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STRUCTURAL GEOL OGY AND STRATIGRAPHY.

BIFURCATION OF THE COLUMBIA RIVER,

SILKIRK MOUNTAINS, B.C.

by

C M ichael John Perkins, B.Sc.

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

Carleton University
Ottawa, Ontario
September, 1983
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The undersigned hereby recommend to the Department of Geology acceptance of this thesis, submitted by Michael John Perkins (B.Sc.), in partial fulfillment of the requirements for the degree of Doctor of Philosophy.

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ABSTRACT

The Hadrynian Horsethief Creek Group that underlies much of the Northern Selkirk Mountains is readily divisible into four map units. The grit unit at the base is correlated to the Lower Grit to the southwest. The overlying semipelitic-amphibolite and marble units are correlated to the lower pelitic and middle marble units respectively. The pelite unit which is correlated to the upper pelitic member is overlain by latest Hadrynian to lower Cambrian rocks of the Hamill Group. The four units correlate with the regionally recognized grit, slate, carbonate and upper clastic divisions of the Horsethief Creek Group. In the southwestern part of the map area, thick carbonates and associated clastics probably correlate to the Badsheet Formation, Lardeau Group and Mohecan Formation of the Hamill Group.

Three phases of folding are recognized. Mesoscopic isoclinal folds of the first phase are rarely preserved, but the presence of a major first phase nappe is confirmed. The regional foliation is axial planar to tight second phase folds. Third phase folds are more upright and open and crenulate the second phase foliation northwest and southwest of a central zone where second and third phase folding is coplanar.

Metamorphic grade increases from the staurolite-out isoograd to the breakdown of muscovite at the second sillimanite isoograd. The map area straddles the metamorphic culmination. The peak of metamorphism and the main growth of porphyroblasts postdate cessation of second phase folding and predate the third phase.
Analysis of the inter-relationships between macrofabrics of the second and third phase of deformation indicates a three-fold zonation. Planar fabrics of the second phase characterise the central zone. To the north and south the regional foliation is crenulated by strong third phase folding. The same fold features are recognized on a mesoscopic and macroscopic scale and have major implications for the structural evolution of the Northern Selkirks.

Underthrusting from the west resulted in westwardly verging second phase folds. Away from the zone of underthrusting, axial surfaces became more upright to slightly overturned as the effects of second phase strain weakened and folds died out. A reversal of underthrusting direction during phase three is probably related to overriding of the supracrustal rocks of the foreland to the east, by the Selkirk Allochthon. Northeasterly overturned third phase folds became upright to the south. The three-fold zonation of fold interference at the Selkirk Fan Axis is the result of superposition of third phase upon variably oriented second phase axial surfaces. A period of crustal relaxation followed third phase folding during which major stratigraphic offsets developed across listric normal faults.
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# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>ABSTRACT</td>
<td>i</td>
</tr>
<tr>
<td>ACKNOWLEDGEMENTS</td>
<td>iii</td>
</tr>
<tr>
<td>TABLE OF CONTENTS</td>
<td>v</td>
</tr>
<tr>
<td>LIST OF FIGURES</td>
<td>viii</td>
</tr>
<tr>
<td>LIST OF PLATES</td>
<td>x</td>
</tr>
<tr>
<td>LIST OF MINERAL ABBREVIATIONS</td>
<td>xi</td>
</tr>
<tr>
<td>1 INTRODUCTION</td>
<td>1</td>
</tr>
<tr>
<td>1.1 Purpose and Scope</td>
<td>1</td>
</tr>
<tr>
<td>1.2 Location and Access</td>
<td>3</td>
</tr>
<tr>
<td>1.3 Previous Work</td>
<td>4</td>
</tr>
<tr>
<td>1.4 Regional Geology</td>
<td>6</td>
</tr>
<tr>
<td>1.5 Tectonic Models</td>
<td>9</td>
</tr>
<tr>
<td>2 STRATIGRAPHY</td>
<td>12</td>
</tr>
<tr>
<td>2.1 Introduction</td>
<td>12</td>
</tr>
<tr>
<td>2.2 Stratigraphy of the Horsethief Creek Group</td>
<td>13</td>
</tr>
<tr>
<td>2.2.1 Nomenclature</td>
<td>13</td>
</tr>
<tr>
<td>2.2.2 Stratigraphy in the Dogtooth Range</td>
<td>14</td>
</tr>
<tr>
<td>2.2.3 Stratigraphy in the Selkirk Mountains</td>
<td>15</td>
</tr>
<tr>
<td>2.3 Detail Stratigraphy within the Big Bend</td>
<td>16</td>
</tr>
<tr>
<td>2.3.1 Introduction</td>
<td>16</td>
</tr>
<tr>
<td>2.3.2 Nomenclature</td>
<td>17</td>
</tr>
<tr>
<td>2.3.3 Internal Map Units</td>
<td>19</td>
</tr>
<tr>
<td>2.3.3.1 Grit Unit</td>
<td>19</td>
</tr>
<tr>
<td>2.3.3.2 Semipelite-amphibolite Unit</td>
<td>21</td>
</tr>
<tr>
<td>2.3.3.3 Marble Unit</td>
<td>23</td>
</tr>
<tr>
<td>2.3.3.4 Pelite Unit</td>
<td>27</td>
</tr>
<tr>
<td>2.3.3.4.1 North of Birch Creek Fault</td>
<td>27</td>
</tr>
<tr>
<td>2.3.3.4.2 South of Birch Creek Fault</td>
<td>27</td>
</tr>
<tr>
<td>2.3.3.5 Quartzite Unit</td>
<td>29</td>
</tr>
<tr>
<td>2.3.3.6 Stratigraphy South of Bighorn Creek Fault</td>
<td>30</td>
</tr>
<tr>
<td>2.3.3.7 Summary</td>
<td>31</td>
</tr>
<tr>
<td>2.4 Local Correlation of Stratigraphy</td>
<td>32</td>
</tr>
<tr>
<td>2.4.1 Introduction</td>
<td>32</td>
</tr>
<tr>
<td>2.4.2 Eastern Assemblage</td>
<td>33</td>
</tr>
<tr>
<td>2.4.2.1 Semipelite-Amphibolite and Marble Units</td>
<td>33</td>
</tr>
<tr>
<td>2.4.2.2 Grit Unit</td>
<td>37</td>
</tr>
<tr>
<td>2.4.2.2.1 Introduction</td>
<td>37</td>
</tr>
<tr>
<td>2.4.2.2.2 Grit Division</td>
<td>38</td>
</tr>
<tr>
<td>2.4.2.2.3 Upper Clastic Division</td>
<td>39</td>
</tr>
<tr>
<td>2.4.2.2.4 Correlation with Upper Clastic Division</td>
<td>41</td>
</tr>
</tbody>
</table>
5.4 Sampling Techniques and Methodology ........................................ 117
5.5 Microfabrics by Metamorphic Zone ........................................... 121
  5.5.1 Chlorite Zone and Chlorite to Biotite Transition ..................... 121
  5.5.2 Chlorite-Biotite Zone ....................................................... 124
  5.5.3 Garnet Zone ................................................................. 128
  5.5.4 Staurolite-Kyanite Zone .................................................. 131
  5.5.5 Kyanite Zone ............................................................... 139
  5.5.6 Sillimanite Zone ............................................................ 141
5.6 Timing of Metamorphism and Deformation ................................... 144
5.7 Distribution of Fabric Elements .............................................. 146
  5.7.1 Zone 1 ........................................................................... 146
  5.7.2 Zone 2 ........................................................................... 148
  5.7.3 Zone 3 ........................................................................... 150
  5.7.4 Summary ......................................................................... 150

6 STRUCTURAL AND TECTONIC IMPLICATIONS OF MICROFABRIC AND
   MESOSCOPIC FOLD ANALYSIS ..................................................... 152
6.1 Introduction ............................................................................. 152
6.2 Previous Tectonic Models ....................................................... 153
6.3 Local Structural Interpretation ............................................... 155
  6.3.1 Structural Fan ................................................................... 159
  6.3.2 Rotation by Simple Shear .................................................. 161
  6.3.3 Composite Model for the Structural Evolution
       of the Northern Mountains ..................................................... 164
    6.3.3.1 Phase Two .................................................................... 164
    6.3.3.2 Interkinematic Phase ................................................... 166
    6.3.3.3 Phase Three ............................................................... 166
    6.3.3.4 Late Stage Faulting ...................................................... 170
  6.3.5 Conclusions ...................................................................... 171
6.4 Tectonic Models ..................................................................... 171

7 SUMMARY AND CONCLUSIONS ..................................................... 176

PHOTOGRAPHIC PLATES ................................................................ 179

LIST OF REFERENCES ..................................................................... 185

APPENDIX I .................................................................................. 193
LIST OF FIGURES

Figure 1  The Big Bend of the Columbia River and the Location of the Study Area  .................................................. 2

Figure 2  Topographic, Geographic and Station Location Map, Big Bend of Columbia River (in pocket)

Figure 3  Detailed Lithologic Map of the Northern Big Bend (in pocket)

Figure 4  Tectonic Setting and Kinematic History, Post-Cambrian to Eocene ............................................................ 8

Figure 5  Detailed Stratigraphic Sections, Northern Big Bend (in pocket)

Figure 6  Structural Cross Sections, Northern Big Bend (in pocket)

Figure 7  Correlation of Horseshief Creek Group Stratigraphy—Northern Dogtooth to Cariboo Mountains (in pocket)

Figure 8  Published Cross Sections, Northern Selkirk Mountains ................................................................. 34

Figure 9  Structural Geology of the Warsaw Mountain Area .......................................................... 44

Figure 10 Geology of the Southern Canoe River Area .......................................................... 46

Figure 11 Correlation of Stratigraphy, Warsaw Mountain Area .......................................................... 50

Figure 12 Schematic Structure Section, Warsaw Mountain Area .......................................................... 52

Figure 13 Structural Geology of the Northern Big Bend (in pocket)

Figure 14 Microscopic Fabrics Associated with the First Phase of Deformation .......................................................... 68

Figure 15 An Example Illustrating the Characteristics of S1; Some Factors Governing its Preservation and Subsequent Transposition .......................................................... 71
Figure 16 Changes in the orientation of second phase structures along Norman Wood Creek...... 81
Figure 17 Schematic 3-D Diagram to illustrate the inter-relationships between second and third phase folds - French Creek to Norman Wood Creek..... 83
Figure 18 Combined stereoplot of second phase data from Domains G and H .................. 90
Figure 19 Correlation of faults, Norman Wood Fault Zone............. 95
Figure 20 Pressure distribution during metamorphism ............. 100
Figure 21 Distribution of Structural and Metamorphic Domains and their relationship to Metamorphic Zones .......... 105
Figure 22 Schematic Diagram to illustrate relationships between porphyroblasts, their inclusion trails and the Axis of Rotation during folding............. 115
Figure 23 Sample location map, South of the Selkirk Fan Axis (in pocket)
Figure 24 Microfabric features of the Chlorite-Biotite zone ....... 123
Figure 25 Microfabric features of the Chlorite-Biotite and Garnet zones .................... 127
Figure 26 Microfabric features of the Staurolite-Kyanite zone ...... 132
Figure 27 Microfabric features of the Kyanite and Sillimanite Zones ..................... 136
Figure 28 Distribution of Fabric Elements from Goldstream River to Columbia Reach (in pocket)
Figure 29 Schematic Representation of Panning of Second Phase Axial Surfaces ............. 160
Figure 30 Deformation by Simple Shear ...................... 162
Figure 31 Schematic Representation of Fold and Fabric Development, N. Selkirk Mountains ......... 165
LIST OF PLATES

Plate 1  Lithologic characteristics of the Grit Unit  179
Plate 2  Lithologic characteristics of the semipelite-
        amphibolite unit  180
Plate 3  Lithologic characteristics of the marble, 
        quartzite and pelite units  181
Plate 4  Lithologic characteristics of stratigraphy 
        south of Bigmouth Creek Fault  182
Plate 5  Folds and fabrics associated with the second 
        phase of deformation  183
Plate 6  Folds and fabrics associated with the third 
        phase of deformation  184
LIST OF MINERAL/ABBREVIATIONS USED IN FIGURES

AND IN APPENDIX 1

chl  chlorite
bi; bio  biotite
mu  muscovite
ga  garnet
St  Staurolite
ky  Kyanite
Sill  Sillimanite
Qtz  Quartz
Plaq  Plagioclase
opq  Opaques
1. INTRODUCTION

1.1 Purpose and Scope

This thesis forms part of a regional program initiated in 1971 by R. L. Brown and graduate students at Carleton University to study the structural geometry and stratigraphy of the Selkirk Mountains, southeastern British Columbia (Figure 1).

The objects of the thesis are to elucidate the stratigraphy, structural geometry and kinematic history of the Selkirk Mountains, north of Bighorn Creek and west of Windy Creek (Figure 2, in pocket).

The study area forms the keystone block between studies by R. L. Brown and graduate students at Carleton University and those of P.S. Simmons, E.D. Ghent and graduate students at Calgary University.

The first objective was therefore to establish, where possible, an internally consistent stratigraphic succession for the thesis area. This was necessary in order to objectively correlate internal stratigraphy with the previously established stratigraphic successions of adjacent map areas.

A second objective was to establish an independent structural geometry and kinematic history of the area. Detailed studies of the folding and fabric development on both mesoscopic and microscopic scales were required to pursue this objective.

A further objective which forms part of the structural study was to establish the time relationships between the growth of metamorphic minerals and deformation.

The final objective was to apply the structural and metamorphic information gained from the microfabric study to the current
FIGURE 1: THE BIG BEND OF THE COLUMBIA RIVER AND THE LOCATION OF THE STUDY AREA

(Geology and Insert Map after a compilation by R.B. Campbell, 1971.)
regional, structural and tectonic models for the southeastern Canadian Cordillera.

The area mapped during the course of this study is underlain by rocks of Kyanite and Sillimanite grades of metamorphism. To understand the relationships between metamorphism and episodes of deformation, it was necessary to compare suites collected by others in the Carleton University group with data collected by the writer. These suites of samples include:

2. J. Van der Leeden (1976): (labelled V)
3. M. Perkins and D. Shaw (collected in 1974): (labelled P)
4. R. Leatherbarrow (1981): (labelled A)
5. M. Perkins (this thesis): (labelled M)

The location of the study areas are shown in an insert on Figure 3 (in pocket).

1.2 Location and Access

The Selkirk Mountains are located in the eastern Canadian Cordillera of southeastern British Columbia. Enclosed by the Big Bend of the Columbia River, the Selkirks are flanked to the east by the Rocky Mountain Main Ranges and to the north and west by the Monashee Mountains (Figure 1, insert).

The five hundred square kilometre area mapped during the course of this thesis is outlined in Figure 1. Bigmouth Creek and its tributary Louis Lee Creek conveniently divide the area (Figure 2) into three northeast-southwest trending ridge systems, from east to west, the
Windy Ranges, the central range containing Mount Chapman and the most readily accessible western ridges centered about Mud Glacier.

Local relief is considerable. Valley bottoms are at 1,000 m while peaks top 3,000 m. Mapping was concentrated above tree line between 1,800 m and 2,500 m. Ice and permanent snow fields cover large areas (Figure 2), but recent deglaciation at their peripheries provides excellent clean ice-polished outcrops. Snow in general is a problem and serious mapping in the high ground can seldom begin before the first of July. The months of July and August are optimal for field work.

Helicopters provide the fastest and most logical access to most of the area and were used extensively for camp moves. Logging roads from Mica Creek townsite to Yellow Creek (Figure 2) offer access to the northern part of the area. Since the completion of field work, stripping in the Columbia Valley, prior to completion of the Revelstoke Power Dam has provided additional access to the area's western periphery.

Field work was carried out in the summers of 1975 and 1976. In 1974 the writer worked with R.L. Brown and D. Shaw in the Windy Ranges and the data collected is incorporated into this thesis.

The study area lies within N.T.S. 1:50,000 map sheets:

-82 M/15, 82 M/16, 83 D/1.

1.3 Previous Work

Reconnaissance mapping at the scale of 1:250,000 in the Rogers Pass and Big Bend map areas (Wheeler 1963; 1965) provides the basic geological information for the region.
More detailed geological information is now available from the results of studies by staff and graduate students at Queen's University, the University of Calgary and Carleton University.

The Illééllewaet area and adjacent Clachnacudainn "Salient" along the Trans-Canada Highway were studied by Thompson (1972), Gilman (1972) and Zwanzig (1973) under the supervision of R.A. Price, Queen's University.

Workers from the University of Calgary, under the supervision of P.S. Simony, E.D. Ghent and others have mapped extensively in the Selkirk Mountains, Dogtooth Range and Esplanade Range (Wind 1967; Ellison 1967; Jones 1969; Ghent et al. 1970; Poulton 1970; Simony and Wind 1970; De Vries 1971; De Vries et al. 1971; Ghent and De Vries 1972; Jones 1972; Poulton 1973; Ghent 1975; Ghent and McKee 1975; O'Neill and Ghent 1975; Simony 1975; Poulton and Simony 1980).

In the Monashee Mountains, workers from the University of Calgary have again been active (Mitchell 1976; Robbins 1976; Ghent et al. 1977; Craw 1978; Knitter 1979; Simony et al. 1980; Morrison 1979 and 1982; Pell and Simony 1980; Pell in progress; and Raeside 1982).

In 1973 students from Carleton University under the supervision of R.L. Brown began working in the area within the Big Bend of the Columbia River. The students are Franzen 1974; Tippett 1976; Van der Leeden 1976; Lane 1977; Psutka 1978; Shaw 1980; Leatherbarrow 1981; and the writer. The thesis areas are illustrated in the insert to Figure 3. The work presented here is an integral part of Carleton University's Northern Selkirks project.
1.4 Regional Geology

The Selkirk and Monashee Mountains (Figure 1) form parts of the Qninea Crystalline Belt (Douglas et al. 1976). Both Mountain ranges are underlain by metamorphic terrane and Brown (1980a) includes both ranges with the Shuswap Metamorphic Complex (Jones 1959; Reesor 1970) in the term "Metamorphic Complex". Following Brown (1980a), the complex is best discussed in terms of an Archean to early Proterozoic Basement and both Autochthonous (reference frame in Basement rocks) and Allochthonous cover rocks (Figure 4A).

Rb/Sr whole rock isochrons from gneissic rocks of Thor-Odin Dome and Frenchman Cap indicate the presence of basement with an age of approximately 3 Ga (Duncan 1978; Armstrong 1979). More recently however, Armstrong and Brown (in preparation) have obtained Rb/Sr dates of 2.17 Ga for Frenchman Cap rocks suggesting that it may not be Archean. Archean ages have also been recorded for the Malton Gneiss (Chamberlain et al. 1978). These core gneisses, excepting the Malton Gneiss, appear to be continous between individual domes (Brown 1980a; Read 1980; Read and Brown 1981), which suggests that the metamorphic complex is underlain by extensive Archean terrane to early Proterozoic terrane.

An autochthonous cover of shallow-marine clastic and carbonate platform sediments overlies the Archean basement (Brown and Psutka 1979). On the basis of stratigraphic similarity these rocks have been correlated with Cambrian strata to the east (Reesor 1970; Wheeler et al. 1979). Brown and Psutka (1979) and Brown and Read (1979b) have argued, on structural grounds, for an early Proterozoic (Belt-Purcell) age. A zircon date of 773±200 m.a. from a nepheline gneiss which intrudes
The cover rocks also suggest that cover rocks are pre-late Proterozoic (Okulitch et al. 1981).

A thick succession of pelite, carbonate and amphibolite rocks is separated from the autochthonous cover by the Columbia River Fault Zone and Monashee decollement (Brown and Psutka 1979, Brown and Read 1979b, Read 1979, Brown 1980, Brown 1980b). These sediments, which form the allochthonous cover, lie above the hanging wall of the Columbia River Fault and have been transported eastwards relative to the Archean basement (Brown 1980a). In the northern Selkirks they consist primarily of westerly prograding miogeoclinal prisms of Hettynian and Lower Paleozoic ages.

Three major episodes of deformation are recognized in the Quineca Belt. These occurred in the Proterozoic, mid-Paleozoic and the Middle Jurassic. The earliest event known in the Kootenay arc is the pre-Mississippian nappe-forming event (Read 1975; 1976), which may extend into the Selkirk and Monashee Mountains.

Polyphase deformation associated with regional metamorphism affected the entire Quineca terrane during the Middle Jurassic. (Read and Wheeler 1976; Pigage 1977; Brown and Tippett 1978; Brown 1980). In the northern Selkirks the middle to upper amphibolite peak of metamorphism separated two fold-forming events. The first event produced structures overturned primarily to the west. Interference between these early folds and the later easterly overturned structures gave rise to the Selkirk Fan Structure (Brown and Tippett 1978). Price (1979), in a discussion of the Brown and Tippett paper (1978), supports Wheeler's (1966) contention that the Selkirk fan structure is best explained in terms of a cleavage fan produced during one protracted
Figure 4: Tectonic Setting and Kinematic History

Post-Cambrian – Eocene.

(Modified after Brown 1980)
episode of deformation. The upward and outward rotation of the Selkirk-Purcell Anticlinorium is suggested by Price (1979) to have involved strong southwesterly overturning whereby existing foliations became rotated into the direction of compressive strain and were overprinted by newly developing foliations.

Granitic plutons were emplaced both during and after tectonism. Rb/Sr dating of post-tectonic intrusive events (Duncan et al. 1979) indicated that regional ductile strain had ceased by the beginning of the Late Jurassic.

Late Mesozoic and Cenozoic sedimentation patterns are related to uplift and stabilization after the Late Jurassic. Initial uplift of the Omineca Complex in Late Jurassic to Early Cretaceous is indicated by the cooling and thermal history data from the Wolverine Complex (Parrish 1976) with renewed uplift associated with an early Tertiary thermal event. K/Ar ages on Late Jurassic plutons (Duncan et al. 1979) indicate reheating at various times up to the Eocene.

High-angle normal faults occur within the metamorphic complex and Rocky Mountains (Price 1965; Bally et al 1966; Leech 1967; Wheeler et al. 1972; Craw 1978; Leatherbarrow 1981; McMechan and Price 1980). Price (1965) and McMechan and Price (1980) have shown some high-angle normal faults in the Flathead Valley, southeastern British Columbia, to have been active in Upper Eocene and later times.

1.5 Tectonic Models

Price and Mountjoy (1970) presented the case for a temporal relationship between the emergence of the core complexes in the west and the generation of foreland thrust and fold belt to the east. The model
involves decollement thrusting over a cratonic basement which remained rigid at least as far west as the Rocky Mountain Trench. Over rocks were contracted over a minimum distance of 320 km, which suggests that metamorphic core complexes may also be allochthonous.

The model proposed by Campbell (1973) involves two distinct tectonic regimes. In the west, high-angle basement-involved thrusting occurred above an essentially stable but uplifted core complex. To the east, westward underthrusting of the craton caused restricted decollement under the Front Ranges and Pootills.

Brown (1980b) expanded the model proposed by Brown and Tippett (1978) to include the Shuswap complex and the Selkirk-Kootenay terranes (Figure 48). The model involves three stratigraphic and structural elements. In the west a platform of Archean to early Proterozoic gneisses was overlain by up to 5 km of shallow marine clastics of probable Precambrian age. To the east lay the Rocky Mountain Miogeosyncline, a westwardly thickening wedge of Proterozoic sediments which unconformably overlay the peneplaned cratonic gneisses of the North American Plate. A sedimentary basin, filled with over 10 km of Proterozoic sediments and volcanics, lay between and to the west of the western platform and the distal edge of the Rocky Mountain foreland (Figure 48). Basement to this basin is thought to be oceanic or transitional rather than continental (Brown 1980b). A westerly prograding wedge of Cambrian shallow marine clastics overlaps the Proterozoic sediments of the basin (Wheeler 1965; Wheeler and Gabrielse 1972; Brown et al. 1978). The basin has been named the Selkirk-Kootenay Basin (Brown 1980b).
A pre-Mississippian (probably Devonian) period of nappe formation in the Omineca Belt is believed by Brown (1980b) to be related to plate convergence (Figure 4B). Subduction of oceanic crust to the west of the Omineca terrane caused plutonism and initiation of collapse of the western margin of the Selkirk-Kootenay Basin leading to nappe formation (Brown 1979). Ghent et al. (1977) and Brown and Read (1979) presented evidence for westward displacement of the nappes, i.e. from the eastern part of the basin onto the Shuswap platform.

In the Middle Jurassic, the early nappes of the Selkirk-Kootenay Basin were folded about westerly overturned folds (Figure 4C, Phase 2). These westerly overturned structures may be accounted for by incipient underthrusting of Shuswap basement beneath the Selkirk-Kootenay Basin (Brown and Tippett 1978; Brown 1980b).

Easterly overturned folds developed as the tectonically thickened and buoyant Omineca belt was underthrust from the east (Figure 4D, Phase 3).

Continued underthrusting and uplift of the Omineca terrane initiated the eastward migrating thrust faulting above the decollement surface in the Rocky Mountain foreland (Figure 4E).

In conjunction with late-metamorphic uplift, cover rocks above the Shuswap platform became detached and were displaced eastwards along the Columbia River Fault Zone (Read 1980; Brown 1980b; Read and Brown 1981). Thrust faulting in the Rocky Mountain foreland continued into the Paleocene (Price and Mountjoy 1970). This fault system was reactivated during a period of Tertiary uplift and extension (Read and Brown 1981).
2. STRATIGRAPHY

2.1 Introduction

In the southeastern Canadian Cordillera, two major post-Aphebian sedimentary sequences have been recognized. The Helikian Belt-Purcell Supergroup is separated from the overlying Hadrynian Windermere Supergroup (Gabrielse 1972) by an unconformity related to the East Kootenay Orogeny (White 1959; Reesor 1973).

In comparison with the generally uniform and fine-grained sediments of the Belt-Purcell Supergroup, the Windermere Supergroup includes a wide range of lithologies. The basal coarse conglomerates of the Toby Formation are overlain by up to 2000 m of volcanic flows, tuffs, breccias and agglomerates of the Irene Formation (Little 1960; Stewart 1972). Higher in the section, the Supergroup is characterized by an overall fining-upwards cycle from coarse sands and grits near the base through slates and silts to a carbonate sequence near the top collectively known as the Horsethief Creek Group. A widespread cross-bedded, coarse grained sand sequence overlies the Supergroup.

This change in the nature of sedimentation is generally correlated with the Hadrynian to Cambrian boundary, although the exact position of the time plane is impossible to define due to lack of faunal control. The Windermere Supergroup and overlying Cambrian sequences occur with apparent conformity in the west, but are separated by a small angular unconformity; the stratigraphic omission across this unconformity becomes increasingly important towards the east (Aitken 1969).

This thesis deals in part with stratigraphic problems in the Horsethief Creek Group within the Big Bend of the Columbia River.
Correlation of the established map units with those in areas where the stratigraphic sequences have previously been defined and named provides the basis for discussion of stratigraphic and structural problems in the thesis area and surrounding region.

2.2 Stratigraphy of the Horsethief Creek Group

2.2.1 Nomenclature

Walker (1926) proposed the name Horsethief Formation for a sequence of clastic sedimentary rocks stratigraphically underlying the Hamill Group. The type section was located along a ridge just east of Mount Law, between Law and Horsethief Creeks (Walker 1929, p.9) within Windermere map area, British Columbia. "Horsethief Formation" was, however, preempted, and Evans (1933) renamed the formation the Horsethief Creek Formation. Walker (1934) renamed the Horsethief Creek Formation the Horsethief Creek Series to indicate the presence of internal subdivisions. Wheeler (1963, 1965) used the term Horsethief Creek Group in accordance with the recommendation of the American Commission on Stratigraphic Nomenclature (1961) that use of the term "Series" be discontinued as a rock-stratigraphic subdivision and the term "Group" be substituted in its place.

Poulton (1970) attempted to show that although similarities in sequence do occur over a large area, detailed stratigraphic correlation within the Horsethief Creek cannot be established except on a local scale. Poulton (1970) considered it advisable, therefore, to retain the name "Horsethief Creek Formation" with local subdivision into members.

The author will attempt to demonstrate that the Horsethief Creek Group is readily divisible throughout the Big Bend and that this
detailed stratigraphy can be extended into those areas mapped by Poulton (1970), Simony and Wind (1970), and de Vries (1971), Brown, Tippett, Lane (1978) to the south; and Campbell (1968), Ghent et al. (1978), Campbell et al. (1973), Pigage (1978) and others to the north. Stratigraphic units previously assigned a formation status in the Cariboo Mountains (Campbell et al. 1973) can be differentiated on a regional scale and it is on this basis that the term Horsethief Creek Group is believed to be the most appropriate nomenclature.

Evans (1933) recognized the following sequence in the northern Dogtooth Mountains, from bottom to top; the Feldspathic Grit Member, the Lower Slate Member, Limestone Member, and the Upper Shale Member. These units were later verified by Poulton (1970) and Simony and Wind (1970). The following is a brief review of the lithologies of the four fold subdivision (above) and their possible stratigraphic equivalents in the Selkirk Mountains.

2.2.2 Stratigraphy in the Dogtooth Range

At the base of the formation are clastic sediments of the Feldspathic Grit Member. Mineralogically immature and poorly sorted "grits" (a local lithological term for coarse sandstones to granule conglomerates) occur together with quartzites, slates and siltstones. The coarsest rocks occur in both graded and non-graded beds. Other sedimentary structures are rare. A thick succession of slate, siltstone and medium-grained sandstone which shows an upwards decrease in sandstone is interpreted by Poulton (1970) to represent the transition into the Lower Slate Member. The bulk of the Lower Slate Member consists of approximately 1000 m of slate, phyllite and schist with
occasional sandstone beds. Finely laminated limestones which occur throughout have been interpreted as reflecting original layers of calcareous sediment, mud and silt (Poulton 1970). Clasts resembling the carbonates of the overlying Limestone Member occur within the upper part of the Lower Slate Member, suggesting that the two units are in part facies equivalent.

The Limestone Member thickens from less than thirty metres in the eastern Dogtooth Range to almost three hundred metres in the western part of the Dogtooth Range (Poulton 1970, 1973). At the former location the member consists of two 2.5-3 m limestone bands separated by 11 m of slate. At the latter, massive to thickly-bedded dolomites predominate with interbedded slate, quartzite and limestone forming the transition to the underlying Lower Slate Member. Occasional lenses of slate occur near the top of the unit.

The Upper Shale Member consists of interbedded slate and quartzite, both variably dolomitic, with minor limestone and dolomites. No clastic feldspar was recognized in the coarse grained rocks (Poulton, 1970).

2.2.3 Stratigraphy in the Selkirk Mountains

In the Selkirk Mountains, Brown et al. (1977) subdivided the Horsethief Creek Group into three lithologically distinct map units and informally named them, from top to bottom, the upper pelitic member, the middle marble member, and the lower pelitic member. The three members have been correlated with the Upper Shale Member, Limestone Member, and Lower Slate Member of the Dogtooth Mountains by Brown et al. (1978).

The lower pelitic member is composed of thinly interbedded
pelite and semipelite with minor psammitic. Distinctive amphibolite bands are both concordant and discordant with respect to compositional layering in the metasediments. The origin and significance of these amphibolites are discussed below.

The middle marble is characterized by the presence of thick grey calcitic marbles. Zones rich in pure and impure marbles alternate with pelites and other clastic rocks in an internally complex stratigraphy (Brown et al. 1978).

The upper pelitic member was subdivided by Tippett (1976) into three assemblages. The lowest assemblage is dominated by thickly interbedded psammites and pelites. The central subdivision includes coarse clastic rocks. The upper subdivision is a sequence of rhythmically laminated pelites with scattered carbonate-rich beds. The stratigraphic column proposed by Tippett (1976) represents the standard reference for workers at Carleton University as it was established in an area where correlation with previously named stratigraphy was possible.

2.3 Detailed Stratigraphy within the Big Bend

2.3.1 Introduction

The stratigraphy of the Horsethief Creek Group within the thesis area will be discussed initially in terms of internal map units. The well-defined stratigraphic succession established elsewhere within the Selkirk Mountains (de Vries 1971; Tippett 1976; Brown et al. 1977; and Brown et al 1978) is then correlated with this internal stratigraphy.

A variety of metasedimentary rock types were encountered in the course of thesis mapping. Sedimentary textures and original
mineralogy have been destroyed or altered to varying degrees by re-crystallization, although some sedimentary structures have been preserved in the more massive rock units. As a result it was concluded that rigorous application of a sedimentary rock nomenclature would not be possible, and descriptive field terminology has been retained for the most part in the thesis. Where, for instance, the term conglomerate is used, there is an interpretive component to the extent that the rock is believed to have been a conglomerate before amphibolite facies metamorphism and penetrative deformation. As well, because all rocks in the thesis area have been metamorphosed, use of the prefix "meta-" was considered to be repetitious and unnecessary.

2.3.2 Nomenclature

One of eight general names have been given to the clastic rocks. Pelites contain more than thirty percent combined muscovite, biotite, and chlorite and were probably muds or mud-rich sediments. Psammites are quartz and feldspar-rich rocks containing between five and ten percent micaceous minerals and were probably siltstones and sandstones. Semipelites are intermediate between the two. Quartzites are quartz-rich psammites with few micas or other impurities. Grain size is not considered in the first four rock-types, although the origin of pelite and psammite is linked to grain size. Grit refers to rocks containing granules from 2 to 8 mm in diameter in a finer-grained matrix. "Graded-grits" also includes that finer-grained part of the graded cycle where maximum grain size could be less than 2 mm. Conglomerates contain clasts greater than 8 mm in diameter.
A problem of slight over-estimation of the mafic component in the field was encountered when percentages were checked in thin section, but no attempt was made to systematically alter original field descriptions except in critical cases.

Carbonates are those rocks which contain greater than fifty percent carbonate. Dominantly clastic rocks with less than 50 percent carbonate are prefixed with the terms dolomitic or calcareous. Calcitic limestone, dolomitic limestone and dolostones are recrystallized to white to buff coarse-grained marbles. The term calc-silicate describes siliceous carbonate rocks.

Lithologies within the thesis area can be grouped into six assemblages, namely:

1) a thick succession of clastic rocks including grits and conglomerates which grade upwards into a dominantly pelitic succession: the grit unit;

2) a semipelite and pelite assemblage with amphibolite bands: the semipelite - amphibolite unit;

3) carbonates, pure and impure, with interbedded clastics: the marble unit;

4) laminated pelites with quartzites and psammites and, less commonly, feldspar-rich amphibolites and hornblende schists, grading upwards into a carbonate sequence similar to that in assemblage 2: the pelite unit; and

5) thinly bedded quartzites with quartz rich calc-silicates: the quartzite unit.

6) the following lithologies are grouped into one assemblage for
structural and stratigraphic reasons which will be discussed later in the text: coarse grained, massive grey and buff marbles underlain by a mixed clastic assemblage and overlain by dark pelitic schist. This assemblage occurs only to the southwest of Bigmouth Creek Fault (Figure 3).

2.3.3 Internal map units

2.3.3.1 The Grit Unit

The southern limit of the grit unit is defined by the trace of Northeast Fault which trends southeast from a position immediately south of Warsaw Mountain to Trident Mountain (Figure 3). North of the fault the grit unit crops out almost continuously to the northern limit of mapping on the northeast-facing slopes above Columbia Reach (Figure 3). Coarse clastic rocks, including grits and conglomerates are interbedded with pelites and minor carbonate and are separated from the amphibolite-bearing semipelites to the south by a thick pelitic succession (Figure 3). Graded beds in the grit horizons allow a stratigraphic rather than structural succession to be presented in Figure 5 - Columns 1, 2 and 3. Plate 1 illustrates some of the more common features of the coarser grained rock units.

At the base of the succession mapped (which probably does not correspond to the base of the unit) is a mixed assemblage of well-bedded psammites, semipelites, pelites and less commonly calc-silicates, quartzites and thin rusty marbles. These are overlain by the thick succession of coarse clastic rocks which characterize the unit. Pebble conglomerates at the base consist of well-rounded quartzite, quartz-feldspar, quartz and psammitic clasts in a matrix consisting of
finer grained equivalents of these same constituents in a quartz-rich pelite. Distinctive green calc-silicate pebbles, rich in actinolite, form an important clastic constituent in the coarser-grained horizons. The percentage of matrix varies such that a complete spectrum from matrix-supported (Plate 1A) to virtually clast-supported conglomerates exist (Plate 1B). Plate 1C shows a graded conglomerate layer. Most of the original clast to matrix relationships have been destroyed by deformation during which the clasts become flattened parallel to the schistosity in the matrix (Plate 1A, C) and elongate parallel to local hinge lines of folds. Only clasts composed of 100 percent quartz have resisted deformation to retain an approximately circular form (Plate 1C). Mitchell (1976) reports the average x/z and y/z ratios are 4.5:1 and 3.9:1 respectively where x, y and z are the long, intermediate and short axes of the pebbles respectively.

