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METAMORPHISM OF PELITIC ROCKS
FROM THE NORTHERN SELKIRK MOUNTAINS,
SOUTHEASTERN BRITISH COLUMBIA

by

Robert Wesley Leatherbarrow, B.Sc., M.Sc.

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

Carleton University
Ottawa, Ontario.

January, 1981

The undersigned hereby recommend to the Faculty
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ABSTRACT

Pelitic rocks from the northern Selkirk Mountains of south-eastern British Columbia have undergone Barrovian type regional metamorphism ranging from chlorite to potassic feldspar - sillimanite grade. Northwesternly trending mineral zones are arranged such that metamorphic grade increases from southwestern (chlorite zone) and northeastern (staurolite-kyanite-garnet parageneses) margins towards a centrally located metamorphic culmination (potassic feldspar - sillimanite zone). Changes in mineral parageneses and textural evidence permit positioning of isograds corresponding to the following mineral reactions:

Isograds corresponding to the first appearance of chlorite, biotite and garnet are also mapped. Sequences of mineral reactions and zones imply that assemblages southwest of the culmination formed at temperatures up to approximately 550°C and at pressures of about 5 kbar whereas assemblages northeast of the culmination

formed at temperatures from approximately 550°C to 670°C and pressures between about 5 and 7 kbar.

Temperatures are estimated using garnet-biotite and calcite-dolomite geothermometers as follows: assemblages southwest of the metamorphic culmination equilibrated at 500°C (staurolite-kyanite-biotite zone) to 550°C (sillimanite-garnet-biotite zone); and assemblages northeast of the culmination equilibrated at 550°C - 590°C (staurolite-muscovite-quartz-kyanite-garnet-biotite zone) to 650°C (potassic feldspar - sillimanite zone). Confidence in temperature estimates is enhanced by duplication of results using independent techniques; replication of results using the same technique; and correlation of estimates with values estimated from phase relations. Pressure estimates using garnet-plagioclase- Al_2SiO_5 -quartz geobarometry indicate that assemblages southwest of the culmination formed at approximately 5 kbar whereas those to the northeast equilibrated at approximately 7 kbar.

A discontinuity in conditions of metamorphism of approximately 110°C and 2 kbar is recorded across the Argonaut Mountain Fault which marks the boundary between the sillimanite-garnet-biotite zone (southwest of the culmination) and the potassic feldspar - sillimanite zone. The vertical component of displacement on this fault is considered to be 7 km.

This information, combined with similar data on the Purcell Thrust, implies that the northeastern part of the study area has moved upward as a fault bounded block relative to the southwestern part of the study area and the Park Range.

Models are presented to outline the tectono-metamorphic

evolution of these blocks. Metamorphism commenced during D_2 as the stratigraphic succession was being tectonically thickened, and peaked in post- D_2 pre- D_3 times when the succession was buried to a depth of at least 25 km. Postmetamorphic uplift probably started during D_3 and culminated in a post- D_3 block faulting event. This sequence is compatible with tectonic models involving subduction, currently proposed for the southeastern Canadian Cordillera. Post- D_3 block faulting may have occurred during a period of extension late in the orogenic evolution of the area.

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I would like to express my appreciation to R.L. Brown for initiating my interest in this project and for the patient and constructive supervision throughout the development of the study. My understanding of metamorphic petrology has profitted from discussions with P.H. Thompson, G.B. Skippen, and J.M. Moore, Jr. A continuing dialogue with M.J. Perkins has kept me aware of structural and stratigraphic developments in Selkirk geology.

Assistance from the following individuals is also acknowledged: J. Hogg, for smoothly guiding the administrative needs of the study; P. Mainwaring and P. Morton, for teaching me the techniques of the electron microprobe; J. Baker, who through X-ray analysis proved that some of the minerals that I was looking at were actually what I thought they were; R. Taylor and R. Conlon, for countless thin and polished sections; N. Stuart, for accompanying me in the field and showing me that there is true art to mountaineering; and D. Lefebvre, E. Chappell, and P. Morton, for providing me with their moral support and understanding.

My wife, Liesbeth, is thanked for two reasons: she has patiently endured the trials of living with a graduate student; and she has expertly edited, criticized, clarified, typed, retyped, and drafted much of this thesis. I dedicate this thesis to Liesbeth in recognition of all her selfless effort.

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in pocket

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LIST OF MINERAL ABBREVIATIONS

ap - apatite	ht - hematite	py - pyrite
bi - biotite	hb - hornblende	qz - quartz
ca - carbonate	il - ilmenite	ru - rutile
cc - calcite	kf - K-feldspar	se - sericite
oz - clinozoisite	ky - kyanite	si - sillimanite
ch - chlorite	mt - magnetite	sp - sphene
ep - epidote	mu - muscovite	st - staurolite
fs - feldspar	Naf - Na-feldspar	to - tourmaline
ga - garnet	op - opaques	zr - zircon
gr - graphite	pa - paragonite	zo - zoisite
	pc - plagioclase	

CHAPTER 1.

INTRODUCTION

(1-1) PURPOSE AND SCOPE

Since 1973, R.L. Brown and graduate students have mapped and studied segments of the Selkirk Mountains in southeastern British Columbia. These studies have emphasized regional structural geometry and stratigraphy, and tectonic models have been proposed (Franzen 1974; van der Leeden 1976; Tippet 1976; Lane 1977; Brown and Tippet 1978; Brown et al. 1977; Brown 1978).

The object of this thesis is to provide a detailed metamorphic analysis of pelitic rocks to complement these structural and stratigraphic syntheses. Until the initiation of this study, a thorough metamorphic study of this region had not been undertaken. Therefore, the initial parts of this study are directed towards the presentation of a regional metamorphic map which outlines mineral zones and isograds in pelitic rocks.

A second objective of this thesis is to estimate temperatures and pressures associated with the peak of metamorphism. This is accomplished by comparing phase relations observed in this region with those derived from a petrogenetic grid; and by using the geothermometers and geobarometer based on chemical compositions of pertinent phases.

A final objective of this work is to apply the metamorphic information obtained to structural and tectonic models currently proposed for this region.

The study area is sufficiently large that sampling of the entire region during the course of one thesis is not feasible. This writer was fortunate to have the opportunity to collate

data from suites of samples collected by others in the Carleton University group with personally collected data. These suites, localities, and prefixes in sample labelling include:

- (1) J. Franzen, east of French Creek in 1973 (labelled F);
- (2) J. van der Leeden, west of French Creek in 1974 (labelled V);
- (3) M. Perkins and D. Shaw, in the Windy Range in 1974 (labelled P);
- (4) M. Perkins and R. Leatherbarrow, near Mica Creek in 1975 (labelled M);
- (5) M. Perkins, near Mount Chapman in 1976 (labelled M);
- (6) R. Leatherbarrow, east of French and Windy Creeks and near McNaughton Lake in 1977 (labelled A).

Station locations are plotted on Map 1 (in pocket).

(1-2) LOCATION AND ACCESS

The Selkirk Mountains of the southeastern Canadian Cordillera are located approximately 300 km northwest of Calgary, Alberta. Their eastern border is marked by the Rocky Mountain Trench.

That part of the Selkirk Mountains examined in this thesis includes most ridge tops found within the region bounded on the western and northeastern flanks by the Big Bend of the Columbia River, on the east by Windy and Norman Wood Creeks, and on the south by Goldstream River (Figs. 1-1 and 1-2 (in pocket)).

Numerous geographic and geologic features are referred to in the study. In order to avoid confusion for the reader all features mentioned are positioned on Figure 1-2 (in pocket). The reader is

Fig. 1-1. Location of study area in southeastern British Columbia.

encouraged to keep this figure at hand throughout the ensuing discussions.

Revelstoke, located on the Trans Canada Highway about 100 km south of the study area, is the nearest major town from which this study area is readily accessible. The village of Mica Creek, situated outside the northwest corner of the study area, provides most basic necessities for provisioning, including a helicopter base.

The boundary of the study region outlines an area approximately 850 km^2 . The maximum relief, measured between the valley of the Columbia River and the highest peaks, is about 2.5 km whereas the average relief, measured between ridge crests and adjacent valley bottoms, is approximately 1.3 km.

Chartered helicopter provides access to most parts of the study area. Continuing logging operations along the flanks of the Columbia River and up major river or stream valleys may provide routes into peripheral parts of the study area in the near future.

Outcrop above tree line is generally abundant. However, valley walls and bottoms are entanglements of brush and trees with poor outcrop exposure. This study is therefore concentrated on data collected along ridge tops.

The Selkirk Mountains are steep and rugged and are adorned by extensive and commonly crevassed icefields. Weather can be harsh and unpredictable.

(1-3) GEOLOGIC SETTING

The northern Selkirk Mountains are associated with the following geologic features (Wheeler and Gabrielse 1972):

- (1) They are part of the zone of transitional structural style between the core zone and the Foreland Thrust and Fold Belt of the Columbian Orogen;
- (2) They form an appendage of the Shuswap Metamorphic Complex;
- (3) They are bounded on the eastern margin by the Purcell Thrust Fault, a major Cordilleran lineament.

(1-4) PREVIOUS METAMORPHIC STUDIES

The northern Selkirk Mountains and neighbouring regions were mapped at a scale of 1 : 250,000 by Wheeler (1963, 1965). Garnet, kyanite, and sillimanite isograds for the study area were documented (Wheeler 1965).

Franzen (1974) and van der Leeden (1976) included treatments of metamorphism for areas around French Creek. Leatherbarrow and Brown (1978) presented a cursory treatment of isograds and metamorphic zones for this study area.

Researchers from the University of Calgary interested in regions adjoining this study area include de Vries (1971), Ghent and de Vries (1972), Ghent (1975), Robbins (1976), Craw (1977), and Knitter (1979).

(1-5) SYNOPSIS OF STRATIGRAPHY OF THE NORTHERN SELKIRK MOUNTAINS

Rocks within the northern Selkirk Mountains belong to the Hadrynian Horsethief Creek Group (which forms part of the Windermere Supergroup) and the late Hadrynian to early Cambrian Hamill Group (Wheeler 1965). Distribution of stratigraphic members is illustrated on Fig. 1-3. Descriptions of units within the study area, outlined below, are abstracted from the stratigraphic analysis that forms part of the Ph.D. thesis of M.J. Perkins.

The Horsethief Creek Group is divided into a lower grit member, a lower pelite member, a middle marble member, and an upper pelite member.

The lower grit member is composed of feldspathic grits and conglomerates which are interbedded with pelites and minor carbonates.

The lower pelite member is characterized by finely interbedded, amphibolite bearing semipelites, pelites, psammites, and carbonates. Outcrops of this member have a striped appearance because of this interlayered nature. Metamorphic index minerals are present but not abundant in this succession.

The middle marble, which lies above the lower pelite, is a sequence of marbles, impure marbles, calcareous pelites, semipelites, and calc-silicates. This sequence is commonly lenticular in nature and shows significant lithologic variation. Facies changes occur both along and across strike.

The upper pelite member lies above the middle marble and represents the uppermost unit of the Horsethief Creek Group. Thick successions of pelite containing abundant metamorphic index minerals

Fig. 1-3. Lithologic map of the northern Selkirk Mountains (modified after Brown and Tippett (1978)).

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characterize this member. Psammites are present near the contact with the middle marble, but these give way to pelites farther away from the contact. Amphibolitic layers, common within the lower pelite member, are lacking in the upper pelite.

Above the Horsethief Creek Group lies the Hamill Group. Where exposed, the contact is invariably a fault. The succession is characterized by thinly bedded quartzites with muscovite rich pelitic partings. Impure carbonates, calcareous quartzite, thin dolomites, and marble bands are also present.

Regional correlations of the Horsethief Creek stratigraphy have been advanced by Brown et al. (1977) and are shown in Fig. 1-4.

Three large bodies of igneous origin are found within or adjacent to this study area. In the northeast part of the study area is an elongate nepheline syenite gneiss. The Adamant Pluton, a monzonite gneiss, lies just east of the study area, and the Bigmouth Creek granitic stock lies just west of the region. Two small granitic stocks are located north of French Creek.

(1-6) SYNOPSIS OF STRUCTURAL GEOLOGY OF THE NORTHERN SELKIRK MOUNTAINS

The structural geometry and succession of deformation events in the Selkirk Mountains have been the subject of much investigation. A synopsis of pertinent features, provided below, and illustrated in Fig. 1-5, is drawn from work by Franzen (1974), van der Leeden (1976), Tippet (1976), Lane (1977), Brown and Tippet (1978), Brown et al. (1977), and Brown (1978). For detailed descriptions of the structures and deformational sequence the reader is referred to these works.

Fig. 1-4. Correlation chart for late Hadrynian and early Cambrian strata in the Dogtooth Range, the Selkirk Mountains, and the Cariboo Mountains. Unit thickness not to scale (from Brown et al. 1977, fig. 10).

Fig. 1-5. Structural map of the northern Selkirk Mountains (modified after Brown and Tippett (1978)).

The earliest recognized phase of deformation resulted in nappe formation (van der Leeden 1976; Lane 1977). Direct structural evidence is sparse; however, the stratigraphic succession west of Downie and French Creeks is overturned, suggesting a minimum limb length of 20 km. Read and Brown (1979) have traced an F_1 axial trace related to this nappe southwest of this study area.

Second phase (F_2) folds are dominant in the southwestern half of this study area. These tight to isoclinal folds are long limbed and possess a well developed penetrative mineral foliation. Axial surfaces trend northwest and dip to the northeast.

Third phase (F_3) folds refold F_2 structures. Although the megascopic expression of F_3 folds dominates the northeast half of the study area, third phase crenulations of F_2 fabric are common throughout the study area. Axial planes of megascopic F_3 folds trend northwest with moderate southwest dips near the Rocky Mountain Trench, becoming progressively steeper until they reach vertical attitudes near Argonaut Mountain.

The superposition of southwest dipping F_3 folds on F_2 structures has resulted in the upward diverging Selkirk fan structure (Brown and Tippet 1978; Fig. 1-2 (in pocket)).

Previous work by Franzen (1974), Tippet (1976), and van der Leeden (1976) has indicated that the peak of metamorphism occurred during and after F_2 but before F_3 . Further fabric analysis is to be contained in a thesis by M.J. Perkins. The Adamant Pluton and nepheline syenite gneiss are thought to be pre- D_2 whereas the granitic stocks are thought to be post- D_2 and pre- D_3 (Brown and Tippet 1978; Brown et al. 1977).

Faults are recognized within the northwest part and along parts of the periphery of the study area, and along the contacts between the Hamill and Horsethief Creek Groups. One fault, which has been documented near the headwaters of Norman Wood Creek (Brown and Tippet 1978), separates two metamorphic zones in this thesis area. The Rocky Mountain Trench, which lies along the northeastern part of the study area, is the erosional expression of the Purcell Thrust (Simony and Wind 1970). This major fault separates the Foreland Thrust and Fold Belt from the transition and core zones of the orogen. The Columbia River Valley Fault, mapped by Brown and Psutka (1979), borders part of the western boundary of the study area.

CHAPTER 2.

PETROLOGY OF PELITIC ROCKS

(2-1) INTRODUCTION

A primary object of interest in this study is to present a metamorphic map (Map 2, in pocket) for the northern Selkirk Mountains. Metamorphic mineral assemblages, which provide the data base for the map, are included in Appendix A. Only assemblages representative of pelitic bulk compositions are considered in this study. All station locations are shown on Map 1 (in pocket). To assist the reader in locating stations of interest, all localities referred to in this chapter are shown on Fig. 2-1.

Throughout this study the term "isograd" denotes the trace of a surface across which a specific change in metamorphic mineralogy takes place (Elles and Tilley 1930, p. 622). These changes in mineralogy must occur in rocks of similar bulk composition and, in this study, are related to metamorphic reactions where possible. For those parts of this study area where boundaries between metamorphic zones are not closely bracketed, the trace is taken to be the median line through the area of uncertainty.

Other aspects of Map 2 (in pocket) require an explanation. The assemblages have been superimposed on a lithologic map. This enables viewers to relate assemblage and sample distribution to regional variations in bulk composition. It also provides geologists with a map that can be used to assess potential for future petrological research. Topography and icefields are also plotted to provide geographic information about the area.

The northern Selkirk Mountains are divided into northwest trending mineral zones indicative of metamorphic grades in the

Fig. 2-1. Location map for stations referred to in the text of Chapter 2.

Barrovian facies series (e.g. Miyashiro 1973) ranging from lower greenschist in the southwest and northeast to upper amphibolite in central parts of the study area (Fig. 2-2). These zones will be discussed as two prograde series. The first succession, referred to as the French Creek section, begins with the lowest grade chlorite zone rocks in the southwest and proceeds northeastward to a normal fault marking the boundary with the Windy Range section. The second sequence, referred to as the Windy Range section, commences in the staurolite-muscovite-quartz-kyanite-garnet-biotite zone in the northeast near the Rocky Mountain Trench and extends southwestward across the potassic feldspar-sillimanite zone to the above mentioned fault.

Each of the metamorphic zones in these two series first will be described. The boundaries between zones and assemblages within the zones will then be discussed for each of the prograde series. The last part of this chapter compares this information with a petrogenetic grid.

It is necessary to outline the criteria used in this study to establish compatibility between mineral phases. Mineral grains are considered to be in equilibrium if they occur within a restricted area (i.e. thin section scale); are within an apparently homogeneous compositional layer (i.e. no changes in mineral assemblage or relative abundances of minerals); and have grain boundaries that lack reaction textures (e.g. coronas). Phases need not be in mutual contact to be considered in equilibrium for the following reasons: first, mineral grains may be in contact and have grain boundaries exhibiting apparent textural equilibrium and yet a

Fig. 2-2. Metamorphic mineral zones in rocks of pelitic bulk composition in the northern Selkirk Mountains.

chemical kinetic barrier may exist, preventing the phases from reacting to attain equilibrium; secondly, it is geometrically difficult, at the two dimensional scale of a thin section, to position more than three mineral phases so that they are in mutual contact; and thirdly, work on reaction textures by Carmichael (1969) and others has shown that reactions can occur between phases not in contact by transfer of ions or by reaction through intermediary phases.

(2-2) DESCRIPTIONS OF METAMORPHIC ZONES

SOUTHWEST OF METAMORPHIC CULMINATION (FRENCH CREEK SECTION)

(2-2-1) Chlorite Zone

A chlorite zone, defined by the assemblage chlorite-muscovite-quartz-plagioclase-ilmenite-epidote-tourmaline-zircon-opaques, is located in the southwestern part of the study area, to the west of French Creek, (Fig. 2-2, Map 2 (in pocket)). Chlorite occurs both as porphyroblasts and as fine grains disseminated throughout the matrix. The porphyroblastic grains are euhedral to subhedral and average about 1 mm in length. Some grains contain straight quartz inclusion trails or are twinned. Matrix chlorite is fine grained, concordant with the dominant foliation, and is in apparent equilibrium with other phases. These textural criteria suggest that chlorite is prograde.

(2-2-2) Chlorite-Biotite Zone

Immediately northeast of the chlorite grade rocks is the chlorite-biotite zone, characterized by the assemblage chlorite-biotite-muscovite-quartz-plagioclase-zircon-epidote-tourmaline-ilmenite-opaques. This zone is about 2.5 km wide, is located west of French Creek, and represents the lowest grade assemblage observed within the limits of mapping on the ridge immediately east of French Creek (Fig. 2-2). Thick successions of calcareous pelites and psammites west of French Creek make placement of a southwestern boundary for this zone difficult.

Chlorite maintains the habit described in the previous section; however, it becomes less abundant towards the high grade boundary of this zone.

Biotite occurs predominantly as porphyroblasts, 1-2 mm in length, which may be oriented at an angle to the dominant matrix foliation. These porphyroblasts commonly contain straight inclusion trails which may curve near the biotite grain boundaries. Chlorite, muscovite, and less commonly biotite, define a matrix foliation that is, in many places, crenulated.

(2-2-3) Garnet-Chlorite Zone

The garnet-chlorite zone, located immediately northeast of the chlorite-biotite zone (Fig. 2-2, Map 2 (in pocket)), is typified by the combination of phases garnet, chlorite, staurolite and biotite as shown in Fig. 2-3. Accessory minerals include

Fig. 2-3. AFM diagram for mineral assemblages observed in the chlorite-biotite and garnet-chlorite zones from the French Creek section.

ilmenite, epidote, zircon, and opaques. This zone is about 2.5 km west of French Creek where zone boundaries are in part obscured by rocks of inappropriate bulk composition, and about 2 km east of French Creek. The eastern part of the garnet-chlorite zone boundary is positioned immediately south of stations A045, A049, A066, and A075.

Garnets from this zone are subhedral to anhedral and are riddled with abundant quartz and less common ilmenite inclusions. Some garnets are poikiloblastic replacements of compositional layers and hence have an elongate habit. These poikiloblasts are up to 3 mm in length.

Within the garnet-chlorite zone two chlorite habits are observed. Individual grains of chlorite lie within the dominant foliation and are similar in habit to chlorite grains from the chlorite and chlorite-biotite zones. These grains are interpreted as representing prograde chlorite. Chlorite also occurs as overgrowths on grain boundaries and cleavage traces of biotite porphyroblasts; as radiating acicular fans that crosscut the rock fabric; and as rims (with lesser biotite or muscovite) around garnet. These textures are interpreted as representing retrograde metamorphism and are common in assemblages ranging from garnet-chlorite to potassic feldspar - sillimanite within the northern Selkirk Mountains. In low grade assemblages, difficulties arise in distinguishing between prograde and retrograde chlorite; however, at higher grades, retrograde chlorite is readily identified. Prograde chlorite has not been identified in some garnet-chlorite zone assemblages.

Samples from localities A066 and A075 contain rare grains of staurolite. At station A060 staurolite may also be present; however, the extremely small grain size and sparse distribution make positive identification difficult. The assemblage staurolite-garnet-chlorite (no biotite) is observed at station A075 and is plotted on Fig. 2-3. Staurolites at A066 and A060 are associated with biotite, inclusion riddled feldspar porphyroblasts, garnet and/or chlorite. These assemblages are plotted on Fig. 2-4.

(2-2-4) Staurolite-Biotite Zone

Five stations west of French Creek (V106, V124, V125, V140, and V144) and seven stations east of French Creek (A019, A027, A060, A066, A093, F041, and F040) contain biotite-garnet-muscovite-staurolite-chlorite (prograde and retrograde) $\pm$ plagioclase $\pm$ quartz $\pm$ tourmaline $\pm$ ilmenite $\pm$ opaques $\pm$ zircon. Mineral assemblages compatible with this zone are outlined in Fig. 2-4.

Staurolite forms anhedral poikiloblasts up to 4 mm in length. Inclusions are generally quartz and may form straight trails. Locally staurolite occurs within biotite grains which rim severely embayed garnets. Some staurolite grains have rims of muscovite. Biotite occurs both as porphyroblasts and as grains within the ground mass. Garnets have anhedral embayed outlines and quartz inclusions. Chlorite occurs mostly as rims on garnet and biotite, or as radiating acicular mats which are oriented obliquely to the dominant foliation. These habits are interpreted as representing retrograde growth. Chlorite is also found as individual, optically

Fig. 2-4. AFM diagram for mineral assemblages observed in the staurolite-biotite zone of the French Creek section.

homogeneous, subhedral grains which lie within the dominant foliation. This habit is interpreted as representing a prograde texture. The distinction between prograde and retrograde chlorite is difficult to make in rocks from this zone. The best examples of prograde chlorite are present in samples from stations A060, V124, and V125.

Ten of the twelve stations mentioned initially in this section lie upgrade of the garnet-chlorite zone (Map 2, in pocket). However, staurolite-biotite assemblages are present at stations A060 and A066 which are located near the southwestern boundary of the garnet-chlorite zone.

The staurolite-biotite zone ranges from about 2 km wide west of French Creek to generally less than 1 km wide east of French Creek. However, at station A018, east of French Creek, staurolite-kyanite-biotite bearing rocks are found within several hundred metres of garnet-chlorite assemblages. Although staurolite-biotite assemblages occur between the garnet-chlorite and staurolite-kyanite-biotite zones about 1 km east and north of this station, the zone is not recognized in the immediate vicinity of station A018. Either the zone is very narrow or entirely absent in the vicinity of station A018, or the assemblage recorded at this station does not represent an equilibrium assemblage within the staurolite-biotite zone.

West of French Creek, the zone of uncertainty extending from the last appearance of garnet-chlorite to the first appearance of staurolite-biotite is 2.5 - 4 km wide. This zone of uncertainty corresponds with a region of psammites and calcareous pelites (Map 2, in pocket).

(2-2-5) Staurolite-Kyanite-Biotite Zone

Northeast of the staurolite-biotite zone is a staurolite-kyanite-biotite zone which ranges from 1 to 6 km in width (Map 2, in pocket). Combinations of the phases staurolite, kyanite, garnet, biotite, muscovite, plagioclase, and quartz are recorded in this zone and shown on Fig. 2-5. Accessory minerals include chlorite (retrograde), ilmenite, tourmaline, opaques and zircon.

Within this zone, staurolite forms subhedral to anhedral quartz riddled poikiloblasts which are generally found near garnet, biotite, and kyanite.

Kyanite occurs as subhedral poikiloblasts with quartz inclusions. Kyanite grains are either inside coarse optically continuous biotite or have grain boundaries intergrown with biotite. Some kyanite grains from samples taken near the high grade boundary of this zone are rimmed by muscovite. Although kyanite is the dominant aluminum silicate, two sillimanite bearing localities (F033 and A103) are recorded 3 km southwest of the sillimanite-in isograd. The sillimanite at these localities occurs as fibrolitic wisps in muscovite or biotite and is not associated with kyanite. These two localities are near carbonate units.

Garnets from the staurolite-kyanite-biotite zone are poikiloblastic. When associated with kyanite or staurolite these grains have embayed grain boundaries and are surrounded by coronas of predominantly biotite with lesser amounts of muscovite and quartz. These coronas may mimic original euhedral garnet outlines. Staurolite is commonly found near these biotite rims.

Fig. 2-5. AFM diagram for mineral assemblages observed in the staurolite-kyanite-biotite zone from the French Creek section.

(2-2-6) Sillimanite-Garnet-Biotite Zone

In the vicinity of French Lake isograds marking two major changes in mineral parageneses are recorded. Proceeding northeastward into the sillimanite-garnet-biotite zone the aluminum silicate phase changes from kyanite to sillimanite, and staurolite bearing assemblages are replaced by assemblages lacking staurolite. Compatible assemblages of sillimanite, garnet, and biotite typical of this zone are outlined on Fig. 2-6. Accessory minerals include tourmaline, chlorite (retrograde), zircon, ilmenite, and opaques.

This zone is about 6 km wide east of French Creek where it is well defined (Fig. 2-2, Map 2 (in pocket)). West of French Creek the southeastern boundary of this zone is not positioned as precisely. Data are not available northeast of Argonaut Mountain and hence it is difficult to position the northeastern boundary of this zone. It is assumed from the data available near the headwaters of Norman Wood Creek that the sillimanite-garnet-biotite zone extends up to or near a northwest trending fault (Fig. 2-2).

The position of the kyanite-sillimanite isograd west of French Lake differs significantly from the position proposed by van der Leeden (1976). According to van der Leeden the sillimanite isograd extends through station V084. Thorough examination of samples by this writer has established that van der Leeden misidentified paragonite as sillimanite. Optical properties, X-ray patterns, and qualitative electron microprobe analyses indicate that this phase is a mica and probably paragonite.

Other paragonite localities have been observed (stations V124,

Fig. 2-6. AFM diagram for mineral assemblages observed in the sillimanite-garnet-biotite zone from the French Creek section.

A019, A150, A165, and A173). At all localities paragonite has an elongate habit and occurs as small grains or weakly radiating aggregates in muscovite. These muscovite grains may enclose either plagioclase or kyanite (Fig. 2-7).

Occurrences of kyanite and staurolite within the sillimanite-garnet-biotite zone are rare. At station F050, about 1.5 km inside the zone near French Lake, the only certain overlap of kyanite bearing assemblages into the sillimanite field is observed.

Staurolite is found at stations A192, A193, A218, F050, F055, F057a, and F154. Only stations A218 and F057a exhibit staurolite that coexists with muscovite. All other stations exhibit muscovite deficient assemblages.

Throughout this zone sillimanite occurs as fibrolitic wisps in biotite, muscovite, or quartz, or as knots of pure fibrolite. Coarse grained sillimanite is rare.

Garnets from this zone are either embayed poikiloblasts or euhedral and relatively inclusion free. The first type of garnet is similar in habit to garnets from the previously described lower grade metamorphic zones and needs not be associated with sillimanite. The euhedral variety of garnet is invariably found rimmed by, or in close association with, sillimanite and biotite. This texture is not observed at lower metamorphic grades.

Garnets with quartz inclusion trails in core regions and inclusion free euhedral rims occur locally. The boundary between rim and core habits is sharp but irregular in shape.

Pelites west of French Lake contain thin white pegmatite dykelets and coarse grained pegmatitic segregations. These

Fig. 2-7. Sketch of paragonite (pa) grains in muscovite as observed in a thin section of sample A165c. Muscovite encloses embayed kyanite.

These dykelets and segregations are composed of quartz, plagioclase, and muscovite. Although some segregations appear to be rootless, others form series of segregations which may represent boudinaged dykes. Staining of a suite of pegmatite samples indicates that potassic feldspar is not present.

(2-2-7) Stauroelite-Muscovite-Quartz-Garnet-Biotite-Sillimanite Zone

Near the headwaters of Norman Wood Creek five localities (A266, A274, A276, A277, and A278) containing the assemblage stauroelite-muscovite-quartz-garnet-biotite-sillimanite outline a narrow band about 2 km long (Fig. 2-2, Map 2 (in pocket)). At two localities (A277 and A278) stauroelite is present in outcrop but was not observed in thin section. Figure 2-8 outlines the assemblage observed in this zone. Accessory minerals include tourmaline, opaques, and zircon.

This zone is bounded along most of its northeast flank by a northwest trending fault. Occurrences of sillimanite-garnet-biotite assemblages are found along the northwestern and southeastern margins of this zone. Distribution of this zone is limited and assigning it to a "zone" may be misleading. However, this paragenesis is anomalous within the immediate region and documentation is therefore considered to be necessary.

Stauroelite occurs as resorbed poikiloblasts which may be closely associated with sillimanite. Garnets are euhedral and commonly occur within knots of intergrown sillimanite and biotite. Sillimanite has two habits. Most of the sillimanite is intergrown

with biotite and forms knots. Also, fibrous mats of sillimanite up to 10 cm in length are found in coarse grained rootless lenses of quartz, plagioclase, and muscovite.

Coarse grained aggregates of quartz, muscovite, and plagioclase form about 5% of some outcrops in this zone. These segregations are about 0.3 m long, lack fabric, and may be surrounded by rinds of biotite. Pegmatite dykes are also present in this region.

The traverse through the French Creek section is now complete; there follows a description of the prograde succession that is recorded in the Windy Range section.

