap Dome

This map is a compilation of detailed mapping by Wheeler (1965) Fyles (1970), Höy and Brown (1980), Journeay (1986), Bosdachin (1989), and this study.

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Contours at 100 m interval

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Projection

Topography from maps 82M.008,
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Symbols

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Fault (Thrust/normal sense)
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feldspar paragneiss
SA Metasedimentary and metavolcanic rocks of the Selkirk Allochthon

Monashee Complex

N Nepheline syenite gneiss, syenite gneiss (740 ± 36 Ma igneous and 59 ± 1 Ma metamorphic; Parrish and Scammell, 1988)

Autochthonous Cover

(Paleoproterozoic to Paleozoic)

5 Sillimanite biotite schist

5 Calc-silicate gneiss and Marble; minor schist

5d Calc-Silicate gneiss: sillimanite schist and biotite gneiss

5c* White Crystalline Marble Strataform Pb-Zn sequence (Cambrian age; Höy and Godwin, 1988)

5b Biotite-hornblende gneiss; minor quartzite, marble, calc-silicate gneiss, mica schist

5a* White quartzite; minor schist

4 Micaceous Schist

3 Basal Quartzite; minor metaconglomerate, white quartzite, schist

Basement Gneiss (Paleoproterozoic)

2 Undifferentiated monzogranitic and granodioritic orthogneiss

1 Mixed Paragneiss and Orthogneiss, Biotite-Hornblende Gneiss, Amphibolite, Granitic Gneiss

* Thin units shown with exaggerated thickness

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Map 4: Structure Map of Southern Frenchman Cap Dome

This map is a compilation of detailed mapping by Wheeler (1965) Fyles (1970), Höy and Brown (1980), Journeay (1986), Bosdachin (1989), and this study.

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North American Datum - NAD83

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Fault (Thrust/normal sense)
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Limit of mapping (See Map 3)

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Limit of mapping (See Map 3)

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viewed to the NW

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Structure Symbols

Attitude of layering

Attitude of transposition
foliation (S2)

Plunge of second generation
hinge line (F2) with/out
attitude of axial surface

Attitude of third
generation axial plane (F3)
and plunge of hinge line
(L3) for S-, Z-, M-, and U-
shaped fold asymmetry (U-
shaped = undetermined)

Bearing and plunge of
mineral, stretch, or
intersection lineation

Axial surface trace,
synform, antiform,
generation indicated by
dots

Axial surface trace of
overtaken synform,
antiform, generation
indicated by dots

Generation of deformation
for axial surface trace
indicated by dots

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Structure Symbols

Attitude of layering

Attitude of transposition foliation (S2)

Plunge of second generation hinge line (F2) with/out attitude of axial surface

Attitude of third generation axial plane (F3) and plunge of hinge line (L3) for S-, Z-, M-, and U-shaped fold asymmetry (U-shaped = undetermined)

Bearing and plunge of mineral, stretch, or intersection lineation

Axial surface trace, synform, antiform, generation indicated by dots

Axial surface trace of overturned synform, antiform, generation indicated by dots

Generation of deformation for axial surface trace indicated by dots

Note: Due to space limitations at camp locations, only representative data points are plotted on the map.

Data with * from Fyles (1970), and data with ^ from Höy and Brown (1980).

Figure 3.1a: Vertical cross section of the Mount Copeland area. Line A-A', view to northwest

The basal quartzite is shown as a continuous unit to highlight folding. See Legend in Map 3 for lithologic units, and Legend in Map 4 for structure symbols. See Map 4 for location of cross section. No vertical exaggeration.

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Figure 3.1b: Vertical cross section of the Mount Copeland area. Line B-B', view to northwest

B

The basal quartzite is shown as a continuous unit to highlight folding. See Legend in Map 3 for lithologic units, and Legend in Map 4 for structure symbols. See Map 4 for location of cross section. No vertical exaggeration.

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Figure 3.1c: Vertical cross section of the Mount Copeland area. Line C-C', view to northwest

C

The basal quartzite is shown as a continuous unit to highlight folding. See Legend in Map 3 for lithologic units, and Legend in Map 4 for structure symbols. See Map 4 for location of cross section. No vertical exaggeration.

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C'

Figure 3.1d: Vertical cross section of the Mount Copeland area. Line D-D', view to northwest

The basal quartzite is shown as a continuous unit to highlight folding. See Legend in Map 3 for lithologic units, and Legend in Map 4 for structure symbols. See Map 4 for location of cross section. No vertical exaggeration.

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D'

Monashee
Antiform

Copeland Synform

Copeland
Creek

Figure 3.1e: Vertical cross section of the Mount Copeland area. Line E-E', view to northwest

The basal quartzite is shown as a continuous unit to highlight folding. See Legend in Map 3 for lithologic units, and Legend in Map 4 for structure symbols. See Map 4 for location of cross section. No vertical exaggeration.

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Monashee
Antiform

E'

Figure 3.1f: Vertical cross section of the Mount Copeland area. Line F-F', view to northwest

The basal quartzite is shown as a continuous unit to highlight folding. See Legend in Map 3 for lithologic units, and Legend in Map 4 for structure symbols. See Map 4 for location of cross section. No vertical exaggeration.

THS

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F'