Although individual horizons are poorly sorted the succession as a whole fines upwards i.e. towards the southwest from pebble to granule conglomerate at the base through grits, quartzites, psammites and pelites, into a dominantly pelitic section. Both graded and non-graded grits occur. Grading is most obvious in the coarser basal grit layers (Plate 1D), being more difficult to recognise upwards as overall grain size decreases. Graded layers fine in the same direction as the fining-upwards trend i.e. to the southwest. They are therefore considered to represent normal rather than reverse-graded sedimentary layers. Although coarse conglomerates are found only in the northeast, grits characterise the succession from Warsaw Mountain eastwards to Trident Mountain.
The transitional sequence from psammites and quartzites to pelites, marked by a gradual upwards increase in the amount of pelite, contains several 25 cm layers of pure or sandy marble and rare calc-silicates. Capping the succession is a dark, coarse grained, biotite-rich "migmatitic" pelite. Interbedded with the pelite and accounting for about ten percent of the volume are 10-25 cm thick psammite beds. The psammites along with the quartz-migmatite outline the strong third phase folding in a highly crenulated pelite (Plate 1E).

The sudden appearance of flaggy semipelites marks the base of the semipelite-amphibolite unit which structurally overlies the thick pelite. Evidence of shearing at or close to the contact (Plate 1, F, G and H) suggests that it is probably faulted (see page 52 for discussion).

2.3.3.2. The Semipelite-Amphibolite Unit

The semipelite-amphibolite unit underlies the central part of the thesis area. The unit occupies the core of a large antiformal culmination, the trend of which controls its areal distribution (Figure 3). From southwest to northeast the unit is exposed in the Windy Range between Mount Neptune and the marbles at the head of Norman Wood Creek; on ridges to the south of and including Mount Chapman; and to the north and south of Mud Glacier respectively (Figure 3). Along its outcrop length the semipelite-amphibolite unit is overlain to the southwest by the middle marble unit.
A typical succession through the unit is presented in Figure 5.- Columns 4 and 5, while some of the characteristic features of the lithologic types are illustrated in Plate 2.

Finely interlayered semipelites, pelites and amphibolites, together with rare psammites and even rarer sandy carbonates and calcareous psammites make up the unit. Semipelite and pelite layers rarely exceed 25 cm and may be as little as 2 cm. Amphibolite layers range in thickness from 2 cm to several tens of metres. Weathered outcrops are characteristically flaggy (Plate 2A), a feature which is due to both the thinly bedded nature of the unit and to the presence of the recessive-weathering micaceous interbeds between the more resistant semipelitic horizons. Ice polished outcrops have a distinctive striped appearance which is accentuated by the dark amphibolite layers (Plate 2B).

Three common types of amphibolites are present.

Type one is composed predominantly of hornblende and contains accessory plagioclase (Plate 2C); it does not contain garnet. A band averaging 2 cm in thickness which is enriched in biotite is generally present where the amphibolites are in contact with adjacent meta-sediments. Type one amphibolites may be discordant or concordant. Thicker layers have coarse homogeneous interiors.

Type two amphibolites contain porphyroblasts of garnet in an amphibole and feldspar matrix which may include 5-15% biotite and minor amounts of quartz (Plate 2D). Poikiloblastic garnets up to 1 cm in diameter occur in darker biotite-rich bands at the margins of these amphibolite layers (Plate 2E). Both type one and two can be distinctly to indistinctly banded due to the lenticular alternation of quartz and
feldspar with amphiboles, usually hornblende. Plate 2F shows a sample from a distinctly banded unit.

Type three amphibolites are rocks which may be more strictly called hornblende-rich calc-silicates. Mafic layers 8-15 cm in thickness alternate with 1-2 cm calc-silicate, or more rarely 8-15 cm calcite bands (Plate 2C). The calc-silicate bands are mineralogically complex and include combinations of calcite, garnet actinolite, hornblende, plagioclase, biotite, quartz, epidote, and titanite.

While the protolith for all amphibolites in the northern Selkirk mountains has not been positively identified, observations on the composition, textures and field relationships agree with those of Poulton and Simony (1981). Both mafic igneous rocks and dolomitic marls are suggested at protoliths. Types one and two were most likely mafic intrusions in the form of dikes and sills. Type three are most probably derived from dolomitic marls, examples of which occur in the semipelite-amphibolite unit in the Rogers Pass area (Poulton and Simony, 1981).

2.3.3.3. The Marble Unit

Throughout the map area the marble unit overlies the semipelite-amphibolite unit. The areal distribution of the unit is shown in Figure 3, while its relationship to the semipelite-amphibolite unit is shown in Figure 6. Plate 3 illustrates some of the more typical features of the unit. Details of the stratigraphy are presented in Figure 5 - Columns 6, 7, 8, and 9.

The marble unit consists of marbles, impure marbles, calcareous and non-calcareous pelites and semipelites, and calc-silicates. Discussion of the unit is hampered by the fact that it
is discontinuous in its distribution and shows great lithological variation both along and across strike. The unit is characterized by the presence of grey marble bands (Plate 3A), which occur at or close to its lower boundary. Correlation of the marble unit across the thesis area is, therefore, based on the presence of grey marble bands and the structural position of the marbles with respect to the amphibolite-bearing sequence.

The grey marble bands within the thesis area crop out as rounded, generally massive horizons within which irregular or lensoidal beds may be poorly defined (Plate 3A). The rough textured look of the weathered surface is due to the coarseness of the grain size which varies from 2 to 6 mm. The remainder of the marble unit weathers to a buff or light rusty brown leading to the descriptive field name "rusty marble" (Plate 3B). In the central part of the unit grey and rusty marble bands alternate in a distinctive lithologic sequence (Plate 3C).

The rusty marbles are siliceous and tend to weather more easily than the grey marbles producing a brown calcareous "sand" (Plate 3D). Although grain size is similar to that of the light grey marbles (2 to 5 mm), individual layers are finer and better defined (Plate 3E) and thus preserve details of the stratigraphy. Early pegmatite and quartz veins have been oxidized as a result of their competent nature during deformation of the ductile marbles. (Plate 3E, F).

Calc-silicate layers occur throughout the unit. The angular fracture surfaces of the weathered outcrop attest to the more resistant nature of these quartz-rich carbonates. Where carbonate is a minor constituent these layers show similar outcrop characteristics to, and may be difficult to distinguish from, calcareous psammites. In the
calc-silicates however the quartz occurs in coalescing patches with garnet and amphibole separated by recessive weathering carbonate rather than as discrete grains. In detail, a complete spectrum from sandy carbonate to calcareous psammite exists. Non-calcareous psammites and pelites also occur throughout.

Southwest of Mud Glacier (Figure 3) the marble unit occurs with structural conformity above the amphibolite-bearing sequence (Figure 5 - Column 6 and Figure 6 - Section 1). Separating the two is a 50 m transition zone of pelitic schist with thin calcareous horizons. The percentage of pelite decreases upwards, being compensated by an increased percentage of carbonate up to the first appearance of grey marble. The remainder of the succession consists of alternations of calc-silicates, rusty calcareous and non-calcareous pelites, buff-coloured impure marbles and thin grey marbles (Figure 5 - Column 6).

South of Mount Chapman (Figure 3) the marble unit (Figure 5 - Column 8) contains a greater number of pelitic horizons with marble and rusty marble constituting only forty percent of the thickness of the section. This trend appears to continue southeastwards to the head of Norman Wood Creek to the west of French Glacier (Figure 3), where the marble unit is composed of three grey marbles in pelite and calc-silicates overlying semipelites with amphibolite bands. The contact is sharp, and transitional lithologies were not observed. The upper part of the marble unit is not exposed at this location and the original percentage of marble is not known.

North of the Mud Glacier (Figure 3) marble bands crop out in the form of a synformal keel, which outcrops high in the ridge to the
west of Yellow Creek (Figure 6). The unit again conformably overlies the semipelite–amphibolite unit. These marbles were not observed in the lower ridges immediately to the east nor in the steep northern slopes of Mount Chapman.

To the south of and including Mount Neptune, however, the marble unit occupies the core of both second and third phase synforms (Figures 3 and 6). Beneath the marble unit is a sequence of semipelites with thin marble bands which probably represents the transition to the marble unit above. This assemblage is similar to the one observed north of Mud Glacier (Figure 3). The fact that both semipelite and marble assemblages occur immediately in fault contact with and to the north of a major antiformal culmination in the semipelite amphibolite unit suggests that they are correlative.

The marble unit or transitional sequence was not observed in the central ridge system, but it is possible that they would crop out under the icefield which covers the upper slopes of Mount Chapman (Figure 3). Plunges of major folds in the northern part of the thesis area are shallow and it can be demonstrated that hinge lines pass through the horizontal (see discussion of Structural Geology). It is equally possible, therefore, that the marble would not crop out in the central section simply due to variations in the amount and direction of plunge of major folds.

The marbles in the northeast-trending ridges are folded into a series of tight to open northeast-southwest overturned structures (Figure 6). The most northerly exposure of the marble unit or transitional sequence on each ridge is synformal. Southwesterly dipping marbles clearly
overlay the amphibolite-bearing semipelite unit before the formation of the synclines.

2.3.3.4. The Pelite Unit

2.3.3.4.1. North of Birch Creek Fault

North of Birch Creek Fault (Figures 3 and 13) a major third-phase synform outlined by the marble unit is cored by interbedded pelite, semipelite and psammites which overlie the calcareous pelites and calc-silicates transitional to the underlying marbles of the marble unit. Coarsely schistose pelites and semipelites are generally buff coloured and stained with the reddish brown to yellow stain so characteristic of pelites within the sillimanite zone. Quartz feldspar leucozomes are common throughout and pegmatite occurs 'lit par lit' with the pelites. Dark amphibolites and garnet-bearing calc-silicates are common in the basal portion.

South of Mount Chapman the abundance of pegmatite makes it difficult to ascertain a true thickness of the pelite sequence. Due to the transitional zone at the base and the limited areal extent of these rocks it is not known whether they constitute a separate stratigraphic unit. Lithologically, however, the bulk of the unit is quite distinct from the underlying marble unit and it is here designated as part of the pelite unit.

2.3.3.4.2. South of Birch Creek Fault

South of Birch Creek Fault a succession of carbonates and calcareous pelites underlain by thick pelites, semipelites and thickly bedded psammites with rare quartzites and mafic rocks lies to the south of and underlies the quartzite unit (described later, page 34). The lithologic details are shown in Figure 5 - Column 10. The contact
between some thinly bedded siliceous marble and a pelite from near the base of the calcareous part of the unit is shown in Plate 3H.

The association of carbonates structurally above pelites, semipelites and psammites is one already recognised several times in this thesis. The obvious conclusion would be that we are dealing with a right-way-up succession where the marble unit overlies the semipelite-amphibolite unit. The carbonate and calcareous pelite sequence outlined in Figure 5 - Column 10 shows a lithologic similarity to the marble unit. The underlying lithologies, however, bear little relationship to any previously described semipelite-amphibolite successions (cf. Figure 5 - Columns 4 or 5 and the lower part of Column 10).

Individual units are thicker, more monotonous and considerably richer in aluminosilicates. Most importantly the garnet amphibolites so characteristic of the semipelite-amphibolite unit are extremely rare. Dark green to black mafic rocks with elongated hornblende crystals (up to 4 cm) do occur locally however.

The succession becomes less pelitic and more psammitic southwards away from the capping marbles until a structural break is reached to the north of the large grey marble bands adjacent to Bigmouth Creek. A further structural discontinuity is indicated at the Birch Creek fault. The Pelite unit south of Birch Creek fault is thus fault bounded and not directly correlative with adjacent stratigraphy. The justification for the correlation of this unit with the pelite unit will be discussed later.
2.3.3.5. The Quartzite Unit

Immediately to the south and southwest of Birch Creek Fault in the south-central part of the map area (Figure 3) the marble unit abuts a blocky to flaggy quartzite sequence.

Individual grey and white quartzite layers (10-15 cm), which make up the bulk of the unit, have micaceous bedding planes or are separated by thin pelitic interbeds (Plate 3G). The pelites are predominantly light coloured and muscovite rich but occasional beds are very dark, biotite-rich and iron stained. Within the sequence, distinctive green semipelite rocks, rich in tremolite-actinolite, are associated with impure carbonates and calcareous quartzites. Two buff-coloured marble bands occur, both of which are finely laminated and contain layers and pods of coarse actinolite.

At the southern margin of the marble unit south of Birch Creek (Figure 3) is a sharp lithologic break from sandy marble, grey marble and pelites to a 5 m section of blocky, slightly calcareous white quartzite and minor psammite (Plate 3G). Lithologic similarity and the nature of the abrupt contact suggest, both that the quartzites may in fact be strike correlative with the quartzites to the north of the rockslide (Figure 3) and that the break represents the western extension of the Birch Creek Fault. The quartzites to the south of Birch Creek are tentatively assigned to the quartzite unit. This proposal would be in agreement with Wheeler (1965).
2.3.3.6 Stratigraphy South of Bigmouth Creek Fault

The most obvious map unit south of the Bigmouth Creek Fault is
dominated by coarse-grained, thickly layered grey marbles (Plate 4A).
Layers may in part be tectonically thickened by early tight to isoclinal
folds (Plate 4B). Interbedded with the massive grey marble layers are
thick sequences of buff, more thinly bedded marble layers (Plate 4C).
Thinly bedded alternations of grey and buff marble bands complete the
carbonate map unit.

Underlying the most northerly of the thick marbles, which
outlines a second phase synform (Figure 3), is an assemblage of
buff-coloured pelites and semipelites with thin psammites, calc
silicates and marble bands (Plate 4D, E, F and G). The contact between
this succession and the overlying marble (as well as the massive nature
of the marbles) is illustrated in Plate 4D. In the core of the same
synform a dark biotite-rich pelite conformably overlies the marble
sequence. The complete succession is illustrated in Figure 5 - Column
12.

The southern of the two thick marbles (Figure 3) cannot be
directly correlated to the northern of the massive marbles due to the
presence of a fault. The trace of the fault lies immediately north of
an antiformal closure in the marble as outlined by a "hook" in the
northern contact of the marble with adjacent metasediments. This tight
antiformal closure is readily appreciated from the air when flying into
the cirque from the west (Figure 2). Above the thick grey marble is a
mixture of pelites, semipelites and thin calc-silicates. Apart from a
slightly greater abundance of psammitic lithologies, this assemblage is
lithologically very similar to that which underlies the northern of the
two marbles. This fact, together with the thick massive nature of the
grey marbles, suggests that the marble-dominated sequences and associated lithologies to the north and south of the fault are stratigraphic equivalents.

The southern marble has been traced almost continuously westwards and finally southwestwards downslope into the valley of the Columbia (Figure 3) where it appears to be correlative with marbles traced northwest from the mouth of Bigmouth Creek by Wheeler (1965) and R.L. Brown and L. Lane (personal communications, 1981). Eastwards from Stn. 229 the southern marble appears to have corollaries along strike with thick marbles south of Bigmouth Creek stock (Van der Leeden, 1976).

The northern marble appears to have strike equivalents to marbles in Argonaut Mountain (Franzen 1974, Brown Tippett and Lane 1978). The correlation of these sequences and the implications for regional stratigraphic interpretation is the subject of a later section — page 65.

2.3.1.7 Summary

The stratigraphy of the map area has been described in terms of six major map units, namely: a sequence dominated by coarse clastic rocks — the grit unit; the semipelite-amphibolite unit with distinctive amphibolite layers; the marble unit dominated by grey and rusty marbles; the quartzite unit, characterised by blocky grey and white quartzites; the pelite unit, a mixed assemblage of clastic and calcareous lithologies; and a carbonate-dominated sequence south of the Bigmouth Creek fault.
The lithologic assemblages can be grouped into four main "blocks" between which the boundaries are thought to be structural rather than stratigraphic. The blocks are:

i) the grit-unit
ii) marble - semipelite-amphibolite block
iii) pelite - quartzite unit block
iv) the zone south and west of Bigmouth Creek Fault

Because of structural complications, correlation between the blocks cannot be established with certainty within the thesis area. Map units within each block will therefore be correlated in turn with known or interpreted stratigraphy outside the map area.

2.4 Local Correlation of Stratigraphy

2.4.1 Introduction

The presence of the Adamant Pluton has caused a deflection of stratigraphy away from the regional northwest-southwest trend (Shaw 1980). Some map units have been deflected eastwards around the eastern and northern margins of the pluton i.e. the grit unit and marble and semipelite-amphibolite block. Other units are deflected westwards to run parallel to Norman Wood Creek before returning to the regional trend. It is convenient to discuss the local correlation of stratigraphy in terms of these two arbitrary groupings, i.e. stratigraphy entering the map area from the east - the eastern assemblage; and stratigraphy entering from the south and southwest - the western assemblage.
2.4.2 The Eastern Assemblage

Those units entering the map area from the east, i.e. north of the Adamant Pluton are: the grit unit, the semipelite-amphibolite unit, the marble unit and those clastic rocks above the marble unit. The semipelite-amphibolite and marble units are directly correlative with units to the east and will thus be discussed first. For identification and correlation of stratigraphic units mentioned in text see Figure 7.

2.4.2.1 The Semipelite-Amphibolite and Marble Units

In areas of lower metamorphic grade where primary structures have not been destroyed by metamorphism, the stratigraphic succession of the Horsethief Creek Group is well established (Tippett 1976; Brown et al. 1978, Poulton and Simony 1980). To the east and southeast of Mount Sir Sandford (Figure 1), amphibolite-bearing rocks of the lower pelitic member at the base are overlain by a marble succession, the middle marble member, overlain in turn by the clastic rocks of the upper pelitic member (Figure 7). East of and including Mount Sir Sandford, Tippett (1976) illustrates the three members exposed right-way-up in an unfaulted section (Figure 8A) with the lower pelite occupying the core of a large anticline, flanked to the west by the middle marble and upper pelitic members. The principal lithologies of the three members are described in the introduction to this chapter.

The stratigraphy of the western limb of the anticline can be traced northward around the eastern end of the Adamant Pluton, where the upper pelitic member disappears against the pluton (Figure 7, insert). At the northern margin of the pluton, the upper pelitic member reappears and graded grits indicate that stratigraphic tops face the pluton (Shaw
FIG 8. PUBLISHED CROSS-SECTIONS, NORTHERN SELKIRK MOUNTAINS
(after Brown and Tippett, 1978)

LEGEND

Paleozoic
6. Nappanee Group
5. Bonneterre Group
4. Portage Group
3. Jessee Group
2. Bonnechere Group
1. Meaford Group

Proterozoic - Paleozoic
1. Upper Pennicak Member
2. Middle Pennicak Member
3. Lower Pennicak Member

Proterozoic
1. Horsehead Creek Group
2. Middle Creek Group
3. Lower Creek Group

For Location of Cross-sections refer to insert, Figure 3.
1980). Down section from the grits are the middle marble and lower pelite members, in that order. Minor structure data indicate that the lower pelitic member occupies the core of a large anticlinal structure (Shaw 1980). Stratigraphic and structural facing directions of the right-way-up or western limb of the anticline have clearly been maintained around the east end of the pluton.

Northwestwards along strike, rocks lithologically equivalent to the lower pelitic, middle marble and upper pelitic members of the western limb of the anticlinal culmination can be traced across the Windy Range (Figure 88) through Mount Chapman to Mud Glacier in the west (Figure 3). No sedimentary tops were observed in the clastic rocks which overlie the marbles. The succession is, however, the same as that described above i.e. clastic rocks above marbles which overlie in turn amphibolite-bearing semipelites and pelites in the core of an antiform. Almost continuous outcrop along strike from Austerity Creek to Mud Glacier and good structural control suggests that the antiformal structure within the thesis area is anticlinal, and that the clastic rocks overlying the marbles, the marble unit and the underlying amphibolite-bearing unit correlate with the upper pelitic member, the middle marble member and lower pelitic member respectively (Figure 7). Stratigraphy therefore, remains right-way-up across the map area and tops face southwest from the anticlinal core situated to the south of Mud Glacier and Mount Chapman (Figure 3).

To the northeast of, and adjacent to, the anticline described above, a large synform is cored by a carbonate-bearing sequence previously assigned to the marble unit (Figure 3). The synform is best exposed in the vicinity of Neptune Peak (Figure 6, Section 3).
Underlying the marbles are rocks assigned to the semipelite-amphibolite unit. The pelite unit which elsewhere overlies the marble unit (Figure 3) was not mapped.

Lithologies which outline the synform are separated from the middle marble and lower pelitic members to the south by a laterally persistent fault of reverse sense downthrown to the northeast (e.g. Figure 6, Section 3). The carbonate sequence and underlying rocks, however, are here correlated to the middle marble and lower pelitic member respectively on the basis of lithologic similarity and the superposition of marbles above amphibolite-bearing clastic rocks. The synform is therefore a syncline and stratigraphy remains right-way-up at least as far north as Northeast Fault (Figure 3).

In the northern Windy Range (Figure 3), a body of nepheline syenite cores a recumbent second phase antiform (Figure 6, Section 3). Wherever observed in outcrop, the nepheline syenite is concordant with the lower pelitic member which is folded about it. The axial surface trace of the antiform extends southeastwards into the area mapped by Shaw (1980). A small outlier of nepheline syenite is separated from the main body by a normal fault (Wheeler 1965), the trace of which is approximately defined by the sharply incised stream valley northwest of Trident Mountain (Figure 3). The fault is downthrown to the east. The nepheline syenite is only found in the vicinity of Trident Mountain, but the lower pelitic member which enclosed it can be traced westwards to Mount Chapman and the northern headwaters of Mica Creek (Figure 3). From stratigraphic relationships established to the south, stratigraphic tops should face outwards from the syenite. The antiform
at Trident Mountain is therefore an anticline (Figure 6 and Cross-section 3).

Thus the central fault-bounded block contains stratigraphically right-way-up lower pelitic member, marble and in a syncline at the southern margin of the block, upper pelitic member. These units have been traced (Wheeler 1963, 1965; Tippett 1976; Brown et al. 1978; Shaw 1980, Poulton and Simony 1980, and the writer) from Rogers Pass (Figure 1) to the ridges immediately east of the Columbia River on the western margin of the map area (Figure 3). Despite the presence of major second and third phase folds, the correct stratigraphic superposition and facing directions have been maintained throughout.

2.4.2.2 The Grit Unit

2.4.2.2.1 Introduction

The middle marble member does not reappear to the north of the anticline cored in the east by nepheline syenite (Figure 3, Figure 80). Instead, the lower pelitic member is in contact with a succession of grits, conglomerates, thin rusty marbles, pelite, semipelites, quartzites and psammites collectively designated the grit unit.

The stratigraphic position of the grit unit in the northern Selkirks is uncertain. The unit can only be traced a relatively short distance eastwards into the area mapped by Shaw (1980) before being lost at the limit of mapping in the heavily timbered slopes above Columbia Reach. The unit, as defined by the writer (this thesis), does not readily correlate to the northwest with stratigraphic units identified by workers from the University of Calgary (Ghent et al. 1977).
Because of the difficulties involved in the direct correlation of this unit, it is pertinent to briefly describe the occurrence of coarse clastic rocks in the Horsethief Creek Group as described by other workers in the southeastern Canadian Cordillera. Coarse clastic lithologies have been described from both the upper and lower Horsethief Creek Group, namely the upper clastic division and the grit division of Young et al. (1973). Each is now briefly summarized, beginning with the grit division.

2.4.2.2.2. The Grit Division

In the Central Purcell Mountains, the entire Horsethief Creek Group is dominated by grits. It also includes conglomerate, quartzite, slate and limestone (Reesor 1973).

In the northern Purcell Mountains (Dogtooth Mountains, Figure 7) where the upper and lower subdivisions are first recognizable, the basal feldspathic grit unit consists of interbedded subarkosic wackestone and pelite, with minor amounts of conglomerate and limestone (Simony and Wind 1970; Poulton and Simony 1980). The middle part of the grit division in the Dogtooth Mountains contains abundant graded beds (Poulton 1973). The basal parts of the grit beds are as coarse as pebble-grade (Poulton and Simony 1980). The division fines upwards into the slate division over a thick interval of interbedded pelite and sandstone.

North of latitude 53N, massive alternating beds of coarse-grained immature grit and phyllite of the Middle Miette and Kaza Groups compose the grit division of the Windermere Supergroup (Campbell et al. 1973).
North of McBride, approximately one-third of the Middle Miette consists of units of coarse-grained and pebbly sandstone (3 to 160 metres thick). Conglomeratic beds grade into or are interbedded with the sandstone (Young et al. 1973). Alternating gritty micaceous quartzite and schists characterise the Kaza Group in the Northern Caribou Mountains (Sutherland-Brown 1963; Campbell et al. 1973). The sandstones are mainly arenaceous, but granules and pebbles are common. In the Miette Group of the same area, cobble-sized clasts of limestone, green phyllite, chert, calcareous sandstone and sandy limestone have been noted.

2.4.2.2.3 The Upper Clastic Division

The upper clastic division in the Dogtooth Mountains (Figure 7) is differentiated from the overlying Hamill Group by the abundance of slate, the lack of cross-bedding and the presence of carbonates. In the northwestern Dogtooth Mountains, a succession of pelitic rocks with quartzite interbeds occurs below a thin carbonate-quartzite unit and above rocks equivalent to the limestone unit (carbonate division). Poulton (1973) compared the carbonate-quartzite strata to a western facies equivalent of the limestone unit, based on a correlation of the uppermost carbonates with those in the northeastern Dogtooth Mountains. Alternatively the uppermost carbonates may be unique, in which case they and the clastic layers below would represent a unit (or units) distinct from and stratigraphically above the limestone unit, i.e. the upper clastic division. The latter alternative was adopted by Poulton and Simony (1980). This situation is analogous to both the succession of limestones above pelites and grits in the western facies equivalent of the upper pelitic member (Lane 1979; Brown et al. 1978), and the
carbonates in the upper clastic division of the Caribou Mountains (Young et al. 1973).

At Rogers Pass (Figure 1), at least 700 m of laminated pelite with minor coarse clastic rocks including feldspathic grits are exposed below the Hamil Group and above the Limestone Member (Poulton 1970).

On the Esplanade Range (Figure 8a), pelites, metasandstones and metagrits overlie the Limestone Member (De Vries 1970; Tippett 1976). Massive grits change along strike into a predominantly pelitic assemblage (Devries 1970) such that two facies of the upper clastic division were thought to be represented. Structural cross-sections through the Esplanades indicate a maximum thickness of 1,000 m for the upper clastic division. An alternative interpretation involving thrust emplacement of grits of the grit division has been proposed by Poulton and Simony (1980).

The upper clastic division thickens westwards to about 2,900 m at Sonata Mountain (Tippett 1976). Conglomerates and grits overlie a basal assemblage of pelites and psammite and are overlain in turn by finely laminated pelites (Figure 7). At the base of the coarse clastic section is a lensoidal quartzite pebble to boulder conglomerate, which also contains pebbles of medium-grained buff carbonate (Tippett 1976). The bulk of the clastic assemblage above the conglomerate consists of fine pebble to granule grits, the coarsest of which is found at the base, such that a crude fining-upwards cycle can be defined (Tippett 1976).

The upper clastic division in the Caribou Mountains may be represented by the Yankee Belle Formation, the basal two-thirds of which consists of two major depositional cycles of interbedded
siltstone, shale and limestone (Young et al. 1973). The carbonates
within and capping the cycles strongly resemble the underlying
Cunningham Formation (Young et al. 1973). Coarse clastics have not been
observed.

From the above discussion it is clear that the coarse clastics
in the thesis area could belong to either the grit or upper clastic
divisions. Indeed both these correlations have been proposed: by
Campbell (1968) in his interpretation of the rocks as Kaza Group (Grit
Division) and by Brown et al. (1977) in their preliminary correlation of
the grit unit to the upper pelitic member. Ghent, et al. (1977)
described another possible correlation. They subdivided
stratigraphically equivalent clastic rocks to the east of Mica Creek
into three lithologic groupings and correlated the respective groupings
with Middle Slate, Carbonate Unit and lower part of the Upper Slate
(upper clastic division) of the northern Purcell Mountains (Simony and
Wind; 1970). Each of the above interpretations will now be examined and
the stratigraphic and/or structural problems inherent in the various
correlations discussed. Conclusions reached via these discussions form
much of the basis for regional correlation of stratigraphy presented in
Figure 7.

2.4.2.2.4 Correlation with Upper Clastic Division

The provisional correlation of the clastic rocks on the
northern margin of the thesis area with the upper pelitic member (Brown
et al. 1977) was based on lithologic similarity to rocks of the upper
pelitic member from the Sonata Syncline (Figure 8b) as described by
Tippett (1970). In Figure 8b, Brown and Tippett (1978) illustrate the
upper pelitic and middle marble members folded about the lower pelitic member. Both upper pelite and middle marble are interpreted to be present at the eastern margin of the Esplanades Syncline as defined by De Vries (1971). Stratigraphy is right-way-up with respect to third phase structures illustrated (Figure 8b).

In the thesis area it has been demonstrated (above) that stratigraphy to the southwest and west of Trident Mountain was right-way-up prior to both second and third phases of folding. Despite faulting, facing directions are maintained throughout (Figure 6, Section 3). The logic of cross-section construction dictates that both middle marble and upper pelitic members should reappear in that order to the north of Trident Mountain Anticline (Figure 6 Section 3). Middle marble was not mapped by the writer but was assumed to be missing at a shear zone represented by Northeast Fault. The presence of coarse clastics to the northeast of the anticlinal culmination, cored by the lower pelitic member (Figure 3; Figure 6, Section 3) was, however, considered by the writer to be additional support for the correlation of these rocks with the upper pelitic member.

Detailed mapping revealed that the younging direction within the grit unit was to the southwest, and not the northeast as expected. North of Northeast Fault, adjacent to Trident Mountain, the grit unit occupies the right-way-up limb of a northeasterly overturned third phase anticline (Figure 6, Section 3). Sedimentary tops face south towards the lower pelite and north as they should if the middle marble member had simply been removed by "ductile shearing" during synmetamorphic strain. A fault of reverse motion which was responsible for "shearing out" the synclinal core (Figure 6, Section 3) is required
to juxtapose southwesterly facing upper pelite to the north of
northeasterly facing lower pelite. Facing directions are also opposed to
the north of Mount Chapman (Figure 6, Section 2). Elsewhere along the
length of Northeast Fault, facing directions concur. A reverse fault is
still necessary however to satisfy the requirements described below.

This situation does not preclude the possibility of the
presence of the upper pelitic member along the northern margin of the
thesis area. Such a correlation would be based simply on lithologic
similarity.

2.4.2.2.5 Correlation with the Grit Division

It has already been pointed out that stratigraphic facing
directions are not constant along the strike length of Northeast Fault.
South of Warsaw Mountain, however, facing directions to both the north
and south of Northeast Fault are to the southwest. Here stratigraphy
might be regarded as a simple monocline sequence younging to the south.
The grit unit would correspond to the grit division at the base of the
Horseshoe Creek Group (Figure 7). This correlation relies on the
assumption that rocks to the north of the fault are stratigraphically as
well as structurally beneath the lower pelite, and that the fault is
normal. These assumptions are investigated below. The area mapped by
Mitchell (1976), Ghent et al. (1977) and by the writer overlap in the
vicinity of Warsaw Mountain. Details of this restricted area are
discussed first to aid later correlation.

Campbell's (1968) correlation of strata around Warsaw Mountain
with the Kasa Group was questioned by Mitchell (1976). Mitchell
Figure 9: Structural geology of the Warsaw Mountain area.

- Second phase cleavage (S2)
- Z, Z indicate sense of vergence
- Third phase cleavage (S3)
- Axial surface trace, third phase antiform
- Synform
- Second phase synform

la, b, etc. - Gris Unit lithologies
2a, b - Semipelite-amphibolite unit lithologies

Pelite
Icefield

See Figure 3 for explanation of symbols.
concluded that the thick pelite (centre, Figure 9) was probably equivalent to the Isaac Formation of the Caribou Mountains (Sutherland-Brown 1963). Clastics north and south of the pelite were considered equivalent and correlated to the Yankee Belle Formation (Holland 1964). The thin lensoidal marble band present only locally near the contacts was thought to represent the Cunningham Formation (Campbell et al. 1973; Figure 7).

Mitchell cites lithologic similarity as evidence for equivalent stratigraphy on either side of the pelite. Repetition of stratigraphic units would indicate the existence of a large, tight fold with the pelite in the core. By measuring dips at, or close to, the contacts and projecting them upwards in cross-section, Mitchell defined an upwards closing first or second phase structure (Mitchell 1976). If the correlation of the pelite with the Isaac Formation, carbonate with the Cunningham Formation and clastic units with the Yankee Belle Formation is correct, then stratigraphy would have been right-way-up before folding and the structure an anticline. This interpreted closure, centred on Mount Fred Laing, is illustrated in Figure 10, Section A-B, from Ghent et al. 1977.

The informal map units of Ghent et al. (1977) are correlated to the stratigraphic subdivisions of the Dogtooth Mountains (Evans 1933; Simony and Wind 1970) rather than to terms used in the Caribou Mountains (Campbell et al. 1973). The correlation is as follows: pelite unit - middle slate; marble unit - carbonate unit; and semi-pelite-amphibolite unit - lower part of the Upper Slate Unit (Figure 7). The areal distribution of the four units is shown in Figure 10.
FIGURE 10:
GEOLOGY OF SOUTHERN CANOE RIVER AREA

(After Ghent, et al., 1977)

LEGEND

- SPA: Semipelite-amphibolite unit
- Marble band unit
- Pelitic unit
- Gneiss unit
- Gneiss
- Major phase I assim surface
- Staurolite-schist
- Disappearance of staurolite line
- Staurolite-kyanite schist
- Thrust fault
- Normal fault
- End points of cross section lines

DIMENSIONS:
0 1 5
0 1 10
MILES
Kilometers
The writer, working in the southeasterly extension of the same lithologies from Yellow Creek to Trident Mountain (Figure 3), has been unable to document either stratigraphic equivalence across the pelite or the existence of an early fold in the area mapped by Mitchell (1976). The contact of the pelite with the amphibolite-bearing semipelites to the south, moreover, shows evidence of shearing (Plate 1G, H) and appears to represent a fault contact (Northeast Fault). Shearing at the contact can be documented along the length of the fault within the thesis area. The presence of a fault would not in itself preclude the existence of a fold, because the significance of the fault and the inferred displacement thereon is dependent upon the interpretation of the stratigraphic position of strata on both sides of the break. Such an interpretation is considered in the following section.

South of the pelite zone (Figure 9), both second and third phase minor folds indicate an antiform, closing to the north. Within the less competent pelites, minor folds are common but are mainly a reflection of strong crenulation developed during the third phase. Second phase fabrics are crenulated while the folds, outlined by thin psammitic bands, are involved in interference folding. So change in the vergence of second phase minor folds can be observed across the pelite zone. North of the pelite, second phase minor structures, although rare, are again clearly overturned and verge to the north (Figure 9). Second phase minor structure data therefore provide no support for the existence of a second phase anticlinal fold about the pelite south of Warsaw Mountain. The anticline in question projects westward to underlie Mount Fred Laing (Figure 10, Section A-B).
In the absence of common, recognisable first phase folds it is not possible on structural grounds to rule out the presence of a first phase structure. Such structures are usually documented by stratigraphic repetition. Columns A and B (Figure 11) illustrate the comparison of lithologies from the north and south sides of the pelite as mapped by the writer. Two different lithologic groupings are represented. The southern column (A), comprises a section through "typical" lower pelitic member where amphibolites are interbedded with semipelites, pelites and occasional psammites, the whole being overlain by the middle marble member. The northern assemblage (Column B) is the grit unit at Warsaw Mountain as described at the beginning of the section. The transitional contact with the pelites in the north contrasts with the sharp southern contact. Lithologically, the columns are distinct and no fold can be justified on the basis of stratigraphic repetition. In addition, graded beds in the grits of the northern assemblage (Figure 9) indicate tops towards the pelite; i.e. facing into the core of the anticline proposed by Mitchell (1976) and Ghent et al (1977). The above arguments lead the writer to conclude that:

1. A major pre-third phase fold does not exist in the vicinity of Warsaw Mountain.

2. Stratigraphy is not repeated across the pelite, and the clastic rocks to the north of the pelite do not therefore belong to the semipelite-amphibolite unit as suggested Ghent et al. (1977). Rather, the grits and pelite are overlain by the semipelite-amphibolite unit, separated from them by Northeast Fault.
In Figure 11, Column C is the total stratigraphy from Columns A and B combined on the premise that no major fold exists, that little or no stratigraphy is missing at the fault, and that the whole is a homoclinal sequence. Column D is the stratigraphic succession from Ghent et al. (1977) and Simony et al. (1980). The lithological similarity between Columns C and D is striking. The upper parts of both columns consist of a semipelite-amphibolite unit and a pelite unit, the mapping of which is agreed upon by workers from the University of Calgary, and the writer. The upper parts of the two columns, including the pelite, are therefore considered to be lithologically and stratigraphically equivalent.