NORTHEAST OF METAMORPHIC CULMINATION (WINDY-RANGE SECTION)

(2-2-8) Staurolite-Muscovite-Quartz-Kyanite-Garnet-Biotite Zone

The lowest grade rock encountered within the limit of mapping on the northeast side of the study area (Fig. 2-2, Map 2 (in pocket)) contains the assemblage staurolite-muscovite-quartz-kyanite-garnet-biotite. Assemblages representative of this zone are depicted in Fig. 2-9. Zircon, opaques, tourmaline, and chlorite (retrograde) may be present as accessory minerals.

Assemblage data for stations A369, A376, A377, and A378 (Fig. 2-1) were determined from hand specimen examination and X-ray identification of staurolite. Attempts to obtain staurolite bearing thin sections from rocks from these localities have not been successful because of the sparse distribution of staurolite.

Fig. 2-9. AFM diagram for mineral assemblages observed in the staurolite-muscovite-quartz-kyanite-garnet-biotite zone from the Windy Range section.

Quartz riddled staurolite porphyroblasts, and less commonly inclusion free staurolite, occur within relatively inclusion free kyanite porphyroblasts and biotite. These quartz inclusions may define straight inclusion trails. Garnets are of two varieties: inclusion free, subhedral to euhedral; and coarse, euhedral to anhedral porphyroblasts with curved quartz and opaque inclusion trails. Euhedral to subhedral garnet porphyroblasts are completely to partially enclosed by some staurolite porphyroblasts.

(2-2-9) Kyanite-Garnet-Biotite Zone

The kyanite-garnet-biotite zone, which is immediately southwest of the staurolite-muscovite-quartz-kyanite-garnet-biotite zone, is approximately 8 km wide (Fig. 2-2, Map 2 (in pocket)).

Assemblages commonly observed in this zone are outlined on Fig. 2-10. Tourmaline, zircon, opaques, and chlorite (retrograde) may be present as accessory phases.

Available data, although sparse, constrain the location of the isograd along the northeastern boundary of this zone reasonably well. The boundary can be extrapolated northwest to coincide with the trace of the staurolite-out isograd through the bend of the Columbia River as proposed by Craw(1978). Only one station in the Windy Range (P147) contains staurolite. R.L. Brown (personal communication, 1979) has observed staurolite bearing assemblages east of Windy Creek near McNaughton Lake. De Vries (1971) maps a staurolite-out isograd in the Esplanade Range, southeast of this region. Since this zone appears to extend to the southeast,

Fig. 2-10. AFM diagram for mineral assemblages observed in the kyanite-garnet-biotite zone from the Windy Range section.

the isograd is positioned through the Windy Range using the limited data available. The effect of the fault immediately south of station Pl47 (Map 2, in pocket) on the position of this isograd is not known. However, as will be discussed later, vertical displacement on this and other faults in the Windy Range is thought to be minor.

Throughout this zone kyanite has a subhedral, relatively inclusion free, porphyroblastic habit and is generally found in close association with garnet and (or) biotite. Garnets are euhedral to rounded and may contain centrally located randomly oriented quartz inclusions. Less commonly, garnets are riddled with quartz inclusions and have embayed grain boundaries. These garnets are generally not associated with kyanite.

Some pelites within this zone contain coarse grained segregations of quartz, plagioclase, and mica which lack fabric. Where present, these aggregates generally form from 5 to 10% of the rocks. Rare aggregates contain coarse kyanite blades, quartz, plagioclase, muscovite, and biotite. Rinds of biotite are found along segregation-matrix contacts.

Pegmatite dykes and bodies are common within most rock types from this zone. Accessory minerals within these pegmatites include tourmaline and garnet. Pegmatites which lie in the axial planes of folds crosscut other pegmatites which have been folded. This evidence indicates episodic emplacement of pegmatites.

(2-2-10) Sillimanite-Garnet-Biotite Zone

Southwest of the kyanite-garnet-biotite zone is the sillimanite-garnet-biotite zone (Fig. 2-2, Map 2 (in pocket)). In the Windy Range, where the boundaries of this zone are best defined, the width of the zone is approximately 2.5 km.

The northerly trending ridge containing Mount Chapman (Fig. 1-2) has been inadequately sampled. It is therefore necessary to draw assemblage data from field notes. Even with these data, the isograd is still poorly constrained. The transition can be interpolated through Mount Chapman using the reasonably well constrained positions mapped at the head of Mica Creek and south of Neptune Peak. This interpolation is consistent with the position of the kyanite-sillimanite isograd proposed by Wheeler (1965).

Assemblages containing sillimanite, garnet, or biotite within this zone are plotted in Fig. 2-11. Opaques, chlorite (retrograde), or zircon may be present as accessory minerals.

With the exception of one kyanite occurrence at station M081 (hand specimen does not appear to correspond with field note description and may be improperly labelled) sillimanite is the only aluminum silicate polymorph observed in this belt. Sillimanite forms fibrolitic intergrowths with muscovite or biotite. Garnet need not be present in sillimanite bearing assemblages; however, where present it has a euhedral, relatively inclusion free habit. These garnets are of the same habit as those discussed in the preceding kyanite-garnet-biotite zone. Garnets encased in sillimanite-biotite knots are present but rare.

Fig. 2-11. AFM diagram for mineral assemblages observed in the sillimanite-garnet-biotite zone from the Windy Range section.

Potassic feldspar is observed in only one sample (P091b) and, since sillimanite is both rare and not associated with potassic feldspar in this specimen, the assemblage potassic feldspar - sillimanite is not considered to be stable. Throughout most of this zone muscovite is euhedral to subhedral; however, locally it may exhibit a vermicular texture. Coarse grained leucocratic segregations and pegmatite relations are similar to those discussed in the kyanite-garnet-biotite zone. Aluminum silicate segregations are not recorded in this zone.

(2-2-11) Potassic Feldspar - Sillimanite Zone

A narrow zone of potassic feldspar - sillimanite assemblages, representative of the highest grade of metamorphism in the northern Selkirk Mountains, extends northwestward from the headwaters of Windy Creek to Birch Creek (Fig. 2-2, Map 2 (in pocket)).

In the Windy Creek area, where it is best defined, the zone attains a maximum width of about 3.5 km. In this vicinity, data suggest that the southwestern boundary of the zone lies along a major fault mapped by Brown and Tippet (1978). Other data along the southern margin of this zone appear to be compatible with this observation. Near Birch Creek, in the western part of the study area, the zone appears to narrow. However, the pelite horizon containing potassic feldspar - sillimanite assemblages is bounded on either side by carbonate units approximately 1 km thick, making definition of boundaries for this zone difficult. The central part of this zone has been positioned by interpolation. The region is

occupied primarily by ice, a large pegmatite body, and marble horizons. The few pelites present have not been properly sampled.

Figure 2-12 outlines the assemblages observed in the presence of muscovite and potassic feldspar (Fig. 2-12A) and muscovite only (Fig. 2-12B). Potassic feldspar bearing assemblages invariably contain muscovite. Accessory minerals include tourmaline, zircon, opaques, ilmenite, and chlorite (retrograde).

Sillimanite developed as monomineralic fibrolitic knots, or as wisps inside muscovite, biotite, or quartz. Muscovite occurs as euhedral to subhedral grains and as anhedral grains which exhibit vermicular texture. The vermicular textures are most common in grains near potassic feldspar. These grains may contain fibrolite. Biotite in juxtaposition with potassic feldspar and euhedral inclusion free garnet may also show a "wormy" texture.

Potassic feldspar forms anhedral grains which are generally associated with sillimanite and muscovite. Lenses rich in potassic feldspar and parallel to compositional layering may be present. Detailed examination generally shows that sillimanite and muscovite are nearby.

Potassic feldspar is also found in rootless coarse grained segregations containing quartz, feldspar, and muscovite. Sillimanite is found within some aggregates. Although these segregations can constitute up to 50% of some outcrops, generally they occupy less than 10% by volume. They are lens shaped and average approximately 0.3 m in length. A melanocratic rind, several millimetres thick, of biotite-tourmaline-garnet-sillimanite surrounds most lenses. Tourmaline and garnet are present inside the segregations near the

Fig. 2-12. AFM diagram for mineral assemblages observed in the potassic feldspar - sillimanite zone from the Windy Range section. Fig. 2-12A represents parageneses containing both muscovite and potassic feldspar and Fig. 2-12B represents muscovite bearing assemblages.

contact with the matrix. Within some pelitic matrices thin quartz-feldspar segregations are found.

Segregations containing fibrolitic mats up to 10 cm in length are common. Potassic feldspar and coarse grained euhedral muscovite have been observed in these segregations.

The above mentioned habits for potassic feldspar and segregations are commonly found in close association. The potassic feldspar present in the matrix of pelites is found near thin lenses of quartz + plagioclase + potassic feldspar + muscovite + sillimanite. These lenses appear to be transitional with the larger coarse grained segregations with melanocratic rinds. Gneissosity is not well developed in these rocks. These segregations are rootless and are locally removed from regions of pegmatitic dyke development.

Nodules of sillimanite-muscovite-quartz set in a matrix of plagioclase-biotite-quartz-feldspar are found in semipelitic and interlayered pelitic and psammitic units. These knots are ellipsoidal in shape and average 3 cm by 2 cm by 0.5 cm in size. They constitute 5-10% of the outcrop and are aligned in the axial planes of some folds. Where present, potassic feldspar is found in equal proportions with plagioclase, quartz, and biotite in the matrix. Potassic feldspar is rarely observed in the knots themselves. Sillimanite, on the other hand, is restricted to the nodules. The knots are surrounded by rims of biotite. A type locality for this feature is station A300.

Each of the metamorphic zones recognized in the northern Selkirk Mountains has been discussed in the preceding sections.

It is now possible to discuss the nature of the individual isograds and boundaries between zones and to assess the significance of assemblages within each zone. As in the preceding presentation the zones will be discussed as two separate prograde sequences.

(2-3) INTERPRETATION OF ISOGRADS AND PHASE RELATIONS ASSOCIATED WITH METAMORPHIC ZONES

SOUTHWEST OF METAMORPHIC CULMINATION (FRENCH CREEK SECTION)

(2-3-1) Chlorite, Chlorite-Biotite, and Garnet-Chlorite Zones

Within this section a brief treatment of the chlorite, chlorite-biotite, and garnet-chlorite zone assemblages is provided. The treatment is hampered by the lack of information on the chemical compositions of the phases under study and the exact nature of the white mica which is presumed to be muscovite. This discussion should be considered with these limitations in mind.

Boundaries between these zones are defined by the first appearance of biotite and garnet (Map 2, in pocket). Possible ranges in position of these boundaries from the available data are presented in Fig. 2-13.

The line on the map marking the first appearance of biotite is difficult to position because of the abundance of psammitic and calcareous pelitic rocks west of French Creek. Data suggest that this boundary may coincide with stratigraphic contacts and may therefore reflect a change in bulk composition (Map 2, in pocket).

Fig. 2-13.

Boundaries of the chlorite, chlorite-biotite, and garnet-chlorite zones defined by the first appearance of index minerals.

- ch
- ch-bi
- ga-bi, ga-ch, ga-bi-ch
- boundary (defined, assumed)
- - - limit of index assemblages
- ⋯ zone of uncertainty

The ridge immediately east of French Creek has been extensively sampled in the vicinity of the first appearance of garnet. Prograde assemblages of this region are plotted on a detailed map (Fig. 2-14). This map outlines at least four distinct pelitic units which are separated by marble bands.

The assemblages chlorite-biotite (B), garnet-chlorite (C), garnet-biotite (c), and staurolite-chlorite-garnet (c'), located on Fig. 2-14, all plot in different composition fields on one AFM diagram (Fig. 2-3). The change in mineralogy in the above assemblages is therefore interpreted to reflect changes in bulk composition of the rock.

The appearance of garnet in this study area cannot be attributed to any one mineral reaction. However, Jones (1972) has proposed a reaction involving ferrous dolomite as a reactant to account for the production of garnet at the garnet isograd exposed along the Trans Canada Highway, approximately 60 km to the south of this study area. Andesine, also formed by this reaction, has abundant quartz inclusions. This plagioclase habit has been observed at two garnet bearing localities (A060 and A075) east of French Creek. Sufficient data are not available to assess the applicability of this reaction to the northern Selkirk Mountains.

The two staurolite-biotite occurrences (A060 and A066) found within the garnet-chlorite zone are discussed in the next section. Variations in mineral assemblages between the chlorite, chlorite-biotite, and garnet-chlorite zones (except at stations A060 and A066) cannot be attributed to mineral reactions which are independent

Fig. 2-14. Distribution of pelitic assemblages with respect to stratigraphy in the southern French Creek section.

of bulk composition. Therefore the boundaries between the zones cannot be described as isograds and must be described only as surfaces marked by the first appearance of an index mineral.

(2-3-2) Staurolite-Biotite Zone

The boundary between the garnet and staurolite-biotite zones (Fig. 2-2, Map 2 (in pocket)) marks the appearance of coexisting staurolite and biotite. A change in topology of the AFM diagrams for the two above mentioned zones is noted across the boundary (compare Figs. 2-3 and 2-4). The garnet-chlorite tie line has been broken and is replaced by the staurolite-biotite tie line. This change in topology can be represented by the discontinuous reaction

Assemblages related to Rx. 2-1 are plotted on Fig. 2-15. Observations of staurolite within biotite grains which in turn surround embayed garnet are compatible with the above reaction.

This change in paragenesis is not dependent on bulk composition and therefore the boundary between the garnet-chlorite and staurolite-biotite zones can be termed an isograd.

A staurolite isograd that corresponds to Rx. 2-1 has been mapped by Carmichael (1970) and others. Van der Leeden (1976) describes Rx. 2-1 as one possible staurolite producing reaction for the area west of French Creek. De Vries (1971) also proposed that the

Fig. 2-15

Position of an isograd corresponding to the reactions:

- Δ assemblage: ch-ga-mu
- assemblage: ch-ga-st-bi
- assemblage: st-bi-ch
- ▽ assemblage: st-bi-ky

--- isograd (defined, assumed)
 - - - limit of reactant or product assemblages
 ... zone of uncertainty

staurolite isograd in the Esplanade Range is compatible with Rx. 2-1.

The staurolite-biotite zone is narrow and has boundaries that are poorly defined. West of French Creek this isograd could be located anywhere within a 2.5 - 4 km wide zone of uncertainty whereas east of French Creek the zone is narrow, and in the vicinity of station A018 the staurolite-biotite zone may be totally absent (see Section 2-2-4).

Two occurrences of staurolite (rare) - biotite (A060 and A066) are recorded about 2 km south of the staurolite-biotite isograd. A sample from station A060 contains both product and reactant assemblages for Rx. 2-1. This writer prefers to consider these two occurrences as assemblages not representing equilibrium conditions within the garnet-chlorite zone. It is considered more reasonable to treat these two assemblages as anomalies within the garnet-chlorite zone (hence the position of the isograd on Map 2 (in pocket)) rather than consider the majority of garnet-chlorite occurrences as metastable occurrences above the staurolite-biotite isograd. A satisfactory explanation for these occurrences of staurolite-biotite, beyond invoking disequilibrium, is not available.

(2-3-3) Staurolite-Kyanite-Biotite Zone

The staurolite-kyanite-biotite zone, which lies immediately northeast of the staurolite-biotite zone, is characterized by the appearance of kyanite. The coexistence of kyanite-biotite indicates that the chlorite-staurolite tie line of the AFM diagram for the staurolite-biotite zone (compare Figs. 2-4 and 2-5) is no longer

stable and is replaced by the kyanite-biotite tie line in the staurolite-kyanite-biotite zone. A mineral reaction such as

can account for this change in topology.

The difficulty in distinguishing prograde from retrograde chlorite (see Section 2-2-4) in the staurolite-biotite zone and the relative scarcity of staurolite-biotite zone assemblages makes recognition of reactant assemblages for Rx. 2-2 problematic. Nonetheless, the isograd marking the appearance of staurolite-kyanite-biotite assemblages is reasonably well defined (Fig. 2-15, Map 2 (in pocket)).

Carmichael (1970) has mapped one isograd on the basis of Rx. 2-2 which lies about 1 km upgrade from an isograd determined using Rx. 2-1. The staurolite-biotite zone in the northern Selkirk Mountains, although poorly constrained, also appears to be narrow.

East of French Creek the staurolite-kyanite-biotite zone is about 6 km wide whereas west of French Creek the sparse distribution of staurolite-kyanite-biotite bearing stations in predominantly semipelitic rocks indicates the zone is less than 1 km wide locally.

(2-3-4) Sillimanite-Garnet-Biotite Zone

Two major changes in mineralogy are observed as the sillimanite-garnet-biotite zone is entered from the staurolite-kyanite-biotite zone.

Kyanite is replaced by sillimanite and staurolite bearing assemblages become extremely rare. As outlined in Section 2-2-6, few kyanite occurrences are recorded in the sillimanite zone. All kyanite localities in the sillimanite-garnet-biotite zone are within 1 km of the kyanite-sillimanite isograd. Most staurolite bearing rocks from this zone exhibit muscovite deficient assemblages.

A marked change in the habit of garnet is observed across this boundary. The appearance of euhedral, relatively inclusion free garnets closely associated with sillimanite and biotite, in contrast with the embayed poikiloblastic habit of garnet observed at lower grades, suggests a new episode of garnet growth. Partially resorbed garnet poikiloblasts found within this zone may represent relict garnets formed at lower grades and are common in assemblages that lack sillimanite. Garnets with straight inclusion trails in central regions and euhedral, inclusion free rims are also found within the sillimanite-garnet-biotite zone. These garnets may represent renewed garnet growth on an early garnet porphyroblast.

The change in mineralogy recorded across the boundary between these zones is represented by two changes in the AFM diagram (compare Figs. 2-5 and 2-6). Kyanite is replaced by sillimanite at the "A" apex of the diagram and staurolite is consumed making the sillimanite-garnet-biotite field stable.

The first of these two changes suggests that the following reaction may have been operative:

(Rx, 2-3)

Figure 2-16 outlines the possible ranges of position for an isograd based on Rx. 2-3. Along the ridge top east of French Creek this isograd is well constrained. Steep topography and calcareous bulk compositions along the eastern flank of the valley of French Creek necessitate a 1 km uncertainty in the position of this isograd. The western segment of this isograd varies from well constrained to a range of locations across a belt 1.5 km wide.

The second change in mineral paragenesis, staurolite disappearance and sillimanite-garnet formation, suggests that the reaction

may have occurred.

Reaction 2-4 has been mapped by Thompson and Norton (1968), Carmichael (1970), and Froese and Gasparrini (1975). Data limiting an isograd corresponding to Rx. 2-4 are presented on Fig. 2-17. Staurolite bearing rocks found upgrade of this isograd and lacking muscovite are not considered within this study since they do not contain complete reactant assemblages. The appearance of the euhedral garnet habit in this zone is compatible with the position of the isograd corresponding to Rx. 2-4.

On the ridge east of French Creek an isograd corresponding with Rx. 2-4 is well defined (Fig. 2-17). The topographical and compositional problem along the east flank of the valley of French Creek, as described above, results in about 1 km uncertainty in isograd location. West of French Creek a zone approximately 4 km

Fig. 2-16

Position of an isograd corresponding
to the reaction:

- Δ reactant assemblage: ky
- ▼ product assemblage: si
- isograd (defined, assumed)
- - - limit of reactant or product assemblage
- ||||| zone of uncertainty

Fig. 2-17

Position of an isograd corresponding
to the reaction:

- product and reactant assemblage
- △ reactant assemblage: st-mu-qz
- ▼ product assemblage: ga-bi-si
- isograd (defined, assumed)
- limit of reactant or product assemblage
- zone of uncertainty

wide separates the disappearance of reactants from the appearance of products. Calcareous rocks occupy about 1.5 km of this 4 km zone. An occurrence of the product and reactant assemblage staurolite-muscovite-quartz-garnet-biotite-sillimanite at station V163 just inside the southern limit of this zone implies that the isograd may pass near the southern boundary of the zone of uncertainty.

From the above discussion it appears that the southwestern boundary of the sillimanite-garnet-biotite zone may represent two mineral reactions (Rx. 2-3 and 2-4). East of French Creek these two isograds are thought to be coincident. West of French Creek the positions are inferred to be roughly coincident. However, further work is necessary to test the position of the staurolite-out reaction.

(2-3-5) Staurolite-Muscovite-Quartz-Garnet-Biotite-Sillimanite Zone

The small enclave of staurolite-muscovite-quartz-garnet-biotite-sillimanite bearing rocks near the headwaters of Norman Wood Creek (Map 2, in pocket) contains both reactant and product assemblages corresponding to Rx. 2-4. The six phases of this paragenesis appear to coexist in textural equilibrium. Euhedral garnets from this zone exhibit textures similar to those in the sillimanite-garnet-biotite zone which are believed to correspond with staurolite breakdown. However, the presence of staurolite suggests that Rx. 2-4 has not gone to completion.

The northwestern and southeastern limits of this zone are

directly bounded by sillimanite-garnet-biotite parageneses which lie within the same stratigraphic package as assemblages from this zone. This suggests that the presence of staurolite need not reflect changes in bulk composition.

This narrow zone may represent a transition to lower grades of metamorphism immediately to the northeast. A northwest trending normal fault along the northeast boundary of this zone makes verification of this proposal impossible. Since faults are mapped along and within the vicinity of this zone it is also reasonable to propose that this zone was emplaced by postmetamorphic faulting.

Discussion of isograds bounding zones southwest of the metamorphic culmination is complete. The prograde succession northeast of the culmination will now be discussed.

NORTHEAST OF METAMORPHIC CULMINATION (WINDY RANGE SECTION)

(2-3-6) Staurolite-Muscovite-Quartz-Kyanite-Garnet-Biotite Zone

The assemblage staurolite-muscovite-quartz-kyanite-garnet-biotite, observed in this zone and represented on Fig. 2-9, may correspond to the reaction assemblage corresponding with the reaction

The euhedral to subhedral nature of staurolite, kyanite, and biotite, and the lack of reaction textures suggest that these

minerals are compatible. The phases staurolite, kyanite, garnet, and biotite are observed in mutual contact and have straight grain contacts.

Two factors make discussion of this zone difficult: the zone is only defined by 11 widely distributed samples; and neither the position of the low grade boundary of this zone nor the evolution of the lower grade assemblages is known. The width of the zone containing this paragenesis is therefore not known.

The coexistence of staurolite-muscovite-quartz-kyanite-garnet-biotite in this zone is problematic. It may be that sluggish reaction rates or trace chemical components, among other factors, allow the reaction assemblage to persist across a wide area. Further work is necessary to determine the significance of this paragenesis.

(2-3-7) Kyanite-Garnet-Biotite Zone

The boundary between the staurolite-muscovite-quartz-kyanite-garnet-biotite zone and the kyanite-garnet-biotite zone is marked by the disappearance of staurolite. Euhedral garnets, which may contain randomly oriented quartz inclusions in central regions of grains, occur with kyanite and biotite.

The change in mineralogy noted across the above mentioned boundary is represented on the AFM diagram by the disappearance of staurolite and the presence of a stable kyanite-garnet-biotite triangle. A reaction such as

can account for the consumption of staurolite at this boundary.

Reactant and product assemblages for Rx. 2-5 are plotted on Fig. 2-18.

Available data in central and northwestern segments of this zone constrain the position of the isograd reasonably well. Southeastern parts of the isograd are poorly constrained. However, for reasons discussed in Section 2-2-9 the isograd is interpolated through this region.

(2-3-8) Sillimanite-Garnet-Biotite Zone

The only change in mineral paragenesis between the sillimanite-garnet-biotite zone and the lower grade kyanite-garnet-biotite zone involves the transition from kyanite to sillimanite (compare Figs. 2-10 and 2-11). The polymorphic transition

ky $\leftrightarrow$ si

(Rx. 2-3)

is correlated with this boundary.

Field locations of these two phases, plotted on Fig. 2-19, indicate that in areas of adequate sampling the isograd is well delineated. The location of the isograd through Mount Chapman is discussed in Section 2-2-10.

(2-3-9) Potassic Feldspar - Sillimanite Zone

The last of the pelitic metamorphic zones warranting analysis is the potassic feldspar - sillimanite zone which represents the

Fig. 2-18

Position of an isograd corresponding
to the disappearance of staurolite by
the reaction:

- product and reactant assemblage
- ▼ product assemblage: ga-bi-ky
- isograd (defined, assumed)
- isograd (defined, assumed)

Fig. 2-19

Position of an isograd corresponding
to the reaction:

Δ reactant assemblage: ky

▼ product assemblage: si

--- isograd (defined, assumed)

--- limit of reactant or
product assemblage

⋯ zone of uncertainty

highest grade of metamorphism observed in pelitic rocks of this study area. The boundary of this zone is marked on the northeast by the first appearance of coexisting potassic feldspar and sillimanite. The southwestern margin is fault bounded (Map 2, in pocket). Muscovite, commonly vermicular in habit, is present in all assemblages containing potassic feldspar and sillimanite.

Within this zone coarse grained, rootless segregations, which lack fabric and contain quartz + plagioclase + muscovite + potassic feldspar + sillimanite + garnet, are common. The habit of these segregations appears to be gradational (see Section 2-2-11) from granoblastic potassic feldspar associated with muscovite, quartz, plagioclase, and sillimanite in the matrix (early formation stages of segregation?), to thin lenses, and finally to well developed coarse grained potassic feldspar - sillimanite bearing segregations with melanocratic rinds. The rootless nature of segregations, and the lack of association with pegmatite dykes, implies that these aggregates are not intrusive in nature. Segregations from the potassic feldspar - sillimanite zone contain potassic feldspar and are therefore different from those observed in the kyanite-garnet-biotite, and sillimanite-garnet-biotite zones of the Windy Range (see Sections 2-2-9 and 2-2-10) which lack potassic feldspar. For this reason the potassic feldspar bearing segregations of the potassic feldspar - sillimanite zone are interpreted as being in situ products of a process unique to this metamorphic zone. The features associated with these segregations are compatible with a partial melting origin.

A reaction that can account for the appearance of potassic

feldspar - sillimanite, muscovite's vermicular habit, and the presence of segregations is

As muscovite is present in all northern Selkirk rocks bearing potassic feldspar and sillimanite it is evident that Rx. 2-6 has not gone to completion. The boundary between reactant, and reactant plus product assemblages associated with Rx. 2-6 is positioned on Fig. 2-20. This boundary is well defined in the Windy Creek area. However, a 3 km wide zone of uncertainty (corresponding with a zone of abundant amphibolite and carbonate) obscures the western limit of this zone.

The coexistence of reactant and product assemblages for Rx. 2-6 may be related to the presence of chemical impurities (most notably calcium) in the system. Thompson and Algor (1977) predict that the presence of anorthite in plagioclase will expand the stability field of the reactant assemblage for Rx. 2-6 to higher temperatures. This implies that partial melting will occur first in pelites with little or no anorthite and the temperatures of reaction will increase as the anorthite content increases. Tracy (1978) has mapped a zone of product plus reactant assemblages (Rx. 2-6) separating a zone of product assemblages from a zone of reactant assemblages in west central Massachusetts. The variation in anorthite composition of plagioclase across the reactant plus product assemblage zone is compatible with the prediction of Thompson and Algor (1977).

Fig. 2-20

Position of an isograd corresponding
to the first appearance of Kf-si
by the reaction:

(Rx. 2-6)

- Δ reactant assemblage: qz-μ-Naf
- product and reactant assemblage
qz-μ-Naf-Kf-si-liq
- isograd (defined, assumed)
- - - limit of reactant or
univariant assemblage
- ⋯ zone of uncertainty

The northeastern boundary of the potassic feldspar - sillimanite zone in the northern Selkirk Mountains may represent an isograd correlative with Rx. 2-6 for rocks of low calcium content.

A comment on the coarse grained segregations of quartz, feldspar and mica encountered from the sillimanite-garnet-biotite zone near French Creek northeastward to the northeastern limit of the kyanite-garnet-biotite zone near the Rocky Mountain Trench is appropriate in this section. They may represent localized, discontinuous metamorphic segregations which have formed in the solid state; have formed by migration of chemical constituents into voids or low pressure zones created during deformation; represent boudinaged dykelets or stringers of pegmatite or gneissosity; or, represent leucosomes formed by partial melting of pelites.

Partial melting of pelites by a reaction such as

as outlined by Thompson and Algor (1977) could produce leucosomes and Al_2SiO_5 at lower temperatures than by Rx. 2-6. This reaction may account for the kyanite and sillimanite bearing segregations observed in the northern Selkirk Mountains.

(2-4) FIELD OBSERVATIONS RESTRICTING THE DIPS OF ISOGRADS IN THE NORTHERN SELKIRK MOUNTAINS

Within the northern Selkirk Mountains the maximum topographic relief, measured between the valley of the Columbia River and the

highest peaks, is approximately 2.5 km; the average topographic relief, measured between ridge crests and adjacent valley bottoms, is approximately 1.3 km. Therefore, sufficient relief exists to enable mapping of isograd surfaces along valley walls and to determine the dip of these isograds.

The two isograds for which some dip information exists are the kyanite-sillimanite isograds in the French Lake and Windy Range regions. Profiles of the ridges containing these isograds, showing the locations of pertinent assemblage data, are provided in Figs. 2-21A and 2-21B. These profiles have been constructed by orthogonally projecting station locations onto ridge profiles which are approximately perpendicular to isogradic surfaces.

Data from the east side of French Creek appear to limit the dip of the kyanite-sillimanite isograd to approximately 15° to the southwest (Fig. 2-21A). When observed in plan view, data from this ridge suggest that the isograd strikes to the northwest, perpendicular to the ridge and parallel to the dominant regional isograd trend. Therefore, the above mentioned dip would reflect a true dip value. However, data from the ridge west of French Creek suggest that the isograd may turn to the west south of French Lake. This would imply that the measured dip reflects an apparent dip value, which would be less than the true dip value. Although some uncertainty exists in the interpretation of this information, the data nonetheless do restrict the dip of this isograd to southwest or vertical directions.

Assemblages at the kyanite-sillimanite isograd in the Windy Range (Fig. 2-21B) limit the dip of the isograd between steep

Fig. 2-21A. Map and cross section showing position and dip of the kyanite-sillimanite isograd near French Lake in the French Creek section.

▽ KYANITE
▲ SILLIMANITE

0 FEET 3000
[Scale bar]

HORIZONTAL AND VERTICAL SCALE

Fig. 2-21B. Map and cross section showing position and range of dips for the kyanite-sillimanite isograd south of Neptune Peak in the Windy Range section.

southeasterly and about 40° to the southwest. The regional strike direction is perpendicular to the estimated range of dips. Although poorly constrained, this range of values is compatible with the dip of 60° to the southwest estimated for the extension of this isograd across the Columbia River Valley as proposed by Ghent et al. (1977) and Knitter (1979).

(2-5) SELKIRK PHASE RELATIONS AND THE PETROGENETIC GRID

In the preceding section of this chapter mineral reactions have been proposed to represent some isograds separating metamorphic zones in the northern Selkirk Mountains. With this information, it is possible to propose metamorphic paths across a petrogenetic grid to represent the sequence of reactions and mineral zones observed in the Selkirk Mountains. This treatment will give a qualitative and semiquantitative assessment of conditions prevailing during the peak of metamorphism in this area.