The coarse psammite and grit bands near the base of the pelite unit (Ghent et al 1977; Simony et al 1980) are probably equivalent to the transitional zone of Column C. The increase in dark mica schist at the top of the grit unit (Column D) would also be part of the same transitional sequence. At the base of both columns are grits, psammites and pebble beds (conglomerates in Column C). The obvious lithological similarity between Columns C and D and the accepted correlation of their upper sections leads the writer to conclude that the lower sections are also probably stratigraphically equivalent. If such a correlation is valid then the bulk of the Horsethief Creek stratigraphy described in detail earlier in the chapter i.e. the pelite unit, marble unit etc. is stratigraphically higher than the bulk of stratigraphy described by Ghent et al (1977) from the Canoe River map sheet. This suggestion, initially presented verbally by the writer at a Cordilleran workshop at Queens University in 1977 has now been accepted in print by Poulton and Simony 1980.
Figure 11:
Correlation of Stratigraphy in the Warsaw Mountain area
It has already been suggested that evidence for the anticlinal structure in Warsaw Mountain is unfounded, and that no closure exists. In the cross-section in Figure 12, drawn immediately to the east of and parallel to section A-B (Figure 10), the fold cored by pelite has been outlined and the two sections fit together as one. The grits at the base of the stratigraphic succession (Figure 11) are continued southwestwards under the major hook-like syncline (Figure 12) to re-appear on the north-facing slopes of Warsaw Mountain where they correlate with the lower part of the grit unit (this thesis). This correlation is not unreasonable because grits are shown to wrap around the southern end of the hook-like structure on Szent et al.‘s map (Figure 14), even though the grits are not represented in cross-section A-B (Figure 10). Strata of the semipelite-amphibolite unit are carried northwards over the intervening anticline, to re-appear in the core of the downwards-closed hook-like structure outlined above. Stratigraphic tops are maintained throughout the section. The belt of stratigraphy from Warsaw Mountain to Trident Mountains which sits structurally beneath the semipelite-amphibolite unit is therefore thought to be equivalent to the grit, brown zone and pelite units (Szent et al. 1977). The westwardly, first phase mappe of Szent et al. (1977; see Figure 10, Section A-B) is not required to explain the distribution of stratigraphy in the vicinity of Warsaw Mountain.

The correlation presented above does not contradict the position of the Esplanades synclise in the south (Figure 3). Peltier and Sumonyi (1980) were able to trace the grit and slate divisions from the western flank of the Dogtooth Range northwestwards through Rogers Pass to underlie the Esplanade Range (Figure 1). Graded bedding
FIGURE 12: Schematic Structure Section, Warsaw Mt. Area.
indicates that the entire succession faces westwards. Both grit and slate divisions were emplaced by eastwards transport on the Explanades Thrust prior to both second and third phases of folding. A later normal fault, downthrown to the west, separates east facing semipelite-amphibolite (lower pelitic member) from the west facing division (Poulton and Simony 1980).

This situation is analogous to the eastern part of the thesis area i.e. Trident Mountain, where a fault separates southwest facing clastic rocks from locally northeast facing lower pelitic member (Figure 6, section 3). If the clastics in the thesis area are indeed equivalent to the grit division, then the fault would be a normal fault downthrown to the west as in the Explanades.

2.4.2.6 Conclusions

From the foregoing it is apparent that the grit unit as defined in the thesis area could belong to either the upper clastic or grit divisions. A correlation with the grit division is favoured by the writer for the following reasons:

1. The semipelite-amphibolite unit is at the top of the succession of Ghent et al. (1977) within the Canoe River and Warsaw Mountain areas (Figure 9), but at the base (excluding the grit unit) of the succession within the northern Big Bend (Brown et al. 1977; 1978).

In addition the thick pelite structurally and conformably beneath the semipelite-amphibolite unit is known to stratigraphically underlie the semipelite-amphibolite in the Canoe River area (Figure 9). It follows that the remainder of the grit unit probably also
underlies the semipelite-amphibolite unit to form the base of the succession.

2. Both second and third phase structures are overturned to the northeast, climbing towards the Rocky Mountain Trench (see cross-sections in Figure 6). Despite the presence of intervening faults, stratigraphy consistently youngs toward the southwest; the oldest part of the succession should be exposed in the northeast.

3. In both the Esplanades Range (Figure 1) and the thesis area, the coarse clastic sequences (grit units) are to the east and northeast respectively of the major anticlinal culmination cored by the lower pelitic member. In the Esplanades the westwards facing grit division is separated from the east facing lower pelitic member by a normal fault, downthrown to the west (Poulton and Simony 1980). North of Trident Mountain, an analogous situation exists where north facing lower pelitic member is separated from south facing grit unit by Northeast Fault. If the grit unit in the map area correlates with the grit division, the fault would have a normal sense of movement. The two areas cannot be directly correlated because the eastwards deflection of stratigraphy about the Adamant Pluton apparently causes the grit unit to be truncated at the Purcell Fault in the Rocky Mountain Trench. Nevertheless, the stratigraphic and structural similarities between the two areas are compelling arguments for their correlation.

From the above it is proposed that the grit unit be correlated with the grit division of Young et al (1973).
2.4.3 The Western Assemblage

The western assemblage consists of lithologies to the south and west of Birch Creek fault, namely the quartzite unit (map unit 5, Figure 3), the pelite unit (map unit 4, Figure 3) and the carbonates and associated stratigraphy southwest of Bigmouth Creek Fault (Figure 3). These units enter the map area west of Adamant Pluton.

Interpretations involving the physical tracing of stratigraphy through the head of Norman Wood Creek are complicated by the existence of several faults and the structure-controlling western margin of the Adamant Pluton. The area at the head of Norman Wood Creek will hereafter be referred to as the Norman Wood fault zone, and although the faults involved are described in detail later, some reference to individual faults within the zone is made to aid correlation.

2.4.3.1 The Quartzite Unit

Tippett (1976, p. 32) has documented the presence of quartzites within the middle marble member, and it is possible that quartzites south of Mount Chapman (Figure 3) are simply their stratigraphic equivalent. If this is the case, then the faulted contact with the marbles could simply be a local feature. Alternatively the quartzites are part of the Hamill Group, as initially suggested by Wheeler (1965). A brief discussion of some known Hamill strata to the south, followed by a comparison of these lithologies with those of the Mount Chapman section, may help resolve the problem.

The Hamill Group is exposed in a series of synclinal 'keels' from the Ventego Syncline (Wheeler 1965) in the east to at least as far
as Downie Creek in the west (Figure 1). At most localities the contact between the Hamill Group and the Horsethief Creek Group is faulted (Wheeler 1964; Lane 1976). In at least one locality where the Basal Hamill Group is preserved, an unfaulted sequence occurs between the two groups (Tippett 1976).

In the Venteqop Syncline, the Hamill Group is characterized by a basal assemblage of finely laminated, crossbedded, chlorite-magnetite-bearing schists, coarse grained quartzite, calcareous quartzite, medium-grained quartz-feldspar grits, dark grey psammites and massive buff dolomitic carbonate (Tippett 1976). Lane (1977) documented westward facies changes in the lower Hamill Group, the most significant factors of which are the disappearance of the crossbedded quartzites and the appearance of metavolcanics in the west. The chlorite-magnetite-bearing schists of the eastern basal assemblages are considered probable lateral equivalents of the metavolcanics (Brown et al 1978).

The western outcrops of the Hamill Group (Figure 3, insert) are perhaps the most useful for comparative purposes. If Hamill Group strata do indeed persist through the structural culmination at the head of Norman Wood Creek, then rocks on strike with western exposures should appear. At the base of these western assemblages, quartzites with thin pelitic interbeds contain two thin bands of dolomite (Lane 1977). A mafic volcanic sequence with ultramafic and sedimentary rocks which are in part volcanogenic, occur above the basal sequence. The sedimentary rocks contain a high proportion of chlorite and actinolite, rock fragments and significant carbonates.

The strata south of Mount Chapman are lithologically similar to those described from outcrops of Hamill elsewhere, particularly those
at Sorcerer Creek (Lane 1976). The green semipelitic schists with actinolite are possible equivalents of the ultramafics. The lithologic similarities, the faulted nature of the contact, and the strike position of these strata with respect to outcrops of Hamill Group in the south leads the writer to conclude that the quartzites to the south of Mount Chapman (Figure 3) are probably part of the Hamill Group and do not belong to the middle marble member of the Horse Thief Creek Group.

2.4.3.2. The Pelite Unit

The pelite unit south of Birch Creek fault is a succession of carbonates, calcareous pelites, pelites, semipelites, psammites and rare quartzites which lie to the south of and underlie the Hamill Group. Psammites and quartzites at the base give way upwards to finer grained clastics and calcareous rocks. Immediately beneath the Hamill the succession includes grey and rusty marbles and calcareous schists. Hornblende schists and amphibolites occur intermittently. The amphibolites contain no garnets and are locally coarse grained and locally crosscut stratigraphy, suggesting an igneous origin.

The succession is illustrated in Figure 5 - Column 10, and the distribution of rock types is shown in Figure 3.

The position of the unit beneath the Hamill Group suggests that it should correlate with the upper pelitic member. Brown et al. (1978) document a progressive increase from east to west in the amount of carbonate in the upper pelite member from a pelitic succession with minor carbonate in the east to a dominantly carbonate succession in the west. Carbonate rocks occur especially immediately beneath the base of the Hamill (Lane, 1977). Figure 5, Columns 10 and 11 illustrate the
similarity between the pelite member of Lane (1977) and that described by the writer.

The sequence of carbonates above a clastic assemblage beneath the Hamill Group suggests that the pelite unit is correlative with the western calcareous assemblage of the upper pelitic member (Brown et al., 1978).

If the correlation of stratigraphy presented above is correct, then the fault at the northern contact of the Hamill Group, previously named the Birch Creek fault (above), is a structure of regional significance.

The whole upper pelitic member and possibly the upper part of the middle marble member have been omitted from the section. The lensoidal nature of the middle marble member is such that an accurate estimate of thickness is unobtainable. The missing section of this unit could vary from 0 - 300 m. An unknown amount of the Hamill Group may also be missing. Assuming an average thickness of 3000 m for the upper pelite member (Tippett 1976; Lane 1977) and recalling that the Hamill Group south of Birch Creek Fault is right way up, the minimum displacement is probably in excess of 3 km. This figure would be increased considerably if a large thickness of Hamill Group is missing across the fault. Birch Creek Fault, as interpreted, is thus a normal fault, downthrown to the south.

2.4.3.3 Correlation of Stratigraphy South of Bignouth Creek Fault

Stratigraphy south of Bignouth Creek Fault cannot be physically correlated to the east, with known stratigraphy from within the map area. The thick grey marbles are on strike with lithologically similar marbles in both Argonaut Mountain and south of Bignouth Stock.
(Figure 3) where marbles have been correlated with the middle marble member of the Horseshoe Creek Group. (Brown et al. 1977; Brown and Tippett 1978; Brown et al. 1978).

To the west the southern of the two thick marbles (Figure 3) has been mapped westwards from the map area downslope into the valley of the Columbia River. Although not mapped in detail south of this point the marbles have been walked out southwards along the power line and new Revelstoke to Mica Creek highway to the mouth of Bighorn Creek. At this point the strike of the marbles turns slightly to the west, i.e., into the Columbia River.

Farther south, Wheeler (1965) traced marbles of the Badshot Formation and overlying Larder stratigraphy westwards from the north flank of Goldstream Pluton (Figure 1 - pluton south of Goldstream River) and into the Columbia Valley. Brown and Psutka (1978) have traced the same marbles across the Columbia River and northwards to a point almost opposite the mouth of Bighorn Creek, where the marbles disappear under the Columbia River. It seems reasonable to suggest that the marbles mapped and walked out by the writer and marbles in the area mapped by Brown and Psutka (1978) are correlative and belong to the Badshot Formation.

In the map area the marbles are overlain by dark pelites and underlain by a mixed clastic assemblage composed of pelites, semipelites and calc-silicates with minor psammitic and marble. Regional considerations suggest that stratigraphy may be inverted south and west of Bighorn Creek Fault and its continuation southwards down Norman Wood Creek (see page 97 for discussion).

If both assumptions are correct i.e., that the marbles are Badshot Formation and that the stratigraphy is inverted, then
lithologies structurally beneath the marbles would correlate with the Lardeau group and the pelites overlying the marbles would probably correlate with the fine grained clastics of the Mohecan Formation of the Hamill Group.

The marbles and adjacent stratigraphy appear to be structurally and stratigraphically correlative with marbles in Argonaut Mountain and south of Bigmouth Creek Stock (Figures 3 and 13). If such a correlation is correct then the stratigraphy which underlies that area between Goldstream River and Bigmouth Creek west of Adamant Pluton would belong to Cambrian Hamill Group, Balschat Formation and Lardeau groups rather than Horsethief Creek Group as suggested previously (Wheeler 1965; Franzen 1974; Van der Leeden 1976; Brown et al 1978). Additional detailed stratigraphic work is needed to test this hypothesis.
3. STRUCTURAL GEOLOGY

3.1 Introduction

With the possible exception of Domain A (Figure 13), the study area lies to the northeast of the Selkirk Fan Axis (Brown and Tippett 1978). Folds are overturned northeastward towards the Main Ranges of the Rocky Mountains. Bedding dips moderately to steeply to the southwest, as do the axial surfaces which fold bedding.

Despite the differences in elevation between valley floors and ridge tops, few major closures can be observed. In addition the shallow plunges of macroscopic folds cause strong parallelism of stratigraphic contacts across the map area (Figure 3) and closures can rarely be walked out on the ground. Where closures are observed, as for example at the head of Wintly and Norman Wood Creeks (Figure 3), they are outlined by the grey marble bands of the middle marble member. Elsewhere, the presence of major folds is indicated by the change in vergence of sets of minor folds across the axial surface of major folds.

Three distinctive phases of folding can be recognized in the northern Selkirk Mountains. The first phase is only rarely represented by mesoscopic folds, but in rocks of low metamorphic grade its presence is indicated by a weak foliation parallel to bedding. Also where stratigraphic tops are overturned to the facing direction of major second phase structures, it is apparent that large-scale folding has occurred prior to the second phase of deformation. The existence of major early structures is more difficult to demonstrate where structural and stratigraphic facing directions concur.
Mesoscopic-scale folds of the second and third phases are relatively common throughout the map area. Second phase folds deform bedding, have an axial planar foliation and vary in style from close to tight. Layer thickening in the hinge areas indicates a flexural flow mode of formation. The folds are genetically related to the strong regional foliation.

Third phase mesoscopic folds, which deform the regional foliation, are common in the northeast and in the southwest parts of the map area only. Elsewhere this phase is expressed as a crenulation cleavage.

Structural terminology used in the text for the description of folds/phases is fairly standard. The prefix D is applied to sequentially numbered deformational events during which macroscopic, mesoscopic and microscopic structures developed. The prefix F denotes folds related to these deformational events e.g. F2 folds were formed during the second deformational event D2. The prefix S refers to a structural deformational surface axial planar to minor folds where, for example, S2 is the foliation formed during D2. The prefix L refers to a linear element and may be the fold axis of a minor fold or the intersection of schistosity and bedding (S0). Thus L2 refers to a linear element developed during D2.

S1 and Sе are special cases used in the description of microfabrics to denote surfaces (S) internal and external to metamorphic crystals, usually in porphyroblastic form. The internal structure, usually defined by elongate quartz grains, is referred to as S1 (S-internal), while the foliation outside the crystal is the Sе (S-external).
Representative structural measurements from these events have been plotted on a structural base map (Figure 13) which also illustrates faults, the axial surface traces of major folds, and the boundaries of structural domains. At the sides of the map (Figure 13), equal-area stereographic projections illustrate mesoscopic fold information for bedding and second and third fold phases respectively. The plots of \( S_0 \) (bedding) also include those measurements taken from isolated first phase folds. Structural cross-sections are plotted separately in Figure 6. Features associated with the various fold phases are illustrated photographically in Plates 5 and 6.

The study area has been divided into ten domains (Figure 13) which correspond to areas dominated by either phase two or phase three folds or areas characterized by interference of the two phases. The boundaries of some of the domains correspond approximately to major faults. Domain A is located on the inverted limb of a first phase nappe; Domains B to F are located on a right-way-up limb.

### 3.2 Recognition and Identification of Fold Phases

#### 3.2.1 Introduction

The dangers of using style and orientation in the interpretation of complexly folded areas have been pointed out by Parker (1969) and Williams (1970). In addition to the problems discussed by these writers, the following aspects make the identification of fold phases in the map area difficult:

a) Metamorphic grade is uniformly high, rocks are coarse
grained and early fabrics have been recrystallised. Fine details of cleavage/nedding relationships cannot be observed in all outcrops.

b) Marker beds are absent from large sections of the map area, and sedimentary tops could be determined only in the grit unit.

c) Lithological units are laterally continuous and closures can seldom be observed or walked out on the ground.

d) Folds for the most part are statistically coaxial. The lack of dispersion of second phase planar surfaces and the similarity of their orientation with those of the third phase makes tedious the discrimination of fold phases by stereographic projection (see Figure 13). In addition the dispersion of third phase planar data is difficult to explain unless the nature of the third phase folds is studied in detail (e.g. Speny et al 1980).

e) The existence of major faults complicates the unravelling of major structures using stratigraphy.

f) Large bodies of pegmatite disrupt layering and transect folds.

Despite the above, the majority of folds within the area can be subdivided into two groupings, namely, those folds which have an axial planar foliation parallel to a strong regional foliation and later folds which fold the foliation. In rare isolated outcrops, however, it is possible to distinguish folds related to an earlier phase. Both limbs of these early isoclines are crosscut by the strong foliation ($S_2$) which is itself crenulated by the later third phase. An axial plane cleavage has not been observed in the axial areas of these early folds.
The fine-grained first-phase muscovite foliation reported elsewhere (Tippett 1976; Lane 1977; Brown and Tippett 1978) has probably been destroyed by recrystallization during metamorphism or by transposition during the second phase of deformation. Evidence to support these conclusions was gathered during a detailed study of microfabric relationships which was undertaken to help unravel the problems of timing and correlation of mesoscopic fabrics (Chapter 5). In the following section it is assumed, therefore, that the mappable foliation is S₂. Details of S₂ and the folds related to D₂ and D₃ are presented after a discussion of the significance, relative abundance and reasons for lack of preservation of fabrics and folds associated with D₁.

3.2.2

Direct evidence of the first phase of deformation is sparse. Rare isolated, intrafolial, long-limited isoclinal folds are preserved in psammite and quartzite in the northern part of the map area. Usually only two folial limbs are present and the sense of vergence is not known. However, in the allochthonous section of the Winkley Ranges (for discussion see page 84), isoclines of quartzite "floating" in a muscovitic pelitic schist preserve long-limited isoclinal minor folds. The isoclines have the opposite sense of vergence to the second phase structures about which they are folded, and therefore represent a previous phase of deformation. An axial planar cleavage has not been observed in the axial areas of these early folds, but both limbs are crosscut by a strong planar fabric related to the second phase of
folding. Planar and linear features associated with $D_1$ have been either destroyed by metamorphism or by transposition during $D_2$.

3.2.2.1 Microscopic Features Associated with First-Phase Folds

Although $S_1$ is described as a regionally pervasive foliation parallel to $S_0$ (Franzen 1974, Van der Leeden 1976, Tippett 1976, Lane 1977), the evidence for $S_1$ in thin section is sparse. Franzen recorded isolated examples in outcrop where garnet and biotite are elongate in $S_1$ and refolded by $S_2$. (Franzen 1974: Plate 23), but noted that the absence of a clear $S_1$ fabric in pelitic rocks, the approximately coplanar nature of $S_2$/$S_3$, together with an abundance of well-developed $S_2$ and $S_3$ makes identification of $S_1$ in thin section difficult. Furthermore, he suggested that although some biotite and garnet porphyroblasts may preserve $S_1$ in the form of straight trails, the abundance of porphyroblasts where $S_1$/$S_0$ indicates that in most cases the growth of biotite is syn- to post-$D_2$ and that the straight trails in the biotite are $S_2$. Van der Leeden (1976) reported evidence for the growth of muscovite and chlorite in $S_1$ before the appearance of biotite in the groundmass foliation. Above the biotite zone, $S_1$ is not observed in thin section but biotite porphyroblasts have straight trails, some of which may represent $S_1$. Lane (1977) described only sporadic evidence of a pre-$D_2$ foliation, where a strongly crenulate muscovite or graphitic schistosity is preserved between $S_2$ spaced-cleavage surfaces (Lane 1977; Plates 8 & 9).

Tippett (1976) reported $S_1$ as a well-developed alignment of muscovite, quartz, epidote and opaque which define a foliation penetrative through the noses of isolated $D_1$ isoclines. Biotite and
garnet porphyroblasts overgrew $S_1$ resulting in straight quartz inclusion trails. (Tippett 1976; Plate 18A). Rocks which had not developed an $S_1$ foliation developed an $S_2$ foliation mineralogically similar to $S_1$. Where $S_1$ previously existed, $S_2$ is a crenulation of that foliation with $S_1$ preserved between $S_2$ spaced-cleavage surfaces (Tippett 1976).

A summary of the above information reveals that $S_1$ has been reported as an initially planar fabric over which biotite and garnet porphyroblasts have grown. It is preserved most often as a curved schistosity between the spaced-cleavage surfaces of $S_2$. Some micaceous minerals previously oriented in $S_1$ have been rotated to lie in $S_2$ (Van der Leeden 1976; Plate 7; Lane 1977; Plates 8 & 9). $S_3$ is described by the same writers, as a crenulation of $S_2$. Where $S_3$ is well developed as a spaced crenulation cleavage, $S_2$ is generally overgrown and preserved as straight to slightly curved inclusion trails within porphyroblasts (Franzen 1974, Plate 22; Tippett 1978, Plate 19).

It would appear that there are two planar fabrics, each of which is overgrown by and preserved as trails within porphyroblasts, and each of which is crenulated by the successive phase of deformation. Without the large number of rock samples and detailed field observations presently available, it would have been a difficult task to distinguish between fold phases in thin section.

Figure 14A, for example, shows a concentric fold with a spaced axial planar cleavage which has folded a previous foliation and preserved it as straight inclusion trails. The porphyroblasts record no evidence of growth during crenulation, but simply deflect the tightly
A. Post D2 biotite porphyroblasts deflect a spaced S3 crenulation cleavage.  
Scale bar 7mm (.7cm)

B. Post D2 biotite has continued to grow into D3. Continued growth into D3 is indicated at the margins of the crystal, by a slight curvature in the otherwise straight quartz inclusion trails.  
Scale bar 5mm (.5cm)

C. Post D2 garnet includes both S2 (bottom right) and S1 quartz inclusion trails. The S1 trails are composed of very fine-grained, elongate quartz grains parallel to compositional layering.  
Scale bar 1mm (.1cm)

D. Chlorite porphyroblast preserves slightly crenulate S1 and thus records an early stage in the crenulation and transposition of S1. Pods in the groundmass preserve highly crenulate S1. The dominant foliation is S2.  
Scale bar 8mm (.8cm)

E. Post-D2 albite porphyroblast has been crenulated by D3. S2 inclusion trails within the porphyroblast show the same curvature as the crystal margins. Pods in the groundmass preserve remnant S1.  
Scale bar 5mm (.5cm)

F. Slight rotation of post-D2 cross-chlorite is due to gentle kinking during D3. Crenulate remnant S1 is preserved in pods in the groundmass.  
Scale bar 3mm (.3cm)
Figure 14: Microscopic features associated with S1
spaced vertical cleavage. Without additional evidence, two obvious possibilities exist.

Either:

1. The biotite overgrew, and preserved as straight trails, a planar S₁. Biotite growth had ceased before later crenulation (in this case S₂) and the porphyroblast deflects the spaced crenulation schistosity surfaces.

or:

2. The biotite preserves an initially-planar S₂ etc. In this case either S₁ was not developed in the sample or S₁ has been completely transposed into S₂ and the alternating quartz and muscovite microlithons are the result of this transposition.

In adjacent, less symmetrical folds, biotite porphyroblasts have continued to grow during rotation such that S₁=Sₑ, where Sₑ is composed of mica flakes and elongate quartz grains which have been rotated to lie in and define the spaced crenulation cleavage (Figure 14B).

Is the early foliation S₁ or S₂?

The following section attempts to describe in detail the features associated with S₁, to critically compare them with S₂ features, and in doing so, to provide reliable criteria by which S₁ can be recognized in thin sections. Only then can comments be made as to the abundance of S₁ in thin sections and to the timing of porphyroblast growth.

Figure 15 contains sketches of three rock slabs cut from the same sample (from Stn F9, Figure 24) which contains a folded psammite-pelite contact. All three phases of deformation are-
represented. The sketches were produced photocopying the dry but unvarnished surface of the smoothly ground cut clads. The main features can then be traced from the photocopy, and details checked against the rock-slab with the aid of a binocular microscope.

The three inserts are detailed sketches of fabric relationships from thin-sections cut from the areas marked on the small insert map. Orientations of the foliations in the insert sketches are the same as in slab 1 from which the thin-sections were cut.

Slab 1 shows the folded contact which is paralleled in the psammite by thin micaceous horizons. \( S_1 \) is represented as a weak muscovite-quartz elongation parallel to \( S_0 \) (insert A). \( S_2 \) is represented in the psammitic part of the hand-specimen as a series of fracture surfaces, and in thin-section as fine elongate muscovite and quartz at an angle to \( S_1 \). Where the angle between \( S_1 \) and \( S_2 \) is high, \( S_2 \) weakly crenulates \( S_1 \) and sinistral offsets of \( S_1 \) can be seen on individual fracture surfaces. \( S_3 \) is penetrative crenulation cleavage which rotates biotite porphyroblasts so that the long axes of the biotites line-up perpendicular to \( S_3 \). \( S_3 \) is penetrative only a short distance into the psammite.

Slab 2 illustrates an increase in the crenulation of \( S_1 \) in the psammite, and the development of small drag-folds between penetrative \( S_2 \) cleavage surfaces in the psammite. In slab 3, \( S_1 \) is highly crenulated in the upper part and almost completely transposed in the lower part of the slab.

Insert A shows \( S_0 \) in the psammite preserved as disrupted layering and \( S_1 \) as an elongation of quartz grains and a parallelism of muscovite flakes within \( S_0 \). \( S_0 \) and \( S_1 \) are considered to be
Fig 15: An Example illustrating the Characteristics of S1; some factors governing its Preservation and subsequent Transposition.
parallel. Within the biotite porphyroblasts $S_1$ is preserved as poorly
defined but essentially straight quartz inclusion trails. The
porphyroblasts are not rotated by $S_2$, but rather the margins of the
biotite partly include quartz elongate in $S_2$.

Insert B is in two parts (B and B') to show $S_1$/$S_2$
relationships in both the psammitic and the pelite adjacent to the
contact. In the psammitic (B') $S_1$ and $S_2$ are shown separated by a
small angle. $S_1$ is only well preserved within the muscovite and
opaque-rich layers. Outside these layers only rare muscovite flakes
indicate the presence and orientation of $S_1$. Biotites which include
quartz in $S_1$ also include quartz elongate in $S_2$. The larger
xenoblastic biotites are situated in the $S_2$/$S_1$ compositional
layering whereas small idiomorphic biotite flakes parallel the strong
muscovite foliation which defines $S_2$. $S_2$ is preserved as straight
inclusion trails in cross-biotites. The $S_2$ inclusion trails are
easier to recognize than $S_1$ trails due to the greater flattening of
quartz grains in $S_2$ before inclusion. $S_3$ is penetrative into the
margin of the psammitite layer, and it is only in areas unaffected by $S_3$
that the above relationships can be observed.

In the pelite (B) the situation is more complex. The
horizontal part of the psammitite/pelite contact is shown. Indisputable
$S_1$ is preserved in biotite porphyroblasts immediately adjacent to the
contact (position illustrated by circle in slab 1). To the right of
this point, biotite porphyroblasts probably include only $S_2$ as
indicated by the continuity of $S_1$ and $S_2$, and the increase in size
and elongation of the inclusions (lower part of B). Away from the
contact all porphyroblasts have been rotated by $S_3$ such that it is
difficult to differentiate porphyroblasts containing \( S_1 \) from the ones containing \( S_2 \). Details of the \( S_3 \) crenulation cleavage are shown in insert C.

While it is difficult to generalize on the basis on one sample, Figure 15 supplies valuable information on the form and mode of preservation of \( S_1 \), and reasons for its disappearance in thin-section. In the hand specimen, \( S_1 \) is easily identified in the psammitic layer on the basis (verified in thin-section) that \( S_1 \) is parallel to \( S_0 \). The competence of the psammitic accounts for the preservation of fine pelitic interlaminae, and the preservation of \( S_1 \) during the deformation. The incompetent pelite responds more easily to deformation, and earlier fabrics are transposed. Even in psammites, however, where second phase folds are tight to isoclinal, rather than being the monoclinal flexure such as that in Figure 15, \( S_1 \) is likely to be completely transposed on the limbs and possibly even in the hinge. At best all that remains of \( S_1 \) may be a series of tight crenulations in the axial zone.

It is also important to note that the sample in Figure 15 comes from low in the biotite zone where prograde chlorite is common. At higher grades of metamorphism, the early foliation is likely to be destroyed by recrystallization. All that may remain are porphyroblasts where \( S_1 \) is unrelated to \( S_0 \).

In the psammitic, biotites which have overgrown \( S_1 \) occur only in the pelitic interlaminae, indicating a compositional control in the siting of porphyroblasts. The same porphyroblasts also include \( S_2 \) in their peripheries. It is possible that biotite began to grow pre-\( D_2 \) to preserve \( S_1 \), and continued to grow during early \( D_2 \). This is not considered very likely, since porphyroblasts record no \( \text{Sym-D}_2 \).
rotational growth either where $S_1$ and $S_2$ are at right angles (Insert A) or where only a small angle exists between them (Insert B).

A rotational component to $S_2$ is indicated by the offsets of $S_1$ on penetrative $S_2$ surfaces. The edges of the biotite, however, simply show the relationship $S_1=S_e$, where $S_e$ represents quartz elongate in $S_2$. It is therefore more likely that biotite is post-$S_2$ and helicidal to both $S_2$ and poorly defined $S_1$.

An upper limit to biotite growth is Syn-D$_3$ (Figure 13c). Biotite porphyroblasts include, as curved margins to straight trails, the rotated groundmass foliation. The central straight part of the trail represents static growth over a pre-existing foliation. Since there are no visible discontinuities in the biotite crystals it is logical to assume that initial growth was post-D$_2$ and the straight part of the trails represent unfolded $S_2$.

Further evidence for a post-D$_2$ porphyroblast growth which includes and preserves $S_1$ and $S_2$ is illustrated in Figure 14c. A large garnet porphyroblast has overgrown compositional layering. The finer grained layer includes quartz and titanite elongate parallel to layering. In the outer part of the garnet, larger elongate quartz inclusions are parallel to, and at the margins continuous with, the external $S_2$ foliation i.e. $S_1=S_e$. Clearly, the garnet grew under static conditions later than the formation of $S_2$ which it includes as straight trails while at the same time preserving $S_1$ in the finer grained compositional layering.

Not all porphyroblast growth is demonstrably post $S_2$. Figure 14d shows a chlorite porphyroblast which includes gently crenulated quartz trails. The strong external fabric is $S_2$. Outside
the porphyroblast, the much finer grained and strongly crenulated S₁ is preserved as remnants between S₂ cleavage surfaces. Chlorite must have grown early in the development of S₂, before S₁ became almost completely transposed into S₂. Figure 14E and F (especially E) illustrate that remnant S₁ is recognizable, on a microscopic scale at low grades of metamorphism, even after two later episodes of deformation. It must be realized however that above the chlorite zone of regional metamorphism S₁ would probably be no longer recognizable in outcrop where later folding is strongly in evidence.

In the absence of mesoscopic D₁ folds, the recognition of S₁ in outcrop is difficult. The following is a list of features by which S₁ may be recognized, in outcrop and in thin sections. Most of these features apply only to areas underlain by rocks of low metamorphic grade.

Firstly in thin section:

1. The S₁ foliation is generally defined by fine flakes of muscovite, quartz, ilmenite and epidote and more rarely chlorite.

2. Where S₁ and S₂ have been observed together, the elongate quartz grains defining S₁ are always much finer grained than those in S₂.

3. Minerals elongate in S₁ are parallel to compositional layering.

Secondly in outcrop:

1. At low metamorphic grade, S₁ is penetrative through the axial regions of long limbed isoclinal first phase folds.
2. In appropriate lithologies, i.e. competent units with fine
pelitic interlaminae and/or finely interbedded lithologies,
S1 can be observed as a fine foliation parallel to bedding.

3. Generally S1 is preserved at lower grades of metamorphism
only. At higher temperatures, S1 is presumably destroyed by
recrystallization which accompanies the deformation.

4. In areas of S2 folding, S1 may be recognized as a
crenulated foliation between S2 cleavage surfaces.

In conclusion, S1 is elusive in thin sections from the
northern Selkirk Mountains. Polyphase deformation combined with high
temperatures, plus a tendency on the part of workers to sample pelitic
rocks, has resulted in there being little record of S1 in assembled
rock suites. Nevertheless S1 can be found, given the right
lithologies and conditions, and provides microscopic evidence of an
early foliation parallel to bedding. Such evidence supports the field
recognition by others (Franzen 1974; Van der Leeden 1976; Tippett 1976)
of mesoscopic long-limbed isoclinal folds considered to be first-
generation structures.

Given the conclusions reached above, it is unlikely that S1
would be preserved in areas underlain by rocks of sillimanite and,
kyanite grades of metamorphism. The dominant foliation is, therefore,
considered to be related to the second regional phase of deformation.
The sequence of folding in the map area has been established by using
S2 foliation surfaces as markers to separate early folds with surfaces,
cut by the foliation from later folds which deform the foliation.
3.2.3 Folds with Axial Planar Foliation (S<sub>2</sub>)

The dominant fabric throughout much of the area is a well-developed foliation defined in micaceous metasediments by coarse-grained muscovite, biotite and elongate quartz grains. Metamorphic grade, which ranges from kyanite to the breakdown of muscovite at the second sillimanite isostructural, has a uniform effect on the development of the foliation. Differences in intensity of foliation are due primarily to changes in lithology. In the pelites, transposition has resulted in schistosity being dominant over bedding. In finely interlayered lithologies, the foliation passes from one lithology to another with only minor deflection. Elongate quartz and feldspar grains define a coarse foliation in quartzites and psammites.

The foliation can be observed to be axial planar to and penetrative through the hinge areas of mesoscopic flexural flow folds (Plate 5A). In the absence of minor folds, the presence of macroscopic folds is recognized by changes in the angles and sense of intersection between schistosity and bedding. This intersection gives rise to a macroscopic intersection lineation which parallels synkinematic fold axes. Tabular minerals e.g. hornblende, as well as pods of sillimanite and quartz (Plate 5B), lie with long axes parallel to the axes of folds to form a mineral lineation (Plate 5C).

Folds associated with the foliation described above are tight to isoclinal and show moderate to strong hinge thickening and limb attenuation which indicates a flexural flow mode of formation (Plate 5D, Plate 2G, Plate 4E). These folds predate the peak of metamorphism and porphyroblasts of biotite, garnet, staurolite, kyanite and sillimanite have overgrown and included quartz elongate in the axial planar
foliation. Pods of silicate initially thought to be syn-F2 can be demonstrated to have grown metamatically on biotite elongate in S2 (Chapter 5).

3.2.4 Folds which Deform S2 Foliation

Mesoscopic folds which deform S2 and earlier fabrics are common in the northeast and southwest parts of the map area. Where the foliation-deforming event is limited to development of crenulation cleavage, competent horizons commonly exhibit open concentric folds, invariably upright in nature (Plate 5E, F). A spaced crenulation cleavage in the interbedded pelitic horizons is only approximately parallel to the axial plane and a slight fanning of crenulation cleavage is common (Plate 5E). Further examples of third-phase folds and associated spaced crenulation cleavage are illustrated in Plate 6A and 6B.

The distribution of both mesoscopic and macroscopic second and third-phase folds is discussed below within the various structural domains. Domains were chosen to highlight structural features i.e. the dominance of either phase two or phase three folds or the interference between the two phases. In some cases domain boundaries correspond to major faults.

3.3 Analysis of Mesoscopic Fold Data by Structural Domain

The outlines of the structural domains along with mesoscopic fold data are shown on Figure 13, while Figure 6 illustrates structural cross-sections referred to in the text.
3.3.1 Domain A

Domain A lies to the north of Bigmouth Creek Stock (Figure 13). Stratigraphy dips steeply to the southwest reflecting the dominant southwest dipping second and third phase axial surfaces. Stratigraphic dip reversal is probably due to several steep northeasterly dipping second phase structures. D3 folds plunge to the southeast and southwest which reflects refolding by the steep southwest-dipping third phase folds. The majority of D3 folds plunge to the west. The southerly dip of axial surfaces of D3 folds indicates that the domain lies to the northeast of the Selkirk Fan Axis (Brown and Tippett 1978).

Included on the stereoplot of third phase data (Figure 13, Domain A) are a few measurements from northeasterly trending structures. These are isolated grains superimposed on major third phase structures. Since they do not cause any reorientation of earlier mesoscopic folds, they are included on the plot of D3 data.

In the northern part of the domain, a large synform outlined by a thick marble sequence (Plate 4H) forms part of the northern overturned limb of a major third phase antiform (Section 1, Figure 6). Only second phase structures change their sense of vergence across the synform; third phase structures consistently climb to the south. The synform is therefore a second phase structure and the apparent closing to the east is probably due to a combination of the effects of topography (Figure 2) and the gentle westerly plunges of later D3 folds.

The second phase synform is probably a continuation of the French Creek Synform which is deflected by third phase folds to run westwards across Argonaut Mountain (Figure 13). The Plunge of third
phase structures becomes shallower to the west, passing through the horizontal over Bignight Creek Stock to become westward plunging in Domain "A". The change of plunge direction of third phase structures would re-introduce the equivalent of the French Creek Synform, as outlined by marble bands, into Domain A.

3.3.2 Domain B

Domain B is a small area which lies immediately to the northeast of the Selkirk Fan Axis. The area is characterized by a lack of structures related to the third phase of deformation. The orientations of $S_0$ and $D_2$ structures are illustrated in stereoplots on Figure 13.

The $D_2$ structures form part of the overturned eastern limb of the French Creek Synform (Figure 8C). To the south of the Selkirk Fan Axis, shallow easterly dipping second phase structures are superimposed on previously inverted stratigraphy (Van der Leeden 1976; Brown et al 1978; Read and Brown 1979). Northwards towards French Glacier, both axial surfaces of second phase folds and bedding are progressively rotated into and through the vertical (Fraunzen 1974) to dip to the southwest in Domain "B" at the head of Norman Wood Creek.

Figure 16 illustrates the change in orientation of second phase structures northwards along Norman Wood Creek. Sub-area one (Fraunzen's Domain I) shows folds with generally easterly dipping axial surfaces and eastern plunges. Southwesterly overturned folds rotate through westerly overturned folds as the structures pass around the end of the Adamant Pluton. In Sub-area 2 (Fraunzen's Domain III) $D_3$ has produced a fan of $D_2$ structures which accounts for the strong steep to vertical orientation of $D_3$ refolded hinges and the scatter of $S_2$. 
Figure 16: Changes in the orientation of second phase structures along Norman Wood Creek.

Data in sub-areas 1 & 2 from Franzen, 1974
Data in sub-area 3 this thesis

KEY TO SUB-AREAS

- POLES TO AXIAL PLANAR (S2)
- PLUNGE F2 HINGE LINE (L2)
eroded and overturned by second phase structures which plunge steeply to the southwest. This progressive rotation and overturning of the structures is illustrated schematically in Figure 17, which represents the change in orientation of the second phase French Creek Synform.

The orogeny in increase in the amount of horizontal motion in the eastern half of the synform was the effect of changing the front, when viewed from a single plane of rock from the overturned limb of a westerly vertical structure to the apparent right up limb of an easterly vertical structure. Simply stated, the sense of vergence of minor folds remains unchanged from French Creek to the southwest end of the French Creek Synform while the inclination of tip of their axial surfaces is rotation through the vertical.

Franzen (1974) suggested that the rotation described above was spatially related to the increased intensity of the large scale D3 folds in the vicinity of the Salinian fault. While Figure 17 illustrates the rotation described above it also shows the progressive steepening from west to east of D3 planes into the vertical.

Franzen (1974) suggested that this was probably the result of superposition on variably dipping limbs of the major second phase French Creek synform. Rather it is suggested that proximity to the Alamal pluton, which acted as a stress block during D2 and D3 (Shaw, 1980) caused steepening of the planes of third phase folds and hence D3 plunges and S3. Further evidence is provided by the opening of D3 structures against the west boundary of the pluton in Norman wood Creek (Figure 17) which accounts for the absence of D3 structures in
domain "B". Large third phase folds were not allowed to form due to the proximity of the adjacent Plate. Thus the rotation of bedding into the vertical is probably better explained with reference to the steepening of layering rather than the increased intensity of Bj. folds.

Domain "C" is approximately bounded to the south by the northwest-southeast trending Bigouth Creek Fault and to the north by the northern margin of the "U" like marble member (Figure 12). Second phase structures are dominant and while rare crenulations and minor folis associated with the third phase are present throughout the domain, strong folding is restricted to the area immediately northwest of Argonaut Mountain. Here east-west trending second phase folis adopt the regional northwest-southeast trend which results in the rotation of the plunges from a south to southwesterly direction (Figure 12).

The large marble bands which outline the folds at the head of Norman Wood Creek (Figure 13) preserve no obvious internal fabric and there is some question as to the age of these structures (Figure 6, Section 3). Pelites in the core of these folds display a strong cleavage which is axial planar to the larger folds. Large flat sillimanite pods elongate in the plane of the foliation (Plate 5A) result in some cleavage surfaces being more apparent than others, the overall effect being that of a spaced cleavage. Spaced or crenulation cleavages are normally associated with third folds, and a cursory examination of the rocks might lead to these structures being assigned to the third phase of deformation.

Careful examination of the sillimanite clots in thin section has revealed that the sillimanite is growing mimetically on biotite.
This clearly defines a strong planar fabric with which the sillimanite was crystallized. This fabric can be traced to the northwest where it is prevalent by thrusts; this thrust is thought to be by S3 (Plate 6). The peak of metamorphism is pre-S3, and it is therefore unlikely that sillimanite would crystallize during its deformation. It is possible that the planar fabric is due to the transposition of S2 into S3 during S3, but this is also considered unlikely since all planar lineations are well preserved in the section and there is no evidence of transposition in this section. The rock layer sequence of fabric development is as follows:

a) thrust grew synkinematically during S2;

b) post-S2 pre-S3 anatectic growth of sillimanite on gneiss during the peak of metamorphism;

c) later crenulation of S2 by S3.

Apart from the central open third phase syncline (Section 1, Figure 6) the foliation in the middle mélange adjacent to Birch Creek fault are therefore considered by the writer to be second phase structures.

3.3.4 Domain 4

Domain 4 encompasses the central parts of the Sandy Range and Mount Chapman sections (Figure 13). Although third phase minor folds are more evident than in the two preceding domains, they tend to be concentrated in small areas which coincide with the core zones of third phase closures. Elsewhere weak crenulations or gentle warping of the bedding are the only signs of D3 activity. Some of the crenulations measured are adjacent to small intrusive pegmatite bodies and therefore may not be strictly D3 in origin.
Second and third phase folds follow the regional northeast-southwest trend. Both sets of northeasterly overturned structures have moderate to steeply dipping axial surfaces and most minor folds plunge to the west.

In the central part of the domain a slice of intensely deformed strata sits above, and in fault contact with, relatively unaltered pelites and amphibolites of the lower pelitic member (Figure 6, Section 3). Above the fault, flaggy grey-green quartzites with schistose bedding planes, grey psammites and thick amphibolite schists are interbedded with minor yellow-stained, semi pelite and impure marble. The rocks are extremely disrupted and intruded by pegmatite. Skarn and coarse grained calc-silicates occur in the marble where it is crosscut by pegmatite. The succession, especially the quartzites, psammites and mafic schists, is lithologically similar to Hamill strata elsewhere in the Selkirk Mountains, but positive identification is not possible at this time.

The fault zone is marked by strongly migmatised pelites in which are floating large, one to two metre size blocks of quartzite (Plate 6D). Thin sections of the pelite reveal an annealed mylonitic texture (Plate 6E) where augen have recrystallized and polygonised whilst retaining their eye shape.

Boulders of quartzite within the migmatite preserve S-sense long-limbed first phase minor folds which maintain an S-sense of vergence whilst second phase minor folds change vergence about second phase antiform-synform couplets. The clastic rocks above the fault zone appear to form part of the overturned limb of a first phase fold and be allochthonous with respect to the rest of Domain D.
Late post-metamorphic minor faults in this part of the domain have resulted in the shearing-out of phase two synclines. Thus second phase anticlines are stacked up against a major fault to the northeast which separates the allochthonous rocks and underlying lower pelitic member from the middle marble member (Figure 6, Section 3 and Figure 8c).

3.4.5 Domain E

Domain E is centered on a large second phase antiformcored and outlined by nepheline syenite (Figure 6, Section 3). South of the nepheline syenite, a major D3 synform in Mount Neptune modifies tight map-scale second phase structures (Plate 6, Figure 6, Section 3). The irregularly shaped nepheline syenite appears to have a controlling influence on the distribution and form of both second and third phase structures. Axial surfaces of both phases dip steeply to both the northeast and southwest as a result of folding about shallow to sub-horizontal axial surfaces, genetically related to the convex northwards exterior shape of the northern contact of the nepheline syenite. Thus opposing dip directions can be measured along a single axial plane. Folds of this nature can be observed in the ridge above stations 175 and 176 (1974 sample locations on Figure 2).

During shortening associated with D3 folding, the nepheline syenite apparently acted as a competent body as second phase and newly forming third folds conform to the external shape of the body. Tightening about the syenite was sufficient to generate crenulation folds with sub-horizontal axial planes (Figure 13, stereonets). Crenulations associated with these folds are superimposed on third phase
folds and crenulations. In outcrops of pelite where erosion has caused an oblique section through both D3 and the later crenulations, both phases approximate a conjugate set and careful measurements must be taken in order to identify the respective fold phases.

South of the nepheline syenite, axial surfaces and limbs of second phase folds conform to the northern right-way-up limb of a large third phase synform centered on Neptune peak (Section 3, Figure 6: Figure 13). Phase two and phase three folds are coaxial and colinear in the southern part of the Domain, but linear data become more scattered northwards towards the nepheline syenite.

3.3.6 Domain F

Domain F is centered on a large 200 m thick unit of amphibolite which appears to have acted as a competent body during folding. Locally folds with shallowly dipping axial surfaces have been generated perhaps in a manner similar to those about the nepheline syenite in Trident Mountain. The dip of bedding increases southwards to become vertical or slightly overturned against the southern margin of the amphibolite. The increase is probably partly a reflection of the refolding of second phase folds by folds with shallow dipping axial surfaces and shallow plunges, and partly due to the tightening and re-orientation of second phase folds against the third phase syncline in the competent amphibolite.

3.3.7 Domains G and H

Domain G and H extend north and east from the amphibolite at Mud Glacier to a position which approximates the location of Northeast
Fault (Figure 13). The area is presented as two domains (G and II) in order to show the details of second and third phases of folding along the strike. Second phase folds are dominant and the tight, nearly isoclinal nature of the structures is shown by the similarity of the stereoplots for $S_2/S_1$ and $S_2$ (Figure 13).

The boundary between the two domains lies in a steep-sided valley formed by a tributary to Louis Lee Creek. On the west side (Domain G) of the valley, second phase folds of S-sense of vergence plunge to the west, indicating a second phase syncline to the north. On the east side of the valley, in strike-equivalent stratigraphy (Domain II), second phase S-sense folds plunge eastwards indicating second phase anticline to the north. This apparent structural contradiction is explained with reference to a combined stereo plot of the second phase data from both domains, shown in Figure 18. The combined plot illustrates a well-defined great circle of linear data (including intersection lineations and hinge lines). Such a well-defined girdle can be formed in one of two ways: passive flow folding or via the second phase hinge lines rotating through the vertical. Passive flow folding involves a later phase of folding distributing previously formed linear data. Passive flow folding is unlikely in this case because a strong $D_3$ folding is absent from this part of section.

A distribution of linear data as seen above can also result from the pitch of the lineation passing through the vertical as illustrated in Figure 18. Such a rotation may be due to shear strain during which the hinges rotated towards the direction of maximum elongation. In support of this mechanism lithologies in the western ridge are displaced southwards relative to the eastern ridge. The sense
Figure 18: Combined Equal-area plot of

Second phase data for Domains G and H

- Poles to second phase cleavage,
  axial planes of second F2 folds

- Lineation direction—hinge lines of F2 folds,
  intersection lineation (L2)
of vergence of the major and minor D2 folds remain unchanged whilst
the hinge lines are rotating. Thus S-sense minor folds with an easterly
plunge becomes S-sense folds with a westerly plunge between domains G
and H.

Rare third phase minor structures show similar trends with
westerly plunges in domain G and easterly to southeasterly in domain H.
The plunge variation may be due to both the doubly-plunging nature of
D3 and to superposition of a weak D3 on variably oriented D2
structures.

3.3.8 Domain I

Domain I approximately corresponds to that area to the north
of Northeast Fault and south of the northern limit of mapping (with the
exception of Domain J). Third phase structures with steep axial planes
are dominant, and fold long-limbed tight to isoclinal second and
possibly first phase folds. The dip of the bedding and axial surface
orientation of second phase folds is dependent upon position relative
to the axial surface of third phase folds which generally have gentle
southwesterly-dipping, long limbs and steep short limbs (Figure 6,
Section 4).

Third phase minor folds are common in the domain. Plate 66
shows a third phase minor fold in thinly bedded psammitic horizons. The
thin pelitic interbeds which separate the psammitic layers are strongly
crenulated in the hinge zone (next to lens cap). The folds are
disharmonic on the outcrop scale and where the folds are sufficiently
tight, the micas in the pelitic layers are rotated into an axial planar
orientation. If only the upper portion of the fold were exposed, one might mistake them for second phase structures.

In the southeastern corner of the domain, a thin band of nepheline syenite occupies the core of a reclined antiform (Figure 6, Section 4). The main body of the nepheline syenite also occupies an F2 antiform (Domain E). The displacement of the two portions of syenite across the valley, northeast from Trident Creek, suggested to Wheeler (1965) the presence of a northeasterly trending fault (Figure 13). This fault with uplift to the west would have the effect of reducing the net plunge of the northwesterly plunging second and third phase folds (Wheeler, 1965) and would account for the left handed offset of the Northeast Fault.

3.3.9 Domain J

The data from Domain J could well have been included in Domain "H", but were separated to illustrate the change in orientation of the third phase axial surfaces northwards towards the Trench. Third folds are strongly asymmetric with sub-horizontal to gently dipping long limbs and steep to overturned short limbs. Axial surfaces become less steep to the north. The Z-sense of vergence has the effect of carrying the same stratigraphic level downslope, as the form surface drops to lower and lower topographic levels. Thus third phase folds are stacked up against the Purcell Thrust in the Rocky Mountain Trench. Second phase folds are refolded in the same fashion described for Domain I.
3.4 Distribution of Phase Two and Phase Three Structures

Structural analysis indicates that the map area can be readily subdivided into three zones, characterized by the influence of either major third or second phase structures. The subdivision highlights those areas characterized by abundant third phase mesoscopic folds and fabrics:

1. That area immediately south of and including the Selkirk Fan Axis dominated by large map scale third phase folds.
2. The central area of the Windy Range and strike equivalent areas characterized by fabrics of the second phase with isolated third phase crenulations and minor folds.
3. The northern area from Warsaw Mountain to Mount Neptune with dominant major third phase folds and type three interference folding involving the second and third phases.

The relationships between these zones, and the roles played by the major folds within them, warrants further discussion. Much of the information for detailed analysis of the role of these phases in the kinematic history comes from an examination of the microfabrics, which forms the subject of Chapter 5. Discussion of the role of the fold phases is therefore left until later (p. 155).

3.5 Faults

Faults with both normal and reverse sense of movement are present in the map area, and examples of both types have already been referred to in the text. Many of the faults take the form of tectonic slides (term reviewed by Hutton 1979). A tectonic slide forms under conditions of ductile strain i.e. during metamorphism, and is an
growth of that mineral postdates the formation of that fold or fold phase. If porphyroblasts at higher and lower grades of metamorphism in the region also contain straight trails then the metamorphic event postdates the deformation phase.

The absence of synkinematic crystal growth implies that we isotherms passed through a rigid mass of rock, rather than straining the mass during metamorphism. Furthermore the system, in which the folds are a part, was held under static conditions long enough for the series of metamorphic reactions responsible for the growth of porphyroblasts to occur.

In addition to simple straight trails many porphyroblasts, in particular biotite and garnet, have central straight trails which show slight to strong "curvature" at the margins of the crystal. Generally, internal and external foliations are continuous. The development of the straight part of the trail must follow the pattern established above, i.e. that the central part of the porphyroblast grew under static conditions after the generation of the foliation. The curved "tails" are most easily explained as the result of one of two processes. Either:

1. Growth of the porphyroblast continued from the static into the following dynamic phase during which the pre-existing foliation was deformed. Slight rotation of the porphyroblast relative to the foliation and/or the foliation relative to the porphyroblast, during continued growth, resulted in the curved ends to the tails. Such single overlaps from static to dynamic phases have been described by Rast and Stuart (1957), Zwart (1962), Ramsay (1962) and Johnson (1963) or:
Figure 19  Correlation of Faults.
Norman Wood Fault Zone.

- Hamilt
- upper pelitic member
- pm
- middle marl/te member
- pm
- lower pelitic member
- pm
- plutonic rocks

Faults - ~ indeterminate
- normal

outcrop width of lithologic assemblages thins drastically and some horizons disappear. Lithologies present to the south of the fault zone do not reappear and vice-versa, which indicates that either they are disappearing at fault surfaces, where such units are fault-bounded, or that faults coalesce (and diverge) locally. A knowledge of the presence of these faults, and their position and structural effects, is necessary before any reasonable correlation of stratigraphy through the fault zone can be made. The following section describes those faults whose position was determined using visual observation of either a surface of discontinuity of an obvious structural nature, or the disappearance or appearance of lithologic units at contacts where shearing was no longer evident, i.e., at a tectonic slide.

Most faults entering from the south coincide with contacts between Hamill and Horsethief Creek rocks. The Hamill crops-out to the south and southwest of the Alamant Pluton in a series of north-tapering tongues (Wheeler 1965) (Figure 1). Faults have been documented at many Hamill-Horsethief Creek contacts south of Goldstream River by Wheeler (1963; 1965), Lane (1977), Brown et al (1978); and north of Goldstream River by Franzen (1974), Brown et al (1978), and Perkins (thesis). Only two outcrops of Hamill south of the fault zone and one to the north are relevant to the following discussion, and for simplicity these will be labelled and referred to as Hamill I, II, and III (Figure 19). Faults are lettered consecutively A to C in order of discussion.

3.5.1.1 Fault A

A regionally significant fault is situated at or close to the western margin of Hamill I. Immediately north of Goldstream River (Figure 19), Hamill I is in close proximity to outcrops of thick grey
marble, which has in the past been correlated to middle marble, Hamill (Brown et al. 1977; Brown et al. 1978; Brown and Tippett 1978). It is possible, as tentatively suggested above (page 59), that these marbles are in fact Badshot Formation.

Fault A is the northwards continuation of a fault developed in the core of a first phase nappe mapped further to the south and correlated northwards by Lane (1977), Brown et al. (1978) and Seal and Brown (1979). The fault postdates the formation of the nappe, west of the fault stratigraphy is inverted, and second-phase structures are superimposed on a succession which was inverted prior to the second phase of deformation (Van der Leeden 1976; Brown et al. 1978). East of the fault stratigraphy is right-way-up at least as far as the Esplanade Ranges (Brown et al. 1978). Brown and Tippett (1978) therefore, position the core zone of a first phase nappe, now represented by a fault surface, at or close to the western margin of Hamill 1. The regional significance of this fault has been previously discussed on page 54. Hamill 1 disappears northwards, but the fault which represents the junction between inverted and right-way-up stratigraphy (Fault A), continues northwards, closely following the trace of the marble until the distinctive grey marble bands disappear (Figure 19). North of this point the position of the fault can be inferred by reference to a distinctive fragmental-tuff marker band which Brown (personal communication, 1978) has traced north and west following the swing in strike of stratigraphy around Argonaut Mountain (Figure 3). The fault must lie to the east and north of the marker horizon and presumably follows the swing into the regional northwest strike of the thesis area. The marker band disappears into the highly disrupted zone.
of pelite and peymitite already mentioned as being the likely
continuation of Bignouth Creek Fault (Figure 3). The writer postulates
that Fault A is the southerly extension of Bignouth Creek Fault. Both
faults separate a dominantly calcareous from a dominantly pelitic
succession. If this correlation is correct, then the stratigraphy in
Domain A is inverted.

3.5.1.2 Fault B

The fault at the northeastern contact of Hamill quartzites
(I[1]) separates Hamill rocks from middle marble member, and in the
writer’s opinion constitutes a structure of regional significance. The
whole of the upper pelitic member and possibly the upper part of the
middle marble member have been removed from section. The lensoidal
nature of the middle marble member is such that an accurate estimate of
thickness is unobtainable. The missing section could vary from 0–300 m.
An unknown amount of the Hamill is missing. Assuming an average
thickness of 3000 m for the upper pelite (Tippett 1976; Lane 1977) and
that the Hamill southwest of the fault is right-way-up, the minimum
vertical stratigraphic offset is in excess of 3 kms. This figure would
be increased considerably if a large thickness of Hamill Group is
missing across the fault. The structure would, therefore, be a normal
fault, downthrown to the southwest. This fault has been previously
named the Birch Creek Fault (this thesis) and is the same as Argonaut

The fault likely defines the southern margin of the middle
marble member southeastwards into Norman Wood Creek (Figure 19), where
it is again recognized at the lithologic contact between two rock
sequences. In the middle marble member, to the east of the contact, minor folds are overturned to the west, whereas in a pelitic succession west of the contact, minor folds are overturned to the east. Opposing vergences of minor folds would indicate an anticlinal structure, but lithologic contrast and lack of lithologic repetition across a discrete planar surface indicate the presence of a tectonic slide.

There is independent evidence to suggest that the interpretation of the continuation of the fault southeastwards is correct. Leatherbarrow (1981), using the grossular-anorthite-alumino-silicate-quartz geobarometer (Ghent 1977), has documented two pressure domains in the northern Selkirks (Figure 20). The relatively high pressure domain (at 7 kb) approximately corresponds to the writer's thesis area, while the relatively low-pressure domain (at 5 kb) lies to the south of and includes Argonaut Mountain and French Glacier region. The sample locations from which pressure data were obtained and the outline of the pressure domains are shown in Figure 20.

A pressure differential of two kilobars can be interpreted to mean a vertical difference in depth of burial during metamorphism of 6 to 7 km. At the head of Norman Wood Creek high and low pressure domains are in close juxtaposition indicating that the boundary between the domains is relatively sharp. Leatherbarrow (1981), postulates that the abrupt junction is the result of faulting, and that the major offsets has occurred along a single post-metamorphic fault surface. Fault B lies between the high and low pressure control points at the head of Norman Wood Creek, and would appear to be in the correct position and have the correct sense of movement to be a candidate for this major post-metamorphic fault.
Fig 20: PRESSURE DISTRIBUTION DURING METAMORPHISM, BIG BEND AREA.

EXPERIMENTALLY DERIVED PRESSURE ESTIMATES

- 7 kb
- 5 kb

10 km:

[Map showing pressure distribution with symbols for Hamill Group, Marbles, Plutonic Rocks, and Faults (major)].

DATA MODIFIED FROM:
R.W. LEATHERBARROW
Leatherbarrow (1981), suggested that the trace of the fault to the west probably approximately coincides with the southwest margin of the zone of muscovite breakdown. This margin in turn approximates the trace of the Hamill Group. The writer feels that the fault can now be accurately positioned at the contact between the Hamill and middle marble member, where field data suggest a minimum vertical offset of 3 km on the Birch Creek Fault.

To the south fault B probably swings sharply southward to run between the eastern margin of Hamill II and the pluton (Figure 19), where a fault is mapped by Shaw (1980).

3.5.1.3 Fault C

On the western margin of Hamill II, to the east of French Creek (Figure 19), Franzen (1974) identified a melange and tectonic breccia zone in the core of a second phase closure. The extensive brecciation of adjacent third phase folds suggested to Franzen that this was the site of a major tectonic break. The break occurs at the lithologic contact between a quartzite-paeanite sequence and a calcareous sequence immediately to the west. The structure has been traced northwards into the Norman Wood Fault Zone where the quartzite disappears.

North of the point of disappearance of the quartzite, an angular discordance in the strike of stratigraphy can be observed (Figure 3 and 13). On the ridge west of Norman Wood Creek, a carbonate, calc-silicate and pelite assemblage strikes north-northeast, while in the creek bottom a similar assemblage strikes north-west. The approximate position of the junction between the assemblages has been
traced northwards towards station P36 (Figure 2). At station P36, a fine-grained, highly granulate, pale blue marble which may represent a shear zone. The granulate marble at the junction between the two lithologic assemblages is in the correct lithologic and structural position to coincide with the extension of Fault C as mapped by Franzen (1974). Lithologic and structural considerations, therefore, suggest that Fault C extends a considerable distance beyond the point of disappearance of the quartzites. The exact position of disappearance of the fault is, however, unknown, but Fault C could extend to the western boundary of Hamill III (Figure 19).

3.5.2 Northeast Fault

The significance of Northeast Fault (Figure 13) has already been discussed during the section on the correlation of the grit unit (Section 2.3.3.1). Briefly, stratigraphic interpretation suggests that the fault be of normal sense downthrown to the south (Figure 6, Section 1).

3.5.3 Minor Faults in the Central Map Area - Domain D

Many small faults (not all map-scale) of reverse sense of motion occur in the central and northern parts of Domain D and westwards towards Domain G and H (Figure 13). These discontinuities are recognized by abrupt changes in the vergence of second phase minor folds and by local repeats of stratigraphy. Minor folds with eastern vergence predominate throughout the zone as a result of the shearing-out of the short overturned limbs of second phase antiforms. Thus second phase
antiforms are stacked-up against a fault of more regional significance in the northeastern corner of Domain D. [Stns. P78, P82]. These faults are illustrated in Figure 6, Section 3.
4. METAMORPHISM

4.1 Introduction

The northern Selkirk Mountains are underlain by rocks which have undergone Barrovian type regional metamorphism. Metamorphic grade ranges from chlorite to sillimanite, with the metamorphic culmination defined by a zone characterized by the coexistence of K-feldspar and sillimanite. Northwest-trending mineral zones are asymmetrically distributed about the metamorphic culmination, with the mineral zones being more widely spaced to the north (Figure 21). The thesis area is underlain by the metamorphic culmination, and only high grades of metamorphism are represented.

The metamorphic petrology of the pelitic rocks in the Big Bend area has been the subject of a doctoral thesis by Leatherbarrow (1981), and the reader is referred to this thesis for a detailed discussion. Metamorphic zones drawn on Figure 21 are taken from Leatherbarrow (1981). The following petrographic observations by the writer are presented as necessary background for Chapter 5.

Four distinct mineral assemblage zones are recognized within the study area (Figure 22).

1. Staurolite - muscovite-quartz-kyanite-garnet-biotite
2. Kyanite - garnet-biotite
3. Sillimanite - garnet-biotite
4. K-feldspar - sillimanite
Figure 21: Distribution of Structural and Metamorphic Domains and their relationship to Metamorphic Zones.
4.2 Staurolite-Muscovite-Quartz-Kyanite-Garnet-Biotite

Assemblage 1 occurs in the northeastern part of the area. The position of the staurolite-out isograd has not been observed in the northwestern part of the area, where the assemblage kyanite-garnet-biotite crops-out to the limit of mapping. Simony (personal communication, 1976) has found staurolite well below tree-line in Yellow Creek, and the isograd therefore presumably runs along the tree-covered slopes immediately above Columbia Reach.

Staurolite takes the form of large xenoblastic to idioblastic polikiloblasts with straight quartz inclusion trails. In the trees, north of station 277 (Figure 2), well-formed shiny, brown staurolite crystals to 6 cm weather out of a muscovite rich pelite. Staurolite becomes smaller, and more difficult to find as the isograd is approached. Well formed kyanite, with few elongate quartz grains defining straight inclusion trails, occurs together with idioblastic garnet without inclusions, and xenoblastic garnet with straight quartz inclusion trails which appear to compatibly coexist with staurolite.

4.3 Kyanite-Garnet-Biotite Zone

The boundary between this and the previous zone is marked by the disappearance of staurolite.

Kyanite is generally well formed, with few inclusions. It may occur with the long axis parallel to, or crosscutting the regional S2 foliation. Quartz inclusions form straight trails which are parallel to and continuous with S2. Garnets in this zone have two forms. They may be idioblastic with a few centrally located "dust-like" inclusions or xenoblastic with many elongate quartz grains. Idioblastic garnets
can be observed within or intergrown with kyanite. Both crystals show straight grain boundaries, and the minerals appear to be existing compatibly.

4.4 Sillimanite-Garnet-Biotite Zone

The sillimanite-garnet-biotite zone occurs to the southwest of the kyanite-garnet-biotite zone. The boundary between the two zones is marked by the polymorphic aluminosilicate transition, kyanite to sillimanite.

Sillimanite occurs as felted knots of fibrolite, often intergrown with muscovite and biotite. Where the fibrolite knots have been flattened on cleavage surfaces they form patches several centimetres across and up to 10 cm long. Rarely, sillimanite-biotite knots "encase" idioblastic inclusion-free garnets. As in the previous zone garnet is present in two forms: idioblastic inclusion free crystals, or remnant quartz-riddled xenoblasts. The latter may have idioblastic inclusion-free rims.

4.5 Potassic Feldspar - Sillimanite Zone

This 3.5 km wide zone corresponds to the metamorphic culmination. The southern boundary of the zone approximately coincides with the major fault at the northern contact of the outcrop of Hamill Group (Figure 3). Apart from the diagnostic mineral assemblage, the zone is marked by an abundance of pegmatite which occurs either concordantly with metasediments or as randomly oriented discordant stringers and dikes.

Sillimanite occurs in the same form as described in the
previous zone. Potassic feldspar forms xenoblastic inclusion-free
grains associated with coarse grained muscovite and fibrolite. Garnets
show the same characteristics as described above. Microfabric analysis
is difficult in this zone due to both the coarse grain size and
abundance of pegmatite.
5. MICROMYPHON

5.1 Introduction

The relationships of growth of metamorphic minerals: episodes of deformation have been described by Rast (1932, 1962, 1965), Voll (1960), Zwart (1960), Johnson (1961, 1962, 1963), Ramsay (1963), Spry (1963a, 1969) and many others. Growth can occur pre-, syn-, or post-deformation, and the textures that may be associated with each growth phase are well illustrated in the literature (see especially Spry 1969). In many cases, for example composite and zonal structures in garnets, it is possible to demonstrate that metamorphic growth which began during deformation continued after deformation had ceased.

Alternatively, growth which began under static conditions continued through the period of deformation which followed (Rast and Sturt 1965; Zwart 1962; Johnson 1963). The implication from the above is that the time during which metamorphism takes place is necessarily longer than the time required for episodes of deformation; however this does not preclude local rapid growth of porphyroblasts.

The recognition of the particular features associated with each phase of growth in composite crystals allows us, by comparison, to recognize similar features in crystals with shorter growth histories, and thus relate their growth period to a particular episode in the deformational history. This is extremely important in areas of polyphase deformation such as the Selkirk Mountains, for it is obvious that a mineral which grows under static conditions between two phases of deformation is post-tectonic to the early phase, and pre-tectonic to the later phase. Fortunately there are features which, in the absence of
composite crystals, allow us to determine whether a crystal has grown pre-, post-, or syn-tectonically and these are (following Spry 1969):

a. Features indicative of pre-tectonic crystal growth.
   1) Twinning extinction in tabular or equant crystals. Kinking or crenulation of platy minerals which process may also cause deformation twins in plagioclase.
   2) Twisted and/or fragmented crystals, especially garnet, which are wrapped around by the foliation.
   3) Rings of a retrograde mineral around deformed crystals e.g. chlorite on garnet.
   4) Mortar texture i.e. where the exteriors of large crystals have broken down into aggregates of smaller crystals of the same composition.

b. Features indicative of syn-tectonic crystal growth.
   1) Rotational - S-shaped inclusion trails preserved in minerals which have been rotated as they grew. The porphyroblasts are called snowball (garnet) if the apparent rotation is greater than 90° and simply rotational if less than 90°.
   2) Non-rotational - these include crystals in which the internal trail is straight in the core but becomes progressively more crumpled towards the rim where it passes out into a crenulated external trail, (Zwart 1961, 1963). Porphyroblasts may also grow during a time when a pre-existing foliation is being progressively flattened. Inclusion trails in the centre of the porphyroblast are widely spaced but become more closely spaced outwards reflecting the flattening of the external foliation.
c. Features indicative of post-tectonic crystal growth.

1) The formation of randomly-oriented crystals replacing a preferred orientation.

11) Growth of porphyroblasts with helicoid structure i.e. where the inclusions may be microfolds, parts of a fold; straight, curved or irregular surfaces, all of which constitute a foliation in existence prior to the growth of the crystal.

111) Polyhedralisation of bent crystals as strain energy is removed.

1v) Mimetic growth and the replacement of early porphyroblasts to give pseudomorphs.

v) Formation of oriented overgrowths to give zones of phylloclastic crystals.

2l) Cross or randomly oriented crystal growth which does not deflect the foliation.

In the list of features associated with pre-tectonic crystallization, point number two describes a feature commonly observed in foliated metamorphic rocks i.e. the foliation-forming matrix minerals appear to be deflected by or wrapped around the porphyroblasts.

Historically, this feature has been controversially attributed to either the mechanical pushing aside of solid particles by the growing crystal, the so-called "Force of Crystallization" (Becke 1904; Harker 1932, 1950; Eskola 1939; Ramsay 1947, 1957, and more recently Misch 1974; Harvey and Ferguson 1973; Sassingon 1974), or to the flattening of the matrix minerals about a pre-existing porphyroblast (Zwart 1962; Rast 1958, 1965; Rast and Sturt 1957; Spry 1969; and others).

The deflection of the phylloclastic minerals around porphyroblasts is the observation most often quoted in support of the
"Force of Crystallization". Rast (1958) and Zwart (1962) pointed out, however, that where such porphyroblasts contain inclusions, the fabric within the porphyroblast can be different from the external fabric. Furthermore, porphyroblasts of the same mineral which have grown statically, replace the surrounding minerals without deflecting them (Rast 1965). The wrapping of accretionary minerals around a porphyroblast can almost invariably be explained by the deformation of matrix minerals against it. In the remainder of the chapter the force that a growing crystal is able to exert on its surroundings is considered negligible.

5.2 Terminology and Some Assumptions Involved in the Study

Under metamorphic conditions, those mineral grains not required by the reaction causing a mineral to grow in a rock are included within the growing crystal. Quartz is the most common inclusion; other minerals found as inclusions being ilmenite and the other opaque, tourmaline, zircon and the other heavy mineral grains, and the micas. Where one mineral is breaking down to form another, fragments of the original mineral may be preserved as inclusions within the enveloping crystal giving a "sievy look" to the porphyroblast or they may define continuous straight or curved trails which represent a structural element i.e. foliation (s) of the host rock.

The internal structure or trail in a porphyroblast is referred to as $S_1$ (S-internal). The S-surface outside the crystal is $S_e$ (S-external). The foliation or bedding may pass without interruption through the crystal ($S_1=S_e$) or $S_1$ and $S_e$ may be discordant ($S_1\neq S_e$). In the latter case, $S_1$ represents an older event and $S_e$ a younger one. Helicitic growth structure refers to the inclusion
of a straight or curved trail which represents a structural element of
the rock older than, and preserved during the growth of, the crystal.
Curved helicitic structures are post-tectonic and should not be confused
with syn-tectonic rotational or snowball structures (Spry 1969), where
the curved or S-shaped inclusion trails are due to the rotation of the
crystal during its growth.

Pre-, syn-, and post-tectonic growth refers to the growth of
a mineral before, during or after a deformation episode, respectively.

A large number of the porphyroblasts, observed in thin
sections from the Northern Selkirks, contain quartz inclusions in the
form of trails, many of which are straight or straight with curved
"tails" in the outer rim of the crystal. Powell and Treagus (1970),
indicate that care has to be exercised in the interpretation of
porphyroblasts whose inclusions show little or no curvature in thin
sections. Such porphyroblasts, usually interpreted as the result of
static crystal growth, may arise in syn-tectonic crystals from certain
cuts parallel to the rotation axis or from cuts near the outer margin of
the crystal. Sufficient porphyroblasts were present, in most thin
sections, to virtually eliminate the possibility of misinterpretation
due to the observation of the outer portions of the crystals only.
Furthermore, the possibility that the thin section paralleled the
rotation axis was not considered very likely due to careful selection of
thin sections relative to the rotation axis in samples, and the large
numbers of thin sections studied. As a precaution, however, random cuts
were made across samples with well defined linear fabrics. Sections
were also cut from rock samples to contain the rotation axis, and the
perpendicular to it. An example of the features observed in such samples is illustrated in Figure 22.

5.3 The Significance of Straight Trails

A porphyroblast which contains a straight trail is here assumed to have grown after formation of the trail, and during a period of no strain. Elliott (1973) has pointed out that porphyroblasts with straight trails could have grown synkinematically if the strain axes remained fixed during deformation. In this case the porphyroblasts would include quartz grains whose long axis parallels the direction of maximum elongation. This case can be discounted in the study area because:

1) the S2 foliation is commonly overgrown by porphyroblasts whose long axis is perpendicular to the foliation e.g. cross biotites i.e. the long axis of the porphyroblast does not lie in the direction of maximum elongation. Inequant crystals attempting to grow in this orientation during deformation would have been rotated resulting in S-shaped inclusion trails. Such trails are rare in samples studied.

2) Second phase folds are strongly asymmetric, which attests to a rotational component during strain. In addition the formation of tectonic slides (documented earlier) involves large shear strains.

Porphyroblasts with straight trails are thus considered to represent growth during a period of no strain.

If the planar foliation over which the mineral has grown, can be demonstrated to be axial planar to a fold or train of folds then the
FIGURE 22: Schematic diagram to illustrate relationships between porphyroblasts, their inclusion trails and the Axis of Rotation.
growth of that mineral postdates the formation of that fold or fold phase. If porphyroblasts at higher and lower grades of metamorphism in the region also contain straight trails then the metamorphic event postdates the deformation phase.

The absence of synkinematic crystal growth implies that we isotherms passed through a rigid mass of rock, rather than straining the mass during metamorphism. Furthermore the system, in which the folds are a part, was held under static conditions long enough for the series of metamorphic reactions responsible for the growth of porphyroblasts to occur.