(2-5-1) Selection of a Petrogenetic Grid

Petrogenetic grids pertaining to phase equilibria of interest in this study are presented by Hess (1969; modification presented in Miyashiro (1973, fig. 3-9)), A.B. Thompson (1976), and Carmichael (1978). Carmichael's grid is constructed using a combination of experimental data and "the sequence and relative spacing of ... isograds in ... well documented metamorphic terrains" (Carmichael 1978, p. 778) whereas Hess' and Thompson's grids are constructed from experimental data.

Successfully accounting for the differences between these grids is a complex task well beyond the scope of this thesis. Carmichael's grid is adopted for use in this study because it makes particular effort to combine sequences and distributions of naturally occurring assemblages with experimental work and geothermometry; and it appears to be compatible with field relations observed in the northern Selkirk Mountains. Revision of this discussion can be made when a grid acceptable to metamorphic petrologists is constructed.

Throughout the ensuing discussions in this thesis the pressure and(or) temperature estimates derived from phase relations will be determined using Carmichael's grid (Fig. 2-22). Reactions considered to represent the isograds separating the garnet-chlorite zone and staurolite-biotite zone, and the staurolite-biotite zone and the staurolite-kyanite-biotite zone, have been added to the grid. The locations of these reactions in pressure-temperature space is approximate.

(2-5-2) Pressure and Temperature Estimates

If the sequence of mineral assemblages for the French Creek section is compared with the sequence in the Windy Range section a striking difference is observed. The French Creek section passes directly from the staurolite-kyanite-biotite zone into the sillimanite garnet-biotite zone, whereas in the Windy Range section a kyanite-garnet-biotite zone is found separating the staurolite-muscovite-quartz-kyanite-garnet-biotite and sillimanite-garnet-biotite zones.

Fig. 2-22. Petrogenetic grid for rocks of pelitic composition (Carmichael 1978) showing relative positions of several metamorphic zones within the northern Selkirk Mountains. Dashed lines represent reactions from D. Carmichael (unpublished grid).

These two sequences are discussed individually.

The transition from a staurolite-kyanite-biotite zone to a sillimanite-garnet-biotite zone in the French Creek section requires the simultaneous consumption of staurolite and kyanite, and production of sillimanite. The reactions

and

have been proposed to account for these changes (Section 2-3-4). The invariant point defined by the intersection of these two reactions in pressure-temperature space occurs at about 550°C and 5 kbar (Fig. 2-22). It therefore seems reasonable to infer that temperatures of approximately 550°C and pressures of approximately 5 kbar accompanied metamorphism.

The presence of andalusite is not recorded within the northern Selkirk Mountains. Therefore the aluminum silicate triple point (approximately 4 kbar using Holdaway's (1971) data) can be taken as a minimum pressure attending metamorphism.

The southeastern part of the French Creek section, where the lowest grades of metamorphism are recorded, should have reached a temperature below 550°C. An approximation of 500°C for the region around the isograd represented by the reaction

may be reasonable.

The maximum temperature estimate for the French Creek section would correspond with the sillimanite-garnet-biotite zone in the northwestern part of the section. Carmichael's (1978) grid (Fig. 2-22) suggests temperatures from about 550°C to approximately 650°C prevail at these conditions. The northwestern margin of this zone is fault bounded; however, the presence of staurolite-muscovite-quartz-garnet-biotite-sillimanite assemblages near the headwaters of Norman Wood Creek (Sections 2-2-7 and 2-3-5) implies that a metamorphic culmination has been reached and lower grade rocks could have existed to the northwest prior to faulting. Therefore, maximum temperatures for the region are probably close to those of the staurolite breakdown reaction

A temperature maximum for the French Creek section of about 575°C in the sillimanite-garnet-biotite zone is proposed on this basis. Postmetamorphic faulting has made direct assessment of this possibility impossible.

The Windy Range section contains assemblages varying from the staurolite-muscovite-quartz-kyanite-garnet-biotite to the potassic feldspar - sillimanite zones. Since the kyanite-garnet-biotite zone separates the sillimanite-garnet-biotite and the staurolite-muscovite-quartz-kyanite-garnet-biotite zones, pressures above 5 kbar are anticipated. Likewise, since the invariant point assemblage muscovite-plagioclase-quartz-potassic feldspar-sillimanite-kyanite-liquid, representative of the intersection of the potassic feldspar - sillimanite forming reaction (Rx. 2-6) and the kyanite-sillimanite

equilibrium (Rx. 2-3), is not observed, pressures below 7.3 kbar are also expected.

Temperature estimates could vary from about 550°C within the staurolite-muscovite-quartz-kyanite-garnet-biotite zone near the Rocky Mountain Trench to approximately 670°C in the potassic feldspar - sillimanite zone. Since muscovite is still present in potassic feldspar - sillimanite parageneses, temperature estimates much above 670°C are not anticipated.

(2-6) CONCLUSIONS

The initial descriptive part of this chapter presents assemblage and textural data currently available for pelitic metamorphic zones in the northern Selkirk Mountains. Where possible, these data are used to propose mineral reactions for the isograds separating mineral zones.

The northern Selkirk Mountains are divided into two sequences of metamorphic zones. The southeastern French Creek section contains the following metamorphic zones: chlorite; chlorite-biotite; garnet-chlorite; staurolite-biotite; staurolite-kyanite-biotite; sillimanite-garnet-biotite; and staurolite-muscovite-quartz-sillimanite-garnet-biotite. The reactions:

are considered to represent the isograds separating some of these zones. Variations in bulk composition could account for the boundaries between the chlorite and chlorite-biotite zones, and chlorite-biotite and garnet-chlorite zones.

The northwestern part of the study area, named the Windy Range section, is divided into the following metamorphic zones: staurolite-muscovite-quartz-kyanite-garnet-biotite; kyanite-garnet-biotite; sillimanite-garnet-biotite; and potassic feldspar - sillimanite. The reactions:

are considered to represent the isograds separating the above zones.

An understanding of mineral reactions and textures enables sequences of mineral growth to be established. For example, the change in garnet habit from embayed poikiloblasts in the staurolite-kyanite-biotite zone of the French Creek section to a euhedral relatively inclusion free habit in the sillimanite-garnet-biotite zone can be related to Rx. 2-4 (Sections 2-2-6 and 2-3-4).

Episodic growth probably occurred for other phases in this study area. Attempts to relate the timing of metamorphism to deformational events using inclusion trails and mineral habits should recognize the possibility of different periods of mineral growth and should attempt to establish textural criteria for recognition of

different generations of minerals.

The application of information on metamorphic zones and isograds represented by mineral reactions to Carmichael's, (1978) petrogenetic grid enables semiquantitative estimates of pressures and temperatures of metamorphism to be made. Two different prograde sequences are recognized within the northern Selkirk Mountains. The southwestern French Creek section is believed to have formed at temperatures ranging from less than 500°C (garnet-chlorite zone) to more than 550°C (sillimanite-garnet-biotite zone) and at pressures of approximately 5 kbar. The northeastern Windy Range section is thought to have formed at temperatures between 550°C (staurolite-muscovite-quartz-kyanite-garnet-biotite zone) and 670°C (potassic feldspar - sillimanite zone) and at pressures between about 5 and 7 kbar. A northwest trending fault separates the two regions.

It is possible to relate changes in mineral parageneses in each of the two sections in the study area to an orderly sequence of mineral zones and accompanying isograds which are indicative of Barrovian type metamorphism. Isograds separating some zones in the northern Selkirk Mountains are sharply defined (i.e. kyanite-sillimanite isograd in Windy Range section) whereas some metamorphic zones (i.e. staurolite-muscovite-quartz-sillimanite(or kyanite)-garnet-biotite) contain parageneses indicative of product and reactant assemblages for a particular reaction (i.e. Rx. 2-4 and 2-5). Further work is necessary to evaluate the significance of these product-reactant assemblages. Nevertheless, it is reasonable to argue that the sequence of mineral parageneses and accompanying isograds indicates that equilibrium conditions were approached during metamorphism.

2

CHAPTER 3.

GEOOTHERMOMETRY AND GEOBAROMETRY

(3-1) INTRODUCTION

Recent advances in experimental studies on mineral assemblages and analytical techniques have enabled calibration of reactions with respect to various intensive parameters. This information can be used to make quantitative assessments of conditions during metamorphism for similar, naturally occurring assemblages. Among the most important of these parameters are those of interest in this study: temperature and pressure.

To ensure confidence in the values obtained through geothermometry and geobarometry, replication of results by independent methods is desirable. Sufficient experimental and field studies have been performed to enable construction of petrogenetic grids which allow comparison of calculated values with values imposed by phase relations.

The geothermometers used in this study involve the mineral pairs garnet-biotite and calcite-dolomite. Pressures have been calculated by the garnet-plagioclase-aluminum silicate-quartz method, using temperatures calculated with the above mentioned geothermometers. The theoretical background, application of data, and results for each method will be discussed separately.

Before accepting the values obtained by these methods one must appreciate some of the factors that can limit the accuracy of the results. First, the accuracy and precision of the initial experimental work upon which the technique is established and the degree to which the experimental system approached equilibrium introduce a component of error. In addition, the accuracy

and precision of the analyses of the natural assemblages, the degree to which equilibrium was approached in the natural system, the effect of other intensive parameters, and the system's response to waning metamorphism may affect the reliability of data applied to any geothermometer or geobarometer.

(3-2) GARNET-BIOTITE GEOTHERMOMETRY

(3-2-1) Theoretical Background

Garnet and biotite coexist at temperatures and pressures that correspond to a wide range in metamorphic grade. Since both garnet and biotite belong to solid solution series, attempts have been made to calibrate a geothermometer applicable to this range of metamorphic conditions based on element partitioning of Fe^{2+} and Mg^{2+} between almandine ($\text{Fe}_3\text{Al}_2\text{Si}_3\text{O}_{12}$) - pyrope ($\text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12}$) garnet, and annite ($\text{KFe}_3\text{AlSi}_3\text{O}_{10}(\text{OH})_2$) - phlogopite ($\text{KMg}_3\text{AlSi}_3\text{O}_{10}(\text{OH})_2$) mica.

Early garnet-biotite geothermometers (e.g. Perchuk 1967; Saxena 1969, among others) did not gain acceptance and it was not until an empirical calibration by A.B. Thompson (1976) and a subsequent experimental calibration by Ferry and Spear (1977, 1978) were made that this method became popular. These recent calibrations are used in this study.

Thompson's empirical calibration is based on a linear least squares analysis of distribution coefficients (K_D) from naturally occurring garnet-biotite assemblages and corresponding temperature

estimates. His distribution coefficient is defined as

$$K_{DT} = (\text{Fe}_{\text{gar}} \cdot \text{Mg}_{\text{bio}}) / (\text{Mg}_{\text{gar}} \cdot \text{Fe}_{\text{bio}})$$

The resulting plot is shown as the dashed line on Fig. 3-1 and is described by the equation:

$$\ln K_{DT} = \frac{2740}{T (^{\circ}\text{K})} - 1.56 \quad (\text{Eq. 3-1})$$

A.B. Thompson (1976, p. 430) suggests that the exchange of cations between garnet and biotite is "virtually independent of pressure... and independent of $\text{a}_{\text{H}_2\text{O}}$ ".

Attempts to determine K_D values experimentally have been hampered by the sluggish reequilibration rates of garnet. Ferry and Spear (1977, 1978) have developed a technique that circumvents this problem and have therefore enabled experimental calibration of $\ln K_D$ vs $1/T$. They define K_D as:

$$K_{DFS} = (\text{Mg/Fe})_{\text{gar}} / (\text{Mg/Fe})_{\text{bio}}$$

and the least squares linear analysis to their data is described by the equation (Fig. 3-1):

$$\ln K_{DFS} = \frac{2109}{T (^{\circ}\text{K})} + 0.782 \quad (\text{Eq. 3-2})$$

Experiments used for the Ferry and Spear calibration were performed at 2.07 kbar. Ferry and Spear (1978) use these data to

Fig. 3-1. Plot of $\ln K_{DFS}$ vs $1/T (^{\circ}K)$ for calibrations of Fe-Mg partitioning between garnet and biotite as proposed by Ferry and Spear (1978) (solid line), Thompson (1976) (dashed line), and Goldman and Albee (1977) (dotted line). (After Ferry and Spear (1978)).

derive the following polybaric, polythermal expression for Fe - Mg partitioning between garnet and biotite;

$$12454 - 4.662T (^{\circ}\text{K}) + 0.057P (\text{bars}) + 3RT \ln K = 0 \quad (\text{Eq. 3-3})$$

Experimental work by Ferry and Spear (1978) demonstrates that over the composition interval $0.80 \leq \text{Fe}/(\text{Fe} + \text{Mg}) \leq 1.00$ iron and magnesium mix ideally in garnet and biotite solid solutions. Mueller (1972) argues that the mixing properties of Fe and Mg in biotite do not deviate greatly from ideal solution.

From Fig. 3-1 it is readily apparent that there is good correlation between Thompson's and Ferry and Spear's calibrations in the range 500°C to 600°C . Goldman and Albee (1977), in a comparison of Fe-Mg partitioning with $^{18}\text{O}/^{16}\text{O}$ isotope data for quartz-magnetite, position the best fit of $\ln K_D$ vs $1/T$ about 50°C lower than either of the above calibrations (see dotted line on Fig. 3-1).

One of the problems in applying data from natural systems to the above calibration results from the effects of element substitution (besides Mg and Fe) in biotite and garnet. A.B. Thompson (1976, p. 430) states that K_D shows "systematic displacements with increasing Ti in biotite" whereas Ferry and Spear (1978, p. 116) state that " K is also a function of Ca and Mn content in the garnet and the Ti and Al^6 content of the biotite". Also, it is not possible to differentiate between ferric and ferrous iron with the electron microprobe and hence all analyzed iron is assigned to the Fe^{2+} state. Greater proportions of ferric iron in biotites compared with

coexisting garnet will increase the value of $\ln K_D$ and so lower the temperature determined by Eq. 3-1.

Dallmeyer (1974) has examined the effects of Ti, Al, and Fe^{3+} substitution in biotite and Ca and Mn substitution in garnet. Octahedral substitution of the above elements in biotite will increase the positive charge on the octahedral layer. This can be accommodated by either increased tetrahedral substitution of Al for Si, or by having unoccupied octahedral positions. Significant Ti substitution causes a preference for substitution of Fe^{2+} over Mg^{2+} in biotite. The net effect of this is to decrease K_{DT} and therefore increase temperature estimates obtained by the garnet-biotite geothermometer. Pronounced Ca and Mn substitution in garnet will favour Fe^{2+} substitution over Mg^{2+} within garnet. This will result in increased K_{DT} values and hence lower garnet-biotite geothermometry estimates.

Ferry and Spear (1978) propose that garnet-biotite geothermometry is reliable up to the following limits for ion substitution within garnet and biotite respectively:

$$(\text{Ca} + \text{Mn}) / (\text{Ca} + \text{Mn} + \text{Fe} + \text{Mg}) \approx 0.20$$

$$(\text{Al}^6 + \text{Ti}) / (\text{Al}^6 + \text{Ti} + \text{Fe} + \text{Mg}) \approx 0.15$$

Temperatures determined by Thompson's method have an estimated error of $\pm 50^\circ\text{C}$. This originates from the error bars associated with A.B. Thompson's data (1976, fig. 1). Publications using Thompson's technique commonly quote errors of $\pm 50^\circ\text{C}$ (e.g. Tracy 1978). Ferry and Spear calculate an error for their method of $\pm 50^\circ\text{C}$ by propagating an analytical error of 0.01% annite, 0.01% phlogopite.

0.01 X almandine, and 0.01 X pyrope through Eq. 3-2.

Garnet-biotite geothermometry is gaining increased acceptance by petrologists, and publications employing this technique are becoming common. Ghent et al. (1978, 1979) and Craw (1977, 1978) have used this method for rocks from both sides of the Rocky Mountain Trench, adjacent to this study area.

(3-2-2) Chemical Compositions of Garnets and Biotites in the Selkirk Mountains

Criteria for grain selection, microprobe operating conditions, and resulting chemical analyses for garnet-biotite assemblages used for geothermometry are listed in Appendix B. As described therein, each tabulated analysis expresses both the average of several good analyses within a thin section and the variation, σ , from this average. A perusal of these chemical analyses will confirm that both phases exhibit reasonably consistent chemistry throughout a sample. Biotite within the matrix is similar in composition to biotite adjacent to garnet. These consistencies support the argument that biotite and garnet approached chemical equilibrium.

For each sample the mole fraction of (Ca + Mn) in garnet and (Ti + Al⁶) in biotite are listed in Table 3-1. Most garnet analyses are well within the limit of 0.20 prescribed by Ferry and Spear (1978). The only analyses near or above this limit are from certain garnet cores. Substitution in biotite fluctuates around the limit of 0.15 suggested by Ferry and Spear (1978). The amount of Al⁶ has been calculated by assuming that (Si + Al⁴) totals to 8.000.

Excess aluminum is assigned to the octahedral site. Since Dallmeyer (1974) predicts increased substitution of Al for Si it is likely that values listed in Table 3-1 are consistently high. Therefore, biotite compositions probably also fall within the substitution limit suggested by Ferry and Spear (1978).

Garnet rim compositions average 77% almandine and 13% pyrope. These values are close to the range over which Ferry and Spear (1978) have demonstrated ideal mixing between Mg and Fe in biotite and garnet solid solutions.

It can therefore be argued that these samples approached chemical equilibrium, that substitution of Ca, Mn, Ti, and Al are within correction limits, and that compositions of these phases are comparable with systems exhibiting ideal mixing. Therefore, application of chemical data from Appendix B to garnet-biotite geothermometry can be made with some confidence.

Garnet can show pronounced chemical zonation and so rims and cores of this phase from twelve samples were analyzed (Appendix B). Spot analyses along traverse lines across garnets from two samples were used to plot chemical zonation profiles for the garnets (Appendix C). Even though these zoning profiles vary, some generalizations can be made. Iron and manganese are strongly and consistently zoned and as iron content increases manganese content decreases. Magnesium is zoned only in one sample and, like calcium, usually shows little variation in content across the grain. One garnet (A345-7) is homogeneous throughout.

Zoning patterns may reflect variations in bulk composition of the system and(or) changes in metamorphic conditions during

garnet growth. Tracy et al. (1976), in a study applying Thompson's calibration of the garnet-biotite geothermometer to zoned garnets, discussed the following problems involved in interpreting estimated temperatures using zoned garnets: modelling of zoning profiles (e.g. Rayleigh fractionation (Hollister 1966)) is successful only for garnets that grew during one continuous reaction; and zoning profiles can be complicated by factors such as retrograde reaction, homogeneous diffusion within garnet, chemical exchange with Fe-Mg inclusions, and reactions involving melt phases at high temperatures.

Therefore, garnet core compositions need not be in equilibrium with biotite in the matrix. Tracy et al. (1976) found garnet-biotite temperatures obtained using garnet core compositions were lower than temperatures obtained using rim compositions. This was interpreted to reflect increasing temperatures during prograde metamorphism. Ghent et al. (1979) found that core temperatures were greater than rim temperatures for samples from the Monashee Mountains, B.C.

Rims and cores of biotite grains do not show either regular or significant variations in chemical composition. It therefore appears that entire biotite grains reequilibrated in response to changing metamorphic conditions.

(3-2-3) Temperature Determinations

Temperatures calculated by garnet-biotite geothermometry using Eq. 3-1, 3-2, and 3-3 are listed in Table 3-1. The

Table 3-1. Temperature estimates using garnet-biotite geothermometry

Station	K_D (Thompson)	$T^{\circ}C$ (Thompson)	K_D (Ferry+Spear)	$T^{\circ}C$ (F+S 1978)	$T^{\circ}C$ (F+S 1978 5 kbar)	A^1	B^2	T_{grid}
A345 -R	4.99	592	0.200	609	620	0.10	0.16	550?
A345 -C	4.31	633	0.232	667	679	0.08	0.16	550?
F061 -R	6.45	527	0.155	524	534	0.11	0.20	550
F061 -C	5.82	552	0.172	556	567	0.15	0.20	550
P57 -R	4.71	608	0.212	631	643	0.05	0.17	625
P57 -C	4.36	630	0.230	663	676	0.08	0.17	625
P87b -R	4.37	629	0.229	662	674	0.08	0.17	625
P87b -C	3.88	666	0.258	713	724	0.08	0.17	625
M63 -R	3.92	662	0.255	709	722	0.06	0.19	625
M63 -C	3.82	671	0.262	721	734	0.07	0.19	625
A300 -R	5.12	585	0.195	600	611	0.08	0.20	650
A300 -C	4.18	642	0.239	680	692	0.09	0.20	650
A350 -R	4.72	607	0.212	631	643	0.08	0.16	550?
A350 -C	7.46	495	0.134	482	492	0.14	0.16	550?
A93a -R	7.40	497	0.135	485	494	0.13	0.14	500
A93a -C	10.12	435	0.099	408	417	0.19	0.15	500

¹ $A = (Ca + Mn)/(Ca + Mn + Fe + Mg)$ in garnet

² $B = (Al^6 + Ti)/(Al^6 + Ti + Fe + Mg)$ in biotite

Table 3-1 (concluded)

Station	K_D (Thompson)	$T^{\circ}C$ (Thompson)	K_D (Ferry+Spear)	$T^{\circ}C$ (F+S 1978)	$T^{\circ}C$ (F+S 1978 5 kbar)	A^1	B^2	T_{grid}
A352 -R	5.17	582	0.194	597	609	0.13	0.20	550?
A352 -C	12.69	396	0.079	362	371	0.21	0.19	550?
A238a-R	6.47	527	0.155	523	534	0.14	0.16	550
A238a-C	6.27	534	0.160	532	544	0.20	0.16	550
A87a -R	7.61	491	0.132	477	488	0.15	0.18	500
A87a -C	12.06	404	0.083	371	380	0.27	0.18	500
A375 -R	4.11	648	0.243	687	700	0.05	0.19	550?
A375 -C	4.43	625	0.226	656	669	0.09	0.18	550?
P95b -R	5.18	582	0.193	596	607	0.09	0.17	625
A278 -R	6.01	544	0.166	546	556	0.10	0.17	550
A284b-R	4.11	648	0.243	687	700	0.10	0.19	670
P147 -R	4.22	640	0.237	676	687	0.06	0.17	550?
M144 -R	5.78	554	0.173	558	567	0.12	0.18	550?
M154c-R	5.26	578	0.190	590	602	0.10	0.20	670
M170a-R	4.39	628	0.228	659	672	0.12	0.20	670
M158b-R	5.62	561	0.178	568	579	0.16	0.19	670

¹ $A = (Ca + Mn) / (Ca + Mn + Fe + Mg)$ in garnet

² $B = (Al^6 + Ti) / (Al^6 + Ti + Fe + Mg)$ in biotite

right hand column of this table lists temperatures estimated for the vicinity from which each sample was taken using the grid of Carmichael (1978) (Section 2-5-2). Sample locations are shown on Fig. 3-2.

As predicted from Fig. 3-1, there is good correlation between the Thompson, and Ferry and Spear calibrations. Although the variation in temperature determinations for all samples is within the $\pm 50^\circ\text{C}$ error bar, the samples within the temperature range 490°C to about 600°C show excellent correlation between methods. The difference in temperatures calculated using the Ferry and Spear calibration developed from experiments at 2.07 kbar (Eq. 3-2) and pressure corrected to 5 kbar (considered a reasonable approximation of pressure of metamorphism for the northern Selkirk Mountains) using Eq. 3-3 is about 10°C .

Rim and core compositions from the same garnet usually calculate to significantly different temperatures. Although the variation is within the $\pm 50^\circ\text{C}$ error bar, temperatures calculated using core compositions do not compare as favourably as those for rim compositions when compared with temperatures anticipated from phase relations (Section 2-5-2). Since garnet rims are more likely to be in chemical equilibrium with biotite than garnet cores it is proposed that only rim compositions be used to calculate temperatures.

The correlation between temperatures calculated using Thompson's 1976 calibration of the garnet-biotite geothermometer and estimates using Carmichael's grid is generally good. Similar results were obtained for samples taken from a fairly restricted area (i.e. A350, A352, A345;

Fig. 3-2. Locations of samples analyzed for garnet-biotite or calcite-dolomite geothermometry, and garnet-plagioclase- Al_2SiO_5 -quartz geobarometry.

F061, A278, A238, A87, A93, P57, P87b) (Fig. 3-2). Discrepancies do exist between grid estimates and garnet-biotite temperatures. These discrepancies are discussed individually below.

Sample M63

Station M63 is located near the kyanite-sillimanite isograd at the head of Micà Creek. The calculated temperature (662°C) corresponds with conditions at or near potassic feldspar - sillimanite grade and yet the nearest assemblages indicative of these conditions are 5 km to the southwest. Although the anticipated temperature is within the error limits allowed by Thompson's method, this value is not considered realistic according to phase relations and is therefore omitted from further discussions. The anomalously high temperature can be explained by high titanium content in biotites from this sample.

Samples A300, M154c, and M158b

Temperatures obtained using garnet rim compositions for samples A300, M154c, M158b, M170a, and A284b (all from the potassic feldspar - sillimanite zone) are 585°C , 578°C , 561°C , 630°C , and 648°C respectively. The first three values are well below the 670°C value estimated for this metamorphic zone using Carmichael's (1978) grid. These values are also lower than garnet-biotite temperatures calculated for sillimanite-garnet-biotite and kyanite-garnet-biotite assemblages to the northeast. The only temperature obtained using garnet core compositions is 642°C (Sample A300).

Tracy (1978) used A.B. Thompson's (1976) calibration of the garnet-biotite geothermometer for potassic feldspar - sillimanite - muscovite bearing rocks from Massachusetts. Temperatures calculated using garnet rim compositions vary from 530°C to 625°C and temperatures calculated using garnet core compositions vary from 580°C to 740°C. Temperature estimates using core compositions are generally 50 to 100 °C higher than those made using rim compositions. Therefore, although three of the five temperature estimates from the potassic feldspar - sillimanite - muscovite-quartz zone of the northern Selkirk Mountains are lower than estimates made using Carmichael's grid, they are comparable to values obtained in Tracy's work. Although an adequate explanation for the difference between grid and garnet-biotite temperature estimates is not available, the garnet-biotite temperatures may represent reequilibration during waning metamorphism.

Samples P147 and A375

Stations A375, A350, A345, A352, M144, and P147 are within or near the staurolite-out isograd corresponding to Rx. 2-5, adjacent to the Rocky Mountain Trench. Carmichael (1978) positions the breakdown of staurolite by this reaction at about 550°C for pressures greater than 5 kbar, whereas A.B. Thompson (1976, fig. 6) positions this reaction at 685°C at $P_{H_2O} = 5$ kbar on the basis of experimental data from Richardson (1968) (Section 2-5-1).

Garnet-biotite temperature estimates for samples from the staurolite-muscovite-quartz-kyanite-garnet-biotite zone near the Rocky Mountain Trench do not conclusively support either of

the above mentioned reaction positions. Three of six samples give rim temperatures of about 600°C , two are 640°C to 680°C , and one value is about 550°C .

Craw (1978) calculated a temperature of 590°C using Thompson's garnet-biotite calibration for a sample from the same grade within the Monashee Mountains, adjacent to this study area. Ghent et al. (1979), using Thompson's garnet-biotite calibration corrected to 5 kbar, estimate temperatures of 615°C to 620°C for staurolite-kyanite assemblages from the Monashee Mountains. Tracy et al. (1976) also calculated maximum temperatures (using Thompson's garnet-biotite calibration) of 580°C to 605°C for kyanite-muscovite-staurolite-garnet-biotite-quartz bearing rocks in central Massachusetts. Three samples from the northern Selkirk Mountains (A345, A350, and A352) have garnet-biotite compositions which calculate to 592°C , 607°C , and 582°C respectively. These values are comparable with garnet-biotite estimates both close to (e.g. Craw 1978) and removed from the study area (e.g. Tracy et al. 1976).

Samples P147 and A375 from the staurolite-muscovite-quartz-kyanite-garnet-biotite zone near the Rocky Mountain Trench have garnet-biotite compositions that calculate to 640°C and 648°C respectively. These values are about 100°C higher than conditions estimated for this grade from Carmichael's grid and approximately 50°C higher than estimates obtained from samples in the same metamorphic zone (A350, A352, and A345). Ghent et al. (1979) report one sample (GM 74-6) from rocks of the same grade within the Monashee Mountains that exhibits two compositional domains characterized by differing biotite compositions. Garnet-biotite

temperatures (corrected to 5 kbar and calculated using spot analyses from juxtaposing garnet-biotite pairs) were estimated as 640°C and 760°C. Ghent et al. (1979) suggest that the lower temperature, which is comparable with nearby estimates of 620°C, is more reasonable and that the higher value may have resulted from disequilibrium fractionation.

Temperature estimates using juxtaposing garnet-biotite pairs individually rather than average compositions (see Appendix B) for samples P147 and A375 were calculated (see Table 3-2). Temperatures for individual pairs were, within error, similar to the estimates calculated using average compositions. Chemical domains are not recognized in either sample. High temperature estimates for samples from the staurolite-muscovite-quartz-kyanite-garnet-biotite zone are not unique to the northern Selkirk Mountains. Temperature estimates for samples P147 and A375 are omitted from further discussions because they are considered to be anomalously high when compared with other nearby values, and with estimates from other studies. As yet, no conclusive explanation for the discrepancy is available.

Temperatures estimated by Eq. 3-1, 3-2, and 3-3 are sufficiently similar that, within error, the results are identical. Thompson's calibration has been preferred over Ferry and Spear's calibration in recent discussions (e.g. Crawford 1978; Ghent et al 1978; Ghent et al 1979; Tracy et al. 1976; Tracy 1978; Craw 1978; Knitter 1979; and Robinson et al. 1978). Temperature estimates quoted in further discussions in this thesis will be those values obtained by Thompson's calibration. Temperatures are rounded to the nearest tens digit.

Table 3-2. Temperature estimates using spot analyses from
biotite and garnet grains in contact with each other

Sample A375 (rims)	x_{Fe}^{bi}	x_{Mg}^{bi}	x_{Fe}^{ga}	x_{Mg}^{ga}	K_D	$\ln K_D$	$T^{\circ}C$ (Thompson 1976)
2-1-1	2.359	2.323	4.512	1.062	4.184	1.43	643
3-1-1	2.360	2.325	4.542	1.082	4.136	1.42	646
4-1-1	2.486	2.206	4.594	0.998	4.085	1.41	650
5-1-1	2.469	2.174	4.556	1.024	3.918	1.37	663
6-1-1	2.369	2.358	4.644	1.068	4.328	1.47	633
7-1-1	2.415	2.338	4.576	1.114	3.977	1.38	660
Sample P147 (rims)							
2-1-1	2.563	2.650	4.443	1.121	4.098	1.41	649
4-1-1	2.545	2.749	4.429	1.139	4.200	1.44	640
6-1-1	2.536	2.717	4.420	1.165	4.065	1.40	653
8-1-1	2.484	2.759	4.397	1.132	4.314	1.46	634
9-1-1	2.533	2.696	4.486	1.098	4.349	1.47	631

(3-3) CALCITE-DOLOMITE SOLVUS GEOTHERMOMETRY

(3-3-1) Theoretical Background

The calcite-dolomite solvus geothermometer is based on the magnesium content in calcite found coexisting with dolomite.