In addition to simple straight trails many porphyroblasts, in particular biotite and garnet, have central straight trails which show slight to strong "curvature" at the margins of the crystal. Generally, internal and external foliations are continuous. The development of the straight part of the trail must follow the pattern established above, i.e. that the central part of the porphyroblast grew under static conditions after the generation of the foliation. The curved "tails" are most easily explained as the result of one of two processes. Either:

1. Growth of the porphyroblast continued from the static into the following dynamic phase during which the pre-existing foliation was deformed. Slight rotation of the porphyroblast relative to the foliation and/or the foliation relative to the porphyroblast, during continued growth, resulted in the curved ends to the tails. Such single overlaps from static to dynamic phases have been described by Rast and Stuart (1957), Zwart (1962), Ramsay (1962) and Johnson (1963) or:
2. The marginal curvature may be the result of crenulation cleavage in the matrix which also affects those surfaces of the crystal adjacent to the cleavage surfaces. This possibility is difficult to demonstrate in isotropic garnets, but biotite in such a system could be expected to show signs of lattice strain. The margins of the crystal adjacent to cleavage planes might show strain, exemplified by extinction angles different to the remainder of the crystal or simple undulose extinction in the marginal areas.

Both processes are thought to occur in the northern Selkirk Mountains and both are described and illustrated in detail in the relevant sections that follow:

5.4 Sampling Technique and Methodology

The study involved the petrographic examination of three hundred and sixty five thin sections from the northern Selkirk Mountains. Because samples from the writer's map area provided information for high metamorphic grade assemblages only, sampling was extended southwards to rock suites of lower metamorphic grade. Samples from French Creek and adjacent areas, collected by Franzen (1974), Van der Leeden (1976), and later by Leatherbarrow (1981), were included in this study. The outlines of the sample areas are shown on Figure 22, while sample locations are shown on Figures 2 and 23 (in pocket). Suites collected by Tippett (1976), and Lane (1977), were studied for comparative purposes only and results are not formally presented here.

The extension of the sampling area south of the Selkirk Fan Axis was also beneficial for the following reasons. Firstly, south of
the Fan Axis the surface trace of isograds is locally perpendicular to stratigraphic strike. Individual pelite horizons have been sampled from low to high metamorphic grade (Leatherbarrow 1981). Reactions which cause the appearance or disappearance of metamorphic minerals are therefore less likely to be affected by changes in the bulk chemical composition of the pelite. Temporal relationships established here between growth of metamorphic minerals and phases of deformation are more reliable, and facilitated a higher level of confidence in interpreting the microfabrics of the higher grade assemblages in the map area.

Secondly, changes in style and complexity of deformation of the microfabrics could also be observed, as both degree of metamorphism and deformation increased with proximity to the Selkirk Fan Axis (Figure 23). The extensive rock suite collected by Leatherbarrow (1981), spanned the fan axis. Access to samples from this suite provided the opportunity to compare and contrast microfabrics from both north and south of the fan axis.

Results obtained from the study are presented in two ways. Firstly, the text involves a discussion of development of microfabrics relative to the grade of metamorphism, beginning with the chlorite zone and ending with sillimanite zones and the breakdown of muscovite. The chlorite, biotite and garnet zones are found south of the Selkirk Fan Axis only (Figure 21). Throughout each zone, porphyroblastic minerals are related to both internal and external foliation, and recognized, where possible, as pro-grade or retrograde. Each new appearance and/or disappearance of a particular mineral is noted but the reactions
involved are not analysed in detail. This was the subject of a separate study by Leatherbarrow (1981).

Secondly, the areas were subdivided into structural domains using information gained from mesoscopic fold analysis (Figure 21). Changes in the relationships between microfabrics and mesoscopic and macroscopic structures could then be documented as changes occurred in the style, orientation, and complexity or intensity of these larger structures.

To do this, reference was made to the original field notes to establish, where possible, the location of the sample with respect to hinge line and axial plane of adjacent minor folds. Care was exercised in selecting the orientation of the site for thin section relative to planar and especially linear features on the hand specimen. In the French Creek area, south of the Selkirk Fan Axis, where L₂ and L₃ are locally at a high angle, thin sections were cut perpendicular to both lineations. North of the fan axis, in the writer's thesis area, D₂ and D₃ folds are co-axial and only one thin section was deemed necessary.

In order to present results from the second part of the study as completely as possible a series of charts were prepared, one for each structural domain. The charts are presented in Appendix I. The reader is referred to one of the charts in Appendix I while reading the following. First, second and third phases of deformation along with post- and/or predeformation phases are recorded in standard format along the horizontal axis. Also in standard format, the vertical axis contains the index mineral zones, chlorite to sillimanite. The charts, however, differ from the standard bar format where each bar represents the growth span of a particular mineral. In the charts developed in
this thesis, each mineral in porphyroblast form is represented by a symbol e.g. a hexagon for garnet. Planar or crenulate fabric elements are represented by straight or curved lines respectively. By superimposing the porphyroblast symbol onto the planar fabric element, relationships between internal and external foliation can be graphically displayed. Different fabric elements are then sequentially lettered, and thin sections within that domain listed at the bottom of the chart. Following each thin section (identified by a reference number) is a series of columns representing the fabric elements displayed in the chart above. The presence of an asterisk in a column means that a particular relationship can be observed in that thin section. In this way not only is the complete microfabric development for each thin section displayed but the microfabric history and style for each domain, as well as the relative abundance of each fabric element is readily observed at a glance. The charts are useful for both petrologists and structural geologists alike because the actual relationships between porphyroblast and foliation, as well as the shape of porphyroblasts e.g. idioblastic, are recorded. Additional notes opposite the chart indicate the size of porphyroblasts and detailed descriptions of new relationships as they occur.

The charts are presented along with a list of symbols and a more detailed description of their use in Appendix I. Reference is made to these charts in the text.
5.5 Microfabrics by Metamorphic Zone

5.5.1 Chlorite Zone and Chlorite to Biotite Transition

Chlorite Zone rocks crop out at the head of McCulloch Creek (Figure 21 and 23). A microfabric analysis of this zone, along with the biotite zone to the north, is presented as Domain VI, Chart I, Appendix I.

Small flakes of chlorite (.2 mm) lie in both S₁ and S₂. Between spaced S₂ cleavage surfaces, S₁ is preserved as a strongly crenulated foliation consisting of muscovite, quartz, and more rarely, chlorite, (Figure 13D). Elongate quartz-rich pods and layers, the remains of original compositional layering, now lie in S₂ (Figure 13E).

The earliest chlorite porphyroblast growth predates the transposition of S₁ into S₂ (Figure 13D). A syn-D₂ chlorite includes gently crenulated fine grained S₁ quartz trails. The dominant S₂ foliation is deflected about the porphyroblast, and randomly oriented chlorite flakes have grown in the pressure shadows.

The majority of chlorite porphyroblasts have grown post-D₂ (Figure 13E) and S₁=Se where Se is S₂. Those elongate in S₂ have been crenulated by S₃ (Figure 13E). Quartz inclusions within deformed chlorite crystals show undulose extinction, and some of the larger inclusions have polygonised into subgrains. Cross-chlorites have been rotated to lie with their long axes in S₃ (Figure 13F).

Indistinct but curved inclusions in the outer margins of some grains may indicate continued growth into D₃ but the evidence is inconclusive.
Albite porphyroblasts elongate in \( S_2 \) postdate \( S_2 \) and include fine straight quartz trails where \( S_1 \) is \( S_6 \). Porphyroblasts elongate in \( S_2 \) have been crenulated by \( D_3 \) (Figure 24A). No syn-\( D_2 \) growth of albite is observed.

Immediately to the south of the biotite isograd (Figure 21 and 23), intergrown clusters of small chlorite grains postdate \( S_2 \). Inclusion trails are not preserved (Figure 24B). Within the cluster, each grain is well defined but each may have a different optical orientation to its neighbour. At the first appearance of biotite, chlorite and biotite grains occur together (Figure 24C). Individual crystals are smaller, less well defined, and randomly oriented within an area of similar dimension to that occupied by the chlorite clusters and single biotite porphyroblasts at lower and higher metamorphic grades respectively. As the garnet zone is approached chlorite disappears at the expense of biotite. The change appears to be accompanied by an increase in the ordering of the constituent grains, which are arranged so that their long axes parallel \( S_2 \). Between the biotite flakes are quartz grains elongate in \( S_2 \). These biotite composites have been later crenulated by \( D_3 \) (Figure 24D) and some flakes have been bodily rotated towards and into \( S_3 \). These rotated grains show strain in the form of undulose extinction which suggests little or no syn-\( D_3 \) growth. Elsewhere biotite appears to have grown in, rather than been rotated into, \( S_3 \).

It is possible (but unlikely) that the sequential ordering above represents one method of formation for the biotite porphyroblasts. Many small grains, nucleating in compositionally favourable areas may have become ordered in response to increasing pressures and/or
A. Post-\(S_2\) albite porphyroblast which included as trails straight \(S_2\), has been deformed by \(D_3\) crenulation cleavage. Sample V308; Scale bar 0.4 mm.

B. Intergrown clusters of randomly oriented chlorite and albite postdate \(S_2\). Sample V310; Scale bar 0.2 mm.

C. Intergrown clusters of randomly oriented chlorite, biotite and fine quartz postdate \(S_2\) and deflect \(S_3\). Note that some biotite flakes within the cluster are aligned \(\parallel S_3\). Sample V322; Scale bar 0.5 mm.

D. A cluster of biotite flakes, crenulated by \(D_3\), occupies an area similar to that of single biotite porphyroblasts at higher grades of metamorphism. Sample V52A; Scale bar 0.3 mm.

E. Chlorite porphyroblasts which cross-cut \(S_2\) have nucleated on ilmenite flakes. Biotite is replacing chlorite along cleavage surfaces in the chlorite. Sample F10C; Scale bar 0.3 mm.

F. "Intergrown" biotite and chlorite. Biotite postdates \(D_3\) and includes \(S_2\) as straight trails. \(S_3\) is deflected about the biotite. Chlorite, elongate in \(S_3\), includes quartz grains oriented in \(S_3\). Note the sharp boundary between the two minerals. Sample F3A; Scale bar 0.3 mm.

G. Continued growth of biotite into \(S_3\) has included quartz grains oriented in or close to \(S_3\). Either the porphyroblasts have rotated relative to the matrix or the matrix relative to the porphyroblast. Sample F9; Scale bar 0.5 mm.

H. Shearing along spaced penetrative \(S_3\) cleavage surfaces, deformed quartz grains continuous from \(S_1\) to \(S_3\) and calcite strain features in the margins of the biotite aligned to the cleavage surfaces. Sample F1G; Scale bar 0.2 mm.
temperatures. The final stage may result in recrystallization into a single crystal of biotite. Alternatively, the porphyroblast clusters like the one illustrated in Figure 24D may be the result of a sequence of pro- and retrograde metamorphic reactions.

5.5.2 The Chlorite-Biotite Zone

The biotite zone of Domain V1, the biotite-chlorite zone of Domain I (Chart 1) and the southern part of Domain 2 (Chart III) are considered together (Figure 23). A complete analysis of the microfabrics is presented in Charts I, II, and III, Appendix 1.

Flakes of chlorite lying in both S2 and S3 foliations occur sporadically throughout the zone. Large flakes in S2 are crenulated by D3. Prograde chlorite porphyroblasts, some of which appear to have nucleated on ilmenite flakes (Figure 24E) clearly postdate the formation of S2. Biotite may rim and replace chlorite along cleavage planes (Figure 24E). Elsewhere prograde chlorite and biotite occur together without visible reaction (Figure 24F). In the example illustrated, chlorite aligned in S3 includes quartz elongate in S3 and clearly postdates the biotite. The boundary between the grains is sharp and clean and neither mineral shows signs of instability, but nonetheless some reaction between biotite and chlorite may have occurred. East of French Creek (Figure 23, Domains 1 and 2), prograde chlorite porphyroblasts disappear well to the south of the garnet isograd. West of French Creek prograde chlorite persists until the appearance of staurolite (Stn 108 - Domain V2, Figure 24). Retrograde chlorite, either as isolated mats in the ground mass or as rims on
biotite and garnet, occurs sporadically throughout the chlorite-biotite zone.

Both equant and inequant biotite porphyroblasts contain elongate quartz grains in the form of inclusion trails which are either straight (Figure 24F), or straight with slightly curved extremities (Figure 24G, H). Inequant biotites lie either with their long axis parallel to foliation or as randomly oriented cross-biotites.

The inclusion and preservation of straight trails is due to the post-\(D_2\) helicitic overgrowth of the \(S_2\) foliation by biotite.

Despite the development of a \(D_3\) crenulation and concomitant rotation of the porphyroblasts, \(S_1\) may still be continuous with \(S_e\) (Figure 24G). Quartz grains in \(S_e\) may be either gently curved in \(S_2\) or rotated to lie in or close to \(S_3\). Syn-\(D_3\) biotite growth during rotation has included curved quartz as inclusions in the margins of the porphyroblasts (Figure 24G).

Where biotites lie between strongly developed \(S_3\) spaced crenulation cleavage surfaces the margins of the porphyroblasts are strained. Partly included quartz grains, which projected beyond the margin of the biotite, have been bodily rotated to lie in \(S_3\) as a result of the shear couple between the strain slip cleavage surfaces. These porphyroblasts can be identified by the sharp change in angle of the quartz grains as they project beyond the edge of the biotite grain (Figure 24H). The margins of the biotite show undulose extinction, and replacement by strain-free chlorite which interdigitates with \(S_e\) (Figure 24H). These strain features appear to be restricted to grains initially elongate in \(S_2\). Inequant biotites of similar orientation in the same thin section, however, may show features compatible with the
syn-D\textsubscript{3} growth described above. Cross-biotites appear to have been readily rotated to lie with their longest dimension in S\textsubscript{3}. Only rarely do cross-biotites show syn-D\textsubscript{3} growth.

Idioblastic to subidioblastic garnets occur in two small circular areas within the chlorite-biotite zone in Domain 1 (Figure 23). Their presence, as well as that of staurolite below the garnet isograde, is due to a local change in the bulk chemical composition of the pelites (Leatherbarrow 1981).

Albite and garnet porphyroblasts occur in the same samples taken from within the garnet outliers in Domain 1. Subidioblastic to idioblastic garnets and idioblastic plagioclase porphyroblasts show a two stage growth. Fine "dust like" particles of quartz and ilmenite define straight trails in the xenoblastic core zones of both minerals. The idioblastic overgrowths contain larger elongate quartz grains which are continuous with S\textsubscript{e} (S\textsubscript{2}). Albite cores may be surrounded by several idioblastic overgrowths which include parallel straight quartz trails where S\textsubscript{i}=S\textsubscript{e}. The overgrowths may have different optical orientations. In Figure 25A, the albite is partly included by the garnet which precluded post-D\textsubscript{2} albite rim growth.

The-grained inclusion trails have been shown to be characteristic of S\textsubscript{1}. The garnet in Figure 26A, which grew pre-D\textsubscript{2} movement to include S\textsubscript{1} as straight trails, may have continued to grow syn-D\textsubscript{2}. Growth involved a small rotation of the garnet or of the external foliation with respect to the garnet. Alternatively, the porphyroblast may have simply been rotated without growth during D\textsubscript{2} causing a deflection of the developing S\textsubscript{2} which was being tightened against the garnet. Note the deflection of S\textsubscript{2} about the albite in
FIGURE 25 - MICROFABRIC FEATURES OF CHLORITE-BIOTITE AND GARNET ZONES

A. Garnet porphyroblast records a two stage growth. The garnet core and adjacent albite include straight very fine grained $S_1$ inclusion trails. The garnet rim records mainly post-$D_2$ growth where $S_1=S_2$ ($S_2$). The slight deflection of $S_2$ about the albite and the slight curvature of $S_1$ indicates a minor rolling of the garnet core prior to the growth of the rims rather than Syn-$D_2$ growth. A post-$D_2$ growth of the sub-idioblastic rims is further indicated by the fact that maximum rim growth occurred perpendicular to $S_2$. Sample A75; Scale bar .5 mm.

B. Post-$D_2$ chlorite includes quartz and opaques elongate in $S_2$. Sample V108; Scale bar .3 mm.

C. Probable post-$D_3$ chlorite (clear) includes quartz elongate in $S_3$. Earlier chlorite (stippled) which grew post-$D_2$ to include $S_2$ is aligned such that included $S_2$ is approximately parallel across the sample. Sample V121; Scale bar .7 mm.

D. Garnet has retrograded (?) to form clusters of biotite, quartz and opaques. Note the corroded margin of the garnet porphyroblast. Sample A6; Scale bar .7 mm.

E. Idioblastic garnet with randomly oriented inclusions has overgrown a cluster of biotite flakes. Large post-$D_2$ cross biotites include quartz oriented in $S_2$. Minor biotite in $S_3$. Sample A14; Scale bar .5 mm.

F. Porphyroblasts of biotite and garnet preserve quartz elongate in $S_2$. $S_3$ is deflected by biotite porphyroblasts but one margin of the garnet may indicate Syn-$D_3$ growth. Rotation of garnet is not suspected because $S_1$ in the garnet parallels $S_e$ in biotite. Sample A159; Scale bar .5 mm.

G. External to the garnet $S_2$ has been transposed into $S_3$ which is deflected about the garnet. Internally the garnet preserve $S_2$ as straight trails perpendicular to the external foliation ($S_3$). Sample A22; Scale bar .6 mm.

H. Hand Specimen. Sketch shows that position and orientation of garnet is compositionally controlled. In thin section garnets contain quartz elongate in $S_2$ and in finer grained crenulated $S_1$. Ga has helicoidally overgrown both foliations post-$D_2$. Sample V106; Scale bar 7 mm.
Figure 25A. Further growth of garnet during post-D2 would helicitically include the curved foliation as trails. The latter explanation is more satisfactory in that the greater development of overgrowths on those margins of the crystal that would have been subjected to the most strain during D2 indicates a mechanism involving post-D2 helicitic overgrowths.

Post-D2 staurolite occurs as small (.5 mm) xenoblastic grains, associated with biotite.

5.5.3 Garnet Zone

The garnet zone in the French Creek section is restricted to a narrow linear belt in Domain 2 (Figure 23). Samples from the area to the north of McCulloch and Old Camp Creeks were studied to ensure a complete representation of the microfabric relationships in the garnet zone. Results are presented in Domain 2 and Domain V2, Charts III and IV, Appendix I.

In the southern part of the zone small laths of chlorite occur as an accessory part of the S2 muscovite-biotite foliation, the formation of which predates the major growth of porphyroblasts. Prograde chlorite porphyroblasts exist throughout the western part of the zone where they clearly postdate S2 (Figure 25B).

Rare probable post-D3 chlorite, oriented in S3, occurs in the French Creek Area but its presence is restricted to those samples which have a strong spaced crenulation cleavage. Where S3 crenulations are open the growth of chlorite is more random with respect to the axial plane of the crenulation and therefore probably postdates D3. Even in samples where chlorite is axial planar to D3, crystal growth may have been helicitic, its position reflecting compositionally
favourable areas only. More convincing evidence of post D3 growth of chlorite is shown in Figure 25C.

Retrograde chlorite occurs throughout the zone as fans and clusters of crystals which postdate S3 crenulations. Chlorite forms rims about biotite and garnet, particularly garnet. Biotite is usually replaced along cleavage surfaces.

Early chlorite porphyroblasts are partly retrograde to muscovite and sericite; remnant chlorite exists within the porphyroblast outline.

Biotite porphyroblasts in the garnet zone have the same microfabric relations as in the chlorite-biotite zone and will not be discussed further here. In the ground mass biotite now rivals muscovite for importance as an S2 foliation forming mineral and varies in abundance, with chemical composition of the pelite, from 20-60%. In the northern part of the garnet zone flakes of biotite define a discontinuous (non-penetrative) S3 foliation which is paralleled by the long axes of rotated cross-biotites. Rotation of the cross-biotites has resulted in a lattice preferred as well as dimensional preferred orientation, which enhances S3. Small amounts of new, and possibly retrograde, biotite have formed at the expense of garnet (Figure 25D) while early biotite porphyroblasts are themselves partly retrograded to chlorite.

Garnet first appears at the isograd as small .2 - .5 mm irregularly shaped porphyroblasts. In sections immediately upgrade, both xenoblastic resorbed garnets with retrograde chlorite and biotite reaction rims (Figure 25D), and rare idioblastic garnet porphyroblasts
which have grown on biotite clusters (Figure 25E), appear to coexist in adjacent samples.

The majority of garnet porphyroblasts are xenoblastic and contain elongate quartz grains in the form of straight inclusion trails. Margins are mildly to strongly resorbed and chlorite rims are invariably present. Garnets can also be observed which have broken-down to form biotite (Figure 25D). This reaction can be observed upgrade in the staurolite-kyanite zone, where a reaction rim of biotite on garnet is surrounded by a rim of staurolite. Both the staurolite and biotite have formed in a reaction involving the breakdown of the garnet (Leatherbarrow, 1981).

The reaction rims have in many cases obscured the relationship between S₁ and S₂. Quartz in S₁ is, however, parallel to and of similar size to quartz elongate in S₂. In only one sample (V120 - Chart IV, Domain V2) is compositional layering preserved but growth appears to have been helicitic to both S₀/S₁ and S₂. Possible early syn-D₃ garnet growth is observed in only one sample (Al5B, Figure 25F) where garnet preserves curved quartz in its one margin. Other syn-D₃ growth rims may have been removed by reaction to chlorite. S₂ however, is usually deflected by the garnet due to flattening and/or crenulation of S₂ during D₃. Where S₂ is transposed into a planar S₃, the new foliation is deflected about both pre-existing biotite and garnet porphyroblasts whose internal trails are parallel across the thin section and make a high angle with S₃ (Figure 25G). Neither the biotite nor the garnet show any syn-D₃ growth.
Regardless of the sequence involved in the appearance and disappearance of garnet all porphyroblasts seem to show a consistent relationship to the formation of foliations i.e. they postdate $S_2$.

Van der Leeuwen (1976), reported garnet growing pre-$D_2$ and folded by $D_2$. Figure 25H illustrates the surface of the hand specimen from which a thin section was cut. $S_0/S_1$ is folded by $D_2$ and a penetrative $S_2$ cuts folded $S_1$ surfaces. However, all but one of the garnets, which are elongated in $S_2$, postdate both $S_1$ and $S_2$. The position of the garnets, which have helicitically overgrown the foliations, is compositionally controlled by the pelitic interbeds within the psammite or by concentration of pelitic minerals in penetrative $S_2$ cleavage surfaces. The main period of garnet growth is therefore considered to be post $D_2$ and pre-$D_3$.

5.5.4 The Staurolite - Kyanite Zone

Domain 3 and part of Domain 2 to the east of French Creek, and Domain J and part of Domain I in the northern part of the thesis area, are underlain by rocks of staurolite-kyanite grade of metamorphism (Figure 21). To the east of French Creek (Figure 23), staurolite and kyanite appear simultaneously. Staurolite coexists with kyanite until the sillimanite isograd, beyond which it coexists with sillimanite for a short distance into the zone of sillimanite grade metamorphism. There is no discrete kyanite zone. In the thesis area staurolite disappears well to the north of the line of first appearance of sillimanite, and a discrete kyanite zone exists (Figure 21). The kyanite zone is discussed separately below. The staurolite-kyanite zones to the north and south of the Selkirk Fan Axis (Figure 22) are characterized by slightly
FIGURE 26 - MICROPORCIC FEATURES OF THE STAUROLITE-KYANITE ZONE

A. Retrograde chlorite rims corroded post-D2 garnet. Quartz in S2 is preserved as straight S1 within the garnet whereas S2 has been tightened and deflected about the porphyroblast during D3. Sample A19; Scale bar 0.6 mm.

B. Coarse inclusion free biotite porphyroblasts and small biotite laths are elongate in S3. Although S3 crosscuts S2 at a high angle S2 has not been crenulated during D3. Staurolite and kyanite have included the S2 foliation as straight trails. Sample F39D; Scale bar 0.9 mm.

C. Retrograde chlorite which has cumenically replaced garnet during a retrograde reaction has been strongly crenulated during D3. Discrete retrograde chlorite occurs as clusters. Sample A134; Scale base 0.7 mm.

D. Idioblastic staurolite and biotite have been formed during a reaction involving the breakdown of garnet. The newly formed biotite and staurolite form concentric rims about the corroded post-D2 garnet. Sample A18; Scale bar 0.9 mm.

E. Sub-idioblastic growth rims on corroded garnet core. Abundant quartz inclusions in the core parallel the external S2 foliation. Retrograde chlorite rosettes in ground mass. Sample A92; scale bar 0.9 mm.

F. Biotite and garnet porphyroblasts contain straight S2 inclusion trails. During D3 crystals have been rotated such that inequant biotites lie with long axes in S3, and quartz grains which protrude beyond the periphery of the garnets have been bent towards S3. Biotite laths define S3 in the groundmass. Sample F39E; scale bar 0.8 mm.

G. Rare multistage garnet growth. Garnet core includes fine S1 trails perpendicular to S2. Post D2 garnet growth is in two stages. An inner zone includes quartz parallel to quartz in S2. The outer zone is inclusion free but appears to have grown pre-D3 because the S2 foliation has been tightened about the outer rim during D3. Sample A127; Scale bar 0.9 mm.

H. A more extreme example of flattening of S2 during D3. Kyanite blades are buckled about garnet porphyroblasts. Note the chlorite rosettes in the pressure shadow. Trails in garnet are probably S1. Sample A125; Scale bar 0.6 mm.
different fabric relationships and in order to simplify discussion are
describale separately. Microfabric information on the staurolite-kyanite
zone can be found in Charts III, V, VI and VII, Appendix I.

To the south of the Fan Axis (Domains 2 and 3, Appendix I)
chlorite flakes in S2 are crenulated by D3. Elsewhere chlorite
forms retrograde rims on garnet and more rarely biotite (Figure 26A) or
exists as abundant retrograde crystals in the matrix.

Biotite flakes, abundant in S2 are crenulated by D3 or
rotated into S3.

Porphyroblasts of biotite, which have overgrown S2, exist
throughout, but are more abundant in Domain 2. Biotite shows all the
features described previously. Most porphyroblasts, in the presence of
D3 crenulations, have been rotated such that S1=S6. Crystals
elongate in S2 deflect S3 and many are noticeably strained. Examples
of syn-D3 growth are documented in Domain 2 (Appendix I, Chart III).

Flakes, larger than those which lie in S2, and new elongate
porphyroblasts with few inclusions define S3 in Domain 3 (Figure 26B).

Early porphyroblasts which overgrew S2 have been replaced by clusters
of smaller biotite crystals with few inclusions. In these cases the
original crystal shape can be observed in plane polarized light,
outlined by an indistinct rim comprised of brown "dust-like" particles.

The relationship between most garnets and the ground mass
foliation has been destroyed by reaction to either chlorite (Figure 26A
and C) or staurolite and biotite (Figure 26D). Most examples, however,
as in the garnet zone, have straight internal quartz trails which are
parallel to and contain quartz grains of similar size to the external
foliation, which suggests a post-D2 garnet growth (Figure 26A).
Additional evidence is provided by a few garnets which have an almost inclusion-free idiomorphic growth rim which precludes D3 crenulation (Figure 26E). The cores of the garnets have straight inclusions trails similar to those described above. Few samples record syn-D3 growth of garnet (see charts). In these samples cross-biotites which grew post-D2 to include \( S_2 \) have been rotated to lie with their long axes in \( S_3 \) (Figure 26F).

Several of the garnets preserve straight, very fine inclusion trains (Figure 26G and H). Figure 26G shows one of the few examples of composite garnet growth observed below the sillimanite isograd. The core of the crystal contains a gently curved, very fine-grained quartz and ilmenite foliation. Surrounding the core is a zone which contains straight quartz trails parallel to but not continuous with \( S_2 \), in this case \( S_2 \). An inclusion-free rim, more characteristic of garnets in the sillimanite zone surrounds these earlier zonal growths. The core most likely contains \( S_1 \) enclosed by a post-\( S_2 \) dominated band. The subidiomorphic rim grew post-D2 but apparently before the flattening of the \( S_2 \) schistosity about the porphyroblasts by D3.

In Figure 26H, a garnet is shown to include fine-grained quartz trails perpendicular to \( S_2 \). Kyanite overgrew \( S_2 \) and included quartz in \( S_2 \) as straight trails. During D3, kyanite was bent about the pre-existing garnet which shows no sign of having been rotated during D3 nor of having included quartz oriented in \( S_2 \). It is therefore likely that the trails represent \( S_1 \).

The appearance of staurolite is abrupt. Well defined sub-idiomorphic to xenoblastic porphyroblasts (1.5 mm) with straight trails exist at the isograd, (Stn. Al8, Figure 24). Their appearance
may be coincident with the breakdown of garnet (Figure 26D). Crystals are both elongate in and at high angles to S₂. The larger crystals generally lie with long axes in S₂ and include straight S₂ inclusion trails (Figure 26D). Cross-staurolite also includes S₂ as straight inclusion trails. Rarely, porphyroblasts are crenulated by D₃ but no syn-D₃ growth of staurolite was observed.

The appearance of kyanite coincides with that of staurolite. Porphyroblasts are generally elongate parallel to S₂ and include S₂ as straight trails. Rare cross-kyanites also include straight S₂ trails.

In the northern part of the zone one thin section contains evidence of post-D₃ kyanite growth (Figure 27B). Note the difference between A and B in Figure 27. In A the lattice of the kyanite has been strained during D₃. In B, however, cleavage surfaces are straight even though quartz inclusion trails are curved and continuous with S₂. S₂ has been crenulated by D₃. The kyanite has overgrown the crest of a D₃ crenulation and is therefore post-D₃. In general, however, the inequant nature of kyanite and its orientation in S₂ has resulted in many crystals being bent, fractured and strained during the D₃ crenulation event (Figure 27). The strain is unannealed and reveals itself as wavy or undulose extinction in the crystal. A post-D₂ but pre-D₃ growth period is indicated for kyanite porphyroblasts.

A staurolite-kyanite zone of metamorphism also underlies the north eastern part of the thesis area, specifically Domain J and the northern part of Domain I (Charts VI and VII, Appendix I).

Remnant chlorite porphyroblasts are included to form the cores of garnet porphyroblasts. The cores contain very fine quartz trails at
FIGURE 27 - MICROFABRIC FEATURES OF THE KYANITE AND SILLIMANITE COPER

A. Post-D3 kyanite bent and fractured by D3 crenulation folding. Both the cleavage plane and the quartz inclusions are warped about the crenulation. Sample A164; scale bar .7 mm.

B. Single example of post-D3 kyanite. Note that in contrast with D4 cleavage is unifomed and quartz inclusions preserve the D3 microfold. Sample A241; scale bar .4 mm.

C. ? post-D3 growth. Large muscovite flakes define the granulated S2 over which the garnet appears to have grown. Inclusions in the garnet core probably represent leftover reactant rather than trails. Sample A165; scale bar .7 mm.

D. Idioblastic, virtually inclusion free garnet has overgrown a fibrolite mat. Crystal faces clearly transect and therefore postdate the sillimanite needles. Inclusions within garnet are random and unrelated to fabric elements. Sample A191A; scale bar .5 mm.

E. Sub-idioblastic garnet has overgrown a fibrolite knot. Fibrolite is elongate in S2 which has been flattened during D3. Note the deflection of S2 about the garnet porphyroblast. Sample P48; scale bar .6 mm.

F. Idioblastic rim on garnet. Position of growth of garnet was compositionally controlled by the pelitic layer. The rim growth postdates the formation of the fibrolite. Note the deflection of S2. Sample A191A; scale bar .6 mm.

G. Carposite staurolite crystal. "Inclusions" are actually quartz infills between crystal faces. Sample F50B; scale bar .3 mm.

H. Pre-D3 sillimanite knot crenulated about D3 microfolds. Sample F55B; scale bar .3 mm.
an angle to $S_2$ such that the relationship between $S_1$ and $S_2$ is not known. For reasons discussed above, however, it is suggested that the fine straight trails represent $S_1$ and that some chlorite therefore grew pre-$D_2$.

There are no recorded examples of chlorite in the matrix either as a pro- or retrograde component. Biotite occurs as a foliation forming minerals only, and along with muscovite defines the strong $S_3$ foliation. Less commonly biotite defines $S_3$ which is axial planar to $D_3$ crenulation. In Domain 1 a late ($D_4$) event crenulates biotite and muscovite in $S_3$, and these crenulated minerals remain unannealed.

The majority of garnet porphyroblasts postdate $D_2$ and either truncate or include $S_2$ as straight trails. Both idiomorphic and xenoblastic post-$D_2$ porphyroblasts exist. A few garnets have nucleated on, or reacted from, chlorite which they include as core zones (see above). Others include very fine grained straight to slightly curved quartz trails oriented at a high angle to $S_2$. In Sample A391 (Chart VII, Appendix I) continued growth of a pre-existing i.e. pre-$D_2$ garnet porphyroblast into $D_2$, preserved a crenulated $S_1$ about a core which includes straight $S_1$ trails. $S_2$ is flattened about all these early garnets which suggests that they grew pre- to earliest syn-$D_2$.

Garnet and kyanite appear to have intergrown. It is possible that in some cases garnet has grown on kyanite in a similar way to its growth on sillimanite at higher grade (see discussion of garnet in the sillimanite zone). In support of this argument is the idiomorphic nature of the garnet “enclosed” by the kyanite. It is likely that if the two had intergrown garnet would only be well formed outside the confines of the kyanite crystal.
In only one sample was syn-D3 garnet growth observed. In the remainder of the samples studied garnet porphyroblasts deflect $S_3$. Where $S_3$ is penetrative in the form of a spaced cleavage, garnets have been rotated such that they disrupt remnant $S_2$ at their margins and $S_1-S_6$. The presence of porphyroblasts of garnet in such rocks is, however, responsible for the preservation of $S_2$. Pressure shadows created by the deflection of $S_3$ preserve a tightly crenulated $S_2$ muscovite-biotite foliation.

There is some evidence of post-D3 growth of garnet. These subidioblastic to idioblastic garnets are small (.5 mm), and contain only a few "dust-like" randomly scattered inclusions in their core. A discrete zoning is not observed, and it is likely that the random quartz in the centre of the crystals represents leftover reactant during the initial growth stage of the garnet, rather than an included fabric element. Post-D3 garnets have overgrown both planar $S_3$ and a crenulated $S_2$ (Figure 27C). Growth of these crystals may have been simultaneous with the formation of idioblastic growth rims on early garnet porphyroblasts. Because of the sitting of the earlier garnets relative to $S_2$ fabric elements rather than those associated with $S_3$, and the inclusion of $S_2$ in idioblastic garnets elsewhere, it is impossible to be conclusive on the timing of idioblastic growth rims. It is equally possible that growth rims grew between D2 and D3. Likewise it is impossible to know the timing of the growth of idioblastic garnet on kyanite, relative to D3.

Both staurolite and kyanite postdate $S_2$ which is either included as straight trails or truncated at the crystal margins. Kyanite has grown with its long axis either parallel to, or at random
angles to $S_2$. Elongate staurolite parallels and has overgrown $S_2$.

Neither staurolite nor kyanite show any evidence of growth during $D_3$.
$S_3$ is deflected by those crystals elongate in $S_2$ and some kyanite
has been crenulated. Cross-kyanites have been rotated such that their
long axes approximately parallel $S_3$. There is one example of a
possible post $D_3$ growth of kyanite (Sample A351, Domain I, Appendix I)
but the example is poor and the confidence level in this interpretation
is low.

5.5.5 Kyanite Zone

Domains E, G, H, and part of Domain I and the northern part of
Domain D are underlain by kyanite grade metamorphic rocks (Figure 21).
Charts VII, IX and X, VII and XI respectively illustrate microfabric
features of the domains above.

$S_1$ is not preserved as a foliation. Rare garnets may
however include $S_1$ as fine inclusion trails (see below). The dominant
foliation in pelitic rocks is a coarse grained planar $S_2$, and $S_1$ has
been obliterated by transposition into $S_2$. This zone, central to the
thesis area, is structurally one of the least complex. $S_2$ is
crenulated only rarely and porphyroblasts simply include $S_2$ as
straight trails. $D_3$ is evidenced, however, as a flattening of $S_2$
about post-$D_2$ porphyroblasts.

Biotite is the dominant foliation forming mineral, except in
Domain E (Figure 21) where muscovite is predominant. Only retrograde
chlorite exists. Porphyroblasts of biotite are not common but those
that do occur postdate $S_2$. $D_3$ folds crenulate biotite and muscovite
in $S_2$. In Domain E (Figure 21) a flattening of second and developing
third phase folds against a nepheline syenite (Figure 3) has caused a crenulation of pre-existing foliations about a sub-horizontal axis. Muscovite and biotite are unannealed across these structures.

Xenoblastic and idioblastic garnets postdate the formation of $S_2$. Both garnet types are inclusion free, have random inclusions, or show the relationship $S_1 = S_e$ where $S_e$ is $S_2$. $S_2$ is later tightened about the garnets by $D_3$ activity.

Only two rock samples include garnets which preserve a possible $S_1$ foliation. The examples are in Domain I. The cores of garnets in sample P147 (Chart VIII) include fine straight trails recognized elsewhere as $S_1$. A developing $S_2$ has been flattened against the garnet. During the post-$D_2$ period a rim of garnet has included the deflected $S_2$. An idioblastic inclusion-free rim envelops the two inner zones. The outer layer may be post-$D_2$ or post-$D_3$. The sample is unaffected by $D_3$ and no criteria exist to be able to distinguish the growth period. Sample P107 (Chart VIII) again records syn-$D_2$ flattening and the deflected $S_2$ has been included by a post-$D_2$ biotite porphyroblast.

Kyanite porphyroblasts elongate in and crosscutting $S_2$ include straight quartz trails where $S_1 = S_e$. In sample M54 an idioblastic garnet is included within the outlines of a kyanite blade. It is not possible to tell if the garnet has been included by, or has grown on, the kyanite. In the light of examples elsewhere a post-$D_2$ period of kyanite growth is indicated.
5.5.6 The Sillimanite Zone

The sillimanite zone underlies a large part of the study area (Figure 21). Sillimanite grade rocks occur in Domains 4 and 5 (Charts XII and XIII, Appendix I) to the south of and including the Selkirk Fan Axis, and in Domains A, B, C, F and part of Domain D (Charts XIV, XV, XVI, XVII and XI respectively) to the north of the fan axis. A zone of potassium feldspar and sillimanite in Domain C (Figure 22) marks the position of the highest grade of metamorphism. Rocks within this zone are coarse grained and generally not suited to microfabric analysis. Nevertheless, samples were collected away from the zones highly disrupted by pegmatite and migmatite, and the information obtained is presented in Chart XVI-Appendix I.

Flakes of biotite and muscovite define a strong planar S2. S3 has folded S2 about 10 to open crenulations. There has been a limited amount of biotite growth oriented in S3 but S3 is non-penetrative on all scales from outcrop to thin section. The large biotite porphyroblasts with straight inclusion trails documented at lower grade disappear in the sillimanite zones to be replaced by clusters of biotite crystals individually only slightly larger than those in the groundmass, except in the potassium feldspar zone where large inclusion free biotite flakes are again common.

Chlorite appears rarely as a retrograde reactant along the cleavage surfaces of biotite.

A new growth of garnet is associated with the appearance of sillimanite. The new garnets are generally idiomorphic, and contain a few randomly distributed quartz inclusions. The best examples are observed growing within and on sillimanite knots (Figure 27D). The face
of the garnet crystals clearly transect fibrolite needles and thus postdate the formation of the sillimanite (Figure 27D and E). The inclusions in the garnet are unrelated to any fabric element, and may simply represent leftover reactant following the formation of garnet. Where these garnets occur in the ground mass, the quartz inclusions are again randomly oriented but tend to be concentrated in the centre of the crystal. This may represent an initially fast growth rate where quartz was included followed by a slower period of growth, during which the crystal was able to rid itself of impurities. Alternatively the quartz may simply be leftover reactant material.

The idiomorphic rims on previously resorbed garnets may be related to the same growth stage. In support of this argument, Figure 27F illustrates an idiomorphic rim on a sub-idiomorphic elongate garnet. The elongate shape of the garnet probably reflects the compositional control exerted by the quartz-rich layer as opposed to the adjacent quartz-poor mica-rich layer. The idiomorphic rim extends into the sillimanite mat where it clearly postdates the growth of sillimanite. The overgrowth is also clearly of the same age as the two smaller garnets at the bottom of the illustration (Figure 27F).

Older xenoblastic garnet porphyroblasts include quartz as straight trails. $S_2$ is tightened against the garnet by $D_3$ activity. A pre-$D_3$, post-$D_2$ timing of growth is indicated.

Staurolite coexists with sillimanite in the southern parts of Domain 4 and 5 only (Charts XII and XIII, Appendix 1).

Staurolite occurs as xenoblastic to idiomorphic single crystals or crystal clusters. A rim of muscovite and/or white mica around the xenoblastic staurolite grains is related to the breakdown of
staurolite. The core of the crystals preserve straight quartz inclusion trails. Elsewhere staurolite, elongate in $S_2$ has overgrown $S_2$ to include it as straight trails. Where $S_3$ is strong rare crystals have been bent or rotated. Staurolite growth is therefore restricted to the interkinematic period between $S_2$ and $S_3$. A word of caution is necessary when dealing with quartz inclusions in staurolite at higher grades of metamorphism. What initially appears to be a poikiloblastic staurolite, which includes random quartz inclusions or quartz at an angle to the external foliation, is found to be a cluster of sub-idioblastic to idiomorphic staurolite grains where the orientation of the quartz is determined by the crystal faces of the staurolite (Figure 27G).

Sillimanite (fibrolite) tends to form in patch-like patches where it appears to be replacing biotite or muscovite aligned in $S_2$. In the association with biotite, however, it is often difficult to tell whether the sillimanite is replacing the biotite or whether the two minerals are growing together as the product of a reaction.

Pods elongate in $S_2$ have been crenulated during $D_3$ (Figure 27H). It is possible that sillimanite is isometrically replacing crenulated biotite to preserve the $D_3$ crenulations. This is considered unlikely in the Selkirk Mountains for the following reasons. Firstly no sillimanite has been observed growing on, or associated with, biotite oriented in $S_3$. Secondly the sillimanite fibers are themselves bent in smooth curves over $D_3$ crenulations and are, in places, unannealed. Finally the idioblastic garnets in the matrix, which are thought to represent the same garnet forming event which produced the idioblastic garnets on sillimanite pods (Figure 28D), can be seen to
postulate \( S_2 \) but not \( S_3 \). The growth of sillimanite is therefore considered to have occurred in late syn-\( D_2 \) and post-\( D_2 \).

5.6 Timing of Metamorphism and Deformation

Despite the range in metamorphic grade, from chlorite to the second sillimanite isograd, and the differences in the intensity of deformation, the development and nature of the regional foliation and the relationships of the porphyroblasts to that foliation are remarkably consistent. The dominant regional foliation has been shown to be \( S_2 \). \( S_1 \) is only locally preserved and its observation in thin section is mainly restricted to rocks of low metamorphic grade. In rocks of low grade the \( S_2 \) foliation is defined by elongate muscovite, quartz and opaques and more rarely chlorite. With increasing metamorphic grade the opaques disappear and biotite becomes a dominant foliation forming mineral along with muscovite and elongate quartz. The percentages of muscovite and biotite vary with composition of the pelites except in the zone of muscovite breakdown where muscovite is less common due to metamorphic reaction.

Porphyroblasts of chlorite, biotite, garnet, staurolite and kyanite have overgrown \( S_2 \). Porphyroblasts are either elongate in \( S_2 \) or randomly oriented with respect to the foliation. All, however, include \( S_2 \) in the form of straight quartz inclusion trails. In most instances porphyroblast growth was completed before the onset of third phase deformation.

\( S_2 \) is either crenulated by third phase folding or the \( S_2 \) foliation planes are flattened and deflected about pre-existing post-\( D_2 \) porphyroblasts. There has been a limited growth of chlorite
and biotite porphyroblasts syn-D3 and rare post-D3 garnet, kyanite, chlorite, and biotite porphyroblast growth.

Micaceous minerals are rotated into S3 adjacent to spaced S3 crenulation cleavage planes and new chlorite and biotite flakes define S3 south of the Selkirk Fan Axis.

Retrograde chlorite occurs sporadically throughout. In the high pressure domain (Figure 20) fabrics are partially recrystallized, but while quartz is commonly polygonised kyanite and biotite blades kinked during D3 show undulose extinction which implies that post-D3 temperatures were insufficient to remove D3-induced strain.

The growth of the majority of porphyroblasts clearly postdates the development of second phase folds and predates third phase deformation (see charts on Figure 28). The peak of metamorphism is, therefore, considered to have occurred post-second phase and pre-third phase of deformation. There is independent evidence to support this suggestion. In the French Creek section the surface trace of isograds, as defined by Leatherbarrow (1981), intersect the axial surface trace of the second phase French Creek synform at a high angle (Figure 21). The formation of the isogradic surfaces therefore postulates second phase folding. There is also independent evidence to suggest that the time span between the two phases may have had considerable duration. Firstly, the static growth period of porphyroblasts had to have been sufficiently long for the growth of large idioblastic porphyroblasts. Secondly, Bignouth Creek stock (Figure 3) truncates phase two folds but contains a foliation related to the third phase of deformation (Van der Leeden 1976). Other smaller granodioritic stocks, which crop out at the head of French Creek, show the same relationships as the above (Brown
and Tippett (1978). And finally Brown and Tippett (1978), further suggest that the spatial distribution and opposed sense of overturning of southwesterly verging phase two and northeasterly verging phase three folds suggest a change in boundary conditions.

5.7. Distribution of Fabric Elements

While the relationships between the development of the second phase foliation and the growth of metamorphic minerals as porphyroblasts are consistent across the study area, remarkable differences occur in both the development of fabrics related to $D_3$, and the relationships of these fabrics to pre-existing ones. Three broad zones can be delineated each characterized by distinctive third phase fabrics. Two of the zones are to the northeast of the Selkirk Fan Axis. The third zone central on French Creek is to the southwest of the fan axis. These zones and the microfabric relationships within them are illustrated on Figure 28.

5.7.1. Zone 1

A strong third phase crenulation of second phase foliation is prevalent throughout Zone 1. The zone encompasses an area underlain by rocks ranging from chlorite to sillimanite grades of metamorphism. The crenulation cleavage varies from a spaced "strain slip" cleavage to a more open crenulation of $S_2$ with increase in metamorphic grade. Where third phase deformation is intense $S_2$ has been transposed into $S_3$ which makes a high angle with inclusion trails in post-$D_2$ porphyroblasts (Figure 26G and 25G).

Equivant porphyroblasts of chlorite and biotite (Domains V1, V2, and 1, 2, and 3) either deflect $S_3$ spaced cleavage without rotation or
show minor rotation and growth into D_3, such that S_1 is continuous with S_e. The lattices of inequant biotite porphyroblasts, where the long axes of the biotite is in S_2, have been strained during D_3. Quartz inclusions which extend beyond the boundaries of the porphyroblasts have been bent towards S_3.

Garnets in zone one are not all of the same growth phase. Early garnets have strongly corroded margins and are surrounded by a reactant halo of chlorite or at higher grades by successive halos of biotite and staurolite. Some of the fabric relationships are therefore masked. Excluding the rare garnets which include fine-grained S_1 most garnets include straight S_2 quartz inclusion trails. Micas in S_3 are deflected about post-D_2 garnet porphyroblasts, and S_1 and S_e are generally at high angles to one another. A few examples exist where S_1 is continuous into S_e (S_3) but the quartz grains which extend beyond the margins of the garnets are bent sharply into D_3. The isotropic nature of the garnets makes it impossible to petrographically detect lattice strain but the similarity between them and biotite porphyroblasts described earlier suggest that the quartz is sheared and/or rotated towards S_3. This motion can occur either as a slight rotation of the garnet relative to the groundmass foliation, or vice versa.

The new growth of garnet in the staurolite-kyanite zone predates and deflects the S_3 cleavage. Kyanite blades have been buckled about garnet porphyroblasts as second phase foliation is flattened and/or crenulated by third phase folding. Pods of sillimanite are slightly elongate in S_2 and either deflect S_3 or are crenulated about D_3 microfoliis (Domains 4 and 5).
Franzen (1974) documented a progressive northwards increase in the intensity of the macroscopic third phase folds from gentle open warps in Domains 1 and 2 to tight major folds in Domains 4 and 5 in the vicinity of the Selkirk Fan Axis. This northward increase in intensity of third phase folding has been illustrated in Figure 17. Fabrics related to the third phase of deformation, however, dominate thin sections from throughout zone 1. "The development of major third phase folds and the crenulation of S2 suggests that tectonic shortening throughout the zone during D3 is likely to have been considerable. It is possible, therefore, that the importance of the third phase of deformation south of the Selkirk Fan Axis has been underestimated in models generated to explain the kinematic history of the area.

5.7.2 Zone 2

The dominant fabric throughout the zone is the regional second phase foliation (Figure 28). Third phase crenulations occur only sporadically throughout most of the zone as do third phase mesoscopic folds, and both seem to be spatially related to hinge areas of isolated mesoscopic third phase folds. The main exception to the above is that area southwest of Bigmouth Creek Fault, specifically Domain A, where third phase folds are strongly in evidence. Domain A may more appropriately belong with Zone 1 (Figure 28).

The zone is underlain by coarse grained rocks of high metamorphic grade. It is probable that the early history of microfabric development has been lost during recrystallization under peak metamorphic conditions. In addition a late idiomorphic growth run on early garnets has masked S1/S2 relationships. Nevertheless
porphyroblasts of biotite and garnet, and in the northern part of the zone kyanite, preserve $S_2$ as straight trails. Sillimanite now is avoid-shaped fibrolite knots slightly elongate in $S_2$. For the most part, fibrolite appears to be replacing biotite.

In both those examples where later growth has involved only garnets and those where later growth is not present, the $S_2$ foliation is tightened and deflected about garnet porphyroblasts. The deflection is not as easy to document as in zone one because the rocks are much coarser grained and some recrystallization of the groundmass foliation has occurred. Third-phase deformation however, appears to be manifested primarily as a flattening of $S_2$ foliation. Where the amount of deflection can be calculated eg. Figure 27, shortening may locally approximate 30%. Relationships between porphyroblasts and the foliation differ from those of zone one in that $S_2$ has not been transposed into $S_3$, and $S_2 \leq S_0$, whereas in zone one $S_2$ has been transposed into $S_3$ (in some samples) and $S_4$ and $S_0$ are generally at a high angle to one another. The dominant foliation in zone two can be traced in areas of third-phase folding where it is precluded by third-phase folds. There is little doubt that the foliation is $S_2$, and this transposition into $S_3$ has not taken place.

In conclusion fabrics related to the third-phase are rare in zone two. Rather, the zone is characterized by a strong coarse grained foliation related to the second phase of deformation.
5.7.3 Zone 3

The zone is dominated by third phase folds which fold second phase folds and fabrics to generate strong third phase crenulation cleavage. The northern part of Domain E, away from the influence of the nepheline syenite body which cores a second phase antiform (Figure 6, Section 3), should probably be included in zone 3. Rocks containing the metamorphic assemblages Ky-Ga-Os and St-Mt-Gz-Ky-Ga (Leatherbarrow 1981), underly the zone.

The well-defined second phase clotsite, muscovite and quartz foliation is folded about tight crenulation folds, and muscays are aligned on spaced $S_3$ cleavage surfaces. $S_3$ is deflected about post $D_2$ garnet and staurolite porphyroblasts which contain $S_2$ trails. $S_2$ inclusion trails make high to right angles with external $S_3$, unlike zone two where $S_1$ and $S_2$ are parallel. Post $P_2$ garnite blades have been kinked or rotated towards $S_3$.

In both Domains I and J there is minor evidence for a post-$D_3$ garnet growth. Small inclusion-free idioblastic garnets postdate the formation of $S_2$ and rare examples appear to be superimposed on $D_3$ crenulation (Figure 27C). There is also one example of a garnet which contains a third phase microfold (Sample A351).

In most respects the microfabrics of zone three resemble those of zone one.

5.7.4 Summary

A study of the microfabrics has allowed the study area to be divided into three zones. A large central zone (zone two) is dominated
by planar second phase fabrics which have been flattened during D1.

The two outer zones (1 and 3) are dominated by macroscopic third phase folding. Second phase folds are refolded resulting in a crenulation of second phase fabrics. The implications of such a zonation are discussed in the following section.
6. STRUCTURAL AND TECTONIC IMPLICATIONS OF MICROFABRIC AND MESOSCOPIC FOLD ANALYSIS

6.1 Introduction

In previous sections (3 and 5) an analysis of both mesoscopic folds and macrofabric data indicate that the study area, including those structural domains along French Creek (Figure 21), can be subdivided into three zones characterized by the influence of either second or third phase folds. The subdivision as illustrated in Figure 28 is:

1. The area south of and including the Selkirk Fan Axis, dominated by third phase folds and fabrics superimposed on macroscopic second phase folds.

2. The central area of the Windy Range and strike equivalent areas to the west, characterized by fabrics of the second phase. Much of the bulk strain in rocks of zone two is the result of further compression of $D_2$ folds and fabrics during an approximately coplanar $D_3$ strain.

3. The northern area from Warsaw Mountain to Mount Neptune, with dominant major third phase folds, and type three interference folding involving second and third phase folds.

An understanding of the relationships between these zones, and the roles played by major folds within them, is crucial before attempting a regional tectonic interpretation. The region is the subject of much controversy and the tectonic models proposed to account for variations in structural style are several (Wheeler 1960; Price and Mountjoy 1970; Roulier 1973; Franzen 1974; Brown and Sappett 1978). The salient features of these models and modifications to the models are
discussed below. Conclusions reached in this chapter will assist in assessing the validity of these models.

6.2 Previous Tectonic Models

Reconnaissance mapping by Wheeler (1963, 1965) in the B.C. and Rogers Pass, areas revealed that the Selkirk Mountains were underlain by one of several structural culminations in the region which exhibited fanning axial surfaces. Folds on the northeastern flank of the fan are overturned northeastwards towards the craton whereas those on the southwestern flank are overturned towards the Shuswap Metamorphic Complex. The Selkirk Fan structure was attributed by Wheeler (1966, 1972) to backfolding and westwards rotation of northeasterly inclined folds, during a late stage wedging between the gneisses of the Shuswap terrane and the edge of the craton.

Price and Mountjoy (1970), interpret the structures of the southeastern Canadian Cordillera, from core zone to thrust and fold, in the east, including the fan structure associated with metamorphic culmination, as a single protracted deformation event caused by the upwards and north-westwards diapiric migration of hot ductile rock from the infrastructure of the Shuswap Metamorphic Complex. The lateral spread under gravity of these upwelling masses resulted in north-eastward translation of the metamorphic infrastructure. Each pulse deformed structures that were formed earlier and produced deformation in the fold and thrust belt to the east.

Roehler (1973) attempted to relate the formation of fans (or rotation zones) to large scale rotation and translation at a plate
boundary, possibly due to plate flip of an ensimatic subduction zone. Folding and rotation is envisaged as occurring around sub-horizontal axes, at the edges or within antithetic fold and thrust belts.

Franzen (1974), in a study of the fan axis in the French Glacier to Argonaut Mountain area established the existence of two major fold phases separated by a metamorphic culmination. Franzen (1974), suggested that the orientation of the early (phase two) folds was comparable to the shallow northeasterly directed imbricate thrust zone of the Rocky Mountains i.e. folds were overturned towards the northeast. Axial surfaces of the later third phase folds vary in dip and fan across the earlier terrane, refolding it and thus forming the Selkirk Fan Axis geometry.

Brown and Tippett (1978), describe the evolution of the fan axis as involving the superposition of the second and third phases of deformation. The fan structure is located where northeasterly dipping second phase structures are overprinted and transposed by steeply dipping third phase axial surfaces. To the southwest of the fan axis, south-westerly overturned second phase folds are dominant while northeasterly overturned third phase folds dominate the northeast flank of the area (Brown and Tippett 1978).

The fan structure is explained by Brown and Tippett (1978), in terms of initial underthrusting of the southwest flank of the Selkirks by crystalline rocks of the Shuswap terrane to form southwesterly-verging phase two folds. Phase three folds, which followed the peak of metamorphism, were due to southwesterly directed underthrusting of the northeastern flank by basement rocks of the craton.
Price (1979), in a critique of Brown and Tippett's (1978) paper reiterates his support of Wheeler's contention (1966), that the Selkirk fan structure is best explained in terms of a cleavage fan produced during one protracted episode of deformation. The upward and outward rotation of the Selkirk-Purcell Anticlinorium is suggested by Price (1979), to have involved strong southwestwards overturning whereby existing foliations become rotated into the direction of compressive strain to become overprinted by newly developing fabrics.

6.3 Local Structural Interpretation

One of the problems facing the writer in the interpretation of the area north of the Selkirk Fan Axis was to determine the pre-phase three orientation of second phase axial surfaces. Previous models had interpreted the second phase folds as initially being either northeasterly directed (Wheeler 1966, 1972; Fransen 1974), or southwesterly directed (Brown and Tippett 1978, Brown 1978).

In Zone One, the dip of second phase axial surfaces varies with the intensity of third phase folding (Figure 17). The dips of the second phase French Creek Synform (Brown and Tippett 1978, Figure 6, G-H p. 555) increase northwards to become vertical across the fan axis. In Domain B (Figure 13), and the remainder of Zone 2, second phase folds dip steeply to the southwest.

In Zone Three, the present dip of phase two axial surfaces reflect folding by southwesterly dipping phase three folds.

Figure 28 illustrates that a different set of microfabric relationships is associated with each of the three zones. Any model generated to explain the structural geometry of the fan axis has to
account for not only the present variation in the attitudes of second phase folds, but also the three-fold distribution of the microfabrics.

In zones one and three the folding of $S_2$ during $D_3$ produces asymmetric crenulations and microfolds. The implication is that axial surfaces of second phase folds, and the axial planar foliation ($S_3$) must have either been initially close too or have been rotated towards the direction of shortening during strain associated with phase three deformation. Locally, in Zone one, $S_2$ either lay in or was rotated into the zone of maximum shortening which resulted in the transposition of $S_2$ into $S_3$. The majority of the porphyroblasts had discontinued growth prior to $D_3$, and therefore most do not record evidence of rotation of the groundmass foliation relative to the porphyroblast. In Zone one a few biotite and garnet porphyroblasts probably record a slight $S_3$-$D_3$ rotation, and $S_1$-$S_e$ where the external trail represents one flank of a $D_3$ microfold (Figure 29). In other cases crystal lattices are noticeably strained and quartz which extends beyond the boundary of the porphyroblast, has been rotated towards $S_3$ without visible rotation of the porphyroblast (Figure 24:1). In these cases the internal trails of the porphyroblast have the same orientation across the width of the thin section. Since it is unlikely that they would all have been rotated by exactly the same amount they presumably preserve the pre-$D_3$ orientation of $S_2$. Similar features can be observed in thin sections from Zone 3.

In Zone 2 the $S_2$ foliation is continuous through the porphyroblast and $S_1$-$S_e$. $S_2$ has been flattened about the porphyroblasts and minor amounts of randomly-oriented, foliation-forming minerals fill the pressure shadows created by the compaction.
Throughout most of the zone S2 must have been oriented close to or have been rotated towards the zone of maximum elongation during D3 strain.

Mesoscopic third phase folds become more abundant in the northern part of Zone 2 as the effects of D3 become more intense northwards. Open gentle crenulations of the S2 foliation give way northwards to the tight crenulations, spaced crenulation cleavages and asymmetric mesoscopic third phase folds of Zone 3 (Plate 6). In a detailed scale a transition can be mapped from Zone 2 where strain during D3 is mostly limited to a flattening of both S2 and a concomitant tightening of second phase folds, to Zone 3 where second and third phase folds are involved in type three interference folding.

The implication from the above is that the initial angle between second and third phase axial surfaces must have been highest next to the Rocky Mountain Trench and in the French Creek area (Zones 3 and 1). Conversely the initial angle between axial planes of the two phases must have been least throughout most of Zone 2.

Zones 1 and 3 should not, however, be directly compared. In Zone 1, Figure 17 schematically illustrates the northwards increase in the dips of the second phase French Creek Synform towards the Selkirk Fan Axis. In this area, the pre-D3 orientation of the second phase axial surfaces is the result of accommodation to the shape and rotation of the Adamant Pluton during deformation. The northeasterly trending second phase axial surfaces were thus at a high initial angle to the northwest trending axial surfaces of the developing third phase folds. Folding of the second phase French Creek Synform becomes more intense northwards as the eastern limb of the synform rotates into the vertical.

across the fan axis. The eastwards increase in the plunge of third phase hinge lines is the result of superposition on the variably dipping limbs of the French Creek Synform.

The situation in the French Creek area (Zone 1) thus differs on a macroscopic scale from that in Zone 3 in that the axial surface traces of second and third phase folds were at a high angle to one another prior to D3. In addition the folding of second phase folds occurs about variably plunging but generally steep third phase fold axes. In contrast second and third phase folds in Zone 3 are co-axial and co-linear, and folding occurs about sub-horizontal axes. It is thus difficult to assess Zone 1 in terms of the typical development of the Selkirk Fan as for example at Sir Sanford Anticline (Brown and Tippet, Figure 7, p. 556).

Away from the area of influence of the Adamant Pluton, second phase folds are southwesterly dipping and involved in interference folding, about sub-horizontal axes, with steeply oriented third phase folds. It should therefore be possible, for the sake of developing a tectonic model for the Northern Selkirks, to ignore the local reorientation of D2 structures about the Adamant Pluton.

The difference in the style of third phase deformation of second phase folds and fabrics might be explained, therefore, by a difference in the pre-D3 orientation of phase two axial surfaces, i.e. a D2 structural fan and/or, by differing amounts of rotation of D2 axial surfaces towards the zone of shortening during strain associated with D3. The two cases are not mutually exclusive and both have probably combined to give the present configuration of folds illustrated in cross sections in Figure 6.
6.3.1 Structural Fan

On the southwest and northeast flanks of the Selkirk Fan Axis, second phase folds are overturned to the southwest and northeast respectively. If we assume, for the moment, that the dip directions represent pre-D$_3$ dip directions, but not necessarily attitudes, a second phase fan structure can be postulated (Figure 29). Second phase axial surfaces would become less steep away from a central area of steeply dipping to vertical axial planes. During D$_3$, where the newly developing folds had steep attitudes, those second phase folds farthest from the centre of the fan might have been oriented with axial surfaces in or close to the zone of D$_3$ shortening, and thus become folded. As the dips of second phase folds become steeper, towards the centre of the second phase fan, the initial angle between second and third phase axial surfaces would decrease until such time that second phase axial surfaces would be close to the zone of D$_3$ elongation. Second phase folds and fabrics would no longer be folded by D$_3$ but would rotate into D$_3$ attitudes such that they would be tightened up during D$_3$ strain. Fabrics transitional between strong crenulation and strong flattening would be observed between the two.

Although such a model might satisfy the observed distribution of microfabrics it requires that the strain axes remained fixed during D$_3$ deformation i.e. pure shear. Pure shear as the major deformational mechanism during D$_3$ can be discounted because of the weakly to strongly asymmetric nature of third phase folds i.e. a rotation of strain axes occurred during D$_3$. Such rotation was possibly the result of simple shear and this mechanism is discussed below.
FIGURE 29: SCHEMATIC REPRESENTATION OF FANNING OF SECOND PHASE AXIAL SURFACES
6.3.2 Rotation by Simple Shear

Since a high initial angle between axial planes is required to develop the type-three interference folding observed on the northeast flank of the Selkirk Fan Axis, Brown and Tippett (1978), inferred that second phase folds dipped to the northeast across the entire area prior to D3. Brown and Tippett (1978) invoked westwardly directed underthrusting of cratonic foreland to generate post-metamorphic third phase folding. Such underthrusting would create a shear couple involving clockwise rotation, (looking northwest along the trench), of stratigraphy, metamorphic isograds and second phase axial planes, towards the Rocky Mountain Trench. Leatherbarrow (1981), suggested such a model to explain the present southwesterly dip of the kyanite-sillimanite isograd in the Windy Ranges on the northeast flank of the metamorphic culmination.

Sections 3 and 4 on Figure 6 are structural cross-sections through the Windy Ranges from Norman Wood Creek to Trident Mountain and beyond. If second phase folds on all scales are unfolded and pre- and post D3 structural configuration compared a shortening of the section of approximately 35% can be attributed to D3 deformation in zone 3. In zone 2, D3 shortening is more difficult to estimate because of the nature of the deformation of second phase fabrics i.e. flattening of the foliation. The amount of curvature of the S2 foliation about porphyroblasts, however, suggests that a D3 shortening of 30-35% for the entire section is probably not unreasonable.

For this model it is assumed that the rotation of second phase axial planes can be attributed to the D3 shear couple and that the accompanying shortening is in the order of 35%. It is also assumed that
FIGURE 30: DEFORMATION BY SIMPLE SHEAR

30A Rotation by Simple Shear

30B Graphical solution for initial orientation of line or surface prior to rotation by simple shear
rotation is via simple shear and that on the scale of the cross-section strain can be considered as homogenous. The final assumption is that the planes of simple shear are horizontal. To the east of the Rocky Mountain Trench the basement dips gently at 3° to the west (Bally, Godly and Stewart, 1966). An approximately horizontal shear couple is therefore considered not unreasonable.

A shortening of 35% as shown by the length of \( \sqrt{11} \), requires an angle of rotation (\( \gamma \)) of 45° (Figure 30A – strained state). The simple shear strain (\( \varepsilon \)) has a value of 1.0 i.e. is equal to one radius of the unit circle. Present orientations of axial planes of second phase folds and \( S_2 \), throughout the Windy Ranges (Figure 6, Section 3), average 60° dip to the southwest. Knowing both the value of the simple shear strain and the present angle between the axial plane of second phase folds and the plane of simple shear (\( \phi \)), it is possible using Figure 30B (Bassay 1967) to calculate graphically the dip of \( S_2 \) etc. prior to \( D_3 \) rotation. Using values of 1.0 and 60° for \( \varepsilon \) and \( \phi \) respectively, initial dips of 67° to the northeast can be calculated for \( S_2 \). Present dips of \( S_2 \) and second phase axial surfaces in Zone Two vary from 55-65° to the southwest which might indicate an initial range of between 65-75° to the northeast i.e. axial surfaces may have rotated by an angle of 40-50°.

A model involving clockwise rotation by simple shear can be invoked to explain the rotation of initially northeasterly-lipping second phase folds into their present moderately-southwest dipping, northeasterly-overturned orientation. A simple shear mechanism is applicable over most of Zone 2.
It is, however, evident that a model involving only a simple shear mechanism falls short of explaining the three fold distribution of fabrics and associated fold interference and that a model which encompasses these phenomena needs to be both composite and complex. The model suggested below includes a partial fanning of second phase axial surfaces synthetic to a zone of northeasterly directed underthrusting; a rotation of steeply dipping second phase axial surfaces and associated reduction in apical angles of second phase folds by a mechanism of simple shear; and the overlap and superposition of axial planes of phase two and phase three folds. The third phase folds are related to a phase of westerly directed underthrusting.

6.3.3 Composite Model for the Structural Evolution of the Northern Selkirk Mountains

Although several assumptions are involved in the construction of the model presented here, most closely follow recently proposed models for the tectonic evolution of the southeastern Canadian Cordillera (Brown and Tippett 1978; Brown 1980b). The early nappe forming event (Phase 1) has been discussed elsewhere and need not be repeated here. It is sufficient to note that stratigraphy had been inverted west of Bighorn Creek Fault prior to the second phase of deformation (Figure 3).

6.3.3.1 Phase Two

Second phase folds were probably formed in response to underthrusting from the west (Brown and Tippett 1978; Brown 1980b). Axial surfaces were oriented synthetic to the zone of underthrusting
FIGURE 31: SCHEMATIC REPRESENTATION OF FOLD AND FABRIC DEVELOPMENT, N. SELKIRK MOUNTAINS
Rather than second phase axial surfaces being overturned to the southwest throughout the zone (Brown and Tippett 1978) it is suggested that axial surfaces fanned as folds died out away from the zone of underthrusting. Figure 31A represents the possible configuration of second phase axial surfaces where folds become steep to vertical to slightly overturned at the position of the trench. Stacking of the folias with resultant crustal thickening accompanied by partial melting and rise of plutonic rocks from subducted crust initiated regional Barrovian metamorphism. A strong muscovite-biotite and elongate quartz grain foliation developed axial planar to developing second-phase folias. Minor porphyroelastic growth may have occurred prior to the cessation of second-phase folding.

6.3.3.2 Interkinematic Phase

During the late stages, and following the formation of second-phase folds, the rock mass continued to heat up such that isotherms rose through an essentially static rock pile. Porphyroblasts of chlorite, biotite, garnet, staurolite, kyanite and fibrolitic sillimanite overgrew and preserved as straight trails the prominent S2 foliation. In the highest temperature areas, foliation-forming minerals continued to grow, resulting in a coarse grained, less well-oriented foliation. During this phase mineral assemblage zones and isogradic or reaction-surfaces were established.

6.3.3.3 Phase Three

Northeasterly directed to upright third phase folds formed in response to the Selkirk mass overriding the cratonic wedge to the east.
Under the Main Ranges Front Ranges and Foothills of the Rocky Mountains, deformation is documented as having occurred above a decollement surface above undisturbed Helikian basement (Sally, et al. 1966). Basement is uninvolved in deformation at least as far west as the present position of the Rocky Mountain Trench. Deformation in the Selkirk Range can be visualized as being due to gentle westwardly underthrusting of infracrustal rocks, during which the supracrustal sedimentary cover rocks were "scraped off" and shortened to form the Fold and Thrust belt to the east. Within the Selkirk Range, third folds formed with axial surfaces synthetic to the underthrusted cratonic plate, parts of which were probably broken off to become involved in eastwardly directed thrusting at depth, e.g. the Malton Gneiss.

The superposition of third phase folds upon second phase folds is illustrated schematically in Figure 31B. For purposes of illustration the eastern edge of the section was remained fixed to show an approximate 35% shortening due to D3. The three fabric zones are delineated and representative fabrics shown beneath each zone. Also shown is the approximate position of the Selkirk Fan Axis.

In Zone three the initial angle between upright second and gently dipping third phase axial surfaces was sufficiently high for hook-like type three interference folds to have formed. An example is shown in Section 4, Figure 6 where a thin flat lying tongue of nepheline syenite outlines a third phase interference fold.

Close to the Trench third phase folds illustrate a progressive decrease in the southwesterly dip of successive third phase axial surfaces northwards, such that stratigraphy drops "step like" into the trench (Figure 6, Section 4).
It is possible to explain observed deformation in the northern part of Zone 3 by considering either displacement along the Purcell Fault or westward underthrusting by the craton. The northeasterly translation of the Selkirk Allochthon (Read and Brown 1981), and concomitant westward underthrusting of the craton (Brown and Tippett 1978), might have caused considerable fault drag on the hanging wall of the Purcell Thrust. During early D3, such resistance to upwards movement adjacent to the fault plane would have the effect of bodily rotating stratigraphy and second phase axial surfaces in a clockwise direction i.e. towards the Trench. A post-metamorphic uplift of 75m for the hanging wall of the Purcell fault has been documented by Craw (1978). Such a large displacement could be envisaged as being accompanied by a local component of body rotation due to drag on the fault plane. A rotation of this type may have been sufficient to locally cause second phase axial surfaces to be rotated into the zone of D3 strain shortening and thus become folded.

Further movement on the fault with continued drag after the main phase of D3 deformation may have caused further changes in orientation of third phase axial surfaces adjacent to the trench.

Alternatively a more regional body rotation and translation of folds and stratigraphy might have occurred due to uplift of the cooling Selkirk mass following the peak of metamorphism and prior to the main D3 folding phase which accompanied westwards underthrusting by rocks of the craton.

In the transition between zone three and zone two where third folds are more upright, the initial angle between second and third phase axial surfaces was lower which resulted in gentle flexure of second
phase axial surfaces to conform to the limbs of large third phase
structures e.g. as at the third phase Mount Neptune syncline (Figure 6
Section 3).

Throughout most of Zone Two the initial angle between second
and third phase axial surfaces was small, and second phase structures
were rotated into third phase orientations. The apical angle of second
phase folds was considerably reduced and second phase foliation
stretched and flattened about post-D2 porphyroblasts.

The sporadic distribution of third phase crenulations and
associated folds throughout the zone might perhaps be explained by
considering the shape of second phase folds prior to rotation. The
limbs of the second folds would have had different initial dip
orientations relative to the developing D3 strain ellipsoid, and would
thus respond differently to strain during rotation. D3 strain on a
local scale is likely to have been inhomogeneous, and this may account
for the distribution and sporadic occurrence of D3 crenulations in
Zone 2.

Clockwise rotation could also explain the present orientation
of the kyanite-sillimanite isograd northeast of the metamorphic
culmination. The present orientation is poorly constrained but the dip
lies somewhere between 70° to the northeast and 40° southwest
(Leatherbarrow 1981; Figure 2–13, p. 69). Leatherbarrow suggests an
initial dip of 45° to the northeast. Using the parameters described
above a 35% shortening and rotation by simple shear would rotate the
kyanite-sillimanite isograd into a vertical orientation i.e. well within
the range documented by Leatherbarrow (1981). An increase in the
percentage of shortening and rotation by simple shear or rotation by a
general shear strain is all that is necessary to rotate the isograd into a southwesterly dipping orientation. The model proposed by Leatherbarrow (1981) to explain the southwesterly dip of the isograd, and the one proposed here to explain the present orientation of second phase folds are thus considered compatible.

Third phase axial planes become steeper until at the Selkirk Fan Axis they are vertical to slightly overturned. Second phase structures are again folded eg. as at Argonaut Mountain and French Glacier (Figure 13), but as previously mentioned it is difficult to assess the true interference of the phases due to both the orientation of the second phase French Creek Synform (Figure 17), and the presence of the Alamant Pluton.

In Zone One a high initial angle was present between axial surfaces of second and third phase folds, but major third phase folding is not observed presumably due to the waning effects of \( D_3 \) strain, and perhaps the presence of the Alamant Pluton. The second phase foliation is strongly crenulated by \( D_3 \) and penetrative crenulation cleavage has resulted in microscopic shearing along discrete \( S_3 \) surfaces. In general a steeply oriented \( S_3 \) is superimposed upon a low angle \( S_2 \). Crenulations are gently assymetric indicating that a rotational component to \( D_3 \) strain was still present at considerable distances from the trench.