Work by Graf and Goldsmith (1955, 1958), Harker and Tuttle (1955), Rosenberg (1963, 1967), and Goldsmith and Newton (1969) demonstrated that the magnesium content in calcite is temperature dependent but relatively insensitive to pressure. Also, the element partitioning is independent of the activity of volatile components. These two qualities are desirable for any reliable geothermometer. A least squares analysis of solvus data (e.g. Sheppard and Schwarcz 1970) can be used to calculate temperatures of formation for naturally occurring calcite-dolomite assemblages. This method has been successfully applied to metamorphic terrains and is gaining increasing acceptance. O'Neil and Ghent (1975) have used this method in the Esplanade Range, about 50 km southeast of this study area.

To say that this geothermometer is totally independent of pressure is not true. Goldsmith and Newton (1969) suggest a mole percent MgCO_3 increase in calcite of 0.12% per kilobar in the 500°C to 800°C range. Ewart (1977), in a treatment of this problem, recalculated solvus data from experimental studies to standard pressure states of 1 bar and 5 kbar. He then calculated the following polynomial equations for the new solvus curves:

$$T (^{\circ}\text{K}) = 477.63 + 88.77X - 6.08X^2 + 0.22X^3 - 0.0032 X^4 \quad (1 \text{ bar})$$

(Eq. 3-4)

$$T (^{\circ}\text{K}) = 474.96 + 82.24X - 5.11X^2 + 0.17X^3 - 0.0032X^4 \quad (5 \text{ kbar})$$

(Eq. 3-5)

where "X" is the mole percent MgCO_3 .

The value "X" is commonly taken from the grain in the sample with the greatest mole percent MgCO_3 (e.g. Hutcheon and Moore 1973; Ewart 1977). As Ewart states (1977, p. 63): "The maximum amount of magnesium possible in calcite is established by the highest temperature reached during metamorphism. While numerous factors ... can reduce this maximum value, none except increasing temperature can increase it."

Substitution of cations such as Fe^{2+} and Mn^{2+} in the Ca and Mg sites may affect the solvus temperature. Sheppard and Schwartz (1970) estimate that the effect of 1 mole percent FeCO_3 will be to increase the calculated temperature by 8°C . Iron rich calcites can be detected by simple staining techniques (potassium ferricyanide in an alizarin red-S solution) and so spurious results can be avoided.

(3-3-2) Collection and Interpretation of Data

Only three localities (F74d, V98a, M64d) with coexisting calcite and dolomite have been found in the study area (Fig. 3-2).

In all samples, exsolution of dolomite lamellae within stained

calcite were recognized. Across any grain featuring exsolution lamellae one could anticipate high MgCO_3 values at the lamellae, low values in diffuse zones of magnesium depletion immediately adjacent to lamellae, and intermediate values in zones lacking lamellae. Individual spot electron microprobe analyses of any grain could give a wide range of temperatures depending on the location of the analysis. Exsolution also makes accurate determination of MgCO_3 in the whole grain difficult and so it is necessary to homogenize the lamellae back into the matrix. This can be achieved statistically by averaging numerous electron microprobe analyses on an individual grain. (See Appendix B for a description of the analytical technique.) The use of a defocused electron beam aided homogenization on a local scale by increasing the surface area coverage of points analysed. The calcite grain with the maximum MgCO_3 average content was then used in Eq. 3-5 to calculate a temperature value.

The amount of Fe^{2+} and Mn^{2+} in the grains used for temperature determinations was either not detectable or of low concentration. The greatest value was 0.87 mole percent FeCO_3 for sample M64d. The error that this introduces is approximately 7°C .

(3-3-3) Temperature Determinations

Temperatures calculated by calcite-dolomite solvus geothermometry for the Selkirk samples are presented in Table 3-3 and located on Figs. 3-2 and 3-3. Comparison of these values with estimates from garnet-biotite geothermometry and phase relations is provided in Section 3-5.

Fig. 3-3. Temperature and pressure estimates considered to be representative of peak metamorphic conditions in the northern Selkirk Mountains.

Table 3-3. Temperature estimates using calcite-dolomite solvus geothermometry

Sample Number	Mole percent MgCO_3	$T^{\circ}\text{C}$
F74d	6.5	563 ⁰⁰ or 560 ⁰
V98a	5.8	538 ⁰ or 540 ⁰
M64d	7.7	602 ⁰ or 600 ⁰

(3-4) GARNET-PLAGIOCLASE GEOBAROMETRY

(3-4-1) Theoretical Background

The assemblage garnet-plagioclase- Al_2SiO_5 polymorph-quartz is common within pelites metamorphosed to conditions corresponding to amphibolite facies or higher. Since both garnet and plagioclase are solid solution phases whose composition is expressed partially in terms of a calcium end member it is possible to relate the phases in the above assemblage by the reaction:

(Rx. 3-1)

This reaction has been studied experimentally by Boyd (1964), Newton (1966), Hays (1967), Hariya and Kennedy (1968) and Perkins et al. (1977), and data from some of these studies have been used to calibrate the above naturally occurring assemblage as either

a geothermometer or geobarometer (Ghent 1975, 1976, 1977).

Temperature estimates from garnet-biotite and calcite-dolomite

solvus geothermometry allow this technique to be used as a geobarometer

in this study. Since Rx. 3-1 does not involve volatiles it is independent of fluid composition and will reflect P_{total} rather than P_{fluid} .

Ghent (1977) has provided a detailed derivation of the theoretical background for this method. A review of this derivation is presented below.

Each phase of Rx. 3-1 obeys the relation:

$$\mu_i^\alpha = \mu_i^0 + RT \ln a_i^\alpha$$

where μ_i^α = chemical potential of component i in phase α

μ_i^0 = chemical potential of pure i at P, T of interest.

a_i^α = activity of component i in phase α

Activity terms ($\ln a_i^\alpha$) are zero if the phases are pure (since $x_i^\alpha = 1 = a_i^\alpha$, $\mu_i^\alpha = \mu_i^0$). Therefore at equilibrium, and when pure phases are studied, Rx. 3-1 can be expressed as

$$\Delta \mu = \mu_{\text{grossular}} + 2\mu_{\text{Al}_2\text{SiO}_5} + \mu_{\text{quartz}} - 3\mu_{\text{anorthite}} = 0$$

At equilibrium Rx. 3-1 (pure phases) can be described by:

$$\Delta G_T^P = 0 = \Delta H_T^0 - T\Delta S_T^0 + (P - 1)(\Delta V_S) \quad (\text{Eq. 3-6})$$

where ΔG_T^P = free energy change at P and T of interest

ΔH_T^O = enthalpy change for pure phases at T and P = 1 bar

ΔS_T^O = entropy change for the pure phases at T and P = 1 bar

ΔV_s = volume change of the pure solids at P and T of interest

An equilibrium constant (K) for Eq. 3-6 is:

$$\log K = 0 = \frac{\Delta H_T^O}{2.303 RT} - \frac{\Delta S_T^O}{2.303 R} + \frac{(P - 1)(\Delta V_s)}{2.303 RT} \quad (\text{Eq. 3-7})$$

If ΔH_T^O , ΔS_T^O , and ΔV_s approximate constants over limited ranges of P and T (Anderson 1977), then Eq. 3-7 can be written:

$$0 = \frac{A}{T} + \frac{B}{T} + \frac{C(P - 1)}{T} \quad (\text{Eq. 3-8})$$

where A is related to H_T^O (enthalpy change of reaction of pure phases at P = 1 bar), B is related to S_T^O (entropy change of reaction of pure phases at P = 1 bar and T of interest), and C is related to the volume change of reaction of the pure solids.

Constants A and B can be derived from either thermochemical data available for individual phases, or from experimental data available for the reaction. Ghent (1977) has solved constants A and B using the experimental data of Hays (1967) and Hariya and Kennedy (1968) and solved for C using data in Robie and Waldbaum (1968) to obtain the following equations for Rx. 3-1 (pure phases) using kyanite and sillimanite polymorphs respectively:

$$0 = \frac{-3272}{T} + 8.3969 - \frac{0.3448 (P - 1)}{T} \quad (\text{Eq. 3-9})$$

$$0 = \frac{-2551.4}{T} + 7.1711 - \frac{0.2842 (P - 1)}{T} \quad (\text{Eq. 3-10})$$

Data for the aluminum silicate polymorph transition are from Holdaway (1971).

The reliability of these equations is dependent on the validity of the experimental data. However, since recent work by Perkins et al. (1977) is consistent with data from Hariya and Kennedy (1968), increased confidence in the experimental data is reasonable.

Naturally occurring garnet and plagioclase are solid solution phases and so Eq. 3-9 and 3-10 must be modified (using $\mu_i^{\wedge} = \mu_i^0 + RT \ln a_i^{\wedge}$) to account for the activity of anorthite and grossular components as follows:

$$0 = - \frac{3272}{T} + 8.3969 - \frac{0.3448 (P - 1)}{T} + \log a_{\text{grossular}}^{\text{garnet}} - 3 \log a_{\text{anorthite}}^{\text{plagioclase}} \quad (\text{Eq. 3-11})$$

and

$$0 = - \frac{2551.4}{T} + 7.1711 - \frac{0.2842 (P - 1)}{T} + \log a_{\text{grossular}}^{\text{garnet}} - 3 \log a_{\text{anorthite}}^{\text{plagioclase}} \quad (\text{Eq. 3-12})$$

If these solid solutions are treated as exhibiting ideal mixing behaviour the $a_{\text{gr}}^{\text{ga}}$ (the activity of grossular in garnet) can be approximated to $(X_{\text{gr}}^{\text{ga}})^3$ (where X is the mole fraction) by mixing Ca randomly on the three eight-fold coordination sites, and

a_{an}^{pl} (the activity of anorthite in plagioclase) can be approximated to (X_{an}^{pl}) if a simple coupled substitution of CaAl for NaSi is accepted. A distribution coefficient (K_D) for Rx. 3-1 can be defined as:

$$K_D = (X_{gr}^{ga})^3 / (X_{an}^{pl})^3$$

Substitution of K_D into Eq. 3-11 and 3-12 yields

$$0 = \frac{-3272}{T} + 8.3969 - \frac{0.3448 (P - 1)}{T} + \log K_D \quad (\text{Eq. 3-13})$$

$$0 = \frac{-2551.4}{T} + 7.1711 - \frac{0.2842 (P - 1)}{T} + \log K_D \quad (\text{Eq. 3-14})$$

In early treatments of this technique, Ghent (1975, 1976) incorporates activity coefficients for anorthite in plagioclase (from Orville (1972)) and for grossular in garnet (from Ganguly and Kennedy (1974)) to account for non-ideal mixing. However, even existing activity data are subject to multiple interpretation (i.e. Kerrick and Darken 1975) and so in the treatment described above Ghent (1977, p. 107) suggests that Eq. 3-13 and 3-14 "can be used as indicators of relative pressure at a given temperature ... Once the effects of non-ideality have been precisely evaluated more quantitative estimates of P can be made...".

Ghent et al. (1979) propose that in the absence of adequate experimental activity data, it may be possible to adjust pressure estimates by using an empirical activity coefficient product (K_γ) that can be added to Eq. 3-13 and 3-14. Ghent et al. (1979)

estimate that an average $\log K_y$ value of -0.4 will adjust pressures for samples taken near the kyanite-sillimanite isograd west of Mica Creek, B.C. to be consistent with experimental data from Holdaway (1971). This approach has not been adopted in this study. The writer would prefer to calculate pressures assuming the ideal behaviour model described above. When experimentally derived activity data become available corrections can be made where necessary.

This geobarometer has been applied to metapelites from area immediately adjacent to this study area such as the Esplanade Range (Ghent 1977), the Park Range and northern Selkirk Mountains (Craw 1977), and the Monashee Mountains (Ghent et al. 1978).

(3-4-2) Pressure Estimates

Pressures calculated by garnet-plagioclase geobarometry according to Eq. 3-13 and 3-14 are listed in Table 3-4. Only temperature values that are believed to accurately represent peak metamorphic conditions have been used in pressure calculations. The calculated pressures are compatible with estimates of the pressure of formation (made using Carmichael's grid) for the metamorphic zone bearing each sample.

Error in pressure determinations can be the result of either uncertainties in temperatures or mineral compositions. Ghent et al. (1979) estimate that the garnet-plagioclase- Al_2SiO_5 polymorph-quartz geobarometer is precise to ± 1.6 kbar (two standard deviations) if a temperature error of $\pm 15^\circ\text{C}$ and a compositional

Table 3-4. Temperatures and pressures considered to be representative of the peak of metamorphism in the northern Selkirk Mountains

Station	T°C	Method	Station	TEMPERATURE				PRESSURE		
				X _{an} (x100)	X _{gr} (x100)	Al ₂ SiO ₅	P _{calc} (kbar)	P _{grid} (kbar)		
A345 -R	590	garnet-biotite	A345 -R	21.6	5.6	kyan	7.1	>5		
F061 -R	530	garnet-biotite	F061 -R	40.9	7.7	sill	5.1	~5		
P57 -R	610	garnet-biotite	P57 -R	23.5	4.8	kyan	6.7	>5		
P87b -R	630	garnet-biotite	P87b -R	21.2	4.1	sill	7.0	>5		
A352 -R	580	garnet-biotite	A352 -R	51.2	12.7	kyan	6.8	>5		
A238a-R	530	garnet-biotite	A238a-R	56.8	9.3	kyan	4.6	~5		
A87a -R	490	garnet-biotite	A87a -R	31.1	10.2	kyan	5.9	~5		
A278 -R	540	garnet-biotite	A278 -R	34.0	6.8	sill	5.5	~5		
A284b-R	650	garnet-biotite	A284b-R	21.2	4.2	sill	7.5	>5		
M170a-R	630	garnet-biotite	M170a-R	13.0	2.4	sill	6.8	>5		
M144	550	garnet-biotite	M144	22.2	6.1	kyan	6.5	>5		
F74d	560	calcite-dolomite	F74d	56.8	9.3	kyan	5.1	~5		
M64d	600	calcite-dolomite	M64d	16.0	3.5	sill	7.0	>5		
P95b	580	garnet-biotite	P95b	22.9	5.1	sill	6.7	>5		
A350 -R	610	garnet-biotite								
A93a -R	500	garnet-biotite								
V98a	540	calcite-dolomite								

error of $\pm 1\%$ grossular and $\pm 1\%$ anorthite is propagated through Eq. 3-13 and 3-14. This value is calculated for samples taken immediately west of this study area.

(3-5) DISCUSSION OF GEOTHERMOMETRY AND GEOBAROMETRY RESULTS

Twenty temperature estimates were calculated using the A.B. Thompson (1976) calibration of the garnet (rim composition) - biotite geothermometer. An additional three temperature estimates were made using Ewart's (1977) treatment of experimental data for the calcite-dolomite solvus geothermometer. Of the twenty-three values obtained, seventeen are compatible with temperatures estimated for the appropriate assemblage using Carmichael's (1978) petrogenetic grid.

A test in reproducibility of temperature estimates can be made by comparing temperatures obtained using different techniques for samples taken from a restricted area. At station A238 a value of 530°C is obtained using garnet-biotite geothermometry, and at station F74, a value of 560°C is obtained using calcite-dolomite solvus geothermometry. Both of these stations are located near French Lake (Figs. 1-2, 3-2, and 3-3). Mineral assemblage data and textural information indicate that in this region two mineral reactions representing the polymorph transition of kyanite to sillimanite, and the breakdown of staurolite by the reaction:

occur across the same isograd (Section 2-3-4). Carmichael (1978) estimates that these reactions intersect at 550°C and 5 kbar (Section 2-5-2).

The calcite-dolomite, and garnet-biotite geothermometers are calibrated using independent phenomena. The correlation of values using these methods, and the similarity with conditions inferred from phase relations allows some confidence in the geothermometers used. Therefore, although the error of an individual technique could be as much as $\pm 50^{\circ}\text{C}$, the reproducibility of results suggests that lower error limits could be realistic for the French Lake area.

Another test of reliability of estimates can be made by comparing the results of one technique for several samples from a restricted area. Reasonably consistent values indicate that the technique may be independent of variables such as bulk composition, fluid composition, and volatile activity.

In the region around French Lake, garnet-biotite temperatures of 530°C , 530°C , and 540°C have been estimated for stations A238, F061, and A278 respectively. South of this area in the kyanite-staurolite-biotite zone temperatures of 490°C and 500°C were obtained for sample A87 and A93. At the kyanite-sillimanite isograd northeast of the potassic feldspar - sillimanite zone, values of 580°C , 610°C , and 630°C were calculated for samples P95b, P57, and P87, and near the Rocky Mountain Trench values of 580°C , 590°C , and 610°C were determined for stations A352, A345, and A350.

These examples demonstrate that, within error, temperature estimates can be reproduced on a local scale using the garnet-biotite geothermometer.

It is not possible to ignore the six garnet-biotite temperatures that were considered to be unrealistic.

One value (M63, 662°C) is considered to be erroneously high due to high titanium content in biotite.

Three temperature estimates from the potassic feldspar - sillimanite-muscovite-quartz zone are about 575°C . These values are lower than garnet-biotite temperature estimates in the lower grade sillimanite-garnet-biotite zone to the northeast. The estimates are not compatible with petrogenetic grids (e.g. Carmichael 1978; A.B. Thompson 1976). Although anomalous, these estimates do compare with values obtained by Tracy (1978) for rocks of the same metamorphic grade. No explanation is available to account for these low estimates.

The remaining two temperatures that are significantly higher than estimates made using Carmichael's grid (640°C and 648°C for samples P147 and A379 respectively) are not consistent with other garnet-biotite temperatures estimated for samples from the same metamorphic zone within the Windy Range section and with estimates from higher grade rocks nearby. Anomalously high temperatures have been recorded for the kyanite-staurolite-garnet bearing rocks elsewhere (Ghent et al. 1979). Again, no adequate explanation for these discrepancies is available.

Of the 20 garnet-biotite temperatures calculated, 14 are compatible with assemblage data, one erroneously high value is accounted for, and five unrealistic values lack an adequate explanation. The empirical calibration of Thompson used in this study is new and needs substantial testing before it can be

thoroughly evaluated. The relatively successful application of this method to the northern Selkirk Mountains, as indicated by the reproducible results and the reasonable comparison of values with estimates from phase relations, is encouraging.

Problems such as chemical zonation of garnets, disequilibrium, effects of retrograde metamorphism, and reequilibration during waning metamorphism could account for these discrepancies. Continued application of this geothermometer and further understanding of the rather nebulous problems outlined above may increase the confidence with which this technique can be used.

All three temperature estimates made using calcite-dolomite geothermometry are compatible with temperatures inferred from phase relations. Similarly, all pressure estimates made using the garnet-plagioclase- Al_2SiO_5 -quartz geobarometer are considered to be reasonable.

If the pressure-temperature coordinates determined for each sample (Table 3-3) are plotted on Carmichael's grid (Fig. 3-4) two clusters of data can be recognized. Symbols are assigned to each sample to enable the reader to relate the calculated pressure-temperature coordinate and the mineral assemblage within the sample to the assemblages determined from Carmichael's grid. Figures 3-3 and 3-4 indicate that two well defined groups of pressure-temperature estimates exist. One group, with temperatures ranging from 490°C to 560°C and pressures about 5 kbar, occupies the southwest half of the study area (French Creek section: see Chapter 2). The other group, with temperatures ranging from approximately 550°C to 650°C and pressures of about 7 kbar, occupies the northeast half

Fig. 3-4. Position of the French Creek section and the Windy Range section temperature and pressure domains compared with phase relations as estimated from Carmichael's (1978) petrogenetic grid.

of the study area (Windy Range section; see Chapter 2). There is no overlap of data from one group into the other. These two pressure-temperature domains are compatible with the estimated temperatures and pressures of metamorphism in the French Creek and Windy Range sections (Section 2-5-2).

The pressure-temperature coordinates determined for individual samples on Fig. 3-4 may not fall within, but are generally close to, the appropriate mineral zone on Carmichael's grid. Differences could be accounted for by the following reasons:

- (1) There is a component of error in the temperature estimates. This affects not only the accuracy of the temperature estimates but also affects the accuracy of the pressure calculation since a value for temperature is needed in the equations used to calculate pressure.
- (2) The value of K_D used in Eq. 3-13 and 3-14 to calculate pressure was calculated assuming ideal mixing behaviour of end member components in garnet and plagioclase. This assumption could introduce significant error if incorrect.
- (3) As yet no single petrogenetic grid for rocks of pelitic bulk composition is universally accepted. This writer considers Carmichael's grid to be the most realistic representation to date for those mineral zones encountered in the northern Selkirk Mountains. Errors in the position of mineral reactions on this grid may be sufficient to allow for the discrepancies between calculated positions and assemblage fields on Fig. 3-4.

(3-6) CONCLUSIONS

This Chapter presents semiquantitative estimates of temperature and pressure for the northern Selkirk Mountains. All of the calcite-dolomite and most of the garnet-biotite temperature estimates are compatible with temperature estimates made using phase relations (Section 2-5-2). All pressure estimates made using the garnet-plagioclase- Al_2SiO_5 -quartz method are compatible with phase relations.

Two temperature-pressure regimes are recognized in this study area. Geothermometry and geobarometry corroborate phase relation evidence presented in Chapter 2 which suggests that rocks from the French Creek section were metamorphosed at lower pressures than those of the Windy Range section. Data presented in this chapter indicate that temperatures from about 500°C (beginning of staurolite-kyanite-biotite zone) to about 560°C (sillimanite-garnet-biotite zone) and pressures of about 5 kbar prevailed during metamorphism in the French Creek section, whereas temperatures from about 550°C (staurolite-muscovite-quartz-kyanite-garnet-biotite zone) to about 650°C (potassic feldspar - sillimanite zone) and pressures of about 7 kbar prevailed in the Windy Range section.

CHAPTER 4.

STRUCTURAL AND TECTONIC IMPLICATIONS OF METAMORPHIC ANALYSIS

(4-1) INTRODUCTION

In the previous chapters an isograd map for pelitic assemblages, and temperatures and pressures believed to be representative of peak metamorphic conditions, are presented. It is now possible to formulate the tectonic implications of this metamorphic information on both a local and regional scale.

Also, by making several assumptions, it is possible to construct several cross sections relating the distribution of isotherms and isobars with respect to isograds for various stages in the orogenic history of the northern Selkirks and adjacent Rocky Mountains (Park Range). Although the cross sections are of a conjectural nature they provide a reasonable metamorphic model which should be tested.

Tectono-metamorphic analysis of this area is particularly valuable since the Selkirk Mountains form a zone of transitional structural style between the core (Monashee Mountains - Shuswap Complex) and the Foreland Thrust and Fold Belt (Rocky Mountains) of the southeastern Canadian Cordillera. This area is the subject of much controversy; three tectonic models are currently proposed to account for the variation in structural style. Hopefully the conclusions set out in this chapter will assist in assessing the validity of these models.

(4-2) LOCAL STRUCTURAL IMPLICATIONS

Brown and Tippet (1978) outline the structural geometry

and stratigraphy of the northern Selkirk Mountains (Figs. 4-1, 4-2, and 4-3). They describe the southwestern part of this region as being dominated by northeast dipping second phase (F_2) folds and the northeastern part of this region as being dominated by southwest dipping third phase (F_3) folds (Fig. 4-3). The boundary between these two structural domains is called the Selkirk fan structure. This fan was first recognized by Wheeler (1965) and was described as a cleavage fan.

The major regional metamorphic event occurred during and after F_2 but before F_3 (Franzen 1974; Tippet 1976; van der Leeden 1976).

From the previous chapter (Fig. 3-4) it is apparent that two zones characterized by different metamorphic conditions can be distinguished in the northern Selkirks. The northeastern domain, which equilibrated at about 7 kbar, is best represented by the Windy Range section, whereas the southwestern domain, which formed at 5 kbar, is exemplified by the French Creek section. If one assumes that 1 kbar pressure corresponds to about 3.5 km of rock strata (i.e. density = 2.85 g/cm^3) then the 2 kbar break between sections can be taken to represent 7 km of relative vertical displacement.

Maximum temperature estimates of about 550°C in the 5 kbar domain are located in the sillimanite-garnet-biotite zone, whereas in the 7 kbar domain maximum temperature estimates of about 650°C are recorded in the potassic feldspar - sillimanite zone. This 100°C difference in temperature, which coincides with the increase in pressure, also divides the study area into two metamorphic domains. Since temperature and pressure estimates are considered to represent peak metamorphic conditions the relative displacement

Fig. 4-1. Lithologic map of the northern Selkirk Mountains (modified after Brown and Tippett (1978)).

Structural Map of Big Bend Area

Fig. 4-2. Structural map of the northern Selkirk Mountains (modified after Brown and Tippett (1978)).

Fig. 4-3. Composite cross section for the French Creek and Windy Range Sections from Brown and Tippet (1978). Cross section lines included on Figs. 4-1 and 4-2.

between the zones must be postmetamorphic.

The two stations (A284, A278) that position the abrupt change in metamorphic conditions are 5 km apart (Fig. 3-3). On either side of this break pressure estimates are reasonably constant which implies that the French Creek section and the Windy Range section can be treated as separate blocks. The Selkirk fan structure lies well within the 5 kbar French Creek section.

A northwest trending fault, documented near the headwaters of Norman Wood Creek by Brown and Tippet (1978), lies between these stations (Fig. 4-2). This steeply southwest dipping fault (Fig. 4-3) is believed to extend along or near the northeastern contact of the Cambrian Hamill Group to the vicinity of the headwaters of Birch Creek (M.J. Perkins, personal communication, 1979). Along most of its length it juxtaposes Hamill Group rocks against the upper pelite member of the Horsethief Creek Group. Near Norman Wood Creek an abrupt change in vergence of minor folds across the fault has been mapped by R.L. Brown (personal communication, 1979). He reports that since stratigraphy southwest of this fault forms part of an inverted limb of an F_1 fold and stratigraphy northeast of the fault forms part of a right way up limb of an F_1 fold, then at least one F_1 closure has been removed by faulting. However, since the geometry of the missing stratigraphy is not known it is not possible to estimate the displacement on the fault from stratigraphic arguments.

The 2 kbar offset documented in this study implies a vertical displacement of approximately 7 km across this fault. Pressure data indicate that the Windy Range section has been uplifted relative to the French Creek section. Since the fault is considered to

dip to the southwest this suggests major normal relative movement on this structure. This fault is herein named the Argonaut Mountain Fault.

The extension of the Argonaut Mountain Fault beyond this study area is not known. To the southeast it may extend around the Adamant Pluton or it may follow stratigraphy down Norman Wood Creek. Further information is required before assessment of these two possibilities can be made. To the northwest this fault may extend across the Columbia River or it may turn sharply southward to coincide with a normal fault that Brown and Psutka (1979) have traced along the Columbia River Valley (Fig. 4-2). Stratigraphy northwest of the Columbia River opposite Birch Creek is assigned to the Proterozoic Kaza Formation by Campbell (1968). Since the stratigraphic change recorded across the Argonaut Mountain Fault is not traced west of the Columbia River, the structure is not thought to extend into the Monashees.

A fault juxtaposing Paleozoic stratigraphy on the east side of the Columbia River against Proterozoic stratigraphy on the west side has been traced north to Ruddoch Creek by Brown and Psutka (1979). North of the Argonaut Mountain Fault isograds and stratigraphy extend uninterrupted across the Columbia River. Therefore, the Columbia River Valley Fault is not thought to extend north of Birch Creek. It may be possible to connect these two faults, either directly or by a series of interconnecting faults. Stratigraphic offsets across these faults demonstrate that the upper Proterozoic - Paleozoic southern block is downthrown relative to the Proterozoic northern block. Vertical displacements of 7 km can only be proposed for the French Creek

area; however, stratigraphic differences along the remainder of the fault support significant offsets. Further work is necessary to test this proposal.

Pressure estimates for the Windy Range section are plotted in plan view in Fig. 3-3 and projected onto cross section in Fig. 4-3. These estimates range from 6.7 to 7.5 kbar. The range of pressure estimates is not significant when possible errors in pressure estimates are considered. It is not possible to recognize postmetamorphic vertical movements due to faulting or folding in the Windy Range with the technique used.

It is assumed that pressure reflects only depth of burial and that tectonic overpressures of more than several tens of bars are generally not realistic (Fletcher 1972). Therefore, the uniform pressure estimates observed throughout the Windy Range section suggest that assemblages presently observed at the surface formed at approximately the same depth of burial of about 25 km. This in turn implies that uplift of the Windy Range section occurred without significant net rotation of isobaric surfaces from their postmetamorphic orientation about the horizontal.

Brown and Tippet (1978) propose that superposition of D_3 folds on D_2 folds should result in exposure of progressively deeper stratigraphic units eastward from the Selkirk fan structure towards the Rocky Mountain Trench (Section 4-5-1). This would be expressed by eastward increasing pressure estimates towards the trench. Available data do not support this proposal.

(4-3) REGIONAL TECTONIC IMPLICATIONS

(4-3-1) The Rocky Mountain Trench - Introductory Discussion

The northeastern part of this study area is bounded by the Rocky Mountain Trench. This pronounced northwest trending lineament separates the Foreland Thrust and Fold Belt from the core and transition zones of the orogen.

In this study area the Rocky Mountain Trench is the erosional expression of the Purcell Fault (Simony and Wind 1970). This structure is a listric thrust fault with southwest directed dips varying from about 35° in the Dogtooth Mountains (150 km southeast of the study area) (Simony and Wind 1970) to about 70° in this region (Craw 1977) to more than 70° approximately 200 km to the northwest (Campbell 1973). Postmetamorphic vertical displacement on this fault is estimated to be about 6 km (Price and Simony 1971; Campbell 1973) or 7 - 10 km (Craw 1978).

Craw (1978) has mapped metamorphic zones across the Rocky Mountain Trench from the Selkirks. Park Range and Monashee Mountains metamorphic zones, as outlined by Craw (1978), are combined with zones determined in this study (Map 3, in pocket).

(4-3-2) Discussion of Craw's Proposed Offset Across the Rocky Mountain Trench

Craw's documentation (1977, 1978) of a 7 - 10 km vertical offset across the Rocky Mountain Trench (southwest side up) is drawn from arguments using phase relations, geothermometry, and

geobarometry. He estimates that metamorphic assemblages presently exposed in the Park Range equilibrated at about 5 kbar whereas the Selkirks - Monashees equilibrated at 7 kbar. Temperature estimates were made using A.B. Thompson's (1976) calibration of the garnet-biotite geothermometer and pressure estimates were made using the equilibrium grossular-anorthite-kyanite-quartz as outlined by Ghent (1976).

Craw (1978, fig. 8) suggests that the Park Range contains assemblages indicative of conditions at or near the invariant point represented by the intersection of the kyanite-sillimanite isograd and the staurolite-out isograd. He states (1978, p. 92) that "the Park Range sillimanite (fibrolite) isograd is approximately coincident with the staurolite disappearance isograd... The sillimanite zone has a very restricted occurrence (upgrade from the staurolite disappearance isograd) ...". Craw also notes that the staurolite-out zone is marked by the appearance of a second habit of kyanite not recognized at lower metamorphic grades.