6.3.3.4 Late Stage Faulting

The 7km offset documented experimentally across the Purcell reverse fault (Craw 1978) is matched in the south by up to 7km of normal
offset across the Birch Creek Fault (Leatherbarrow 1991; this thesis).

While it is not possible to determine the exact age relationships
between these faults a likely sequence would be as follows. Reverse
movement along the Purcell and associated thrust faults is most likely
related to third phase underthrusting during which basement blocks were
emplaced e.g. the Valton Gneiss, followed in the late stages of tectonism
by listric normal faulting as a relaxation phenomena. Similar late
stage listric normal faulting is known from elsewhere in the
southeastern Canadian Cordillera e.g. the Flathead Fault in the Flathead
valley of B.C. which allowed for the accumulation of up to 3000m of
Kishenehn Formation against the fault.

6.3.3.5 Conclusions

The distribution of second phase folds and the effects of
their subsequent deformation during D3 can be explained by invoking
the superimposition of a late northeasterly directed to upright fold
phase and accompanying clockwise rotation, upon an early southwesterly
directed to upright second fold phase. The effects of the shortening
and folding of the section during D3 strain, along with a body
rotation associated with post-metamorphic uplift and cooling would
combine to produce the zonal distribution of folds and fabrics described
above.

6.4 Tectonic Models

The structural style of the Selkirk Mountains is transitional
between the intensely deformed and metamorphosed core complex and the
Poreland thrust and fold belt. Three tectonic models which attempt to relate structures formed in the core of the orogeny with those formed in the thrust belt to the east have been proposed. An understanding of the tectonic evolution of the Northern Selkirks may help in assessing the validity of these models. The models are summarized in Brown (1978) and are therefore only briefly discussed here.

The allochthonous core zone model of Price and Mountjoy (1970) involves the upwards and northeastwards migration of hot ductile rock from the infrastructure of the Shuswap metamorphic complex. Lateral spreading of those rocks, under the effects of gravity, resulted in northeastwards translation of the metamorphic infrastructure, and in eastwards thrusting and folding of the over rocks in the Rocky Mountains, above a basement decollement. The translation and uplift of the complex are considered to be syn-metamorphic and syn-deformational, and structures are considered to be the result of one protracted deformation where each pulse deformed structures that were formed earlier.

The autochthonous model of Campbell (1973) involves an 11km post-metamorphic vertical uplift of the metamorphic core zone and the western Main Ranges. The Poreland thrust and fold belt may have resulted from westwards underthrusting of the craton.

Brown and Tippett (1978) and Brown (1978) envisage a three stage evolution of the structures which underly the Northern Selkirk Mountains. The earliest stage involved the formation of westward facing nappes, some of which have limb length in excess of 20km (Van der Leeden 1976) and may have evolved in the Paleozoic. The Selkirk fan axis evolved by superposition of two distinct phases of deformation whereby
early southwesterly overturned second phase axial surfaces were
overprinted by later vertical to northeasterly overturned third phase
axial surfaces (Brown and Tippett 1978). The peak of metamorphism
separates the second and third phases of deformation in the northern
Selkirks (Franzen 1974; Van der Leeden 1976; Tippett 1976; Brown and
Tippett 1978; Leatherbarrow 1981 and this thesis).

Brown (1980b), expanded the earlier model of Brown and Tippett
(1978) to include the Shuswap complex and Selkirk-Kootenay terranes.
Three structural and stratigraphic elements comprise the framework upon
which the model is built (Figure 4B). A platform of Archean to early
Proterozoic gneisses overlain by up to 5km of shallow marine clastics of
probably Precambrian age forms the western element. The eastern element
is represented by the Rocky Mountain Miogeocline filled with a
westerly thickening wedge of Proterozoic sediments which unconformably
overlie the cratonic gneisses of the North American Plate (Price and
Mountjoy 1970). The central element is a sedimentary basin filled with
over 10km of proterozoic sediments above an oceanic or transitional
basement. This basin has been named the Selkirk-Kootenay Basin (Brown
1980b).

A pre-Mississippian period of nappe formation in the Omineca
Belt is believed by Brown (1980b) to be related to plate convergence.
Subduction of oceanic crust to the west of the Omineca Terrane caused
uplift, plutonism and the collapse of the western margin of the Selkirk
Kootenay Basin leading to nappe formation. Ghent et al (1977), and
Brown and Read (1979) present evidence for westward displacement of
these nappes onto the Shuswap platform.

In the Middle Jurassic, the phase one nappes were folded about
southwesterly overturned phase two structures (Figure 4C).
Underthrusting of the Shuswap basement beneath the sediments of the Selkirk-Kootenay basin may have accounted for the generation and orientation of these second phase structures (Brown and Tippett 1973; Brown 1980b).

Tectonic thickening of rocks of the Selkirk-Kootenay Basin resulted in crustal thickening associated with the onset of high-grade metamorphism in the infrastructure (Brown 1978). The 25km depth of burial proposed for the peak of metamorphism can be accounted for by combining estimates for the original thickness of the underformed sedimentary wedge with evidence for tectonic thickening (Leatherburrow 1981).

Following the peak of metamorphism, third phase northeasterly overturned folds developed in response to the westerly underthrusting of the tectonically thickened and buoyant Qminean belt by the craton. The late to post-metamorphic northeastward and upward translation of the Selkirk terrane took place by movement on the Purcell and other related thrust faults (Brown 1978).

Information obtained in this study on the timing of metamorphism with respect to deformation and on the distribution of fold phases and associated fabrics is pertinent in assessing the validity of the three models.

In most cases the growth of metamorphic porphyroblasts clearly postdates the formation of S_2. The S_2 foliation is included as straight trails where S_1=S_6. The peak of metamorphism therefore postdates any movement associated with the formation of second phase folds and fabrics, and porphyroblasts have grown during a static
period. The time span between the second and third phases of deformation was long enough not only to have allowed for porphyroblastic mineral growth, but also long enough for the intrusion of small plutons. The Bigmouth Creek Stock, for example, crosscuts second phase folds but includes a weak foliation related to the third phase. The inference drawn from this data is that the time span between fold phases is likely to have had a considerable duration. Clearly a model involving one protracted phase of deformation where translation and uplift of the rock mass are considered to be syn-metamorphic is inapplicable. The allochthonous model of Price and Mountjoy is therefore considered inappropriate for the Northern Selkirk Mountains.

Similarly the autochthonous model of Campbell (1973) is inappropriate for this portion of the Cordilleran fold belt because it does not allow for the generation of southwesterly overturned folds prior to the metamorphic peak. In addition a vertical post-metamorphic uplift of 11 km is insufficient to expose sillimanite and higher metamorphic grades at present topographic levels.

The model which best fits the measured and available data is that proposed by Brown and Tippett (1978), and further modified by Brown (1980b). Further modifications to this model suggested by data from this study would involve a partial fanning of second phase axial surfaces, a clockwise rotation of these axial surfaces to become southwesterly dipping during third phase deformation by a mechanism involving simple shear, and finally some form of body rotation of folds adjacent to the Rocky Mountain trench.
7. SUMMARY AND CONCLUSIONS

The following conclusions were reached during the course of this study.

1) Most of the study area is underlain by Proterozoic Horsethief Creek Group. The southwest corner (Domain A) is in part probably underlain by Cambrian Badshot Formation and Lardeau Group.

2) The Horsethief Creek Group is divisible into 4 map units, from bottom to top, the grit unit, semipelite-amphibolite unit, marble unit and pelite unit. The three upper units are correlated with the lower pelitic member, the middle marble member and upper pelitic member respectively of Brown and Tippett (1978).

The grit unit along the northern margin of the area is tentatively correlated with the Lower Grit Unit of Sumony and Wind (1970).

The four map units of the Horsethief Creek Group therefore correlate regionally with the grit division, slate division, carbonate division and upper clastic division of Young et al (1973).

3) Mapping of the thick carbonate horizons in the southwest of the area (Domain A), and regional considerations suggest that the carbonates belong to the Badshot Formation. Clastics below the Badshot marble probably belong to the Lardeau Group rather than to the Horsethief Creek Group as suggested previously (Wheeler 1965; Brown et al 1977; Brown and Tippett 1978). Strike correlation eastwards suggests that the large marbles which underlie Argonaut Mountain and French Glacier (Figure 3) may also be Badshot Formation.
If the correlation suggested here is correct, then the Badshot Formation and Lanleau Group are in reverse stratigraphic order. The Fault which represents the core of the first phase nappe to the west of French Creek and which separates inverted stratigraphy in the west from right-way-up stratigraphy in the east (Brown and Tippett 1978; Read and Brown 1979) swings across French Glacier and would thus, as suggested, correlate with the Bighorn Creek Fault.

4) Stratigraphy can be readily extrapolated northwards into the Canoe River area where P.S. Simony, E.D. Ghent and graduate students of the University of Calgary have mapped and interpreted Horsethief Creek Group stratigraphy. Stratigraphic mapping by the Calgary group overlaps that of the writer north of Birch Creek Fault, and in the vicinity of Warsaw Mountain. Reversal of younging directions about the axial surface of an inferred first phase nappe (Ghent et al 1977) is not observed in the map area, and is not required to explain the mapped distribution of stratigraphy. Stratigraphy therefore, remains right-way-up across the map area from Bighorn Creek Fault to the north-facing slopes above Columbia Reach.

5) Metamorphic grade increases from staurolite-kyanite zone in the northeast to breakdown of muscovite at the second sillimanite isograd which underlies the southern and central parts of the map area.

6) Folds belonging to the first phase of deformation are rarely preserved. Where found only two limbs of the fold remain and the sense of vergence is unknown. First phase fabrics are only rarely preserved within garnet porphyroblasts. Elsewhere S\textsubscript{1} has been transposed into S\textsubscript{2}.
7) A threefold zonation of mesoscopic folds and fabrics has been outlined. Each zone is characterized by a different set of microfabric relationships generated during third phase folding of second phase fabrics. Differences are due to the different orientations of second phase axial surfaces prior to D3 strain.

The central zone is characterized by second phase planar fabrics (S2). Third phase deformation is restricted to a flattening of pre-existing S2 foliation. The two outer zones are characterized by strong crenulation of S2 during D3, and the generation of type-3 interference folding between second and third phase folds.

8) Analysis of the microfabric and mesoscopic fold data indicates that a partial fanning of second phase axial surfaces probably existed prior to D3. During D3 second phase axial surfaces north of the present position of the fan axis were rotated during simple shear to become southwesterly dipping. Second folds were tightened-up and S2 deflected about pre-existing porphyroblasts. Adjacent to the Rocky Mountain Trench further rotation of second phase axial surfaces placed them in the zone of shortening during D3 strain and they became folded. Later movements on the Purcell and associated thrust faults may also have accentuated local fanning of third phase axial surfaces northeastwards towards the Trench.
PLATE ONE

A. Matrix-supported pebble conglomerate. Whiter clasts to right of hammer handle are quartzite. Flat elongate clasts are psammitic and rarely calc-silicate. Grit unit, Stn. M340.


C. Strongly deformed pebble conglomerate. Original clast to matrix relationships have been mainly destroyed. Photograph shows a local coarse fining-upwards gradation. Photograph by R.L. Brown. Stn. M348.

D. Graded Grits. Above lens cap a 10 cm layer of grit with clasts of quartz and quartz/feldspar to 1 cm grades upwards to coarse sand at top of photograph. Lens cap sits on underlying fines. A complete gradation is difficult to observe due to recrystallization of quartz and feldspar. Photograph by R.L. Brown. Stn. M348.

E. Coarse grained Kyanite-Garnet pelite at top of grit unit. Lighter bands are psammitic and quartz migmatite. Rocks in the outcrop show third phase crenulation. Stn. M343.

F. Second phase folds in pelite at top of grit unit exhibit sheared-out limbs with local injection of migmatite. Stn. M337.

G. Highly sheared zone close to contact of grit unit with overlying semipelite-amphibolite unit. Beds above and below shear appear undisturbed. Stn. M113.

H. Detail of above showing quartz/feldspar augen in pelitic matrix. Stn. M113.
A. Characteristic weathered outcrop of flaggy semipelite from the semipelite-amphibolite unit. Note arrow which points to figure for scale. Stn. M96.

B. Finely interbedded pelite, semipelite and amphibolite exposed on ice-polished surface. Outcrop width approximately 1.5 m. Stn. M128.

C. Photomicrograph of type one amphibolite, showing interlocking mosaic of hornblende and plagioclase. Field of view 24 x 36 mm. Stn. M38.


E. Photomicrograph of biotite rich band with garnet porphyroblasts from margin of type two amphibolite. Field of view 24 x 36 mm. Stn. M127.

F. Photomicrograph of type two amphibolite showing fine banding due to alternating amphibole and feldspar rich layers. Field of view 24 x 36 mm (Plane light). Stn. M47.

G. Folded type three amphibolite. Lighter bands are silicate rich and include quartz, calcite, garnet, epidote and titanite. Dark bands are composed almost exclusively of amphibole and feldspar. 100m west of Stn. M127.
PLATE THREE

A. Rounded outcrop of massively bedded coarse grained grey marble of the marble unit. Stn. M185.

B. Thick succession of thinly bedded buff (rusty) marble bands with interbedded white to grey marbles. Grey marble in centre of cliff 2 m thick. Photograph taken looking south from Stn. M159.

C. Distinct interbedded sequence of grey and more recessively weathering buff marble bands. Layers are approximately .5 m thick. Stn. M185.

D. Recessive weathering rusty marble which weathers to a coarse grained calcareous "sand" (bottom left) along the ridge tops. Stn. M162.

E. Fine laminations in buff marbles outline early intrafolial folds. Pegmatite layers have been severely boudinaged and marbles can be seen to have "flowed" into the spaces between boudins. Stn. M161.


G. White quartzite layers with thin pelitic interbeds or micaceous bedding planes typify the quartzite unit. Beds are 10-15 cms thick. Note: flowers for scale. Stn. M164.

H. Contact of coarse grained sillimanite-garnet pelite and finely laminated siliceous marbles from the upper pelite unit. Stn. M220.
A. Thick sequence of coarse-grained, coarsely bedded grey marbles characterise the marble sequence south of Bigmouth Creek Fault. Outcrops show rounded weathered layers with recessively weathered bedded surfaces. Stn. M226.

B. Siliceous stringers and layers define early intrafolial folding in otherwise massive grey marble. Pencil is aligned parallel to both cleavage of early folds and present bedding. Stn. M229.

C. Thick sequence of yellow-buff thinly bedded marbles which form sharp angular ridge tops and scree in contrast to the rounded grey marbles. Small pines on outcrop 1-1.5 m in height. Stn. M237.

D. Contact of massive grey marbles with underlying pelitic succession (lower right). From contact on ridge to top of peak on right is 150 m. Photograph taken from Stn. M227.

E. Thinly bedded carbonate (in core of fold) pelite and psammite (lower right) outline a phase two fold. Part of sequence of lithologies which underlies northern of two large marbles. Stn. M238.


H. Dark pelite in the core of a major second phase synform outlined by grey marbles. Photograph taken looking west from Stn. M211.

I. Looking east from Stn. M211 towards Argonaut Mountain where two peaks are underlain by massive grey and buff marbles. Thin to thickly bedded marbles of Stn. 211 in foreground.
PLATE FIVE


B. Pods of sillimanite and quartz define the $S_2$ foliation which crosscuts compositional layering at a shallow angle. The pencil parallels $S_2$ while the magic marker parallels compositional layering. Stn. M261.

C. Elongate pods of sillimanite and quartz define $l_2$. The elongation direction parallels the hinge line of adjacent $F_2$ folds. Stn. M261.

D. Second phase folds in psammites. Note thickening in the hinge area and attenuation on limbs which indicates a flexural flow mode of formation. Ice axe for scale. Stn. M23.

E. Third phase folds in interbedded pelite and semipelite. Note the slight fanning of axial planes of the minor crenulation folds outlined by the thin semipelite in the core of the larger antiform. Stn. M215.

F. Asymmetric third phase folds in psammitic rocks of the grit unit. Note figure for scale. Stn. M279.
PLATE SIX

A. Upright third phase folds developed in a thinly bedded pelite and
schistose sequence. The fold is sufficiently tight for the pelites
to have developed a spaced crenulation cleavage. Stn. M204.

B. Close-up of S3 spaced crenulation cleavage in pelites. Pen is
aligned parallel to S3 which is non-penetrative through the
psammitic layer (left of centre). Stn. M114.

C. Crenulated sillimanite-bearing pelitic schist. White sillimanite-
quartz pods are folded about open third phase crenulations. Stn.P50.

D. Strongly migmatized and foliated pelite includes an isolated one
metre block of quartzite which preserves two limbs of an early fold
(F1). Stn. P78.

E. Photomicrograph of pelite close to fault zone in Domain D. The
original mylonitic texture is partly recrystalized such that while
the quartz in the augen is polygonised, original augen shapes are

F. Tight map-scale second phase folds. The axial planes of the folds
are gently warped by a major third phase synform, the core zone of
which is preserved on the horizon. The waterfall in the upper right
is approximately 20m high. Stn. P121.

LIST OF REFERENCES


190


APPENDIX I

Appendix I is a complete list, in chart form, of all macrofabric relationships observed in the study. Each chart represents one domain (Figure 21) and includes all thin sections taken from samples within that domain. Seven charts, Domains 1-5 and Domains VI and V2, correspond to Franzen's (1974) Structural Domains and part of Van der Leeden's thesis area (1976) and are arranged in order of both increase in metamorphic grade and amount of deformation. The remainder of the charts A-J correspond to structural domains within the thesis area. As the charts involve an unorthodox presentation of the relationships between mineral growth and phase of deformation some words of explanation are necessary. The following instructions are best read while viewing one of the charts.

The vertical axis on the left hand side of the chart lists minerals observed within the domain and thus provides an indication of metamorphic grade. Note that although the charts have the same framework and that minerals are always listed in the same order, from top to bottom, variations in metamorphic grade result in a particular mineral occupying different spaces in the column from chart to chart. The horizontal axis is standard i.e. $D_1$, $D_2$, etc., with the blank spaces representing interkinematic periods. In this respect the charts are standard in that the appearance of a mineral in one of the vertical columns indicates the presence of that mineral during that phase in the deformational history.

The charts are meant to be of use to a petrologist as well as a structural geologist and therefore rather than the standard bar
symbol each species of mineral was individually symbolized. Both
porphyroblasts and matrix forming minerals are represented and the
relationships between them also symbolized. Where possible the symbol
has a physical resemblance to the outline of the minerals as they
appear in thin section e.g. an idioblastic garnet is represented by a
hexagon and a xenoblastic one by a small circle. All symbols are listed
below. The user is therefore able to tell whether idioblastic garnet,
for instance, grew syn- or post-D2 and whether or not it included S2
foliation-- forming minerals. Similarly the growth periods of the
other minerals can be simply illustrated. A complete history of the
timing of mineral growth versus deformation in a particular domain can
be readily appreciated by looking at one chart. Domains can easily be
compared and changes in the constituent fabric elements which accompany
an increase or decrease in either metamorphic grade or deformation
become obvious.

When looking at the standard "bar chart" in publications it is
rare to find listed the number of examples of a fabric element upon
which the authors have based their conclusions. To avoid the bias
inherent in this sort of study and to obtain a true statistical balance
between the various microfabric elements, a list of thin sections is
provided at the bottom of each chart. Next to each thin section listed
is a series of asterisks in columnar form under a set of letters. Quite
simply each index letter refers to a particular fabric element or
relationship illustrated in the chart above, and the letter appears next
to that fabric element. Where five or less thin sections show a
particular feature no letter is assigned. Rather sample numbers are
printed next to the fabric symbol. By running one's eye vertically down
the columns, one can see at a glance whether a particular fabric element is commonly or rarely observed. By accurately describing each thin section and faithfully recording the fabrics much of the bias mentioned above is hopefully removed.

An additional attraction of the system outlined above is that the user is instantly aware of which thin section he must turn to if he wishes to view a particular feature for himself. All thin sections used in the study are housed at Carleton University with R.L. Brown.
LIST OF SYMBOLS USED IN MICROFABRIC CHARTS

1. FOLIATION ELEMENTS

/   mineral aligned in S₁

→  minerals aligned in S₂

++  crenulated S₁ preserved between spaced S₂ foliation

↓   minerals aligned in S₃

\\  S₂ foliation crenulation by D₃

2. PORPHYROBLASTS

□  Chlorite, biotite

□  Plagioclase

○  Garnet (idioblastic)

○  Garnet (xenoblastic)

○  Staurolite (idioblastic)
Staurolite (xenoblastic)
Kyanite
Sillimanite

3. RELATIONSHIPS BETWEEN POMPHOROBLAST AND $S_2$ FOLIATION

- Post-$D_2$ biotite has overgrown but not included $S_2$

- Post-$D_2$ minerals have overgrown and included $S_2$ as straight trails

- Post-$D_2$ biotite (cross-biotite)

- Post-$D_2$ biotite also includes fine $S_1$ trails

- Post-$D_2$ biotite includes crenulate $S_1$

- Post-$D_2$ biotite clusters

- Post-$D_2$ garnet with random inclusions
garnet has overgrown composition layering. The finer inclusions are preserved from the mica rica as opposed to the quartz rich microlithons.

Two stage growth. Inner zone contains fine S1 trails at an angle to external S2. Outer zone is inclusion free.

Two stage garnet growth. Inner xenoblastic core postdates and includes S3. Outer idioblastic rim is inclusion free.

Post-D2 kyanite includes or is overgrown by an idioblastic ga. (timing of growth not certain).

Post-D2 intergrowth of kyanite and garnet.

Syn-to post-D2 sillimanite pod has been overgrown by idioblastic garnet. Sillimanite fibres are truncated by the margin of the garnet porphyroblast.

4. RELATIONSHIPS BETWEEN POST-D2 PORPHYROBLASTS AND D3/S3

biotite: cross-biotite rotated without continued growth during D3

biotite and garnet show some growth and rotation during D3

quartz outside porphyroblast has been bodily rotated towards S3

S2 crenulated by D3 outside post-D2 porphyroblast
Post-D₂ porphyroblasts crenulated during D₃

S₂ tightened about post-D₂ porphyroblasts

S₃ deflected about post-D₂ porphyroblasts

**POST-D₃ MINERAL GROWTH**

- Post-D₃ chlorite porphyroblast

☑ Post-D₃ growth includes crenulate S₂

☑ Post-D₃ idioblastic garnet has overgrown a D₃ microfold

☑ Retrograde chlorite has overgrown D₃ microfold

R Retrograde mineral growth - timing of growth may not be clear
<table>
<thead>
<tr>
<th>Chart</th>
<th>Domain</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chart I</td>
<td>Domain V1</td>
<td>201</td>
</tr>
<tr>
<td>Chart II</td>
<td>Domain 1</td>
<td>203</td>
</tr>
<tr>
<td>Chart III</td>
<td>Domain 2</td>
<td>205</td>
</tr>
<tr>
<td>Chart IV</td>
<td>Domain V2</td>
<td>209</td>
</tr>
<tr>
<td>Chart V</td>
<td>Domain 3</td>
<td>211</td>
</tr>
<tr>
<td>Chart VI</td>
<td>Domain J</td>
<td>213</td>
</tr>
<tr>
<td>Chart VII</td>
<td>Domain I</td>
<td>215</td>
</tr>
<tr>
<td>Chart VIII</td>
<td>Domain E</td>
<td>218</td>
</tr>
<tr>
<td>Chart IX</td>
<td>Domain G</td>
<td>221</td>
</tr>
<tr>
<td>Chart X</td>
<td>Domain H</td>
<td>223</td>
</tr>
<tr>
<td>Chart XI</td>
<td>Domain D</td>
<td>225</td>
</tr>
<tr>
<td>Chart XII</td>
<td>Domain 4</td>
<td>227</td>
</tr>
<tr>
<td>Chart XIII</td>
<td>Domain 5</td>
<td>229</td>
</tr>
<tr>
<td>Chart XIV</td>
<td>Domain A</td>
<td>231</td>
</tr>
<tr>
<td>Chart XV</td>
<td>Domain B</td>
<td>233</td>
</tr>
<tr>
<td>Chart XVI</td>
<td>Domain C</td>
<td>235</td>
</tr>
<tr>
<td>Chart XVII</td>
<td>Domain F</td>
<td>237</td>
</tr>
</tbody>
</table>
CHART I - DOMAIN VI
(Chlorite-Biotite Zone).

The domain lies below the appearance of ga, wholly within the chl and chl-bio zones of metamorphism.

Grain Size:

Matrix - ranges from .1-.4 mm; av .2 mm
Porphyroblasts - Bio 1-2 mm, Chl 1-1.5 mm, Plag 1-3 mm

Microfabric Development

D₁ (S₁) - S₁ is defined by mu, qtz and rare chl and preserved as a crenulate foliation between spaced S₂ cleavage planes. Generally S₁ has been transposed into S₂.
Pre-D₂ - One thin section (V6NB) shows chl porphyroblasts which include fine straight trails.
D₂ (S₂) - Defined by mu, qtz, chl and more rarely bio.
Syn-D₂ - Rare chl porphyroblasts grew syn-D₂ to include crenulate S₁ trails. S₂ is deflected about chl. Figure 14D shows pressure shadows at the chl margins.
Post-D₂ - Main porphyroblast growth period. Bio, chl and intergrown clusters of bio/chl include S₂ as straight qtz trails.
D₃ (S₃) - Strong crenulation during D₃. S₃ in some cases is a penetrative strain-slip cleavage. Elsewhere S₃ is not penetrative on the scale of the thin section. D₃ has crenulated S₂ and some inequant post-D₂ porphyroblasts. Cross-micas were rotated so that their long axis is approximately parallel to S₃. Equant porphyroblasts deflect S₃.
Syn-D₃ - Rare examples of continued porphyroblast growth into D₃ (V29A). Some bio flakes appear to have grown in S₃.
Post-D₃ - One example of a (?) prograde post-D₃ chl. Some retrograde chl.
<table>
<thead>
<tr>
<th>CHART I</th>
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</thead>
<tbody>
<tr>
<td><strong>DOM VI</strong></td>
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<tr>
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</tr>
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<tr>
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</tr>
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<tbody>
<tr>
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<tr>
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</tr>
<tr>
<td><strong>V1A</strong></td>
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<td><strong>V26C</strong></td>
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<td><strong>V29A</strong></td>
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CHART II

DOMAIN 1
The domain lies wholly within the chl-bio zone of metamorphism: ga and st. appear due to favourable bulk chemical composition (R.W. Leatherbarrow 1981).

Grain Size:

Matrix - grain size ranges from .1-.4 mm, av. .2 mm. Slight increase northwards towards the ga. isograd.

Porphyroblasts - Bio, 1-3 mm
    Chl, 1-1.5 mm
    Ga, 2-4 mm
    Plag, 1-3 mm
    St, 5 mm (av. size)

Microfabric Development

$D_1$ ($S_1$) - Observed only rarely in thin section. Preservation of $S_1$ is discussed in the text and need not be repeated here except to point out that bio porphyroblasts which preserve $S_1$ also preserve $S_2$.

$D_2$ ($S_2$) - Defined by qtz, mu, opq and more rarely chl and-bio.

Synt-D$_2$ - No evidence of synt-D$_2$ porphyroblast growth.

Post-D$_2$ - Porphyroblasts of bio, chl (ga and st) overgrew and included $S_2$ as straight qtz trails.

$D_3$ ($S_3$) - $S_3$ is a spaced penetrative strain-slip cleavage defined by chl, bio and mu. Elongate mineral grains adjacent to $S_3$ surfaces have been rotated to lie approximately in $S_3$. Chl and bio grains which have grown in and not been rotated into $S_3$, are "cleaner" and larger than rotated grains.

Syn-D$_3$ - Bio porphyroblasts appear to have continued to grow into $D_3$. Curved ends to centrally straight inclusion trails have resulted from continued growth of the porphyroblast during rotation of the porphyroblast relative to the ground mass or vice versa. The amount and direction of rotation and presumably strain appears to have been homogeneous within the limits of the thin section in that $S_1$ lines up across the section.

Elsewhere porphyroblasts deflect $S_3$. The margins of some of the crystals adjacent to $S_3$ cleavage surfaces are strained and it is possible that curved ends to inclusion trails are the result of deformation by strain-slip motion on $S_3$.

Post-D$_3$ - No evidence of post-D$_3$ porphyroblast growth.
### Chart II

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**Bio**

- a
- b
- 13a
- d
- e
- f3
- f10a
- a77b

**Ga**

- a6b
- a75
- a69

**St**

- a64

**Upq**

- as for Mu.

**Mu**

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- a60b
- a75
- a66

**List of Thin Sections**

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- A49
- A53
- A55
- A56
- A75
- A60b
- A59
- A23b
- A60b
- A59
- A55
CHART III

DOMAIN 2
Metamorphic grade within the Domain ranges from chl-bio at its southern end to st-ky in the north (Figure 21).

Grain Size:

Matrix - composed of qtz, bio, mu and more rarely opq., ranging in size from .1 mm to 1 mm; av. size increases from .2 mm to .4 mm upgrade.

Porphyroblasts - Bio, 1-1.5 mm; rare 2-3 mm grains Ga, 2-7 mm; av 4 mm St, <1-5 mm; av 2-3 mm Ky, 1 mm-1 cm; av 2-4 mm

Microfabrics

\(-D_1 (S_1)\) - Ga includes as straight trails fine qtz in S1. Straight fine trails indicate that ga began to grow before the heating episode responsible for the coarsening of fabric and before the onset of D2. Syn-D2 ga growth preserves mildly to strongly crenulated S1. One post-D2 ga preserves both S0/S1 and S2 (Figure 14C).
- Foliation defined by bio, mu and rare chl.

Syn-D2 - Only the ga's described above preserve any Syn-D2 growth.

Post-D2 - Post-D2 bio, ga, st and ky preserve S2 as straight inclusion trails. Ga with random inclusions and st and ky with no inclusions truncate and therefore also clearly postdate S2. Relationship between S1 and S3 is obscured in some garnets by:
- a) reaction to bio or bio and st. b) retrograde chl.

D3 (S3) - S3 is generally a well defined mu-bio cleavage which forms as a spaced penetrative (strain-slip) foliation which is deflected about pre-existing porphyroblasts. Some porphyroblasts are bent between S3 cleavage surfaces. Elsewhere D3 has simply crenulated D2 and a penetrative S3 is not developed.

Syn-D3 - Bio porphyroblasts probably continued to grow into (early) D3. S3=S2 where S3 is comprised of a straight central section with curved tails which are continuous with S2 across crystal boundaries.

Elsewhere S2 is preserved in porphyroblasts only as straight trails while S3 is crenulated by D3 folding.

St and ky show no D3 growth but rather are bent during D3. Strain is manifested by
bending of crystal cleavage surfaces accompanied by sweeping extinction or by cracked and broken porphyroblasts across D₃ crenulations.

Post-D₃

One e.g. of a post-D₃ biotite which has grown across a D₃ crenulation. Some chl clusters may have grown post-D₃.
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### RETROGRADE CHLORITE:

- **Post D3 clusters** - A14, A23, A27.
- **Retro. on Bio** - A28, A134, A127, A125A, F41A.

CHART IV
DOMAIN V2
The domain is underlain by rocks of garnet grade of metamorphism.

**Grain Size:**

- **Matrix**
  - composed of bio, mu, chl and qtz ranging in size from 1-4 mm

- **Porphyroblasts**
  - Bio, 1-2 mm, rare 4 mm grains
  - Chl 1-2 mm
  - Ga, 5-3 mm

**Microfabric Development:**

- **D₁ (S₁)**
  - One qa includes fine qtz trails at an angle to Se (S₃)

- **D₂ (S₂)**
  - Foliation composed of bio, mu, chl and elongate qtz.

- **Syn-D₂**
  - No evidence of Syn-D₂ porphyroblast growth.

- **Post-D₂**
  - Bio, chl and qa porphyroblasts overgrew S₂ to include it as straight qtz trails. In some cases S₁=Sₑ but later rotation during D₃ or retrograde replacement by chlorite on both qa and bio has masked internal and external foliation relationships.

- **D₃ (S₃)**
  - One example of new biotite growth in S₃.

  Elsewhere D₃ crenulates S₂ foliation.

  Post-D₂ chl porphyroblasts have been rotated such that S₁=Sₑ. Qtz grains which protrude beyond bio porphyroblasts are rotated towards S₃. Bio are strained and replaced by chl.

  In two thin sections S₂ is tightened about post-D₂ garnet.
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CHART V

DOMAIN 3
The domain lies wholly within the st-ky zone of metamorphism.

Grain Size:

Matrix  - composed of bio, mu and qtz ranging in size from .1-1 mm; av .2-.4 mm. In the south mica flakes rarely exceed .4 mm but in the northern part of the domain isolated examples of both mu and bio attain 1 mm in length.

Porphyroblasts - Bio, 1-2 mm; av 1.5 mm
St, .4-3.5 mm; av 1-2 mm
Ga, 1-5 mm; av 3 mm
Ky, <1-5 mm - many sizes in any one thin section.

Microfabrics

D₁ (S₁) - S₁ is only preserved within Syn-D₂ ga porphyroblasts as a crenulated fine-grained qtz schistosity.

D₂ (S₂) - S₂, a planar foliation, is defined by bio, mu and elongate qtz.

Syn-D₂ - Ga have grown Syn-D₂ to preserve a crenulated S₁. Outside these ga’s no evidence of S₁ remains suggesting transposition into S₂.

Post-D₂ - Bio, ga at and ky porphyroblasts grew statically preserving as straight trails a planar S₂. Generally S₁=S₉ except where relationships are destroyed by D₃ activity.

D₃ (S₃) - Defined by bio and rotated mu. D₃ bio is larger and cleaner than S₂ bio and shows no sign of unannealed strain. S₃ is deflected by pre-existing porphyroblasts. Some porphyroblasts have been bent and strained during D₃ especially bio and ky.

Sample A151A records possible ga growth continuous from the D₂/D₃ interkinematic period.

Post-D₃ - No evidence of post-D₃ growth except for possible retrograde chl.
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**LIST OF THEIR SECTIONS:**

<table>
<thead>
<tr>
<th></th>
<th>A155</th>
<th>A144A</th>
<th>A144B</th>
<th>A151A</th>
<th>A151B</th>
<th>A149</th>
<th>A148A</th>
<th>A161</th>
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<tbody>
<tr>
<td></td>
<td>***</td>
<td>**</td>
<td>***</td>
<td>***</td>
<td>***</td>
<td>**</td>
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</tr>
</tbody>
</table>

**DE DE G**
CHART VI

DOMAIN J
The domain lies mostly within the st-mu-qz-ky-bi-ga zone of metamorphism (Figure 21).

Grain Size:

Matrix
- Bio, 1-3 mm; av 2 mm
- Mu, 1-3 mm; av 2.5-3 mm
- Qtz, up to 3 mm
- Plaq, 1-2 mm

Porphyroblasts
- St, .5 mm
- Ky, 1-7 mm, av 3 mm
- Ga, 1-5 mm, av 2.5-3 mm

Microfabric Development:

D1 (S1)
- no evidence of S2 in thin section unless fine trails in chl are S1. Chl now included as core to inclusion free ga porphyroblasts.
- Relations of S1 to S3 not known.

D2 (S2)
- Defined by musc with lesser amounts of bio and qtz (except for section A358). Crenulate S2 preserved between spaced S2 cleavage - original planar foliation preserved next to porphyroblasts only.

Syn-D2
- No evidence of syn-D2 porphyroblast growth.

Post-D2
- Ga, st and Ky porphyroblasts postdate S2 and either truncate or include as straight qtz trails.

D3 (S3)
- A strong spaced D3 cleavage axial planar to mesoscopic folds is defined by mu and bio much of which has been rotated to lie in or approximately in, S3. Residual strain remains in crenulate grains. S3 is deflected about equant porphyroblasts e.g. ga, but ky is bent and crenulated between S3 surfaces. Where the long axis of ky grains were initially at a high angle to S2, ky has been rotated to lie approximately in S3. Strain effects are visible in both rotated and crenulated grains.

Syn-D3
- No evidence of Syn-D3 porphyroblast growth.

Post-D3
- Late euhedral inclusion free ga appears to postdate S3. Some of these small (1 mm) ga are associated with qtz "sweats" and may be hydrothermal in origin.

D4 (S4)
- Manifested as a weak to strong crenulation of planar S3. No associated mineral growth.
<table>
<thead>
<tr>
<th>Dom J</th>
<th>D1</th>
<th>D2</th>
<th>D3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chl</td>
<td>A358 A358</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bio</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ga</td>
<td>A357 A357</td>
<td></td>
<td></td>
</tr>
<tr>
<td>St</td>
<td></td>
<td>A358 A353A</td>
<td></td>
</tr>
<tr>
<td>Ky</td>
<td></td>
<td>A358 A357 A353A</td>
<td></td>
</tr>
<tr>
<td>Sill</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mu</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Qtz</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**List of Thin Sections**

- A355A- a b c
- A357 - **
- A358 - ***
- M270A- **
- A377A- *
- M275 - **
- A369 - *
CHART VII

DOMAIN I
DOMAIN I: ADDITIONAL MICROFABRIC INFORMATION

The domain straddles the staurolite-out isograd (Figure 21).