Evidence for both kyanite and sillimanite in the staurolite-out zone of the Park Range makes identification of the representative stable assemblage for this zone difficult. Craw's term "staurolite-out" is adopted for this zone because of this lack of a well defined mineral assemblage for this zone (Map 3, in pocket).

(4-3-3). Block Faulting Adjacent to the Rocky Mountain Trench

Two faults which parallel one another and which have major components of vertical offset have been described. The Argonaut

Mountain Fault separates the southwest part of the study area (5 kbar domain) from the northeast part (7 kbar domain). The Purcell Fault separates the northeast part of the study area from the Park Range (5 kbar domain). Pressure estimates suggest that the Windy Range section has been uplifted 7 km relative to the adjacent regions.

The extensions of the high pressure domain observed in the Windy Range section to the northwest and southeast of this study area are not known. Craw (1978) assigns the Monashee Mountains adjacent to the trench to the 7 kbar domain. E. Ghent et al. (oral presentation, 1978) estimate pressures from 6.0 to 6.7 kbar for the extension of the Windy Range section kyanite-sillimanite isograd northwest of the Columbia River. The pressures from Ghent et al. (1978) have been adjusted using an empirical activity correction for calcium components in garnet and plagioclase. Incorporation of this activity term will reduce pressures calculated by Eq. 3-13 and 3-14 by about 400 bars (E.D. Ghent, personal communication, 1978). If this is taken into account, then Ghent's data are approximately correlative with pressure estimates for the Windy Range section from this study. The Esplanade Range, 50 km southeast of this study area and adjacent to the Rocky Mountain Trench, formed at about 7 kbar (E.D. Ghent, personal communication, 1981).

The high pressure regime exemplified by the Windy Range section may therefore extend along the Rocky Mountain Trench from the Malton Gneiss to the Trans Canada Highway, a distance of 150 km.

Although the Argohaut Mountain Fault and Purcell Fault are postmetamorphic, relative timing of displacement cannot be

determined for this study area. Without this specific information it is simplest to assume that the movement path of the Windy Range section involved simultaneous uplift along both faults.

It is nonetheless possible to envisage more complex movement paths. Price and Mountjoy (1970) and Bally et al. (1966) propose that normal faulting postdates thrust faulting in the Rocky Mountains. If this sequence applies to the northern Selkirk Mountains then it is conceivable that movement on the Purcell Thrust preceded that on the Argonaut Mountain Fault.

In consultation with R.L. Brown, one possible model has been formulated on the basis of the above sequence. Initial movement of the northern Selkirk Mountains along the Purcell Fault may have resulted in counterclockwise rotation of isobaric surfaces. Subsequent normal displacement on the Argonaut Mountain Fault during the early stages of an extensional event allowed isobaric surfaces in the French Creek and Windy Range sections to dip towards each other. Continued extension may have caused normal movement on the Columbia River Valley and Purcell (?) Faults to readjust isobaric surfaces to their initial horizontal attitudes.

The final configuration, depicted schematically in Fig. 4-4, defines a horst and graben system which is suggestive of a period of extension late in the evolution of the orogen.

(4-3-4) Review of Previous Work on Normal and Block Faulting in the Southeastern Canadian Cordillera

Normal faults have been recognized throughout the southeastern

Fig. 4-4. Schematic block diagram outlining the series of fault bounded blocks within the northern Selkirk Mountains and adjacent regions.

Canadian Cordillera. On a scale local to this study area Price and Mountjoy (1970) infer southwest dipping normal faults from the offset of prominent structural markers along the Columbia and Canoe River segments of the Rocky Mountain Trench. Crow (1977) describes northwest trending late normal faults with offsets in the order of a few hundred metres (west side down) in the Park Range. One normal fault, shown in Crow (1977, fig. 1) displays almost 2 km of vertical offset. He postulates that since most of the faults die out towards the southeast there is a complementary decrease in offset in this direction.

On a regional scale Bally et al. (1966) document early Tertiary thrust faults which precede late Tertiary normal faults. From seismic reflection data they propose that these normal faults, which are steep at the surface, become listric with depth. They also claim that both block faulting and normal faulting are observed along the Rocky Mountain Trench. Wheeler (1966) states that the Purcell Anticlinorium and Shuswap Complex are cut by steep normal faults, and Simony and Wind (1970) describe normal faults which parallel thrust faults in the Dogtooth Mountains. Price and Mountjoy (1970) describe northwest trending normal faults as common in the central Rocky Mountains and along the Rocky Mountain Trench. These faults have the southwest side thrown down and hence are opposite in sense of slip to thrust faults in this area. Price and Mountjoy (1970) assign these normal faults to a post-thrusting relaxation event which implies minor offsets. This hypothesis may apply to structures in the Rocky Mountains but for this study area wherein normal vertical

offset approximately equals reverse vertical offset, such an explanation is not considered to be realistic.

Campbell (1973) has recognized general vertical movement of Precambrian blocks along steep faults in the Columbia Mountains 220 km to the northwest of this study area (Fig. 4-9). The tectonic model proposed by Campbell to account for the development of the southeastern Canadian Cordillera is contingent upon vertical postmetamorphic movement of these blocks with minor rotation about the horizontal.

(4-4) TECTONO-METAMORPHIC MODEL FOR THE SELKIRK MOUNTAINS - PARK RANGE

(4-4-1) Introduction

P.H. Thompson (1976) has outlined a means of determining pressure and temperature distributions in regionally metamorphosed terrains from isograd patterns and isograd dips. This treatment enables calculation of both orientation of isotherms with respect to isobars (the angle between them being α), and temperature gradients (change in temperature per kilometre measured perpendicular to isothermal surfaces).

This section outlines the application of the technique to the northern Selkirk Mountains and adjacent Rocky Mountains (Park Range). Certain assumptions, as listed below, are necessary to complete the cross sections relating isothermal and isobaric traces to isograd distribution. The accuracy of these cross sections is therefore

limited by the validity of these assumptions.

(4-4-2) Assumptions

Temperature

If temperatures for three localities, two of which are at different elevations, are known then it is possible to geometrically solve this triangle for attitudes of isotherms in the same manner as elementary strike lines are constructed for three point problems (Figs. 4-5A - 4-5D).

Construction of isotherm profiles has been made as follows. Temperature estimates for two surface localities are chosen from either garnet-biotite geothermometry or from estimates inferred from phase relations using Carmichael's grid (Fig. 2-22). Rocks presently exposed at the surface are considered to have formed at depths corresponding to 5 kbar, 7 kbar, and 5 kbar in the French Creek section, Windy Range section, and Park Range section respectively. One of the above surface temperatures is positioned near an isograd whose dip is known or can be restricted to a range of values. If this isograd is extrapolated to a depth considered to correspond with a pressure 1 kbar greater or less than surface conditions (assuming that 1 kbar pressure is equivalent to a 3.5 km depth of rock strata) then it is possible to estimate a temperature for this isograd at some new depth using Carmichael's petrogenetic grid. Using the three point problem approach, cross sections illustrating isotherm dips and spacing can be constructed.

cross section views

Fig. 4-5. Geometric technique for estimating isotherm dip and spacing.

Cross sections have been constructed in the above manner for the French Creek and Windy Range sections of the northern Selkirk Mountains and for the Park Range section in the adjacent Rocky Mountains. The temperature estimates used to plot the isotherm profiles are highlighted on Fig. 4-7A (in pocket). The Park Range treatment is derived from P.H. Thompson's (1978) discussion of Craw's (1978) paper. Isotherms have been interpolated between sections where possible. This interpolation is made on the assumption that isothermal surfaces vary in a regular manner.

The effects of errors in temperature estimates used to construct profiles are illustrated in Figs. 4-5A - 4-5D. Consider Fig. 4-5A to be a hypothetical profile constructed with "correct" data; Fig. 4-5B and 4-5C were constructed incorporating an error of $\pm 50^{\circ}\text{C}$ in one temperature estimate (400°C). The effect of this error is to change the dip of the isotherm by up to 7° from the correct version of Fig. 4-5A and to change the spacing between the isotherms. The direction of dip of the isotherms remains unchanged. The only way to change the direction of isotherm dip is to reverse surface temperatures so that temperatures decrease rather than increase from left to right (Fig. 4-5D).

The above discussion suggests that errors in temperature estimates will not affect the direction of dips of isotherms but will only alter the magnitude of dip and the relative spacing of isotherms. Therefore, errors in temperature estimates used to construct these tectono-metamorphic models do not necessarily invalidate the model.

Pressure

As outlined elsewhere in this chapter, pressures are assumed to reflect depth of burial only, the rate being 1 kbar per 3.5 km depth, and therefore in cross section isobaric traces are fixed as horizontal lines parallel to the earth's surface.

Dip of Isograds

The two isograds in this study area for which some information on the dip direction and magnitude is available are the kyanite-sillimanite isograds in the French Creek section and the Windy Range section (Section 2-4). The French Creek section isograd dips to the southwest at about 15° . Since the true strike for this isograd is not accurately known this value may represent an apparent dip rather than a true dip. Variation of the strike direction could increase the dip magnitude to approximately 90° . However, only southwest dips are possible.

The kyanite-sillimanite isograd in the Windy Range section dips anywhere from approximately vertical to about 40° to the southwest. Although poorly constrained, the range is comparable to the 60° southwest value presented for this isograd immediately west of the Columbia River (Ghent et al. 1977; P.S. Simony, communication to Cordilleran Symposium, Queen's University, January 1979). De Vries (1971) describes the garnet-in and staurolite-in isograds in the Esplanade Range as dipping from 33° to 46° southwest. A post- D_3 dip of 60° southwest for the kyanite-sillimanite isograd for the Windy Range section

is proposed on the basis of work by Ghent et al. (1977) and P.S. Simony (communication to Cordilleran Symposium, Queen's University, January 1979).

The effect of errors in magnitude of isograd dips can be considered in the same manner as error in temperature estimates discussed above. A steeper isograd dip will cause isotherms to dip more steeply. Therefore, magnitudes of dip of isotherms may be affected; however, the direction of isotherm dip is not likely to be affected.

Effect of Postmetamorphic Deformation

Temperature and pressure distributions indicative of peak metamorphic conditions can be related to isograd patterns only for areas where major postmetamorphic penetrative deformation is not observed. If deformation has occurred it is necessary to make assumptions to restore the area to a configuration representative of the peak of metamorphism.

The Windy Range section has undergone postmetamorphic D_3 penetrative deformation (Brown and Tippet 1978). Ghent et al. (1977, p. 16) describe the effect of this deformation on the sillimanite isograd near Mica Creek which extends southeast through the Windy Range as follows: "it is clear that the sillimanite 'inlier' in Columbia Valley, north of Mica Creek, is related to phase 3 folding and that the metamorphic surfaces are folded during phase 3".

The determination of a pre- D_3 isograd dip for the kyanite-sillimanite isograd in the Windy Range section presents a problem.

From Brown and Tippet's (1978) cross section (Fig. 4-3) it is estimated that the Windy Range section was shortened horizontally by about 50% during D_3 . This not only changes the relative spacing of isograds but also permits rotation of surfaces that were not originally parallel to the direction of maximum shortening towards the direction of maximum elongation (Fig. 4-6A).

Brown and Tippet (1978) also propose that D_3 folds were formed by westerly, shallowly dipping underthrusting of crust transitional between oceanic and continental beneath the Selkirk Mountains. The shear couple established by this event would rotate markers clockwise, looking northwest across the trench (Fig. 4-6B).

It is therefore conceivable that the combined effect of shortening and shearing has been to rotate the kyanite-sillimanite isograd from an original northeasterly dip through the vertical to steep southwesterly dips presently observed.

Second phase fold limbs in the French Creek section dip to the northeast (Fig. 4-3) whereas those in the Windy Range section dip to the southwest. Clockwise rotation of fold limbs through 75° is compatible with the above argument. However, until a detailed strain analysis of the Windy Range is accomplished, it is not possible to quantify the amount of rotation.

It has therefore been necessary to assign an arbitrary 45° northeast dip to the pre- D_3 kyanite-sillimanite isograd in the Windy Range section. This represents a 75° rotation during postmetamorphic deformation which coincides with the rotation postulated for second phase fold limbs. Although the initial configuration may have been steeper or shallower, the error

Fig. 4-6. Change in orientation of horizontal and inclined lines in an element subjected to (A) pure shear, and (B) simple shear.

introduced by variation in isograd dip affects only the magnitude and not the direction of the isotherm dip.

Pressure estimates suggest that isobaric surfaces in the Windy Range suffered neither significant folding nor rotation (Section 4-2). The present orientation may represent the resultant movement during several faulting events (Section 4-2). However, it is still necessary to describe a deformational model that allows for the rotation of isotherms, isograds, and stratigraphy with no apparent rotation of isobaric surfaces during D_3 .

If horizontal shortening of 50% occurred during D_3 then only surfaces which were originally parallel to the direction of shortening would not have rotated (Fig. 4-6A). Isobaric surfaces fulfill this requirement.

If gently dipping westward subduction of crust transitional between oceanic and continental under the Selkirk terrain (R.L. Brown, in press) resulted in shearing of the supracrustal succession, then it is conceivable that the angle between the direction of shear and isobaric surfaces may have been small. Since a marker originally parallel to the direction of shear will not rotate it is possible that isobaric surfaces in the Selkirk Mountains were in an orientation favourable for maintaining a horizontal attitude (Fig. 4-6B).

Theoretical consideration of shearing and compression can therefore account for rotation of some surfaces and not others.

Data used to construct the cross section for the French Creek section are all located southwest of the Selkirk fan structure. Since this area has not been deformed by D_3 on a megascopic scale the pattern proposed has remained essentially unchanged since the peak of metamorphism.

Isograd Distribution in the Park Range

The cross section for the Park Range is adapted from P.H. Thompson's (1978) discussion of Craw's (1978) paper. Craw (1978) realizes that postmetamorphic penetrative deformation has affected metamorphic patterns but from his data he can only assume "that the configuration of the thermal antiform ... was determined at depth assuming that 'half' the strain of macroscopic folding occurred during F_1 " (Craw 1978, pp. 92-93).

Postmetamorphic deformation is assigned to D_2 in the Park Range (Craw 1977) as opposed to D_3 in the Windy Range section. Sufficient information is not yet available to unravel the effects of this postmetamorphic event. Therefore, one should accept P.H. Thompson's (1978) treatment of Craw's data, included in cross sections later in this study, in light of this limitation.

Prethrusting Proximity of the Selkirk Mountains and Park Range Metamorphic Terrains

The displacement along the Purcell Fault which separates the Selkirk Mountains and Park Range is not known. It is assumed that the displacement is sufficiently small to allow these two regions to represent the same regional metamorphic event. However, it is possible that large horizontal displacements have fortuitously juxtaposed the two distinct metamorphic terrains of similar characteristics.

Since the Purcell Fault dips steeply southwest near the Big Bend of the Columbia River (Craw 1978) it is conceivable that the vertical displacement predominates over horizontal movement.

Therefore, the two terrains may be part of the same metamorphic complex. The Purcell Fault is thought by some to become listric at depth (Price and Mountjoy 1970) whereas others extend it to depth at steep angles (Campbell 1973). If it does become listric with depth then the horizontal component of displacement could be increased significantly.

Summary Statement of Assumptions

Armed with these assumptions a sequence of steps beginning with the peak of metamorphism and ending following the block faulting event can be outlined (Figs. 4-7A, 4-7B, and 4-7C, in pocket). There are a sufficiently large number of major assumptions and unfortunately vague solutions to problems to warrant criticism of the validity with which these cross sections accurately reflect the tectono-metamorphic history of the area. Although this work is of a conjectural nature, it does provide a model that accounts for the observed isograd pattern. It has therefore been included in this study. Further field studies are required to test the predictions made using this model.

(4-4-3) Tectono-metamorphic Model

Peak of Metamorphism

Figure 4-7A represents a possible isotherm-isobar-isograd pattern correlative with conditions during the peak of metamorphism.

Isotherms outline two thermal anticlines. The French Creek section and Windy Range section lie on opposite flanks of one

thermal dome whose isotherms dip outwards at about 15° . The Park Range section lies on the northeast flank of a second dome whose isotherms dip outwards at about 35° .

These domes, by necessity, are separated by at least one thermal trough (Thompson 1978). Although the simplest solution is the placement of one thermal trough between the anticlines it is conceivable that other thermal anticlines and synclines existed between the Windy and Park Ranges.

Evidence for a trough between the Selkirks and Park Range has been removed, either during D_3 or by displacement on the Purcell Fault.

At pressures of interest in this study dehydration reactions have greater slopes than solid solution reactions in orthogonal (i.e. $\alpha = 90$) pressure-temperature space. Slopes of dehydration reactions are also subparallel to isotherms. When α is set at 15° or 35° , isobaric surfaces are set at horizontal, and pressures are set as increasing with depth, dehydration reactions dip at shallower angles than solid solution reactions. Therefore, the staurolite-in, staurolite-out, and muscovite breakdown isograds should all dip outwards from the centre of the thermal anticline at shallower dips than the kyanite-sillimanite isograd.

Within the French Creek section the assemblages staurolite-kyanite-muscovite-quartz and sillimanite-garnet-biotite bracket an invariant point defined by the intersection of the kyanite-sillimanite isograd and the staurolite-out isograd determined by Rx. 2-4 (Section 2-5-2). Other assemblages isobarically bracketing this invariant point (Fig. 2-22), sillimanite-staurolite-muscovite-quartz and kyanite-garnet-biotite, are rarely observed.

These two isograds intersect at a low angle in the cross section. Therefore, the paucity of the two assemblages discussed above may reflect the restricted size of their respective stability fields. Hence, the isograd pattern around this invariant point as proposed in the cross section is compatible with field assemblage data.

Temperature gradients representative of the peak of metamorphism can be calculated from Fig. 4-7A. Confidence in temperature gradient estimates depends on the accuracy of the dip of the isotherms (which in turn depends on the accuracy of the temperature coordinates for the original "three point problem" and the dip of the isograds). Greater isotherm dips result in lower temperature gradients. Therefore, it is best to consider these gradients as first approximations.

In the French Creek section a gradient of about $50^{\circ}\text{C}/\text{km}$ is proposed for the $450 - 550^{\circ}\text{C}$ temperature interval; in the Park Range section a gradient of $25^{\circ}\text{C}/\text{km}$ is proposed for the $500^{\circ} - 600^{\circ}\text{C}$ interval; and in the Windy Range section a gradient of about $15^{\circ}\text{C}/\text{km}$ is proposed for the $600^{\circ} - 650^{\circ}\text{C}$ interval. By interpolation, a gradient of about $25^{\circ}\text{C}/\text{km}$ can be advanced for the $500^{\circ} - 600^{\circ}\text{C}$ interval in the Windy Range section.

Even though the accuracy of these estimates is questionable, the magnitude of these values tends to decrease with increasing temperatures. This observation is consistent with Thompson (1977).

Post-penetrative Deformation - Pre-faulting Configuration

Figure 4-7B outlines a possible relationship between isotherms, isobars, and isograds following the final episode of penetrative deformation but before the block faulting event. The assumed effect of D_3 in the Windy Range section has been to deform isograds through about 75° and to shorten the section by about 50%. The Park Range pattern is left as in Fig. 4-7A because of the lack of data on the effects of postmetamorphic deformation (D_2) on isograds. The French Creek section also remains unchanged since it has not been megascopically affected by D_3 folds. Figure 4-7B has been constructed such that isotherms, isobars, and isograds that were originally near horizontal have been shortened but not significantly rotated whereas those that originally had a component of dip have been rotated.

The assumed D_3 rotation of the Windy Range section infers a near vertical dip for the dehydration equilibrium staurolite-out and sillimanite - potassic feldspar isograds and a southwest dip for the solid-solid equilibrium kyanite-sillimanite isograd.

Ghent et al. (1977, fig. 3.2) suggest that the staurolite-out isograd dips steeply to the southwest in the vicinity of Warsaw Mountain, just inside the Big Bend of the Columbia River. They also map a postmetamorphic normal fault southeast of Mount Nagle which trends approximately perpendicular to isograds (Ghent et al. 1977; fig. 3.1; Mitchell 1976). Offset of the staurolite-kyanite-biotite isograd across this fault is small implying that the isograd dip is very steep. Therefore, evidence from mapping on the periphery of this study area supports the

prediction of very steep dips for isograds defined by dehydration equilibria as proposed by this model.

Effects of postmetamorphic deformation of the thermal trough separating the Windy and Park Ranges can only be speculated: Brown and Tippet (1978) suggest that upward movement occurred along the eastern flank of the Selkirk Mountains during D_3 . They also suggest that D_3 structures could have resulted from underthrusting of basement material. This underthrusting may have drawn the thermal trough downwards. With the limited data available the question marks in Fig. 4-7B are appropriate.

Block Faulting

The final episode in this model, represented in Fig. 4-7C, details the resultant configuration of the 7 km vertical uplift of the Windy Range section along the Argonaut Mountain and Purcell Faults. This is accomplished without penetrative deformation or significant net rotation of isobaric surfaces. Therefore, the profiles for the individual sections remain as in Fig. 4-7B. The resulting sequence of metamorphic zones, and temperature and pressure distributions is consistent with field observations and available data.

Normal faulting along the Argonaut Mountain Fault has brought sillimanite-garnet-biotite assemblages of the French Creek section against potassic feldspar - sillimanite assemblages of the Windy Range section. Isograd, isotherm, and isobar patterns are readily accounted for by normal faulting.

Thrusting along the Purcell Fault juxtaposes garnet and staurolite-muscovite-quartz-kyanite-garnet-biotite zone rocks

of the Windy Range against kyanite-garnet-biotite assemblages of the Park Range. If the thermal trough still existed after D_3 , it was uplifted and eroded during this thrusting event (Thompson 1978).

Since muscovite is still present in the potassic feldspar - sillimanite zone of the Windy Range section it is possible that these pelites are at or just above the potassic feldspar - sillimanite isograd. This is compatible with the position of this isograd on the cross section.

(4-5) TECTONIC MODELS

(4-5-1) Introductory Discussion

The variation of structural styles across the southeastern Canadian Cordillera has prompted the proposal of three unique tectonic models which relate structures formed in the core of the orogeny with those formed in the Foreland Thrust and Fold Belt. Summaries of these models are presented in Brown (1978) and are discussed below. The Selkirk Mountains lie within the transition zone between the core and Foreland Thrust and Fold Belt of the Cordillera. It is hoped that an understanding of the tectono-metamorphic evolution of this area will aid in assessing the validity of these models.

The allochthonous core zone model proposed by Price and Mountjoy (1970) (Fig. 4-8) involves diapiric upwelling of hot mobile core zone gneisses which, through lateral spreading under the effect of gravity, resulted in eastward thrusting and folding

Fig. 4-8. Simplified structural cross section of the southeast Canadian Cordillera proposed by Price and Mountjoy (1970). (After Brown and Tippett (1978)).

of cover rocks in the Rocky Mountains. Hence, the structural and metamorphic events in the Selkirk Mountains are thought to be supracrustal manifestations of this diapiric rise. As a result, northeastward translation and uplift of the metamorphic complex are considered to be syn-metamorphic and syn-deformational.

Campbell's (1973) autochthonous core zone model (Fig. 4-9) involves an uplift of about 11 km for the metamorphic core zone complex and main ranges of the Rocky Mountains. The Foreland Thrust and Fold Belt of the Front Ranges and foothills may have resulted from westward underthrusting of the craton. Uplift of the metamorphic complex occurred along fault bounded blocks. This uplift was postmetamorphic since isograds are offset by faults.

Brown and Tippet (1978) and Brown (1978) propose a three-stage evolution for the Selkirk Mountains: Phase I (D_1 structures) resulted in east facing (?) nappes; Phase II (D_2 structures) resulted in southwest verging folds and regional metamorphism; and Phase III (D_3 structures) resulted in northeast verging refolding of the eastern flanks of the Selkirk Mountains (Fig. 4-10). The superposition of Phase III structures on Phase II structures formed the Selkirk fan structure.

Phase II deformation is thought to be related to eastward underthrusting of core zone rocks resulting in crustal thickening whereas Phase III deformation is related to westward underthrusting of the northeastern flank by basement from beneath the Rocky Mountains. The Foreland Thrust and Fold Belt developed through continued westward directed underthrusting.

Northeastward and vertical displacement of the metamorphic

Fig. 4-9. Simplified structural cross section of the southeast Canadian Cordillera proposed by Campbell (1973). (After Brown and Tippett (1978)).

Fig. 4-10. Generalized structure profile and interpretation of pre-Phase III geometry for the northern Selkirk Mountains (from Brown and Tippett (1978)).

complex occurred along fault systems in post-Phase III times.

Recently R.L. Brown (in press) has suggested that the northern Selkirk Mountains formed part of a Selkirks-Kootenay marginal basin that originated during the Proterozoic. To the west of this basin was Aphebian sialic crust. However, the basin itself was underlain by oceanic to intermediate crust (Fig. 4-11A). Collapse of the basin by subducting oceanic crust began during the Paleozoic and culminated during Middle Jurassic with formation of Phase III folds as described above (Fig. 4-11B). Uplift of tectonized basin was completed during Late Jurassic by attempted westward underthrusting of the cratonic margin. Eastward post-metamorphic movement of the allochthonous cover rocks occurred across the Aphebian basement (Fig. 4-11C). Northeast-southwest extension occurred during the Tertiary (Eocene) following the completion of westward underthrusting of the Rocky Mountain foreland.

Of particular interest in this study is the nature of the heat source for regional metamorphism proposed for each of these models. Price and Mountjoy (1970, p. 14) propose that "the diapiric rise of a tongue of hot, ductile gneissic rocks from the infrastructure" resulted in "penetrative deformation and development of a metamorphic halo in the suprastructure around this tongue". Campbell (1973) infers that the heat source for regional metamorphism was depth of burial. Brown (1978, p. 146) proposes "that deformation in the northern Selkirks ... resulted in crustal thickening associated with onset of high grade metamorphism in the infrastructure".

Fig. 4-11. Middle Jurassic to Eocene kinematic evolution of the northern Selkirk Mountains (from R.L. Brown (in press)).

(4-5-2) Depth of Burial and Timing of Metamorphism

Pressure estimates for the northern Selkirk Mountains suggest that metamorphic assemblages that are presently exposed at the surface formed at depths of about 18 - 25 km (i.e. 1 kbar = 3.5 km of rock strata). The Park Range and French Creek sections are composed of units that straddle the Proterozoic-Cambrian stratigraphic boundary and were formed at pressures indicative of depths of about 18 km. The Windy Range section is composed of Proterozoic strata only and appropriately was formed at pressures suggestive of about 25 km depth of burial. Throughout this region the base of the supracrustal stratigraphic sequence is not exposed.

Price and Mountjoy (1970) estimate that the total thickness of the undeformed sedimentary wedge above the inferred western limit of basement (i.e. Park Range) was about 15 km. West of the Rocky Mountain Trench the exact thickness of the stratigraphic sequence is not known. Campbell et al. (1973) report that the Proterozoic and Lower Cambrian formations of the Cariboo Mountains west of the trench are in excess of 8 km thick. The thickness of eroded Paleozoic rocks above this sequence is not known. Campbell et al. (1973) also indicate that the sedimentary succession in the Cariboo Mountains thickens westward. Monger and Price (1979) suggest that the trench marks the eastern cratonic margin of a zone of pre-orogenic rifting, and may accordingly have provided an environment suitable for rapid thickening of the stratigraphic column.

Tectonic thickening of the Selkirk sequence during the

formation of megascopic Phase I nappes or during Phase II could conceivably have doubled the thickness of the succession. Pigage (1978) also has argued that tectonic thickening occurred in the Azure Lake region, 120 km west of this study area.

The depths of burial proposed for the peak of metamorphism can therefore be reconciled by combining the available estimates, the evidence for the westward thickening sedimentary wedge, and the evidence for tectonic thickening.

Regional metamorphism has been described above as being syn- to post- D_2 . It is therefore conceivable that metamorphism was initiated during D_2 as the stratigraphic sequence was being tectonically thickened, but that maximum temperatures and pressures were probably attained at the end of D_2 when westward verging second phase folds were at their greatest depths.

R.L. Brown (in press) suggests that westward underthrusting of oceanic basement resulted in uplift and postmetamorphic D_3 deformation of the Selkirk terrain. Continued differential uplift occurred along the Purcell and Argonaut Mountain Faults.

(4-5-3) Heat Source

On a regional scale the Selkirk metamorphic terrain can be viewed as an appendage of the Shuswap Metamorphic Complex. This would imply the existence of a fairly broad heat source. If the Selkirk terrain was transported eastward over part of the Shuswap terrain along a postmetamorphic décollement as proposed by R.L. Brown (in press) then the heat source may have been far

removed from the present location of its metamorphic expression.

The data presented in this study indicate that maximum temperatures of about 650°C occurred at depths of about 25 km. This corresponds to an average geothermal gradient of about $25^{\circ}\text{C}/\text{km}$.

The metamorphic culminations described in Section 4-4-3 correspond to two thermal anticlines which are separated by a trough. The shallow to moderate dips of these domes describe a gently undulating thermal topography.

This topography may parallel the outline of the heat source for regional metamorphism or it may reflect differing thermal conductivities from a heat source at depth. However, the shape of the isotherms does suggest that the heat source at least is as wide as the metamorphic complex within this study area. This would not aid in distinguishing between burial-type or rising ductile heat source proposed in tectonic models.

Equilibration of mineral assemblages at pressures equivalent to 25 km of depth and postmetamorphic uplift of part of the complex along faults do not support the syn-metamorphic, syn-deformational upwelling heat source proposed by Price and Mountjoy (1970).

Campbell (1973) also criticizes the Price and Mountjoy model's heat source because of the problems of laterally transporting a mass of rock with sufficient heat to cause metamorphism and penetrative deformation of the suprastructure. The Malton Gneiss is proposed by Price and Mountjoy (1970) to represent an exposed infrastructural diapir and hence should have a concomitant metamorphic halo. However, Ghent et al. (1977), by showing that the metamorphic grade decreases towards the gneiss, provide field evidence contradictory to Price and Mountjoy's model.

(4-5-4) Summary Discussion on Tectonic Model

From the arguments presented in this discussion it is evident that any tectonic model that accurately reflects the evolution of the northern Selkirk Mountains must account for burial of the succession with concomitant deformation and metamorphism, followed by vertical uplift of the deformed sequence as fault bounded blocks. The block faulting may be related to an episode of post-orogenic extension.

Models proposed by Campbell (1973) and R.L. Brown (1978, in press) and Brown and Tippet (1978) are compatible with the above sequence. However, the Price and Mountjoy (1970) model is not considered to be consistent with the limitations imposed by this study.

CHAPTER 5.

SUMMARY OF CONCLUSIONS

SUMMARY OF CONCLUSIONS

Contributions and conclusions in this thesis are outlined below.

(1) The northern Selkirk Mountains are divided into the following metamorphic mineral zones (proceeding from southwest to northeast): chlorite; chlorite-biotite; garnet-chlorite; staurolite-biotite; staurolite-kyanite-biotite; sillimanite-garnet-biotite; staurolite-muscovite-quartz-sillimanite-garnet-biotite; potassic feldspar - sillimanite; sillimanite-garnet-biotite; kyanite-garnet-biotite; and staurolite-muscovite-quartz-kyanite-garnet-biotite.

(2) Mineral reactions that define isograds separating some of the above zones are:

These reactions have been identified through recognition of changes in mineral parageneses and textures across isograds.

(3) Isograds trend northwest and parallel most major faults, stratigraphy, and axial surface traces. Mineral zones are arranged such that metamorphic grade increases towards a centrally located metamorphic culmination. Field evidence and modelling suggest that isograds in the French Creek section dip outwards from the core of the metamorphic culmination whereas those in the Windy Range section dip vertically to steeply to the southwest.

(4) Comparison of sequences of mineral assemblages on both sides of the culmination with Carmichael's (1978) petrogenetic grid suggests that the French Creek section assemblages formed at about 5 kbar whereas Windy Range section assemblages formed at greater pressures.

(5) Geothermometry by garnet-biotite and calcite dolomite methods provides temperature estimates ranging from 500°C to 650°C. These estimates are compatible with phase relations.

(6) Geobarometry confirms the presence of two different pressure regimes. Assemblages in the Windy Range section formed at about 7 kbar whereas those in the French Creek area formed at about 5 kbar.

(7) Changes in temperature and pressure occur abruptly across the Argonaut Mountain Fault. Vertical displacement on this fault is about 7 km.

(8) The northern Selkirk Mountains are divided into two fault bounded blocks; the Windy Range block contained between the Argonaut Mountain and Purcell Faults; and the French Creek block located between the Columbia River Valley and Argonaut Mountain Faults. Postmetamorphic movement on these faults resulting in the horst and graben arrangement probably corresponds with an Eocene(?) period of extension.

(9) A model describing the evolution of isothermal, isobaric, and isogradic patterns for the northern Selkirk Mountains and adjacent Park Range is presented. This model is compatible with field observations of mineral assemblages.

(10) Tectonic models that describe the evolution of this part of the southeastern Canadian Cordillera should account for the following sequence of events: burial of a Proterozoic sedimentary succession to depths of up to 25 km with concomitant metamorphism and deformation; and subsequent uplift of blocks along faults following equilibration of peak metamorphic assemblages. Models proposed by Brown (1978, in press) and Campbell (1973) are compatible with this sequence of events.

Suggestions for Further Work

This study has hopefully established a foundation for future petrological studies in the Selkirks. Several pertinent problems deserving of further work are outlined in this section.

The trace of the Argonaut Mountain Fault outside this study area is presently a matter of conjecture. It may be possible to extend this structure by mapping discontinuities in conditions necessary for formation of metamorphic assemblages.

The metamorphic map presented in this thesis is the best representation of the data available. Further work in areas of sparse sampling could improve this map. Also, now that isograd traces have been somewhat restricted in plan view it would be possible to attempt to trace the dip surfaces of the isograds. Such a study could test some of the conclusions made in the tectono-metamorphic model presented in this study.

This study has been restricted to rocks of pelitic bulk composition. Calc-silicate and carbonate rocks are widespread and it may be possible to outline metamorphic mineral zones on the basis of assemblages from these bulk compositions. Comparisons of conditions of metamorphism determined from such a study with results from this work could be extremely informative.

Paragonite has been identified locally in the Selkirk Mountains. Further study on the significance of this phase may be beneficial.

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APPENDIX A.

MINERAL ASSEMBLAGE DATA

This appendix contains a compilation of assemblage data available for rocks of pelitic bulk composition in this study area.

The left hand column of the table provides sample numbers.

Samples from suites studied are prefixed as follows:

- A - collected by R. Leatherbarrow
- F - collected by J. Franzen
- M - collected by M. Perkins and R. Leatherbarrow
- P - collected by M. Perkins and D. Shaw
- V - collected by J. van der Leeden

The centre column of the table describes the mineral assemblage for the sample of interest. The mineral abbreviations are listed in Table A-1. Before each phase listed in the assemblage is a digit from 1 to 10 or an "X". These indicate the approximate mineral modes: e.g. "X" refers to less than 1% by volume; 1 refers to 1 - 10%; 2 refers to 11 - 20% etc. Following each mineral abbreviation are one or two digits from 1 to 8. These refer to the following textures observed within the phases:

- 1 - retrograde
- 2 - phase not in textural equilibrium with rest of assemblage
- 3 - garnets embayed poikiloblasts with biotite coronas
- 4 - sillimanite-biotite intergrowths containing euhedral garnets
- 5 - phase occurs in muscovite
- 6 - vermicular texture
- 7 - poikiloblastic
- 8 - relatively inclusion free

As an example, the abbreviation lga32 means that 1 - 10% of the slide is made up of garnets that are poikiloblastic with ragged outlines and biotite rims. These garnets are not in textural equilibrium.

The right hand column of the table denotes the compatible mineral assemblage present in the sample. The convention for diagnostic mineral assemblages, as outlined below, is also used on the metamorphic map (Map 2, in pocket).

A - chlorite

B - chlorite-biotite

C - garnet

c - chlorite-garnet

c' - chlorite-garnet-staurolite+biotite

D - staurolite-kyanite-biotite

d - staurolite-biotite (no Al_2SiO_5)

E - staurolite-muscovite-quartz-kyanite-garnet-biotite

E' - staurolite-muscovite-quartz-sillimanite-garnet-biotite

F - kyanite-garnet-biotite

f - kyanite-biotite

G - sillimanite-garnet-biotite

g - sillimanite-biotite

H - potassic feldspar - sillimanite - muscovite

Z - staurolite-muscovite-quartz-sillimanite

o - assemblage not diagnostic

Table A-1. Mineral abbreviations

ap - apatite	ht - hematite	py - pyrite
bi - biotite	hb - hornblende	qz - quartz
ca - carbonate	il - ilmenite	ru - rutile
cc - calcite	kf - K-feldspar	se - sericite
cz - clinozoisite	ky - kyanite	si - sillimanite
ch - chlorite	mt - magnetite	sp - sphene
ep - epidote	mu - muscovite	st - staurolite
fs - feldspar	op - opaque	to - tourmaline
ga - garnet	pa - paragonite	zr - zircon
gr - graphite	pc - plagioclase	zo - zoisite

Sample Number	Assemblage	Metamorphic Zone
A003	2bi-Xcc-1ch-1ep-3mu-1pc-3qz-Xto-Xzr-lop	B
A006a	1bi-1chl-1ga7-5mu-1qz-Xzr-lop-lil	C
A008	3bi-1chl-1ga-4mu-1pc-2qz-Xsp-lop	C
A010b	2bi-Xcc-1ch-1ga-7qz-lop	c
A013	2bi-Xchl?-Xep-4mu-1py-3qz-Xto-Xzr	O
A014	3bi-1chl-1ga-5mu-1pc-1qz-Xto-Xzr-Xop-lil	C
A015b	2bi-1ch-1ga7-3mu-3qz-Xto-lop	c?
A016b	2bi-2ch-Xep-1ga7-2mu-3qz-Xzr-lop	c
A018	2bi-Xep-1ga32-1ky-2mu-2pc-2qz-1st-Xto-lil	D
A019a	2bi-1chl-1ga7-5mu-pc?-1qz-1to-1zr-lil	C
A019d	2bi-1chl-1ga7-3mu-Xpc-3qz-Xst5-Xto-Xpa?	O
A022	3bi-2ch-1ga7-2mu-3qz-lop	c
A023a	2bi-1chl?-Xep-Xga-4mu-Xpc-2qz-Xto-lop-lil	C
A023b	1bi-1ch-1ga7-1mu-7qz-lop	c
A023c	1bi-1ch-4mu-4qz-1to-lop-Xfs	B
A026	2bi-1ch-Xep-1ga-1hb-1pc-6qz-lop (no muscovite)	c?
A027	1bi-1chl-1ga7-3mu-5qz-1st-lop-lil	d
A028b	2bi-1ca-1chl-1ga7-3mu-3pc?-3qz-lil	C
A029	Xap-3bi-Xcc-1ch-1ep-1ga-1pc-1py-5qz-Xto	c?
A036b	2bi-1ch-7mu-Xpy-Xzr-lil	B
A038	2bi-Xch-1ep-3mu-3qz-Xto-lop	B
A045	1chl-1ga-2mu-6pc-2qz-lop-lil?-1cz	C
A046a	3bi-Xca-1ch-4mu-1py-3qz-lil	B
A049b	2bi-1ch-Xga7-4mu-3qz-1to-1zr-lop-1fs?	c
A050a	2bi-1ch-4mu-4qz-1to-lop	B
A053	1bi-1ch-Xep-3mu-1pc-5qz-Xzr-lop	B
A054	2bi-1ch-1ep-3mu-1py-4qz-1zr-lop	B
A055	3bi-1ch-3mu-Xpc-4qz-1to-lop	B
A056	1bi-2ch-3mu-5qz-1to-lop	B
A059	2bi-1ch-Xep-4mu-Xpc-1py-3qz-1to-1zr	B

A060b	2bi-lch-Xga7-2mu-2qz-1st-lto-2fs-lil	c'
A066	3bi-Xchl-Xep-lga-4mu-lqz-Xst-Xto-lfs-lil	c'
A071a	2bi-lch-3mu-4qz-Xzr-lop	B
A073a	2bi-lch-lep-3mu-2pc-3qz-lto-Xzr-lil	B
A075	1ch-lga7-4mu-3qz-Xst-lfs-lil	c')
A075	1ch-lga7-4mu-3pc-3qz-Xst2-lto-lil-lcz	C
A077b	2bi-lch-gr?-4mu-Xpy-3qz-Xzr	B
A083c	2bi-lchl-2mu-3pc-5qz	O
A085a	2bi-lch2?-lga7-2mu-3pc-3qz-Xzr-Xil	O
A085d	3bi-lga32-lky-3mu-lpc-3qz-1st-lzr-lop-lil	D
A087a	2bi-lga32-lky-lmu-lpc-4qz-1st7-lto-lzr-lop	D
A092a	3bi-Xga32-lky-3mu-2pc-lqz-3st-Xzr-lil	D
A093a	2bi-Xchl-lga32-2mu-2pc-4qz-1st7-Xzr-lop	d
A093b	3bi-Xchl-lga32-2mu-Xpc-3qz-2st-lzr-lop-lil	d
A103b	2bi-lky-2mu-pc?-3qz-lsi-2st7-lop-lil	D-g
A104	2bi-lky-lmu-6pc-1st-lil	D
A106c	3bi-lga-lky-2mu-lpc-3qz-1st-lto-lzr-lop	D
A108a	2bi-lga32-lky-3mu-lpc-2qz-1st-lto-Xop	D
A108b	3bi-lga-lky-3mu-lpc-4qz-1st-lop-Xil	D
A113b	1bi-4mu-5qz-lto-lop	O
A114	3bi-lca-3mu-lpc-4qz-Xto-lzr-lop	O
A116a	1bi-lchl-lga7-4mu-4qz-lto-lzr-lil	O
A121a	3bi-lchl-lga32-lky7-lmu-lpc-5qz-1st-lto-lzr-lil	D
A121b	3bi-lga32-lky-4mu-lpc-lqz-1st	D
A125	3bi-lchl-lga32-2ky-3mu-lpc-2qz-1st-lzr-lil	D
A125a	2bi-lchl-Xep-lga7-4mu-lpc-2qz-lto-lzr-lil-cz?	O
A125b	2bi-lchl-lga7-lky-6mu-lpc-lqz-lil	O
A127	3bi-lchl-lga7-3mu-4pc-qz?-lto-lzr-lop-lil-lcz	O
A133a	2bi-lchl-lga32-lky-2mu-2pc-2qz-1st-lzr-lop-lil	D
A134	1bi-lca-lchl-lga7-2mu-2pc-4qz-Xto-Xzr-lzo-lop-lil	C

A134a	1bi-1ca-1chl-1ga-2mu-2pc-4qz-Xto-Xzr-Xzo-lop-lil	C
A137a	1bi-1chl-1ga-1mu-6pc-qz?-1se-Xzr-lop	C
A144a	3bi-1chl-1ga7-1ky-3mu-Xpc-4qz-Xzr	O
A144b	3bi-1chl-1ga32-1ky-3mu-1pc-3qz-Xst	D
A146	2bi-1chl-1ga32-2ky-2mu-1pc-2qz-1st-lzr-lop-lil	D
A147b	3bi-1chl-1ga3-2mu-1pc-4qz-Xzr	O
A147b	2bi-1ga32-1ky-4mu-4qz-1st	D
A148a	3bi-1ga32-1ky-2mu-Xpc-3qz-1st-Xzr-lil	D
A148b	3bi-1chl-1ga32-1ky-2mu-Xpc-3qz-1st-lzr-lop	D
A149	3bi-Xchl-1ga3-1ky-4mu-3qz-Xto-Xzr	f?
A150a	2bi-1chl-1ga3-2mu-1pa-1pc-3qz-Xst-lto-lop	d
A151a	3bi-1ga32-2ky-2mu-1pc-2qz-1st-Xto-Xzr-Xil	D
A151b	3bi-1ch-1ga32-1ky-2mu-1pc-3qz-1st-Xto-lop	D
A155	3bi-1ga32-1ky-3mu-1pc-3qz-Xsi?-1st-Xzr-lil	D-g?
A161a	3bi-1ga-3mu-1pc-3qz-lzr	O
A161b	3bi-1chl-1ga-1ky-3mu-2pc-2qz-1st-Xto-lil	D
A162a	1bi-1chl-5mu-3qz-lto-lop	O
A163	2bi-1chl-1ga32-1ky-3mu-4qz-1st7-Xto-ca?	D
A164b	3bi-1chl-1ga32-3ky-2mu-3qz-1st7-Xzr-lop-lil	D
A165b	2bi-1chl-1ga-1ky-2mu-1pa5-1py?-2qz-1st-lzr-lil	D
A169	2bi-1chl-1ep-1ga-4mu-2pc-2qz-Xzr	O
A172a	2bi-1ga-1ky-2mu-6qz-Xst-Xto-Xzr	D
A172b	2bi-1ch-1ga7-3mu-1pc-4qz-1st-lto-lop	d
A173	3bi-1ga-1ky-3mu-Xpa?-Xpc-3qz-1st-Xzr-Xop	D
A177b	2bi-1ga-1mu-4qz-2si-Xzr-lop	G
A180b	2bi-2mu-2pc-3qz-1si-Xto-lop-lil	g
A181a	3bi-1mu-6pc-qz?-1si-Xzr-lop-lil	g
A183	4bi-4mu-1pc-qz?-lto-lop-Xcz?	O
A183b	3bi-Xchl-1ga-Xmu-3pc-1qz-Xzr-lop-lil	O
A184	2bi-1ga7-1mu-2pc-2qz-2si-lto-Xzr-lop	G

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A188b	2bi-lchl-lmu-lpc-3qz-lsi5-lto-Xzr-lop-lil	g
A189	2bi-lchl-lga34-2mu-2pc-2qz-lsi-Xto-Xzr-Xop	G
A190b	2bi-lchl-lga-3mu-2pc-2qz-lzr-lop	O
A191a	3bi-lga4-3mu-lpc-2qz-2si-lto	G
A191b	3bi-lga-3mu-2pc-3qz-lsi5-lop	O
A192	3bi-Xchl-ky?-lpc-4qz-3si-lst-lto-lzr-lil	g
A193	3bi-lga4-lop-4qz-2si-lst-lto-Xzr-Xil	G
A193b	Xap-2bi-lga-2mu6-lpc6-3qz-lsi5-Xzr-Xop	g
A194a	3bi-Xga4-lmu-lpc-3qz-2si-lto	G
A218a	3bi-lchl-lga-lmu-lpc-4qz-lsi-lst	E
A238a	3bi-lchl-lga-lky-3mu-lpc-3qz-Xst-lop-lil	F
A240	2bi-Xchl-lga7-3mu-lpc-3qz-Xto-Xzr-lop	D
A241a	2bi-lchl-lga-lky-4mu-2qz-lst-lto-lzr	D
A241b	2bi-lchl-2ga7-lky-2mu-lpc-3qz-lst-Xzr-lop	D
A242	3bi-lchl-lga-lky-3mu-3qz-Xst2-Xto-Xop	f
A252	2bi-lchl-lga4-2mu-2pc-3qz-lsi-lto-lzr-lop	G
A253	2bi-lchl-lga4-2mu-2pc-3qz-lsi-lto-lzr-lop-lil	G
A255b	2bi-lga4-2mu-3pc-2qz-lsi-lto-Xop	G
A262	3bi-lchl-lga4-2mu-3qz-lsi-Xto-Xzr-Xop	G
A265	3bi-lga4-2mu-3qz-lsi-lto-Xzr	G
A266a	3bi-Xchl-lga4-2mu-3pc-Xqz-lsi-Xst-Xto	E?
A274b	3bi-lga8-2mu-4qz-lsi-lst-lto-Xop	E
A276a	2bi-lga4-2mu-lpc-3qz-lsi-lst-lto-Xzr-lop	E
A276b	3bi-lga4-3mu-lpc-3qz-lsi-lst-lop	E
A277	2bi-lga4-2mu-lpc-4qz-lsi-Xto-Xzr (st in outcrop)	E?
A278	2bi-lchl-lga4-lmu-lpc-4qz-lsi-Xzr-Xop (st in outcrop)	E?
A281b	1bi-lkf-2mu-3pc-4qz-Xzr	O
A282b	1bi-lchl-lkf-lmu6-lpc-6qz-lsi-Xzr-lil?	H
A284a	1bi-lga-lkf-lmu6-lpc-6qz-lsi-Xzr	C
A284b	2bi-lga-lkf-lmu6-lpc-4qz-lsi	H

A289	lbi-3kf-1mu-1pc-3qz-1si-Xzr-lop	H
A291	lbi-1ga-1kf-1mu-3pc-3qz-Xzr	O
A298b	1mu-5pc-5qz-1si5	O
A300a	2bi-1ch-1ga-1kf-1mu-1pc-5qz-1si	H
A300b	2bi-1ga-2kf2-1mu6-Xpc-6qz-1si5-Xzr	O
A301a	2bi-1chl-Xga-3kf2-1mu-4qz-1si5	O
A305b	lbi-1chl-1mu-1pc-7qz-2si-lop	O
A308c	lbi-1kf-1mu6-3pc-3qz-1sel-2si	H
A310b	2bi-1kf-1chl-1mu-2pc-2qz-2si-lto-lil	H
A312b	lbi-1chl-1ga-2kf-Xmu6-3pc-4qz-Xsi-Xop-Xil	H
A314a	3bi-Xkf-1mu-1pc-4qz-1si-Xzr-Xop	H
A331	4bi-1ga-1ky-4mu-1pc-1qz-Xst	E
A333	3bi-1ga7-1ky-1mu-1pc-4qz-Xzr	f
A344	3bi-1ga8-1ky-3mu-2qz-1st-lil	E
A345	3bi-Xchl-1ga78-1ky-2pc-3qz-1st-sp?-Xto-Xop	E
A346	ap?-3bi-1ga7-1ky-3mu-2qz-Xst-lto-Xzr	E
A347	Xap-2bi-1ga-1ky-2mu-1pc-3qz-lto-lop-lil	F
A349	3bi-1ga7-1ky-3mu-3qz-1st7-lil	E
A351	3bi-1chl-1ga7-1ky-3mu-1pc-2qz-sp?-1st-Xto-Xzr-lop	E
A352	3bi-1chl-2ga8-2ky-3mu-1pc-1qz	f
A353	lbi-1ga7-1ky-3mu-1pc-4qz-1st7-Xto-Xzr-lil	E
A355a	Xap-3bi-1ga8-1ky-2mu-3qz-Xto-Xzr-Xop	F
A357	3bi-1ga-1ky-3mu-3qz-lto-lop	F
A358	2bi-1ga-1ky-1mu-3pc-3qz-lto-Xru?	F
A369a	2bi-1chl-1ga-3ky-2mu-3qz-lsp-Xst (hand specimen)	E
A375	2bi-1ga-1ky-2mu-1pc-3qz-Xsp-Xto-Xzr-lop	f
A376a	3bi-1ga-1ky-3mu-2qz-Xst (hand specimen)	E
A377a	2bi-1ga-1ky-2mu-1pc-5qz-Xst-Xto	E
A378	3bi-1ga-2ky-2mu-2qz-Xst (hand specimen)	E

P006a	2bi-1chl-Xep-1ga7-1mu-3pc-3qz-1zr-lop	O
P006b	2bi-1ga-1mu-1pc-4qz-1si5-1to-1zr-lop	G
P007	3bi-1kf-1mu6-1pc-5qz-1si-1zr-lop	H
P009	2bi-1ga4-3mu-1pc-3qz-2si-Xto-Xzr-Xop	G
P015	2bi-Xchl-1ga7-2mu-1pc-4qz-Xsi-1zr-lop	G
P016	3bi-Xga-3mu-Xpc-3qz-3si-Xzr-lop	G
P019	2bi-1ga-3mu6-1pc-5qz-1si5-1zr	G
P024	2bi-1ga8-1kf-1mu6-1pc-6qz-1si5-lop	H
P029a	1bi-2kf-2mu6-2pc-5qz-lop	O
P034a	1bi-2chl-1ga-1kf-2mu-2pc-4qz-lop	O
P034b	2bi-1ga-2mu-1pc-4qz-1si-1to-1zr-lop	G
P043b	2bi-1kf-1mu-2pc-4qz-Xsc-1si5-Xzr-lop	H
P050b	3bi-1ga8-3mu-1pc-1qz-2si-Xzr-lop	G
P051	1bi-1chl-1ga-1mu-3pc-5qz-1zr-lop	O
P057	2bi-1chl-1ga8-1ky-2mu-2pc-3qz-Xzr-lop	F
P060a	2bi-1chl-1ga-1ky-2mu-2pc-4qz-Xzr-Xop	F
P060b	2bi-Xchl-1ga8-1ky-2mu-2pc-4qz-Xop	F
P063a	1bi-1chl-1ga-1ky-1mu-2pc-5qz-1zr-lop	F
P065b	2bi-Xchl-Xky-2mu-2pc-5qz-Xzr-lop	f
P069a	1bi-1ga-1mu-3pc-6qz-1zr-lop	O
P069d	1bi6-1mu6-2pc6-6qz-1si5-lop	g
P069d	2bi-Xga-1mu6-2pc-5qz-1si-Xzr-Xop	G
P078	3bi-1ga8-1gr?-1ky-2mu-5qz-1zr-lop	F
P079a	2bi-Xchl-Xep-1ga-Xky-1mu-1pc-5qz-1zr-lop	F
P080	1bi-Xchl-1ga-1mu-2pc-6qz-1zr	O
P082b	2bi-Xchl-1ep-Xmu-3pc-5qz-1sp-lop	O
P084	3bi-Xchl-3mu-1pc-3qz-1si-lop	g
P085a	1bi-Xchl-Xep?-1mu6-1pc-8qz-Xsp?-1to	O
P087	2bi-1chl-1ga-1mu-2pc-5qz-1to-1zr	O
P087b	1bi-Xga8-2mu-1pc-5qz-2si5	G

P091b	1bi-1chl-1kf-1mu-1pc-8qz-Xsi2-Xzr-lop	g
P095a	1bi-1chl-1ga3-1kf-1mu-2pc-7qz-Xsc-Xzr-lop	O
P095b	2bi-1ga-Xgr?-1ky-1mu-2pc-4qz-Xto-Xzr	F
P097	2bi-1ga8-1ky-2mu-1pc-3qz-1to-Xzr-lop	F
P098	1bi-1ga3-1ky-1mu-Xpc-6qz-Xzr	F
P100	2bi-1ga8-2ky-2mu-1pc-4qz-Xzr	F
P101c	2bi-2ga-2ky-2mu-1pc-2qz-1to-1zr-lop	F
P102a	3bi-1ga8-1ky-1mu-3pc-3qz-1to-Xzr-lop	F
P102b	2bi-1ga8-1ky-2mu-1pc-4qz-Xzr	F
P107b	2bi-1ga-1ky-2mu-2pc-5qz-1to	F
P107b	2bi-1ga3-Xky?-2mu-1pc-3qz-1to-Xzr-Xop-Xil	O
P113a	3bi-1gr?-3mu-3pc-3qz-1zr	O
P147	Xap-3bi-Xch-1ga3-1ky-2mu-1pc-2qz-1sp-Xst-1to-lop	D
P151b	3bi-1ga-3mu-2pc-2qz-1to-1zr-lop	O
P161	3bi-1ga8-3mu-Xpc?-3qz-1sp-1to-1zr	O
P169a	1bi-1chl-1ga78-Xmu-2pc-7qz-1zr-lop	O

M001a	1bi-1chl-1ga3-1mu-2pc-7qz	O
M004	1bi-1ga8-1ky-4mu-1pc-1qz-Xto-1zr-lop	F
M005	2bi-1chl-1ga7-1ky-2pc-4qz-1sp-1zr-lop	F
M009	1bi-1chl-1ga8-1ky-1mu-4pc-4qz-1to	F
M010a	2bi-Xchl-1ga8-1ky-1mu-3pc-3qz-Xto-Xzr-Xop	F
M021	2bi-1ga-1ky-1mu-2pc-4qz-1sp	F
M023	3bi-1ep-1ga-3mu-1pc-2qz-lop	O
M026	3bi-1ga8-2ky-3mu-3qz-Xzr-lop	F
M028	2bi-Xchl-1mu-4pc-1py?-3qz-1sp-Xzr	O
M031b	2bi-1chl-1mu-2pc-6qz-Xzr-lop	O
M035	3bi-2ga-2pc-3qz-1sp-Xzr-Xop	O
M041	2bi-1ch-Xkf-4pc-4qz-Xto-1zr	O
M054	2bi-1ga8-1ky-3mu-1pc-3qz-lop	F
M057a	1bi-1ga8-1mu-3pc-4qz-1si-1zr	G
M063	2bi-1ga8-2mu-3pc-3qz-1si5-1zr-lop	G
M065	3bi-Xchl-4pc-3qz-Xse-si?-Xzr-Xop	g
M065	2bi-1ga-2mu-2pc-3qz-1si-Xto-Xzr-Xop	G
M081	2bi-1chl-1ga-1ky-1mu-3pc-3qz-1sp-Xst-1zr-1il	D
M083a	2bi-1chl-1mu-3pc-4qz-1sf-1zr	g
M083b	1bi-1chl-1ga-2mu-2pc-3qz-1si-Xop	G
M085	3bi-2mu-2pc6-3qz-1si-Xzr-lop	g
M086	2bi-1chl-Xep-2mu-3pc-3qz-1si-lop	g
M091	2bi-1chl-1ga8-2mu-3pc-3qz-Xzr-lop	O
M096	2bi-1ga-1ky-4pc-1qz-lrt	f
M099	3bi-1ga8-1ky-3mu-2qz-Xto	F
M103	2bi-1chl-1ga-2mu-2pc-4qz-Xst2-lop-Xrt	O
M106	3bi-Xchl-1ga8-1ky-3mu-1pc-2qz-Xzr	F
M108	3bi-Xchl-1ga8-1ky-5mu-1qz-Xrt	F
M110a	3bi-1ga-1ky-3mu-1pc-4qz-Xzr-lop-lrt?	F
M111	1bi-Xchl-ep?-1pc-7qz-1sp-lop	O

M114	3bi-Xchl-2ga-1ky-2mu-3qz-Xsp-Xzr-lop	F
M135	2bi-1ch-Xep-1mu-2pc-7qz-Xop	O
M144	2bi-1ga8-1ky-2mu-1pc-4qz-Xzr-lop-Xrt	F
M150b	2bi-1ch1-1ga-1mu-2pc-6qz-1zr-1op	O
M154b	2bi-1ga8-1mu-2pc-4qz-1si-1zr-lop	G
M154c	1bi-1ga8-1kf-2mu6-2pc-3qz-1si-1zr	H
M158b	2bi6-1ga8-1kf-2mu6-2pc-3qz-2si-Xzr-lop	H
M158c	2bi-1ga8-1mu6-3pc-3qz-1si-Xzr-lop	G
M161a	3bi-2kf-3mu6-2pc-1qz-1si-1to-1zr-lop	H
M164c	3bi-gr?-2mu6-2pc-3qz-1si5-1to-1zr	g
M170a	2bi-1ga8-1mu6-3pc-4qz-1si-1zr-Xop	G
M215a	3bi-Xchl-Xga-Xgr?-2mu-2pc-2qz-2si-Xto-Xzr	G
M233b	4bi-4mu-2pc-2qz-Xsi5-Xzr	g
M255	4bi-1ga3-Xmu-3pc-3qz-Xzr	O
M236	3bi-2mu-1pc-py?-2qz-2si-Xzr	g
M268	2bi-Xchl-Xga8-1mu-2pc-4qz-1si-1zr-lop	G
M275	2bi-1ch1-1ga8-1ky-2mu-4qz-Xsp-Xst-Xzr	D?

F003	1bi-1ch-Xep-5mu-4qz-Xto-Xzr-lop-lil	B
F003b	3bi-3ch-2mu-3qz-Xto-Xzr-lop	B
F003c	1bi-1ch-Xep-3mu-1pc-4qz-Xto-lop	B
F005	1bi-1ch-Xep-6mu-1py-3qz-lto-lzr-lfs	B
F006	2bi-Xep-5mu-1pc-3qz-Xto-Xzr-Xop	O
F009	Xap-1bi-2ch-Xep-4mu-1pc-4qz-Xsp-Xto-Xop	B
F009b	1bi-1ch-6mu-4qz-Xto-Xzr-Xop	B
F010c	1ch-Xep-fs?-4mu-3qz-lil	A
F014	1bi-2ch-3mu-1pc-3qz-Xto-Xzr-lop-Xil	B
F018	2bi-1ch-4mu-pc?-3qz-Xto-Xzr-lop	B
F019	Xap-3bi-1ch-Xep-3mu-3qz-Xto-lzr-lop	B
F023b	1bi-2mu-8qz-Xzr-lop-lca	O
F026	2bi-1ch-4mu-3qz-Xto-lop	B
F028	2bi-1ga-1ky-3mu-2pc-3qz-1st-Xto-Xzr-lil	D
F030	1bi-3cc-1chl-3mu-3qz-Xsp-lop	O
F031a	1bi-1chl-Xep-ht?-1mu-1pc-8qz-Xop	O
F033a	2bi-1chl-1ga-2mu-1pc?-4qz-1st-lzr-lop	O
F033b	3bi-6mu-1pc-1qz-Xsi5-1st-Xto-lop	g
F033c	3bi-1chl-1ga32-1ky-3mu-1pc-3qz-Xsi5-1st-Xto-lzr-lop	D-g
F033e	3bi-1chl-1ga32-1ky-2mu-1pc-4qz-1st-Xto-Xzr-lop	D
F034a	2bi-1ga32-1ht?-1ky-3mu-Xpc?-3qz-1st-lil	D
F039d	3bi-1ga32-1ky-3mu-1pc-3qz-1st-lto-lzr	D
F039e	2bi-1ga32-ht?-1ky-2mu-1pc-4qz-1st-Xto-lzr-lop	D
F040a	3bi-1chl-1ga-3mu-3qz-1st7-Xto-Xzr-lop-lil	O
F041a	2bi-1chl-1ga32-4mu-3qz-1st-Xto-lop-lil	O
F041b	2bi-1ca-1chl-1ga7-1mu-pc?-6qz-lzr-lop	O
F048a	4bi-1ga-1mu-2pc-4qz-Xzr-lop-lil	O
F050a	2bi-1chl-1ga-2mu-3pc-3qz-lsi5-Xst?2-Xzr-lop	G
F050b	3bi-Xchl-1ga4-3mu-2pc-3qz-lsi-Xto-lop	G
F050d	5bi-1chl-1ky-3qz6-1st-lfs-lil	D

F050e	5bi-1ky-3pc-3qz-1st-lop	D
F050f	ap?-4bi-1chl-1ga-1ky-3pc-2qz-1sel-2st-Xop-Xil	F
F050g	5bi-1chl-1ga-ky?-2qz-1sel-2st-lzr-lfs	O
F050h	5bi-1chl-1ky-Xmu-1pc-2qz-2st7-Xzr-lop	D
F050i	8bi-1ch-1ga-1qz-1st2	O
F053b	2bi-Xchl-1ga4-lmu-1pc-lpy-3qz-3si	G
F055a	4bi-1ga-5pc-1qz-1se-1st2-lop	O
F055b	3bi-1ga8-lmu-1pc-4qz-1si-Xto-Xzr-lop	G
F056b	2bi-1chl-1ga4-lmu-1pc-4qz6-1sel-1si-Xto-Xzr-lop	G
F056c	2bi-1ga-2mu-2pc-4qz-1si-Xto-Xzr-lop	G
F057a	3bi-lmu-2pc-2qz-1si-1st-1il	Z
F057b	2bi-1ga-4mu-1pc-3qz-lop	O
F058	2bi-1ga-2mu-3pc-3qz-1si-Xzr-Xop	G
F060a	2bi-lmu-2pc-4qz-Xto-Xzr-Xop	O
F061	3bi-1chl-1ga8-lmu-2pc-3qz-1si-1zr-lop	G
F062a	2bi-1ga4-2mu-2pc-3qz-1si-Xto-Xzr-lop	G
F065a	3bi-1ga4-2mu-1pc-3qz-2si-Xto-Xzr-lop	G
F065b	3bi-1ga4-3mu-1pc-3qz-1si-Xto-Xzr	G
F066a	2bi-1chl-1ga38-lmu-2pc-4qz-1si-Xto-Xzr-lop	G
F068d	1bi-1ch-1ep-2mu-1pc-6qz-1sp-lop	O
F076a	3bi-1ep-2kf-lmu-1pc-5qz-1zr-lop	O
F077a	1bi-1chl-2kf-lmu-6qz-1zr-lop	O
F077b	Xap-3bi-1chl-2kf-3mu-3qz-Xzr-lop	O
F077c	2bi-ep?-2mu-4pc-4qz-lop	O
F078a	3bi-1chl-1ga7-2kf-3mu-3qz-Xzr-1il-lop	O
F078d	3bi-1chl-1ga-3mu-4qz-Xto-Xzr-1il	O
F078e	3bi-1chl-3mu-3qz-1st-lop	O
F078f	3bi-1chl-1ga7-lmu-2pc-4qz-1zr-1il	O
F078g	1bi-1chl-1ga7-lmu-5qz-Xto-lop-lfs-1il	O
F078h	2bi-1chl-1ga-2mu-1pc-3qz-1si-1st-lto-1zr-lop-1il	E

F078i	3bi-1chl-1ga-3mu-3qz-1si5-Xzr-lop	G
F084	3bi-1chl-1ga4-2mu-1pc-3qz-1si-1zr-lop	G
F090b	3bi-3mu-1pc-4qz-1si-1zr-lop-lil	g
F090c	3bi-3mu-1pc-3qz-1si-Xto-lop	g
F092a	3bi-3mu-1pc-4qz-1zr-lop	O
F0981	3bi-1chl-Xga-4mu-3qz-Xto-lil	O
F103b	3bi-1ga7-4mu-3qz-Xzr-lil	O
F110a	1bi-Xchl-Xep-Xga8-1mu-1pc-6qz-1si5-Xzr-lop	G
F113	2bi-1chl-1ga-1mu-3pc-4qz-Xsp-Xzr-lop	O
F120	2bi-1ga4-2mu-Xpc-3qz-1si-Xzr-lop	G
F122	2bi-1chl-1ga7-3qz-Xzr-3fs-lil	O
F122rc	2bi-2chl-3mu-2pc-2qz-1si5-Xzr-lil	g
F128a	2bi-1ga-2mu-1pc-5qz-1si-1zr-lop	G
F139a	2bi-1chl-2ga7-Xmu-2pc-4qz-lil	O
F139d	3bi-1chl-1mu-2pc-4qz-1si-sp?-Xto-Xzr-lop	g
F141b	1bi-1ga47-2mu-2pc-4qz-1si-lto.	G
F149b	3bi-2mu-2pc-3qz-1si5-1zr	g
F151a	2bi-1chl-1ga7-2mu6-2pc-3qz-1si-1zr-lop	g
F151b	2bi-1chl-1ga7-2mu-3pc-2qz-1si5-lop	g
F151c	3bi-1chl-1ga7-2mu-3pc-2qz-1si-Xzr-Xop	g
F151d	2bi-1chl-1ga-1mu-4pc-1qz-1si-Xzr-lop	G
F154b	2bi-1chl-2ga-4pc-1qz-1st-lop	O
F154d	3bi-Xchl-1ga8-2pc-3qz-1si-1st-Xzr-lil	G
F154e	3bi-1ga7-3pc-3qz-Xsi-1st-Xzr-lil	G
F160	3bi-3mu-1pc-3qz-1zr-lop	O
F163	3bi-2mu-1pc-4qz-1si5-Xzr-lop	g
F163d	3bi-Xchl-1ga-1mu-2pc-2qz-2si-1zr-Xop	G

V001a	2ch-1ep-4mu-1pc-3qz-lop-lca	A
V006n	1ch-2mu-1pc-5qz-lop-lca	A
V006nb	1ch-6mu-1pc-2qz-Xto-Xzr-lop	A
V006q	5mu-pc?-4qz-lop	O
V012h	1ch-Xep-2mu-1pc-6qz-Xto-Xop	A
V014a	gr?-7mu-3qz-Xto-il?-lop	O
V024b	1bi-1ch-3mu-4qz-lzr-lop	B
V026c	Xap?-1bi-1ch-3mu-1pc-5qz-lto-Xzr-lop-Xcz	B
V029a	1ch-2mu-5qz-2ca-lop	A
V030b	1ch-lgr?-lkf?-7mu-2qz-il?	A
V031a	Xch-9mu-1qz-Xto-lil	A
V031d	1ch-6mu-2qz-Xzr-lop-lil	A
V032d	1bi-1ch-3mu-4qz-Xzr-lop	B
V032g	1bi-1ch-1mu-5qz-Xzr-lop-fs?	B
V032j	2bi-2ch-7mu-Xqz-Xzr-lop	B
V032l	2bi-3ch-4mu-3qz-Xzr-Xop	B
V032m	1bi-1ch-kf?-1mu-1pc-6qz-Xzr-lop	B
V033b	2ch-3mu-5qz-Xto-lzr-il?-lca	A
V033c	Xbi-2ch-2mu-6qz-Xto-lop	B
V037e	Xep-1mu-2pc-7qz-Xto-Xop-lca	O
V037f	2ch-2mu-1pc-5qz-Xto-Xzr-lop-Xca	A
V038a	1ch-3mu-1pc-6qz-Xto-lop	A
V038a	Xbi-Xch-4mu-6qz-lop	B?
V047b	1bi-1ch-1mu-1pc-7qz-Xzr-lop	B
V048b	2bi-1ch-7mu-lop	B
V048c	2bi-2ch-7mu-pc?-qz?-Xzr-lop-lca	B
V049a	1bi-1ch-1mu-2pc-1qz-Xto-lop	B
V049b	1bi-1ch-Xga7-1mu-1pc-7qz-Xto-lop-lca	c
V051c	1ch-2mu-6qz-Xto-lop-lca	A
V051cb	Xbi?-Xch?-1mu-5qz-lsel-Xto	B?

V051f	1bi-1ch-1mu-1pc-7qz-Xzr-lop	B
V052a	1bi-1ch-4mu-4qz-Xto-Xzr-lop	B
V053a	1ch-6mu-3qz-Xto-lop-lil	A
V058c	3ch-3mu-5pc-qz?-Xto-Xzr-lop	A
V060a	6ch-1mu-3qz-lop-1ca	A
V060b	1ch-3mu-6qz-Xto-lop-lil	A
V060c	1ch-Xep-1mu-1pc-6qz-Xop-2ca	A
V062a	1ch-1mu-2pc-6qz-Xto-lop	A
V068d	Xap-2bi-1ch-1mu-1pc-7qz-lzr-lop	B
V069a	1bi-1ch-4mu-5qz-Xto-lzr-lop-lil	B
V070	1bi-1ch-2mu-6qz-lto-Xzr-lop-lil	B
V071a	2bi-2ch-Xga7-3mu-Xpc-4qz-lzr-lop	c
V073b	Xap-1bi-1ch-Xga7-Xmu-1pc-7qz-lzr-lop	c
V074b	1bi-1ch-Xga-1mu-2pc-6qz-Xzr-lop	c
V075a	2bi-1ch-Xga-2mu-Xpc-6qz-Xto-lzr	c
V075b	1ch-5mu-4qz-lop-lil	O
V078a	1bi-1ch-2mu-5qz-Xto-Xzr-lop	B
V084	2bi-1ch-5mu-1pa5-Xzr-lop (garnet in hand specimen)	C
V084	2bi-1chl-Xga7-5mu-2qz-Xzr-lop	C
V085c	1bi-1ch-1ep-1mu-8qz-lsp-lop	O
V086a	1bi-Xep-1mu-1pc-7qz-Xzr-lop	O
V094	1bi-1chl-Xga-6mu-2pc-Xqz-Xto-Xzr-lop	C
V100c	Xap-2bi-Xep-2mu-2pc-5qz-Xto-lzr-Xop	O
V106a	2bi-1chl-1ga-Xmu-2pc-5qz-lzr-lop	C
V106c	1bi-1chl-Xga7-kf?-3mu-3qz-2st7-lop-lil	d
V108c	1bi-1ch-Xep-Xga7-2mu-5qz-lzr-lop-lil-Xca	C
V114	1bi-2chl-1ga-4mu-1pc-2qz-lto-lzr-lop-lil	C
V115	2bi-1chl-Xga-1mu-1pc-5qz-lzr-lop	C
V120	2bi-1chl-Xga-2mu-1pc-4qz-Xto-Xzr-lop	C
V121	2bi-2chl?-3mu-2qz-Xto-lop	O

V124	1bi-1chl?-lga-1mu-6qz-1st-Xto-Xzr-lfs-lil	d
V125b	2bi-1chl?-lga7-1mu-lpc-4qz-1st5-lzr-lop	d?
V131a	1bi-1chl-lga-Xmu-2pc-6qz-lzr-lop	C
V135a	Xap-1bi-1chl-lga-1mu-2pc-6qz-Xto-Xzr-lop-lil	C
V140b	2bi-1chl?-lga-4mu-Xpc-3qz-1st-lto-lop	d
V142a	2bi-1chl-lga3-Xky5-lpa5-lpc-2qz-1st5-lto-lzr-lop	D
V142b	1bi-1chl?-3mu-lpc-lqz-lto-lop	O
V144a	2bi-1chl-lga4-2mu-Xpc-5qz-1st7-Xto-Xzr-lop	d
V145a	2bi-1chl-lga3-Xky-2mu-3qz-2st-lto-lzr-lop-lil	D
V146b	2bi-Xchl-2mu-6qz-lto-lzr-lop-lfa	O
V148	1bi-Xchl-Xga-1mu-4pc-4qz-Xto-lop	O
V154c	Xap-2bi-Xchl-2mu-lpc?-5qz-lsi5-lto-Xzr-lop	g
V159b	2bi-1chl-Xga-1mu-lpc-5qz-lsi5-Xzr-lop	g
V160a	Xap-2bi-2chl-5mu-lqz-lto-lzr-lop	O
V162a	2bi-1chl-lga7-lky-2mu-lpc-2qz-1st-Xto-Xzr-lop	D
V163b	2bi-1chl-lga7-3mu-2pc-2qz-lsi5-1st-Xto-Xzr-il?	E
V173a	2bi-lga4-2mu-3pc-2qz-2si-lto-lzr-lil	G
V176	2bi-2mu-2pc-4qz-lsi5-lzr-lop	g

APPENDIX B.

CHEMICAL DATA

The chemical data used for geothermometry, geobarometry, and textural studies are contained within this appendix. All analyses were done on the two channel wavelength dispersive Cambridge Instruments Microscan electron probe at Carleton University. Operating conditions were 15 kV and 0.5×10^{-7} A specimen current.

Calcites from the northern Selkirk Mountains are not stable under a focussed beam at 15 kV so, to avoid decomposition of the sample, a defocussed beam was used. The increased surface area coverage accompanying a defocussed beam aided in homogenization of exsolved lamellae of dolomite.

Samples chosen for analysis had to possess as many of the following characteristics as possible:

- (1) euhedral grains (particularly for garnet);
- (2) juxtaposition or close proximity of mineral grains of interest;
- (3) textural equilibrium;
- (4) no optical zoning;
- (5) no retrograde effects.

A list of standards used for analysis is given in Table B-1.

Probe data were processed by a computer program after Rucklidge and Gasparrini (1969):

Numerous analyses that had both good total oxide weight percents and good ion totals were averaged, and these averages were used in subsequent calculations. The only exception to this was for calcite analyses for which the average MgCO_3 value within individual grains was of interest.

The average chemical analyses of garnet, biotite, and plagioclase have been compiled in a manner that expresses the variation of values

about the average. Beside each sample number is either an R (rim) or G (core) followed by a number which represents the number of samples in the average. Following each value for oxide weight percent is a bracketed number which corresponds to the difference between the average weight percent and a value of the average weight percent plus sigma. Sixty-five percent of the analyses lie between average plus sigma and average minus sigma. For example, a value of 36.0 (0.20) means that 65% of all analyses for a particular oxide are between 35.8 and 36.2 wt.%. By this method the variation of chemistry for any phase can be recognized at a glance.

Content of MgCO_3 is variable within individual grains of calcite due to the presence of exsolution lamellae of dolomite. In order to assess the overall MgCO_3 content an average of numerous spot analyses within each grain is necessary. The average of individual grains is tabulated below.

Table B-1. Standards used in electron microprobe chemical analysis

Element	Phase Analysed	Standard	Phase Analysed	Standard	Phase Analysed	Standard	Phase Analysed	Standard
K								
Si		Bi1, Bi3		Evans		Anor, Alb		
Ti		Bi1, Bi3						
Al		Bi1, Bi3		Evans		Anor, Alb		
Fe		Bi1, Bi3		Evans				
Mn	Biotite	R1596	Garnet	R1596	Plagio- clase		Calcite	Boett
Mg		Bi1, Bi3		Evans				Boett
Ca				Evans		Anor		
Na		Jade				Alb		
K		Bi1, Bi3				Phlogo		

Electron Microprobe Chemical Analyses of Biotite on the basis of 22 (O) anhydrous (24 (O) hydrous)

	A87a		R		3		A87a		C		2		A375		R		7		A375		C		6		P95b		R		4	
SiO ₂	36.3	(0.20)					36.2	(0.00)					36.4	(0.40)					36.5	(0.30)					35.8	(0.30)				
TiO ₂	2.04	(0.36)					1.82	(0.16)					1.64	(0.22)					2.06	(0.18)					2.65	(0.06)				
Al ₂ O ₃	19.3	(0.20)					19.3	(0.00)					19.5	(0.40)					18.9	(0.30)					18.5	(0.30)				
FeO	18.5	(0.20)					18.7	(0.30)					19.2	(0.40)					19.2	(0.40)					19.2	(0.60)				
MnO	0.07	(0.01)					0.07	(0.01)					0.06	(0.02)					0.08	(0.01)					0.06	(0.02)				
MgO	10.6	(0.20)					10.7	(0.10)					10.3	(0.30)					10.3	(0.20)					10.3	(0.30)				
Na ₂ O	0.22	(0.05)					0.20	(0.03)					0.16	(0.02)					0.15	(0.02)					0.15	(0.02)				
K ₂ O	9.23	(0.06)					9.32	(0.04)					8.67	(0.16)					8.61	(0.10)					9.28	(0.15)				
Total	96.3						96.3						95.9						95.8						95.9					
Calc H ₂ O	4.01	(0.05)					4.00	(0.02)					4.00	(0.07)					3.99	(0.06)					3.97	(0.07)				
Total	100.3						100.3						99.9						99.8						99.9					
Si	5.423	8.000					5.415	8.000					5.455	8.000					5.481	8.000					5.403	8.000				
Al	2.577						2.585						2.545						2.519						2.597					
Al	0.825						0.815						0.897						0.816						0.687					
Ti	0.229						0.205						0.184						0.233						0.300					
Fe	2.307	5.734					2.336	5.758					2.406	5.786					2.413	5.770					2.422	5.737				
Mn	0.009						0.009						0.008						0.010						0.008					
Mg	2.364						2.393						2.290						2.300						2.320					
Na	0.063	1.821					0.059	1.839					0.046	1.703					0.045	1.693					0.043	1.829				
K	1.758						1.779						1.657						1.648						1.786					
H	4.000						4.000						4.000						4.000						4.000					

Electron Microprobe Chemical Analyses of Biotite on the basis of 22'(O) anhydrous (24 (O) hydrous)

	A345	R	9	A345	C	11	F061	R	10	F061	C	4	P57	R	6
SiO ₂	36.7	(0.30)		36.9	(0.20)		36.8	(0.30)		36.7	(0.30)		35.6	(0.50)	
TiO ₂	1.69	(0.30)		1.86	(0.12)		2.77	(0.27)		2.61	(0.29)		2.83	(0.20)	
Al ₂ O ₃	18.3	(0.20)		18.0	(0.10)		18.5	(0.40)		18.8	(0.50)		18.2	(0.30)	
FeO ₃	18.6	(0.30)		18.7	(0.20)		20.9	(0.20)		20.4	(0.20)		19.9	(1.30)	
MnO	0.09	(0.10)		0.09	(0.10)		0.01	(0.00)		0.02	(0.02)		0.03	(0.02)	
MgO	11.2	(0.20)		11.2	(0.10)		8.76	(0.23)		8.90	(0.35)		9.67	(0.60)	
Na ₂ O	0.52	(0.06)		0.47	(0.09)								0.20	(0.04)	
K ₂ O	8.76	(0.15)		8.84	(0.15)		8.62	(0.14)		8.63	(0.17)		9.29	(0.20)	
Total	95.9			96.1			96.4			96.4			95.7		
Calc H ₂ O	3.99	(0.06)		4.00	(0.04)		3.99	(0.06)		3.99	(0.08)		3.95	(0.10)	
Total ₂	99.9			100.1			100.4			100.4			99.7		
Si	5.505	8.000		5.530	8.000		5.522	8.000		5.505	8.000		5.409	8.000	
Al	2.495			2.470			2.478			2.495			2.591		
Al	0.730			0.700			0.801			0.828			0.671		
Ti	0.191			0.210			0.312			0.294			0.323		
Fe	2.329	5.776		2.336	5.760		2.622	5.698		2.594	5.711		2.521	5.706	
Mn	0.012			0.011			0.001			0.003			0.004		
Mg	2.513			2.504			1.961			1.991			2.186		
Na	0.151	1.827		0.138	1.826			1.651			1.652		0.059	1.857	
K	1.677			1.688			1.651			1.652			1.798		
H	4.000			4.000			4.000			4.000			4.000		

Electron Microprobe Chemical Analyses of Biotite on the basis of 22 (O) anhydrous (24 (O) hydrous)

	P57	C	S	P87b	R	7	P87b	C	S	M63	R	4	M63	C	4
SiO ₂	35.9	(0.10)		34.7	(0.20)		34.8	(0.20)		35.0	(0.20)		35.0	(0.20)	
TiO ₂	2.84	(0.09)		3.10	(0.44)		3.43	(0.21)		3.46	(0.17)		3.50	(0.05)	
Al ₂ O ₃	18.0	(0.10)		17.7	(0.20)		17.4	(0.20)		18.0	(0.10)		18.3	(0.10)	
FeO	19.6	(0.20)		22.9	(0.30)		23.0	(0.30)		21.1	(0.60)		20.7	(0.10)	
MnO	0.03	(0.02)		0.08	(0.01)		0.05	(0.02)		0.04	(0.01)		0.05	(0.02)	
MgO	9.83	(0.14)		7.80	(0.30)		7.59	(0.31)		8.15	(0.32)		8.18	(0.13)	
Na ₂ O	0.22	(0.03)		0.12	(0.03)		0.12	(0.01)		0.16	(0.01)		0.14	(0.15)	
K ₂ O	9.25	(0.14)		9.63	(0.10)		9.67	(0.06)		9.34	(0.15)		9.52	(0.11)	
Total	95.7			96.0			96.1			95.3			95.4		
Calc H ₂ O	3.95	(0.03)		3.88	(0.06)		3.88	(0.05)		3.89	(0.05)		3.90	(0.03)	
Total	99.6			99.9			100.0			99.1			99.3		
Si	5.444	8.000		5.356	8.000		5.377	8.000		5.386	8.000		5.373	8.000	
Al	2.556			2.644			2.623			2.614			2.627		
Al	0.658			0.578			0.543			0.647			0.676		
Ti	0.324			0.360			0.399			0.401			0.404		
Fe	2.482	5.694		2.962	5.704		2.972	5.668		2.717	5.640		2.656	5.615	
Mn	0.004			0.010			0.007			0.006			0.007		
Mg	2.225			1.795			1.748			1.871			1.872		
Na	0.064	1.856		0.035	1.932		0.035	1.941		0.047	1.880		0.042	1.907	
K	1.792			1.897			1.906			1.833			1.864		
H	4.000			4.000			4.000			4.000			4.000		

Electron Microprobe Chemical Analyses of Biotite on the basis of 22 (O) anhydrous (24 (O) hydrous)

	A93a	C 3	A352	R 4	A352	C 4	A238a	R 5	A238a	C 5
SiO ₂	35.6	(0.00)	36.9	(2.10)	36.8	(1.10)	35.2	(0.30)	35.5	(0.20)
TiO ₂	1.60	(0.04)	1.85	(0.29)	1.75	(0.27)	1.95	(0.17)	1.98	(0.15)
Al ₂ O ₃	18.2	(0.10)	19.7	(1.00)	18.6	(0.40)	18.1	(0.10)	18.3	(0.40)
FeO	19.9	(0.10)	18.2	(0.90)	18.4	(0.60)	20.0	(0.40)	20.2	(0.40)
MnO	0.05	(0.01)	0.04	(0.03)	0.05	(0.01)	0.03	(0.00)	0.02	(0.01)
MgO	10.7	(0.10)	10.6	(0.50)	10.4	(0.10)	10.1	(0.30)	10.2	(0.30)
Na ₂ O	0.20	(0.04)	0.15	(0.03)	0.13	(0.05)	0.13	(0.05)	0.11	(0.02)
K ₂ O	9.34	(0.05)	8.07	(1.48)	8.88	(0.06)	9.17	(0.03)	9.35	(0.23)
Total	95.6		95.5		95.0		94.7		95.7	
Calc H ₂ O	3.93	(0.02)	4.02	(0.25)	3.97	(0.11)	3.89	(0.06)	3.94	(0.06)
Total	99.5		99.5		99.0		98.6		99.6	
Si	5.414	8.000	5.500	8.000	5.555	8.000	5.417	8.000	5.406	8.000
Al	2.586		2.500		2.445		2.583		2.594	
Al	0.677		0.952		0.860		0.694		0.690	
Ti	0.183		0.207		0.199		0.225		0.226	
Fe	2.531	5.832	2.262	5.775	2.321	5.717	2.568	5.798	2.574	5.799
Mn	0.006		0.005		0.006		0.004		0.003	
Mg	2.435		2.348		2.331		2.308		2.306	
Na	0.060	1.875	0.044	1.577	0.039	1.748	0.039	1.838	0.032	1.847
K	1.815		1.533		1.708		1.799		1.815	
H	4.000		4.000		4.000		4.000		4.000	

Electron Microprobe Chemical Analyses of Biotite on the basis of 22 (O) anhydrous (24 (O) hydrous)

	A300		A300		A350		A350		A93a	
	R	4	C	4	R	4	C	5	R	5
SiO ₂	35.0	(0.20)	34.8	(0.20)	35.8	(0.20)	35.9	(0.50)	35.3	(0.40)
TiO ₂	3.67	(0.05)	3.58	(0.10)	1.39	(0.37)	1.65	(0.20)	1.58	(0.10)
Al ₂ O ₃	18.3	(0.10)	18.3	(0.09)	18.9	(0.40)	18.6	(0.40)	18.1	(0.10)
FeO	21.6	(0.20)	21.9	(0.30)	20.1	(0.50)	20.2	(0.40)	20.5	(1.10)
MnO	0.05	(0.02)	0.05	(0.01)	0.03	(0.01)	0.01	(0.01)	0.05	(0.00)
MgO	7.21	(0.13)	7.23	(0.09)	10.5	(0.10)	10.3	(0.20)	10.6	(0.50)
Na ₂ O	0.13	(0.01)	0.12	(0.01)	0.42	(0.02)	0.43	(0.03)	0.21	(0.04)
K ₂ O	9.31	(0.12)	9.57	(0.07)	8.59	(0.08)	8.51	(0.09)	9.35	(0.14)
Total	95.3		95.6		95.7		95.6		95.7	
Calc H ₂ O	3.89	(0.03)	3.89	(0.03)	3.96	(0.07)	3.95	(0.08)	3.92	(0.08)
Total	99.2		99.5		99.7		99.6		99.6	
Si	5.385	8.000	5.364	8.000	5.416	8.000	5.443	8.000	5.389	8.000
Al	2.615		2.636		2.584		2.557		2.611	
Al	0.714		0.691		0.785		0.759		0.637	
Ti	0.425		0.414		0.158		0.188		0.182	
Fe	2.786	5.589	2.825	5.598	2.543	5.848	2.555	5.823	2.619	5.861
Mn	0.007		0.007		0.004		0.002		0.006	
Mg	1.656		1.660		2.358		2.318		2.418	
Na	0.037	1.867	0.037	1.917	0.123	1.781	0.126	1.771	0.063	1.884
K	1.830		1.880		1.658		1.645		1.820	
H	4.000		4.000		4.000		4.000		4.000	

Electron Microprobe Chemical Analyses of Biotite on the basis of 22 (O) anhydrous (24 (O) hydrous)

	A278	R	4	A284b	R	5	P147	R	2	M144	R	3	M154c	R	5
SiO ₂	39.6	(0.30)	34.9	(0.10)	36.4	(0.00)	35.8	(0.30)	34.8	(0.10)					
TiO ₂	2.55	(0.05)	3.97	(0.10)	1.12	(0.08)	2.10	(0.03)	3.72	(0.16)					
Al ₂ O ₃	18.5	(0.20)	17.8	(0.20)	19.3	(0.30)	19.7	(0.10)	18.9	(0.50)					
FeO	19.6	(0.40)	22.3	(0.30)	18.6	(0.00)	18.1	(0.30)	21.5	(0.20)					
MnO	20.01	(0.82)	0.05	(0.01)	0.04	(0.01)	0.07	(0.01)	0.08	(0.01)					
MgO	10.1	(0.10)	7.01	(0.13)	11.2	(0.05)	11.2	(0.10)	7.69	(0.13)					
Na ₂ O	0.20	(0.01)	0.14	(0.07)	0.26	(0.00)	0.23	(0.00)	0.12	(0.02)					
K ₂ O	9.17	(0.05)	9.82	(0.10)	8.73	(0.01)	8.54	(0.15)	9.65	(0.12)					
Total	95.7		96.0		95.7		95.7		96.5						
Calc H ₂ O	3.95	(0.05)	3.89	(0.03)	4.00	(0.02)	4.01	(0.03)	3.94	(0.04)					
Total	99.7		99.9		99.7		99.7		100.4						
Si	5.389	8.000	5.374	8.000	5.460	8.000	5.355	8.000	5.301	8.000					
Al	2.611		2.626		2.540		2.645		2.699						
Al	0.693		0.606		0.869		0.821		0.698						
Ti	0.291		0.460		0.126		0.236		0.425						
Fe	2.478	5.751	2.877	5.561	2.328	5.834	2.264	5.828	2.738	5.617					
Mn	0.002		0.006		0.006		0.009		0.011						
Mg	2.287		1.612		2.505		2.498		1.745						
Na	0.058	1.831	0.042	1.973	0.076	1.746	0.066	1.693	0.036	1.910					
K	1.773		1.932		1.670		1.627		1.874						
H	4.000		4.000		4.000		4.000		4.000						

Electron Microprobe Chemical Analyses of Biotite on the basis of 22 (O) anhydrous (24 (O) hydrous)

	M170a	R	S	M158b	R	S
SiO ₂	34.8	(0.20)		35.2	(0.10)	
TiO ₂	4.45	(0.12)		3.93	(0.20)	
Al ₂ O ₃	18.1	(0.10)		18.0	(0.50)	
FeO	21.5	(0.20)		20.6	(0.20)	
MnO	0.16	(0.05)		0.26	(0.09)	
MgO	7.56	(0.10)		8.12	(0.10)	
Na ₂ O	0.10	(0.03)		0.05	(0.01)	
K ₂ O	9.73	(0.15)		9.62	(0.14)	
Total	96.4			95.8		
Calc H ₂ O	3.92	(0.04)		3.92	(0.05)	
Total	100.3			99.7		

Si	5.319	8.000	5.379	8.000
Al	2.681		2.621	
Al	0.577		0.628	
Ti	0.512		0.452	
Fe	2.743	5.575	2.635	5.597
Mn	0.021		0.033	
Mg	1.722		1.849	
Na	0.030	1.927	0.016	1.890
K	1.896		1.874	
H	4.000		4.000	

Electron Microprobe Chemical Analyses of Garnet on the basis of 12 (O)

	A345	R	9	A345	C	7	F061	R	5	F061	C	13	P57	R	8
SiO ₂	37.4	(0.20)	37.6	(0.10)	37.0	(0.10)	37.0	(0.10)	37.0	(0.10)	37.0	(0.10)	37.4	(0.10)	
Al ₂ O ₃	21.5	(0.20)	21.6	(0.10)	21.1	(0.30)	21.1	(0.30)	21.1	(0.10)	21.1	(0.10)	21.3	(0.10)	
FeO	33.3	(0.20)	33.2	(0.10)	35.7	(0.20)	35.7	(0.20)	33.6	(0.20)	33.6	(0.20)	35.1	(0.20)	
MnO	1.78	(0.56)	1.12	(0.09)	1.45	(0.39)	1.45	(0.39)	3.71	(0.23)	3.71	(0.23)	1.30	(0.06)	
MgO	4.04	(0.33)	4.63	(0.09)	2.32	(0.11)	2.32	(0.11)	2.48	(0.10)	2.48	(0.10)	3.63	(0.06)	
CaO	1.95	(0.22)	2.11	(0.14)	2.68	(0.16)	2.68	(0.16)	2.28	(0.06)	2.28	(0.06)	1.68	(0.12)	
Total	100.0		100.3		100.3		100.3		100.2		100.2		100.4		
Si	2.986		2.980		2.986		2.986		2.986		2.986		2.990		
Al	2.022		2.022		2.008		2.008		2.008		2.008		2.006		
Fe	2.224		2.203		2.407		2.407		2.266		2.266		2.346		
Mn	0.121	2.994	0.075	3.005	0.099	3.016	0.099	3.016	0.254	3.016	0.254	3.016	0.008	3.010	
Mg	0.481		0.548		0.279		0.279		0.299		0.299		0.432		
Ca	0.167		0.180		0.231		0.231		0.198		0.198		0.144		
Alm	74.3		73.3		79.8		79.8		75.1		75.1		77.9		
Sps	4.0		2.5		3.0		3.0		8.4		8.4		2.9		
Pyr	16.1		18.2		9.2		9.2		9.9		9.9		14.4		
Grs	5.6		6.0		7.7		7.7		6.6		6.6		4.8		

Electron-Microprobe Chemical Analyses of Garnet on the basis of 12 (O)

	PS7	C	7	P87b	R	6	P87b	C	6	M63	R	5	M63	C	5
SiO ₂	37.5	(0.10)	37.0	(0.00)	37.0	(0.20)	36.7	(0.30)	36.8	(0.10)	36.8	(0.10)	36.8	(0.10)	36.8
Al ₂ O ₃	21.3	(0.10)	21.1	(0.10)	21.1	(0.00)	21.3	(0.10)	21.3	(0.10)	21.3	(0.10)	21.3	(0.10)	21.3
FeO	34.5	(0.10)	36.2	(0.30)	35.9	(0.10)	35.6	(0.50)	35.1	(0.30)	35.1	(0.30)	35.1	(0.30)	35.1
MnO	1.71	(0.11)	1.75	(0.08)	1.82	(0.03)	1.30	(0.03)	1.71	(0.15)	1.71	(0.15)	1.71	(0.15)	1.71
MgO	3.98	(0.03)	2.82	(0.06)	3.05	(0.16)	3.51	(0.08)	3.64	(0.11)	3.64	(0.11)	3.64	(0.11)	3.64
CaO	1.58	(0.05)	1.44	(0.02)	1.50	(0.09)	1.21	(0.02)	1.22	(0.04)	1.22	(0.04)	1.22	(0.04)	1.22
Total	100.6		100.3		100.4		99.6		99.8		99.8		99.8		99.8
Si	2.989		2.984		2.984		2.966		2.970		2.970		2.970		2.970
Al	1.998		2.003		1.998		2.031		2.022		2.022		2.022		2.022
Fe	2.299		2.443		2.415		2.405		2.366		2.366		2.366		2.366
Mn	0.116	3.023	0.124	3.026	0.124	3.035	0.089	3.021	0.117	3.025	0.117	3.025	0.117	3.025	0.117
Mg	0.473		0.339		0.366		0.423		0.437		0.437		0.437		0.437
Ca	0.135		0.119		0.129		0.105		0.105		0.105		0.105		0.105
Alm	76.0		80.7		79.6		79.6		78.2		78.2		78.2		78.2
Sps	3.8		4.0		4.1		2.9		3.9		3.9		3.9		3.9
Pyr	15.7		11.2		12.1		14.0		14.5		14.5		14.5		14.5
Grs	4.5		4.1		4.3		3.5		3.5		3.5		3.5		3.5

Electron Microprobe Chemical Analyses of Garnet on the basis of 12 (O)

	A300	R	3	A300	C	4	A350	R	4	A350	C	5	A93a	R	3
SiO ₂	36.4	(0.20)	36.6	(0.10)	37.0	(0.10)	36.9	(0.20)	36.6	(0.40)					
Al ₂ O ₃	20.7	(0.20)	20.9	(0.00)	21.0	(0.10)	20.8	(0.20)	20.9	(0.20)					
FeO	36.4	(0.10)	35.3	(0.60)	34.2	(0.60)	34.3	(0.60)	33.6	(0.50)					
MnO	2.03	(0.04)	2.44	(0.36)	0.23	(0.14)	1.00	(0.80)	1.88	(0.17)					
MgO	2.37	(0.14)	2.79	(0.06)	3.76	(0.22)	2.34	(0.63)	2.36	(0.16)					
CaO	1.24	(0.03)	1.28	(0.05)	2.72	(0.51)	3.94	(0.50)	3.06	(0.68)					
Total	99.1		99.3		98.9		99.3		98.4						
Si	2.982		2.980		2.990		2.994		2.994						
Al	2.003		2.006		2.002		1.992		2.017						
Fe	2.491		2.409		2.311		2.327		2.300						
Mn	0.141	3.031	0.169	3.029	0.016	3.016	0.069	3.022	0.130	2.986					
Mg	0.289		0.339		0.454		0.283		0.287						
Ca	0.109		0.112		0.236		0.343		0.268						
Alm	82.2		79.6		76.6		77.0		77.0						
Sps	4.7		5.6		0.5		2.3		4.4						
Pyr	9.6		11.2		15.0		9.4		9.6						
Grs	3.6		3.7		7.8		11.3		9.0						

Electron Microprobe Chemical Analyses of Garnet on the basis of 12 (O)

	A93a	C	4	A352	R	6	A352	C	4	A238a	R	10	A238a	C	6
SiO ₂	36.5	(0.20)		37.3	(0.20)		36.6	(0.20)		36.9	(0.10)		37.0	(0.10)	
Al ₂ O ₃	20.8	(0.20)		21.4	(0.10)		21.1	(0.10)		20.8	(0.10)		20.7	(0.20)	
- FeO	32.6	(0.90)		32.9	(0.30)		33.1	(0.50)		33.1	(0.30)		31.0	(0.60)	
MnO	2.65	(0.96)		0.25	(0.07)		2.03	(0.57)		2.23	(0.15)		2.64	(1.07)	
MgO	1.73	(0.13)		3.71	(0.45)		1.47	(0.05)		2.58	(0.10)		2.48	(0.44)	
CaO	4.47	(0.13)		4.51	(0.59)		5.84	(0.19)		3.19	(0.63)		4.84	(1.36)	
Total	98.8			100.1			100.2			98.8			98.7		
Si	2.985			2.970			2.963			3.002			3.006		
Al	2.007			2.013			2.010			1.998			1.983		
Fe	2.229			2.195			2.236			2.252			2.107		
Mn	0.184	3.017		0.017	3.039		0.139	3.058		0.154	2.998		0.182	3.011	
Mg	0.212			0.441			0.177			0.313			0.301		
Ca	0.392			0.386			0.506			0.279			0.422		
Alm	73.9			70.3			73.1			75.1			70.0		
Sps	6.1			6.6			4.6			5.1			6.0		
Pyr	7.0			14.5			5.8			10.5			10.0		
Grs	13.0			12.7			16.5			9.3			14.0		

Electron Microprobe Chemical Analyses of Garnet on the basis of 12 (O)

	A87a	R	3	A87a	C	4	A375	R	6	A375	C	4	P95b	R	2
SiO ₂	37.0	(0.10)		37.1	(0.30)		37.2	(0.20)		37.1	(0.20)		37.2	(0.10)	
Al ₂ O ₃	21.1	(0.30)		20.7	(0.20)		21.5	(0.20)		21.5	(0.20)		21.2	(0.10)	
FeO	32.5	(0.20)		29.5	(1.10)		33.9	(0.30)		33.1	(0.50)		33.6	(0.00)	
MnO	2.23	(0.18)		3.68	(1.56)		0.44	(0.03)		1.38	(0.95)		1.78	(0.01)	
MgO	2.45	(0.11)		1.41	(0.15)		4.41	(0.17)		4.00	(0.53)		3.48	(0.02)	
CaO	3.47	(0.25)		6.48	(0.13)		1.35	(0.22)		1.82	(0.22)		1.73	(0.01)	
Total	98.8			98.9			98.8			98.9			99.0		
Si	3.008			3.015			2.993			2.988			3.006		
Al	2.016			1.986			2.035			2.041			2.021		
Fe	2.205			2.002			2.285			2.231			2.265		
Mn	0.153	2.958		0.253	2.990		0.030	2.961		0.094	2.962		0.122	2.956	
Mg	0.297			0.170			0.529			0.480			0.419		
Ca	0.302			0.564			0.116			0.157			0.150		
Alm	74.6			67.0			77.2			75.3			76.6		
Sps	5.2			8.5			1.0			3.2			4.1		
Pyr	10.0			5.7			17.9			16.2			14.2		
Grs	10.2			18.9			3.9			5.3			5.1		

Electron Microprobe Chemical Analyses of Garnet on the basis of 12 (O)

	A278	R	4	A284b	R	3	P147	R	6	M144	R	3	M154c	R	5
SiO ₂	37.3	(0.10)		36.7	(0.20)		37.5	(0.10)		37.1	(0.20)		36.6	(0.10)	
Al ₂ O ₃	21.1	(0.10)		21.6	(0.20)		21.6	(0.20)		21.3	(0.10)		20.8	(0.10)	
FeO	34.3	(0.20)		34.4	(0.50)		33.2	(0.20)		32.2	(0.30)		34.8	(0.40)	
MnO	1.53	(0.15)		2.56	(0.08)		0.83	(0.07)		2.63	(0.22)		2.75	(0.17)	
MgO	2.95	(0.09)		2.63	(0.11)		4.75	(0.09)		3.46	(0.17)		2.37	(0.09)	
CaO	2.34	(0.08)		1.41	(0.02)		1.46	(0.16)		2.09	(0.09)		1.35	(0.05)	
Total	99.5			99.3			99.3			98.8			98.7		
Si	3.007			2.977			2.995			3.000			3.003		
Al	2.008			2.064			2.032			2.028			2.006		
Fe	2.312			2.333			2.214			2.179			2.386		
Mn	0.105	2.973		0.176	2.950		0.056	2.960		0.180	2.956		0.191	2.985	
Mg	0.355			0.318			0.565			0.416			0.289		
Ca	0.202			0.122			0.125			0.181			0.119		
Alm	77.8			79.1			74.8			73.7			79.9		
Sps	3.5			6.0			1.9			6.1			6.4		
Pyr	11.9			10.8			19.1			14.1			9.7		
Grs	6.8			4.2			4.2			6.1			4.0		

Election Microprobe Chemical Analyses of Garnet on the basis of 12 (O)

	M170a	R	2	M158b	R	4
SiO ₂	36.9	(0.10)		37.2	(0.20)	
Al ₂ O ₃	20.6	(0.20)		20.7	(0.20)	
FeO	33.6	(0.10)		32.5	(0.40)	
MnO	4.22	(0.10)		5.56	(0.49)	
MgO	2.69	(0.00)		2.26	(0.10)	
CaO	0.83	(0.01)		1.10	(0.02)	
Total	98.8			99.1		
Si	3.016			3.029		
Al	1.986			1.988		
Fe	2.295			2.203		
Mn	0.292	2.988		0.384	2.958	
Mg	0.328			0.275		
Ca	0.073			0.096		
Alm	76.8			74.5		
Sps	9.8			13.0		
Pyr	11.0			9.3		
Grs	2.4			3.2		

Electron Microprobe Chemical Analyses of Plagioclase on the basis of 8 (O)

	A345	R	10	A345	C	6	F061	R	5	F061	C	10	P57	R	S
SiO ₂	63.8	(0.70)	63.9	(0.80)	58.5	(0.40)	58.4	(0.10)	44.9	(0.30)			44.9	(0.30)	
Al ₂ O ₃	23.3	(0.50)	23.0	(0.50)	26.5	(0.20)	26.3	(0.20)	23.8	(0.30)			23.8	(0.30)	
CaO	4.45	(0.40)	4.27	(0.43)	8.40	(0.27)	8.42	(0.14)	4.99	(0.19)			4.99	(0.19)	
Na ₂ O	8.87	(0.27)	8.93	(0.20)	6.65	(0.26)	6.91	(0.12)	8.91	(0.10)			8.91	(0.10)	
K ₂ O	0.11	(0.04)	0.12	(0.02)	0.08	(0.01)	0.10	(0.01)	0.10	(0.00)			0.10	(0.00)	
Total	100.5		100.2		100.1		100.1		100.7				100.7		
Si	2.802	4.005	2.812	4.005	2.610	4.002	2.607	3.994	2.765	4.000			2.765	4.000	
Al	1.203		1.193		1.393		1.386		1.235				1.235		
Ca	0.209		0.202		0.402		0.403		0.235				0.235		
Na	0.756	0.971	0.762	0.971	0.575	0.981	0.598	1.007	0.760	1.000			0.760	1.000	
K	0.006		0.007		0.005		0.006		0.005				0.005		
Ab	77.8		78.5		58.6		59.4		76.0				76.0		
An	21.6		20.8		40.9		40.0		23.5				23.5		
Or	0.6		0.7		0.5		0.6		0.5				0.5		

Electron Microprobe Chemical Analyses of Plagioclase on the basis of 8 (O)

	P57	C	6	P87b	R	3	P87b	C	5	M63	R	4	M63	C	4
SiO ₂	62.9	(0.50)		62.9	(0.10)		63.0	(0.40)		62.7	(0.40)		62.6	(0.40)	
Al ₂ O ₃	23.5	(0.20)		23.2	(0.10)		23.4	(0.20)		23.4	(0.10)		23.9	(0.20)	
CaO	4.72	(0.05)		4.46	(0.13)		4.59	(0.14)		3.43	(0.07)		3.70	(0.23)	
Na ₂ O	9.01	(0.07)		9.05	(0.07)		9.02	(0.08)		9.82	(0.16)		9.47	(0.19)	
K ₂ O	0.12	(0.02)		0.18	(0.02)		0.20	(0.03)		0.20	(0.04)		0.26	(0.04)	
Total	100.3			99.8			100.2			99.6			99.9		
Si	2.776	3.999		2.788	3.999		2.783	3.999		2.784	4.010		2.770	4.017	
Al	1.223			1.212			1.217			1.226			1.247		
Ca	0.223			0.212			0.217			0.163			0.175		
Na	0.770	1.001		0.778	1.000		0.772	1.000		0.845	1.020		0.813	1.003	
K	0.007			0.010			0.011			0.012			0.015		
Ab	77.0			77.8			77.2			82.9			81.0		
An	22.3			21.2			21.7			16.0			17.5		
Or	0.7			1.0			1.1			1.1			1.5		

Electron Microprobe Chemical Analyses of Plagioclase on the basis of 8 (O)

	A352	R	6	A352	C	5	A238a	R	6	A238a	C	6	A87a	R	3
SiO ₂	55.2	(0.50)		54.8	(0.20)		53.9	(0.70)		53.2	(0.90)		61.7	(0.10)	
Al ₂ O ₃	28.4	(0.40)		28.9	(0.20)		29.5	(0.70)		30.2	(0.60)		24.5	(0.30)	
CaO	10.6	(0.30)		11.0	(0.20)		11.7	(0.90)		12.5	(0.70)		6.59	(0.14)	
Na ₂ O	5.51	(0.48)		5.29	(0.10)		4.88	(0.44)		4.48	(0.43)		8.03	(0.16)	
K ₂ O	0.11	(0.01)		0.10	(0.01)		0.07	(0.02)		0.08	(0.02)		0.08	(0.01)	
Total	99.8			100.1			100.1			100.5			100.9		
Si	2.488	4.000		2.469	4.000		2.432	3.999		2.396	4.000		2.717	3.989	
Al	1.511			1.531			1.568			1.603			1.272		
Ca	0.512			0.531			0.568			0.604			0.311		
Na	0.482	1.000		0.462	0.999		0.428	1.000		0.392	1.000		0.686	1.002	
K	0.006			0.006			0.004			0.005			0.004		
Ab	48.2			46.3			42.8			39.2			68.5		
An	51.2			53.2			56.8			60.4			31.1		
Or	0.6			0.6			0.4			0.5			0.4		

Electron Microprobe Chemical Analyses of Plagioclase on the basis of 8 (O)

	A87a	C	2	A375	R	4	A375	C	3	P95b	R	2	A278	R	3
SiO ₂	62.5	(0.10)	64.1	(0.20)	64.3	(0.30)	61.4	(0.90)	59.8	(0.20)			59.8	(0.20)	
Al ₂ O ₃	24.2	(0.10)	22.9	(0.30)	23.1	(0.30)	24.7	(0.40)	26.7	(0.20)			26.7	(0.20)	
CaO	6.22	(0.10)	4.62	(0.04)	4.47	(0.07)	4.83	(0.12)	7.13	(0.16)			7.13	(0.16)	
Na ₂ O	8.03	(0.06)	9.28	(0.05)	9.26	(0.06)	8.89	(0.03)	7.61	(0.05)			7.61	(0.05)	
K ₂ O	0.07	(0.01)	0.10	(0.02)	0.12	(0.03)	0.14	(0.01)	0.07	(0.01)			0.07	(0.01)	
Total	101.0		101.0		101.3		100.0		101.3				101.3		
Si	2.742	3.994	2.806	3.988	2.807	3.994	2.723	4.015	2.632	4.014			2.632	4.014	
Al	1.252		1.183		1.187		1.292		1.382				1.382		
Ca	0.293		0.217		0.209		0.229		0.336				0.336		
Na	0.684	0.981	0.788	1.011	0.784	1.000	0.764	1.002	0.649	0.989			0.649	0.989	
K	0.004		0.006		0.006		0.008		0.004				0.004		
Ab	69.7		78.0		78.4		76.3		65.6				65.6		
An	29.9		21.5		20.9		22.9		34.0				34.0		
Or	0.4		0.6		0.7		0.8		0.4				0.4		

Electron Microprobe Chemical Analyses of Plagioclase on the basis of 8 (O)

	A284b	R	3	P147	R	3	M144	R	4	M154c	R	5	M170a	R	3
SiO ₂	61.3	(0.20)	62.7	(0.20)	63.1	(0.20)	63.7	(0.30)	65.9	(0.40)	65.9	(0.40)	65.9	(0.40)	65.9
Al ₂ O ₃	24.6	(0.20)	24.7	(0.20)	23.5	(0.00)	23.2	(0.20)	22.6	(0.18)	22.6	(0.18)	22.6	(0.18)	22.6
CaO	4.52	(0.17)	4.91	(0.07)	4.92	(0.20)	4.56	(0.18)	2.75	(0.15)	2.75	(0.15)	2.75	(0.15)	2.75
Na ₂ O	9.12	(0.08)	8.99	(0.05)	9.42	(0.20)	9.54	(0.14)	9.95	(0.10)	9.95	(0.10)	9.95	(0.10)	9.95
K ₂ O	0.25	(0.01)	0.15	(0.03)	0.13	(0.01)	-0.25	(0.01)	0.30	(0.03)	0.30	(0.03)	0.30	(0.03)	0.30
Total	99.8		101.5		101.1		101.3		101.5		101.5		101.5		101.5
Si	2.727	4.014	2.738	4.011	2.771	3.985	2.790	3.986	2.857	4.011	2.857	4.011	2.857	4.011	2.857
Al	1.287		1.273		1.214		1.195		1.153		1.153		1.153		1.153
Ca	0.215		0.230		0.231		0.214		0.128		0.128		0.128		0.128
Na	0.786	1.015	0.761	0.999	0.802	1.041	0.810	1.038	0.837	0.982	0.837	0.982	0.837	0.982	0.837
K	0.014		0.008		0.007		0.014		0.016		0.016		0.016		0.016
Ab	77.4		76.2		77.1		78.1		85.3		85.3		85.3		85.3
An	21.2		23.0		22.2		20.6		13.0		13.0		13.0		13.0
Or	1.4		0.8		0.7		1.3		1.7		1.7		1.7		1.7

Electron Microprobe Chemical Analyses of Plagioclase on the basis of 8 (O)

	M158b	R	2
SiO ₂	64.3	(0.30)	
Al ₂ O ₃	22.6	(0.40)	
CaO	4.12	(0.06)	
Na ₂ O	9.49	(0.06)	
K ₂ O	0.25	(0.01)	
Total	100.8		
Si	2.821	3.991	
Al	1.170		
Ca	0.194		
Na	0.808	1.016	
K	0.014		
Ab	79.5		
An	19.1		
Or	1.4		

Electron Microprobe Chemical Analyses of Calcite on the basis of 6 (0)

M64d

F74

	(1)	(2)	(3)	(4)	(1)	(2)	(3)	(4)	(5)	(6)
CaO	50.1	50.3	49.5	51.1	52.3	52.3	53.3	54.0	52.7	52.5
MgO	2.90	2.74	2.99	2.77	2.62	2.36	2.33	2.06	2.13	2.15
FeO	0.68	0.60	0.60	0.64	nd	nd	nd	nd	nd	nd
MnO	0.23	0.20	0.21	0.22	nd	nd	nd	nd	nd	nd
CO ₂	43.9	43.9	43.9	43.9	43.9	43.6	44.4	44.6	43.6	43.5
Total	97.8	97.7	97.2	98.6	98.8	98.3	100.0	100.7	98.4	98.2
Ca	1.802	1.811	1.787	1.830	1.870	1.883	1.886	1.899	1.894	1.893
Mg	0.145	0.137	0.150	0.138	0.130	0.117	0.115	0.101	0.107	0.108
Fe	0.019	0.017	0.017	0.018	-	-	-	-	-	-
Mn	0.007	0.006	0.006	0.006	-	-	-	-	-	-
C	2.014	2.015	2.020	2.004	2.000	2.000	2.000	2.000	2.000	2.000

Mole%
MgCO₃

7.3 7.0 7.7 6.9

5.4

nd = not detected

Electron Microprobe Chemical Analyses of Calcite on the basis of 6 (0)

V98a

	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)
CaO	52.1	52.4	52.1	52.3	51.8	51.3	53.5	51.1
MgO	2.08	1.38	1.90	1.84	1.85	2.30	1.27	2.20
FeO	0.09	0.09	0.10	0.10	0.12	0.10	0.13	0.13
MnO	0.26	0.26	0.25	0.23	0.26	0.23	0.24	0.24
CO ₂	43.9	43.9	43.9	43.9	43.9	43.9	43.9	43.9
Total	98.4	98.0	98.3	98.4	97.9	97.9	99.0	97.6
Ca	1.869	1.888	1.872	1.877	1.865	1.845	1.916	1.842
Mg	0.104	0.069	0.095	0.092	0.093	0.115	0.064	0.111
Fe	0.003	0.003	0.003	0.003	0.003	0.004	0.004	0.004
Mn	0.007	0.007	0.007	0.007	0.008	0.007	0.007	0.007
C	2.009	2.017	2.012	2.012	2.016	2.015	2.005	2.019

Mole%
MgCO₃

5.3 3.5 4.8 4.7 4.7 5.8 3.2 5.7

APPENDIX C.

GARNET ZONING PROFILES

118° 30'
52° +

Maximum elevation of McNaughton Lake (reservoir contained by Mica Dam)

331
332
333
334
335
336
337
338
339
340
350
351
352
353
354
277
278
279
280

40f

118°
+ 52°

506

80F

SYMBOL

MAP 1

STATION LOCATION MAP

906

SYMBOL

PREFIX

COLLECTOR

7

F

J P Franzen

V

MAP 1

LOCATION MAP

1005

COLLECTOR

DATE

J P Franzen

1973

118°
+ 51°45'

140F

F

J P Franzen

V

J van der Leeden

P

M Perkins & D Shaw

M

M Perkins & R Leatherbarrow (in part)

A

R Leatherbarrow

Contour Interval 1000 ft

119°
+ 52° 117°
+ 52°

150F

J P Franzen

1973

J van der Leeden

1974

M Perkins & D Shaw

1974

M Perkins & R Leatherbarrow (in part)

1975

R Leatherbarrow

1977

Contour Interval 1000 ft

mi

0

1

2

3

1

0

1

2

3

4

km

160F

118°30'

1°45'

7000

5000

3000

[Handwritten signature]

170F

180F

190F

118°
+ 51°45'

119°
+ 51°

2000

21 of 21

Fig. 1

1-2 Location map of selected geologic and geographic features
Selkirk Mountains and adjacent areas

[308]

geographic features in the

2

North Th

Monashee Mountains

52° + 119°

Ruddoch
Creek

Columbia River Valley Falls

40f

ms

60F

118°
+52°

Fault

7015

mi

km

Selkirk Mountains

Downie Creek

Revelstoke

118°
+51°

Fan Structure

Esplanade
Range

Trust

Trans-Canada Highway

118°
+51°

9 of 9

105

118°30'
52°

20F

Columbia River

Maximum elevation of McNaus

McNaughton Lake (reservoir contained by Mica Dam)

1 of 1

118°
+52°

MAP 2

METAMORPHIC A
NORTHERN SELKIRK
BRITISH CO

1505

MAP 2

METAMORPHIC MAP OF THE
NORTHERN SELKIRK MOUNTAINS
BRITISH COLUMBIA

METAM NORTHER BR

90F

MI

A

B

C, c, c' garnet, garnet-chlorite

D, d staurolite-kyanite

E staurolite-muscovite

E' staurolite-muscovite

F, f kyanite

G, g sillimanite

H potash

Z sillimanite

F, G mineral assemblages

O assemblage

— — — isograd correspondence

METAMORPHIC MAP OF THE NORTHERN SELKIRK MOUNTAINS BRITISH COLUMBIA

104

MINERAL ASSEMBLAGES

chlorite

chlorite - biotite

garnet, garnet-chlorite, garnet - chlorite - staurolite + biotite

staurolite - kyanite - biotite, staurolite - biotite $\pm$ chlorite

staurolite - muscovite - quartz - kyanite - garnet - biotite

staurolite - muscovite - quartz - sillimanite - garnet - biotite

kyanite - garnet - biotite, kyanite

sillimanite - garnet - biotite, sillimanite

potassic feldspar - sillimanite

sillimanite - staurolite - muscovite - quartz

mineral assemblages, as outlined above, from field notes

assemblage not diagnostic

isograd corresponding to a mineral reaction (approximate, assumed)

- +++++ + + boundary, mark
- mineral or group
- ~~~~~ ~ ~ fault (a
- — — lithologic con

-
-
-
-
-
-
-
-
-
-
-

boundary, mark
mineral or group

fault (a

lithologic con

LITHO

igneous, met

ca

cont

140F

— + — boundary marking the appearance or disappearance of a mineral or group of minerals (approximate, assumed)

— fault (approximate assumed)

— lithologic contact (approximate, assumed)

LITHOLOGIC UNITS

amphibolite

igneous, meta-igneous or pegmatite

marble

impure marble

pelite

semipelite

calcareous pelite

psammite

quartzite

icefield

contour interval 1000 ft

A geological map showing a central 'CHLORITE ZONE' outlined by a dashed line. The zone is intersected by several linear features: a thick solid line with cross-ticks, a dashed line with cross-ticks, and a solid line with cross-ticks. Labels 'A' and 'B' are scattered within the zone, often with subscripts like 'AA', 'BB', 'Aa', and 'Ab'. Outside the zone, labels include 'C', 'D', 'E', 'G', 'H', 'I', 'J', 'K', 'L', 'M', 'N', 'O', 'P', 'Q', 'R', 'S', 'T', 'U', 'V', 'W', 'X', 'Y', 'Z', 'AA', 'BB', 'CC', 'DD', 'EE', 'FF', 'GG', 'HH', 'II', 'JJ', 'KK', 'LL', 'MM', 'NN', 'OO', 'PP', 'QQ', 'RR', 'SS', 'TT', 'UU', 'VV', 'WW', 'XX', 'YY', 'ZZ'. A scale bar at the top left indicates '30'. A large, dark, irregular shape is located in the upper right. A wavy line at the bottom is labeled '160F'. The word 'GARN' is partially visible on the right side.

[illegible]

A geological map showing a central 'CHLORITE ZONE' outlined by a dashed line. The zone is intersected by several thick, solid black lines representing faults or structural boundaries. Some of these lines have small perpendicular tick marks. The map includes various labels: 'A', 'AA', 'Aa', 'AB?', 'ABB?', 'Bc', 'BB', 'B', 'A', 'B', 'C', 'CC', 'C', 'D', 'E', 'd', 'D', 'P', 'G', 'ST', 'CH', 'GA', 'B', 'GARN', 'CHLORITE', and '160F'. There are also small circles and a wavy line at the bottom left. A scale bar at the top left indicates '30'.

A geological map showing a central 'CHLORITE ZONE' outlined by a dashed line. The zone is intersected by several thick, solid black lines representing faults or structural boundaries. Some of these lines have small perpendicular tick marks. The map includes various labels: 'A', 'AA', 'Aa', 'AB?', 'ABB?', 'Bc', 'BB', 'B', 'A', 'B', 'C', 'CC', 'C', 'D', 'E', 'd', 'D', 'P', 'G', 'ST', 'CH', 'GA', 'B', 'GARN', 'CHLORITE', and '160F'. There are also small circles and a wavy line at the bottom left. A scale bar at the top left indicates '30'.

[illegible]

psammite

quartzite

icefield

contour interval 1000 ft

Lithology from:

Franzen 1974

van der Leeden 1976

Perkins, in progress

200P20

MAP 3. METAMORPHIC MAP OF MONASHEE MOUNTAINS

MAP OF THE NORTHERN SELKIRK MOUNTAINS, MOUNTAINS AND ADJACENT PARK RANGE

ROCKY
MTS

20F

N SELKIRK MOUNTAINS, CENT-PARK RANGE

PARK RANGE
(Craw 1978)

4 4
OF / DE

MONASHEE MTS

COLUMBIA RIVER
VALLEY FAULT

COLUMBIA RIVER

SELKIRK

MT SONA
630

ST-B
CH/CH

404

WINDY RANGE

FRENCH CREEK SECTION

60F6

1 of

FRENCH CREEK SECTION

4 kbar

LEGEND

20F

308

WINDY RANGE SECTION

40f

PARK RANGE SECTION

LEGEND

- isograd
- - - isotherm
- . - isobar

Fig. 4-7a Cross section of the n

60F

the northern Selkirk Mountains and adjacent Park Range illustrated

7015

ustrating distributions of isotherms, isobars and isograds at t

at the peak of metamorphism. Temperature profiles were const

constructed using temperatures highlighted on figure by stars.

FRENCH CREEK SECTION

LEGEND

- isograd
- - - isotherm
- - - isobar

20F

30F

WINDY RANGE SECTION

4 of 1

PARK RANGE SECTION

4 kbar

5

st-mu-qz
ga-bi-ky

6

ch-st-mu
ky-bi

500°

si ky

600°

7

0

LEGEND

- isograd
- - - isotherm
- - - isobar

Fig. 4-7b Cross section of the

50F

of the northern Selkirk Mountains and Park Range illustrating

ating distributions of isotherms, isobars and isograds after D

D₃

1 of

FRENCH CREEK SECTION

4 kbar

20F

WINDY RANGE SECTION

30F

ION

PARK RANGE

5kbar

136

6

7

st - mu - qz
99 - bi

500°

404

PARK RANGE SECTION

4kbar

EROSION SURFACE

500°

mu-qz

ga-bi-ky ^{550°} ~~st-mu-qz~~ ^{ky} ~~si~~

Fig. 4-7c Cross section of the

of the northern Selkirk Mountains and Park Range illustra

trating distribution of isotherms, isobars and isograds after

70F

, isobars and isograds after faulting

END

02H09H82

FIN