Grain Size:

Matrix
- Bio, 5-6 mm; av 1-2 mm
- Mu, 1-5 mm; av 2-3 mm
- Qtz, 1-3 mm; av 2 mm
- Plaq, 5-4 mm; av 1-1.5 mm

Porphyroblasts - Ga, 1-8 mm; av 3 mm except late ga - av .5-1 mm
- Ky, 1 mm-1 cm; av 2.5-3 mm
- St, 1 mm-2 cm

Microfabric Development

D₁ (S₁)
- Some evidence to suggest that S₁ is preserved in pre-D₂ ga as very fine-grained qtz trails. Chl porphyroblasts which form the core to ga in M278 may also preserve S₁.

D₂ (S₂)
- Defined by mu, bio and qtz.

Syn-D₂
- Very fine trails which probably represent S₂ were slightly warped and crenulated prior to inclusion in xenoblastic ga. Both these and pre-D₂ ga (above) deflect S₂; the deflection being greater in the latter case. In sample A347 idioblastic ga growth has included the deflected S₂ as curved trails.

Continued growth of ga in the form of syn-D₂ rims on pre-D₂ ga lead to the inclusion of crenulate fine-grained S₁ trails. Other ga began to grow syn-D₂ during crenulation of S₁ and include curved qtz trails; qtz in S₁ is in all cases much finer grained that in S₂.

In all above cases S₁=S₂.

Post-D₂
- Ga, st and ky grew post-D₂ to include S₂ as straight trails. Where S₂ is not included it abruptly terminates against porphyroblast margins. Both idioblastic and xenoblastic ga include S₂.

In addition inclusion-free idioblastic ga postdate S₂. The growth of these idioblastic ga may have immediately followed cessation of D₂ movements.

In support of this idea ga in section A344 have a bio reaction rim which in some cases is surrounded in turn by st. A similar st forming reaction was observed south of the Selkirk Fan Axis. Some euhehedral ga appear to be included in ky.
D$_3$ (S$_3$)  - Where present S$_3$ is defined by mull, bio and qtz. D$_3$ manifests itself as a strong crenulation of S$_3$. Ky is both kinked and crenulated and both ky and qtz deflect S$_3$.

Syn-D$_3$  - In Section M114 straight central trails in ga become curved towards the margins of the porphyroblasts, S$_1$=S$_e$, which is a crenulate S$_2$. Some ga continued to grow into D$_3$.

Post-D$_3$  - Another idioblastic ga growth episode - small (0.5-1 mm) ga postdate the formation of D$_3$ crenulations. It is difficult to tell these from post-D$_2$ idioblastic ga. In sample M108, ga can be seen growing on the crests of D$_3$ crens (Figure 27C in text). One ky grain also postdates crens.
<table>
<thead>
<tr>
<th>DOM I</th>
<th>D1</th>
<th>D2</th>
<th>D3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chl</td>
<td><img src="image1.png" alt="Image" /></td>
<td><img src="image2.png" alt="Image" /></td>
<td><img src="image3.png" alt="Image" /></td>
</tr>
<tr>
<td>Bio</td>
<td><img src="image4.png" alt="Image" /></td>
<td><img src="image5.png" alt="Image" /></td>
<td><img src="image6.png" alt="Image" /></td>
</tr>
<tr>
<td>Ga</td>
<td><img src="image7.png" alt="Image" /></td>
<td><img src="image8.png" alt="Image" /></td>
<td><img src="image9.png" alt="Image" /></td>
</tr>
<tr>
<td>St</td>
<td><img src="image10.png" alt="Image" /></td>
<td><img src="image11.png" alt="Image" /></td>
<td><img src="image12.png" alt="Image" /></td>
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<tr>
<td>Ky</td>
<td><img src="image13.png" alt="Image" /></td>
<td><img src="image14.png" alt="Image" /></td>
<td><img src="image15.png" alt="Image" /></td>
</tr>
<tr>
<td>Mu</td>
<td><img src="image16.png" alt="Image" /></td>
<td><img src="image17.png" alt="Image" /></td>
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<tr>
<td>Qtz</td>
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<td><img src="image20.png" alt="Image" /></td>
<td><img src="image21.png" alt="Image" /></td>
</tr>
</tbody>
</table>

**CHART VII**

**LIST OF THIN SECTIONS**

- M96
- M99
- M106
- M108
- M114
- M278 (also M286)
- A345
- A347
- A349
- A346
- A344
- A333
- A331
- A351
CHART VIII

DOMAIN E
The domain lies mostly in the ky-qg-bio zone. Although the st-out isoedd is projected to cut the N.E. corner no staurolite was identified in thin section.

Grain Size:

Matrix
- Bio, 0.3-3 mm; av 1-1.5 mm
- Mu, 0.2-3 mm; av 1.5-2 mm
- Qtz, 0.2-4 mm; av 0.5-1 mm

Porphyroblasts - Ga, <.5-8 mm; av 1-2 mm (xenoblastic), .5 mm (idioblastic)
- Ky, 1-8 mm; av 1.5-3 mm

Microfabric Development:

D1 (S1) - In one thin section (P147) zoned qa includes very fine grained qtz trails in their core. Enveloping the core is a zone containing curved trails, the curvature of which is due to tightening of trails against a pre-existing qa before inclusion in the enveloping layers. The outer idioblastic inclusion free rim postdates S2. The middle growth ring probably represents a late syn-D2 or more likely a post-D2 growth where S2, deflected about a pre-existing qa was included as trails. If this is the case then the fine trails in the core may represent S1. In support of this argument post-D2 qa in the same and adjacent thin sections which show the relationship S1=S3 includes qtz grains in S1 which are the same size as qtz in S3.

Alternatively it may be suggested that the trails in the core of the qa represent an early stage in the development of S2. With increasing temperature and continued growth under conditions of irrotational strain the qa grew to include S2 which was being tightened by continued deformation against the developing garnet. (note the garnet under this system increases its size by including the matrix not by pushing it aside). Finally a post-D2 inclusion-free rim developed. The main reason for not favouring the above is that there is a distinct boundary between the core and enveloping rims which suggests a break in time during which qa did not grow.

D2 (S2) - Defined by bio, mu and qtz.
Syn-D2

- (See above discussion)

Post-D2

- Ga, ky and rare bio porphyroblasts have overgrown S2 and include it as straight trails. The idioblastic inclusion free ga are much smaller than xenoblastic ga and may represent as in other domains a later growth episode. In Domain H, late idioblastic ga has grown across the crest of D3 crens but this microfabric relationship was not observed in Domain E. Ga growth is therefore simply called post-D2 (except in the example described above).

D3 (S3)

- No observed development of an S3 foliation. D3 crenulations are open with a large wavelength (2 cms) but little amplitude (.5 cm).
<table>
<thead>
<tr>
<th>DOM E.</th>
<th>D1</th>
<th>D2</th>
<th>D3</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chl</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bio</td>
<td></td>
<td>→ d</td>
<td>= E = P107</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ga</td>
<td></td>
<td></td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ky.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sill</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mu</td>
<td></td>
<td>→ b</td>
<td>∨ f</td>
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**LIST OF THIN SECTIONS**

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<tr>
<th></th>
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<th>b</th>
<th>c</th>
<th>d</th>
<th>e</th>
<th>f</th>
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<td>*</td>
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<td>*</td>
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<td>*</td>
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</tr>
</tbody>
</table>
CHART IX

DOMAIN G
Domain G lies wholly within the ky-ga-bio zone.

Grain Size:

Matrix and porphyroblasts are listed together:

- **Bio**, 0.5-1.5 mm
- **Mu**, 1-4 mm; av 2 mm
- **Qtz**, 0.2-4 mm; av 2 mm
- **Plaq**, 0.5-3 mm; av 1-1.5 mm
- **Ga**, 1-8 mm; av 1.5-3 mm
- **Ky**, 0.5-4 mm; av 2 mm

Microfabric Development

<table>
<thead>
<tr>
<th>Event</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>D1 (S1)</strong></td>
<td>- No evidence for S1 in thin section.</td>
</tr>
<tr>
<td><strong>Pre-D2</strong></td>
<td>- ? early ga growth - see below.</td>
</tr>
<tr>
<td><strong>D2 (S2)</strong></td>
<td>- Defined by bio, mu and qtz. S2 reflected about pre-existing ga. This deflection can be distinguished from a D3 tightening of S3 because post-D2 idioblastic rims have included and preserved the deflected trails.</td>
</tr>
<tr>
<td><strong>Syn-D2</strong></td>
<td>- No evidence of syn-D2 porphyroblast growth.</td>
</tr>
<tr>
<td><strong>Post-D2</strong></td>
<td>- Major porphyroblast growth period. Ga and ky either truncate or include S2 as straight trails. Rarely ky includes small xenoblastic to idioblastic ga. In some cases the two are intergrown suggesting simultaneous growth as opposed to later ga growth as on sill porphs at higher grade. Where ga is completely surrounded by ky timing is not certain however.</td>
</tr>
<tr>
<td><strong>D3 (S3)</strong></td>
<td>- Bio flakes axial planar to D3 crenes define a non-penetrative D3. D3 either crenulates the pre-existing foliation or simply tightens it about pre-existing porphyroblasts.</td>
</tr>
<tr>
<td><strong>Post-D3</strong></td>
<td>- No evidence of post-D3 mineral growth, unless ga growth described above (D2) is post-D3.</td>
</tr>
<tr>
<td>DOM CAT</td>
<td>D1</td>
</tr>
<tr>
<td>---------</td>
<td>----</td>
</tr>
<tr>
<td>Chl</td>
<td></td>
</tr>
<tr>
<td>Bio</td>
<td>→ a</td>
</tr>
<tr>
<td>Ga</td>
<td></td>
</tr>
<tr>
<td>St</td>
<td></td>
</tr>
<tr>
<td>Ky</td>
<td></td>
</tr>
<tr>
<td>SILL</td>
<td></td>
</tr>
<tr>
<td>Mu</td>
<td>→ b</td>
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</table>

**LIST OF THIN SECTIONS:**

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<thead>
<tr>
<th></th>
<th>a</th>
<th>b</th>
<th>a</th>
<th>b</th>
</tr>
</thead>
<tbody>
<tr>
<td>M25-</td>
<td>• •</td>
<td>M54-</td>
<td>• •</td>
<td>M26-</td>
</tr>
<tr>
<td>M35-</td>
<td>• *</td>
<td>M91-</td>
<td>• *</td>
<td>M41-</td>
</tr>
<tr>
<td>M50-</td>
<td>•</td>
<td>M150-</td>
<td>•</td>
<td></td>
</tr>
</tbody>
</table>
CHART X

DOMAIN H
The sub-domain lies wholly within the ky-ga-bio zone (Figure 21).

**Grain Size:**

**Matrix**
- Bio, .3-1 mm;
- Mu, <.5-4 mm;
- Qtz, 1-4 mm; av 2.5 mm
- Plaq, .5-2 mm; av 1 mm

**Porphyroblasts**
- Bio; 1-4 mm; av 1.5-2 mm
- Ga, .3-4 mm; av 1.5-2.5 mm
- Ky, .5-4 mm; av 2 mm

**Microfabric Development**

**D₁ (S₁)**
- No evidence for S₁ in thin section.

**D₂ (S₂)**
- Defined by bio, mu, qtz and plaq.

**Syn-D₂**
- No evidence of any syn-D₂ porphyroblast growth.

**Post-D₂**
- Bio, ga and ky include S₂ as straight trails or truncate S₂ at their margins. A late? post-D₂ growth of ga is suggested by clusters of small, clean, inclusion-free, xenoblastic to sub-idioblastic ga (.5 mm). Clusters postdate S₂.

**D₃ (S₃)**
- No evidence of mineral growth in D₃. S₂ has been tightened about pre-existing ga in M30.
<table>
<thead>
<tr>
<th>DOM H</th>
<th>D1</th>
<th>D2</th>
<th>D3</th>
</tr>
</thead>
<tbody>
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<td>Chl</td>
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<td><img src="image1" alt="image" /></td>
<td><img src="image2" alt="image" /></td>
</tr>
<tr>
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<td><img src="image3" alt="image" /></td>
<td><img src="image4" alt="image" /></td>
<td><img src="image5" alt="image" /></td>
</tr>
<tr>
<td>Ga</td>
<td><img src="image6" alt="image" /></td>
<td><img src="image7" alt="image" /></td>
<td><img src="image8" alt="image" /></td>
</tr>
<tr>
<td>St</td>
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<td><img src="image10" alt="image" /></td>
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<tr>
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<td><img src="image12" alt="image" /></td>
<td><img src="image13" alt="image" /></td>
<td><img src="image14" alt="image" /></td>
</tr>
<tr>
<td>Sill</td>
<td><img src="image15" alt="image" /></td>
<td><img src="image16" alt="image" /></td>
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<td><img src="image22" alt="image" /></td>
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</tbody>
</table>

**LIST OF THIN SECTIONS**

- M1
- M4
- M5
- M8
- M9
- M10A
- M21
- M30
- M31
CHART XI.

DOMAIN D
The domain straddles the sill isograd. Domain D is structurally homogeneous and dominated on the mesoscopic and microscopic levels by D2 folds.

Grain Size:

The list below includes matrix and porphyroblast sizes for respective minerals.

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Size</th>
<th>Average</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bio</td>
<td>&lt;1-3 mm</td>
<td>1.5 mm</td>
</tr>
<tr>
<td>Mu</td>
<td>&lt;1-5 mm</td>
<td>2 mm</td>
</tr>
<tr>
<td>Qtz</td>
<td>2-5 mm</td>
<td>1.5 mm</td>
</tr>
<tr>
<td>Ga</td>
<td>&lt;1-7 mm</td>
<td>2 mm</td>
</tr>
<tr>
<td>Ky</td>
<td>1-7 mm</td>
<td>2 mm</td>
</tr>
<tr>
<td>Sill</td>
<td>1x2 mm - 1x5 mm range</td>
<td></td>
</tr>
</tbody>
</table>

**Microfabric Development**

- **D<sub>1</sub> (S<sub>1</sub>)** - No evidence of S<sub>1</sub> in thin section.
- **D<sub>2</sub> (S<sub>2</sub>)** - Dominant foliation; defined by Bio, Qtz and Mu.
- **Syn-D<sub>2</sub>** - No evidence of syn-D<sub>2</sub> porphyroblast growth.
- **Post-D<sub>2</sub>** - Bio, Ga, Ky include S<sub>2</sub> as straight trails.

Ga has many forms. Xenoblastic Ga with no inclusions, random inclusions or straight trails, all appear to postdate D<sub>2</sub> i.e. they truncate S<sub>2</sub> foliations. Xenoblastic Ga are preserved as cores to later idioblastic inclusion-free Ga growth. Growth rims probably represent the same Ga-forming reaction as small idioblastic inclusion free Ga in the matrix. Although late Ga are basic inclusion free, under high power a core containing fine randomly oriented dust-like particles bearing no relationship to S<sub>e</sub> can be observed.

Some Ky blades include Qtz as straight trails. S<sub>1</sub>=S<sub>e</sub>=S<sub>2</sub>.

Sill forms in pods in association with Musc/Qtz and Bio and has probably grown metasomatically. Pods crenulated by D<sub>3</sub> are unannealed; growth of Sill predates D<sub>3</sub>. 

...
**CHART XI**

<table>
<thead>
<tr>
<th>DOM D.</th>
<th>D1</th>
<th>D2</th>
<th>D3</th>
<th>R</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chl</td>
<td></td>
<td>a</td>
<td></td>
<td>R P65A P82A</td>
</tr>
<tr>
<td>Bio</td>
<td><img src="image" alt="Diagram" /></td>
<td><img src="image" alt="Diagram" /></td>
<td><img src="image" alt="Diagram" /></td>
<td>P50B</td>
</tr>
<tr>
<td>Ga</td>
<td><img src="image" alt="Diagram" /></td>
<td><img src="image" alt="Diagram" /></td>
<td><img src="image" alt="Diagram" /></td>
<td>P87C</td>
</tr>
<tr>
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**LIST OF THIN SECTIONS:**

| M256  | - | - | - | - | - | - |
| M255B | * | * | * | * | P57 | - |
| P19   | * | P60A | * | * | - | - |
| P24   | * | P60B | - | * | - | - |
| P29   | * | P63A | - | * | - | - |
| P34A  | * | P63B | - | * | - | - |
| P34B  | * | P63D | - | * | - | - |
| P50   | - | * | * | * | - | - |
| P51   | * | P65 | - | * | - | - |
| P54A  | - | - | - | - | - | - |

P57 - P79 | P80 - P84 | P87 - P97 | P95B - P97 | P69D -
Domain IV straddles the sillimanite isograd (as defined by the first appearance of sillimanite).

Grain Size:

Ground Mass - defined by bio, mu and qtz. Bio and Mu range in size from .2-1 mm while qtz varies from .1-.6 mm.

Porphyroblasts - Bio, 1 mm
Ga, 1-3 mm, av 1.5-2 mm
St, 1 mm    Ky, 1x2 2x5 mm
Sill, 1-3 mm  Mu, 1-1.5 mm

Microfabric Development

$D_1$ ($S_1$) - probably syn-$D_2$ ga's preserve crenuluated very fine-grained qtz trails where $S_1$=$S_e$ and $S_e$ is $S_2$. Ga's are partly resorbed and so positive identification of $S_1$ is difficult but qtz in the tightly crenulated $S_1$ is much finer grained than qtz in $S_2$ which is bent about larger open crenulations. The two crenulation phases are obviously different and qtz grain size indicates that the foliations are also of different ages. Since $S_e$ is a crenulated $S_2$ it seems reasonable to suggest that ga preserves a crenulated $S_1$. Ga grew syn-$D_2$ prior to the final transposition of $S_1$.

$D_2$ ($S_2$) - comprised of bio, mu and elongate qtz. $S_1$ transposed into $S_2$ leaving mica rich and qtz rich microlithons.

Syn-$D_2$ - ga grew syn-$D_2$ to preserve a crenulated $S_1$.
Post-$D_2$ - Ga, st, ky and sill postdate the formation of $S_2$. Bio porphra, so prevalent at lower grade have disappeared to be replaced by clusters of smaller flakes which generally have no inclusions. Where $D_2$ activity is not strong $S_1$=$S_e$. Elsewhere post-$D_2$ porphyroblasts deflect $S_3$.

$D_3$ ($S_3$) - defined by bio and mu. Where $S_3$ is a spaced cleavage mu has been rotated to lie in $S_3$.
Post-$D_3$ - An apparent late growth of kyanite can be observed immediately below the sillimanite isograd. An unstrained ky blade clearly overgrew and preserved a crenulate schistosity. Elsewhere kyanite porphyroblasts postdate $S_2$ and those elongate in $S_2$ have been strained during $D_3$.

$D_4$ - A second phase of crenulation, at a higher angle to $D_3$ thought to represent local $D_4$. Away from the areas of high grade metamorphism where heat during metamorphism must have been less and abated more quickly, $D_4$ is manifested by kink bands.
<table>
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<tr>
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**LIST OF TIN SECTIONS**

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CHART XIII

DOMAIN 5
Although the Domain straddles the sillimanite isograd, all samples were collected from the sill-ga-bio zone.

**Grain Size:**

*Matrix* - composed of qtz, bio and mu. qtz, .1-1.5 mm, av .5 mm; bio, .2-1 mm, av .5 mm; and mu, .3-1.5 mm, average size slightly larger than bio.

*Porphyroblasts* - Bio, 1-2 mm
  - Ga, 1-8 mm, av 3-6 mm
  - St, 1-6 mm, av 3-4 mm
  - Ky, 1 mm
  - Sill, 1 mm-1 cm, av 5-6 mm

**Microfabric Development**

*D₁ (S₁)* - no remaining evidence outside of ga porphyroblasts (see below).

*D₂ (S₂)* - defined by mu, bio and qtz

*Syn-D₂* - Ga preserves fine crenulate S₁ trails. It is suggested that trails are S₁ for the same reasons outlined in Domain III.

*Post-D₂* - Clusters of bio and ga, st and sill porphyroblasts postdate the formation of S₂. Rare xenoblastic kyanite persists above the sillimanite isograd. Both xenoblastic and idioblastic ga postdate S₂. The formation of the idioblastic ga's in the groundmass may be related to and contemporaneous with growth of idioblastic ga's on pods. These ga's obviously postdate the formation of sill. Their is no evidence to suggest a D₃ or post-D₃ growth period and so time of growth is thought to be late post-D₂. North of the Selkirk Fm Axis there is some evidence to suggest a post-D₃ growth of idioblastic garnets.

*D₃ (S₃)* - S₃ is not a strong penetrative cld at thin section scale probably because of the coarseness of the grain size in thin section. Elongate cross-biotite, axial planar to D₃ crenulations defines S₃.

*Post-D₃* - No evidence of post-D₃ mineral growth.
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**LIST OF THIN SECTIONS**

- A181A -
- F68A -
- F68B -
- F68D -
- F50A -
- F50B -
- F50D -
- F50G -
- F53B -
- A218 -
- F55B -
- A192 -
- A193 -
- A194 -
- F56B -
- F66A -
- A191A -
- A191B -
- A190 -
- F84 -
CHART XIV

DOMAIN A
Domain A lies wholly within the sill-ga-bio zone of metamorphism immediately to the north of the Selkirk Fan Axis.

Grain Size:

Rocks in Domain A are coarse grained and there is little difference between matrix minerals and porphyroblasts.

Matrix - formed by mu, bio and qtz; av grain size 2 mm with individual flakes to 4 mm.

Porphyroblasts - Bio, 3 mm; Ga, 1-8 mm, av 4 mm; Sill, 4-5 mm patches and pods

Macrofabric Development

$D_1$ ($S_1$) - no evidence of $S_1$ in thin section

$D_2$ ($S_2$) - formed by bio, musc and elongate qtz

Syn-$D_2$ - One thin section contains evidence for a syn-$D_2$ ga growth (Note a pre-$D_2$ age of growth cannot be ruled out). Sievy xenoblastic garnets deflect $S_2$ which is in turn overgrown by post-$D_2$ bio. (Section M225).

Post-$D_2$ - Bio, rare mu, ga and sill porphyroblasts postdate the formation of $S_2$. Post-$D_2$ ga's occur in several forms. Idioblastic ga's have grown on sill patches and postdate the formation of sill. Idioblastic ga's in the matrix may represent the same generation of ga. The relationship of the inclusion free xenoblastic ga's to idioblastic ga is not known.

Sill appears to be mimetic on biot and/or musc/Qtz patches. The timing of the growth of sillimanite is thus not exactly known. Since there is no evidence for a $D_3$ age of the ga porphyroblasts which grow on sill patches an early post-$D_2$ age for the growth of sill is suggested. In the field sill pods are flattened in the $D_2$ schistosity and syn-$D_2$ growth cannot be positively ruled out.

$D_3$ ($S_3$) - A weak $S_3$ is defined by cross-biotite.

Syn-$D_3$ - $D_3$ crenulates $S_2$ about open crenulations with a wave length of 2-3 cm and an amplitude of 1 cm. Sill pods are themselves bent.

Post-$D_3$ - no evidence of post-$D_3$ mineral growth.
### Chart XIV

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**List of Thin Sections:**

- M215- ⬤ ⬤ ⬤ ⬤
- M216- ⬤ ⬤
- M217- ⬤ ⬤ ⬤ ⬤
- M221- ⬤ ⬤ ⬤ ⬤
- M225- ⬤ ⬤
- M233- ⬤ ⬤ ⬤
- M236- ⬤ ⬤ ⬤ ⬤
CHART XV

DOMAIN B
Domain B lies wholly within the Sill-Ga-Bio zone.

Grain Size:

There is little difference in size between bio and mu in the matrix and porphyroblast growth of the same minerals. Matrix and porphyroblast minerals are therefore listed together.

Bio, 2-3 mm
Mu, up to 4 mm
Ga, 2-5 mm, av 2 mm
Sill, 2-4 mm pods

Microfabric Development

\[D_1 \text{ (S1)}\] - Two thin sections preserve a weak S1, poorly defined by mu flakes.

\[D_2 \text{ (S2)}\] - Defined by bio, mu and qtz. S2 is generally undisturbed by later tectonism.

\[\text{Syn-D}_2\] - No evidence of Syn-D2 porphyroblast growth except perhaps for sill (see below).

\[\text{Post-D}_2\] - Bio, mu, ga and sill postdate S2. Ga and bio include qtz in S2 as straight trails. Sill appears to be nematic on biot and musc/qtz concentrations. Remnant micas can be seen within sill masses. The fibrous nature of sill makes it difficult to pinpoint the time of growth. Idioblastic ga's have grown on sill pods and postdate the formation of sill. Growth of these ga's was probably simultaneous with the appearance of small inclusion free idioblastic ga's in the ground mass which postdate S2. There is no evidence of D3 growth in this Domain. Timing of porphyroblast growth must therefore be stated simply as post-D2, with the possibility of some syn-D2 sill.

\[D_3\] - Manifested as a tightening of the groundmass foliation about post-D2 porphyroblasts.
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**List of thin sections:**
- P4A
- P4B
- P6A
- P6B
- P7
- P9C
- P15
- P16
CHART XVI

DOMAIN C
The Domain lies wholly within the Sill-Ga-Bio zone of metamorphism.

**Grain Size:**

The matrix is comprised of coarse grained bio, mu and qtz. Foliation forming matrix minerals are often larger than "porphyroblast" minerals.

Bio, 1-4 mm, 1 mm most common
Mu, 1.5-8 mm, av 3 mm
Ga, 5-3 mm, av 1.5-2 mm
Sill, 3 mm - 4.5 cm, 4-5 mm most common
Qtz, 2-1.5 mm

**Microfabric Development**

| S2 (S1) | - no evidence in thin section
| S2 (S2) | - S2 is not as well defined (in thin section) as previously due to the very coarse grained nature of the micas and a slight randomness of the fabric which was probably the result of thermal activity outlasting the deformation. Under a planar lens however, the strong penetrative nature of the cleavage, obvious in outcrop, is apparent.
| Syn-D2 | - no evidence of syn-D2 porphyroblast growth.
| Post-D2 | - Ga and sill growth postdates S2. Both xenoblastic and idioblastic ga's include S2 as straight trails. Two growth episode of garnet are apparent, both of which postdate D2. The early stage is represented by xenoblastic ga's with no inclusions, random inclusion or straight inclusions. These early ga's are themselves included in a few instances by idioblastic inclusion-freq rims. The development of these rims probably represents the same growth phase as the discrete idioblastic ga in the matrix.
| Sill appears to have grown mimetically on mu/qtz and more rarely bio. Pods are elongate in S2. Unanealed pods are folded about D3 crenulations. Growth of sill is therefore pre-D3 and probably post-D2.
| D3 (S3) | - only observed in one thin section - defined by mu blades. Bio, mu and sill folded about D3 crenulations.
| Post-D3 | - in one thin-section large mu blades have grown post-D3 to include crenulated qtz trails.
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**LIST OF THIN SECTIONS:**

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| M154C- | * * * *
| M158A- | * * * |
| M158B- | * * |
| M161   |     |
| M164C- | * * * |
| M165   | * * |
| M170   | * * * |
| M186   | * * * |
| M221   | * * |
| M245A- | * * |
| M246   | * * * |
| M247   | * * * * |
| M261   | * * * * |
| M268   | * * * * |
CHART XVII

DOMAIN F
Domain F lies wholly within the sill-ga-bio zone of metamorphism. A large amphibolite causes a local reorientation of folding as axial planes wrap around the amphibolite mass.

Grain Size:

| Matrix        |     
|---------------|-----|
| bio, <1 mm    |     |
| mu, <1 mm     |     |
| qtz, <.5-1 mm |     |

| Porphyroblasts|     
|---------------|-----|
| Bio, 1-4 mm;  |     |
| av 2-3 mm     |     |
| Mu, 1-3 mm;   |     |
| av 2 mm       |     |
| Ga, 1-5 mm;   |     |
| av 2-3 mm     |     |
| Sill, 1-4 mm; |     |
| av 2-3 mm     |     |

Microfabric Development

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<tr>
<th>D₁ (S₁)</th>
<th>no evidence of S₁ in thin section</th>
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<td>D₂ (S₂)</td>
<td>defined by mu, bio and qtz</td>
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<tr>
<td>Syn-D₂</td>
<td>Two samples indicate the possibility of syn-D₂ growth of ga i.e. rotational growth (snowball). These resorbed xenoblastic ga are enclosed by later idioblastic growth. The relationships between S₁ and S₂ are not known nor is the phase of rotation of the ga. Similarity of grain size between qtz in S₁ and S₂ suggests that growth may have been syn-D₂.</td>
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</tr>
<tr>
<td>Post-D₂</td>
<td>Bio, mu, ga and sill have grown post-D₂ to truncate or include S₂ as straight trails. Both xenoblastic and idioblastic ga exhibit straight qtz trails. Early small idioblastic ga random inclusions are enclosed by idioblastic rims. Several ga forming reactions must have occurred. The idioblastic rims may represent the same phase of growth as idioblastic ga on sill porphyroblasts.</td>
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<tr>
<td>D₃ (S₃)</td>
<td>bio flakes axial planar to D₃ crenulations define S₃. D₃ crenulates S₂ foliation including sill pods.</td>
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<tr>
<td>Syn-D₃ to</td>
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Date
FIGURE 2: TOPOGRAPHIC, GEOGRAPHIC AND STATION LOCATION MAP, BIG BEND, COLUMBIA RIVER

LEGEND

STATION LOCATIONS: • M Perkins 1975-76
                 • P Brown, Perkins and Shaw 1974
FIGURE 2-1 TOPOGRAPHIC, GEOGRAPHIC AND STATION LOCATION MAP, BIG BEND, COLUMBIA RIVER

LEGEND

STATION LOCATIONS:

- M Perkins 1975-76
- P Brown, Perkins and Shaw 1974
- V Van der Leeden 1974
- F Franzen 1973
- A Leatherbarrow 1977

(Date refers to year of mapping)

Ice or permanent snow

Contour lines, contour interval 100'

Scale 1:50,000

miles
kilometres

4 OF/DE
FIGURE 3: DETAILED LITHOLOGIC MAP OF THE NORTHERN

STRUCTURE SYMBOLS

--- axial surface trace of second phase folds
--- axial surface trace of third phase folds
--- normal fault

schist; 3d rusty pelitic schist; bedded pelite semipelite an
LITHOLOGIC MAP OF THE NORTHERN BIG BEND

LEGEND

STRUCTURE SYMBOLS
- axial surface trace of second phase folds
- axial surface trace of third phase folds
- normal fault

LITHOLOGIC UNITS (in text)

STRATIGRAPHIC CORRELATION
CAMBRIAN AND (?) LATE ARCHEAN

LARDEAU GROUP: 7.gp=psammite
7.hthinly interbedded pelite and psammite
7.c=pelitic schist; 7.d=grey marble and psammite; 7.f=calc-silicate

snowfield
Northern Big Bend

Legend

Symbols

Lithologic Units (in text)

Stratigraphic Correlation

Other Symbols

Cambrian and (?). Later

Larreau Group: 7psammitite and quartzite,
7b thinly interbedded pelite and carbonate;
7c pelitic schist; 7d grey marble; 7e pelite
and psammite; 7f calc-silicate
CAMBRIAN AND (?) LATER

GROUSE MOUNTAIN SULFUR DISTRICT

LARDEAU GROUP: 7a psammite and quartzite; 7b thinly interbedded pelite and carbonate; 7c pelitic schist; 7d grey marble; 7e pelite and psammite; 7f calc-silicate

BADSHOT FORMATION: 6a grey marble; 6b rusty marble; 6c pelitic schist; 6d psammite; 6e calc-silicate.

HAMILL GROUP: 5a quartzite and psammite; 5b calcareous quartzite, calc-silicates and minor marble with mafic pods.

HORSETHIEF CREEK GROUP

PELITIC MEMBER: 4a dark pelitic schist; 4b rusty pelitic schist; 4c thinly interbedded pelite and carbonate; 4d impure marble; 4e grey marble; 4f psammite and quartzite; 4g amphibolite; 4h semipelite.

MIDDLE MARBLE MEMBER: 3a grey marble; 3b impure and rusty marbles; 3c pelitic schist; 3d rusty pelitic schist; 3e thinly interbedded pelite, semipelite and marble-transition series; 3f mafic rocks; 3g quartzite.

LOWER PELITIC MEMBER: 2a thinly interbedded pelite and semipelite with concordant amphibolite bands; 2b amphibolite; 2c pelitic schist; 2d semipelite; 2e marble; 2f quartzite; 2g psammite

GRIT MEMBER: 1a psammite; 1b pelite; 1c interbedded pelite, psammite and semipelite; 1d interbedded psammite and calc-silicate; 1e rusty
LOWER PELITIC MEMBER: 2a thinly interbedded pelite and semipelite with concordant amphibolite bands; 2b amphibolite; 2c pelite schist; 2d semipelite; 2e marble; 2f quartzite
2g psammite

GRIT MEMBER: 1a psammite; 1b pelite; 1c interbedded pelite, psammite and semipelite; 1d interbedded psammite and calc-silicate; 1e rusty marble and/or grey marble; 1f calc-silicate; 1g feldspathic grits; 1h conglomerate; 1i semipelite; 1j quartzite

Scale 1:50,000
(middle marble member)

- psammite
- quartzite
- feldspatic grits
- conglomerate

↑ graded beds indicating sedimentary tops
SEMIPELITE-AMPHIBOLITE UNIT
(lower pelitic member)

For location of columns refer to cross-sections in Figure 6.

FIGURE 5: DETAILED STRATIGRAPHIC COLUMNS, NORTHERN BIG
quartzite

feldspathic grits

conglomerate

graded beds indicating sedimentary tops

garnet-bearing amphibolite

base or top of unit not exposed

top in Figure 6

after Lane (1977)

PELITE UNIT

(upper pelite member)
PELITE UNIT

(upper pelite member)

BADSHOT FORMATION?
SECTION 1

△ : Stratigraphic Top Determination (from graded bedding)

N : Nepheline Syenite

Granitic Stocks

HORIZ. and VERT. SCALE 1:50,000

Km 10000

Miles 1 2 3 4 5 6 7 8 9 10
LEGEND

STRUCTURE SYMBOLS

--- Axial plane of first phase-fold

--- Axial plane of second phase antiform (anticline)

--- Axial plane of second phase synform (syncline)

--- Axial plane of third phase antiform (anticline)

--- Axial plane of third phase synform (syncline)

---------- Fault with normal sense of movement

---------- Fault with reverse sense of movement
Fault with normal sense of movement
Fault with reverse sense of movement

STRATIGRAPHY

LARDEAU GROUP: 7a psammite and pelite, 7b thinly interbedded
7c pelitic schist, 7d grey marble, 7e pelite
7f calc-silicate

BADSHOT FORMATION: 6a grey marble, 6b rusty marble, 6c pelitic
6e calc-silicate.
Fault with normal sense of movement

Fault with reverse sense of movement

STRATIGRAPHY

LARDEAU GROUP: 7a psammite and pelite, 7b thinly interbedded pelite and carbonate,
  7c pelitic schist, 7d grey marble, 7e pelite and psammite,
  7f calc-silicate

BAGSHOT FORMATION: 6a grey marble, 6b rusty marble, 6c pelitic schist, 6d psammite,
  6e calc-silicate.
BADSHOT FORMATION: 6a grey marble, 6b rusty marble, 6c calc-silicate

HAMILL GROUP: 5a quartzite and psammite, 5b calc-silicate and minor marble with mafic pods

UPPER PELITIC MEMBER: 4a dark pelitic schist, 4b rusty pelite and carbonate, 4d impure and 4f psammite and quartzite, 4g amphibolite and schist

MIDDLE MARBLE MEMBER: 3a grey marble, 3b rusty marbles, 3c thinly interbedded pelite, 3e calc-silicate, 3g quartzite

LOWER PELITIC MEMBER: 2a thinly interbedded pelite and schist, 2b amphibolite, 2c pelitic gneiss, 2f quartzite, 2g psammite
BADSHOT FORMATION: 6a grey marble, 6b rusty marble, 6c pelitic schist, 6d psammite, 6e calc-silicate.

HAMILL GROUP: 5a quartzite and psammite, 5b calcareous quartzite, calc-silicates and minor marble with mafic pods.

UPPER PELITIC MEMBER: 4a dark pelitic schist, 4b rusty pelitic schist, 4c thinly interbedded pelite and carbonate, 4d impure and rusty marble, 4e grey marble, 4f psammite and quartzite, 4g amphibolite, 4h semipelite.

MIDDLE MARBLE MEMBER: 3a grey marble, 3b rusty marbles, 3c pelitic schist, 3d rusty pelitic schist, 3e thinly interbedded pelite, semipelite and marble, 3f calc-silicate, 3g quartzite.

LOWER PELITIC MEMBER: 2a thinly interbedded pelite and semipelite with concordant amphibolite bands, 2b amphibolite, 2c pelitic schist, 2d semipelte, 2e marble.
MIDDLE MARBLE MEMBER: 3a grey marble, 3b rusty marbles, 3c pelitic schist, 3d rusty pelitic schist, 3e thinly interbedded pelite, semipelite and marble, 3f calc-silicate, 3g quartzite.

LOWER PELITIC MEMBER: 2a thinly interbedded pelite and semipelite with concordant amphibolite bands, 2b amphibolite, 2c pelitic schist, 2d semipelite, 2e marble, 2f quartzite, 2g psammite.

GRIT MEMBER: 1a psammite, 1b pelite, 1c interbedded pelite; psammite and semipelite, 1d interbedded psammite and calc-silicate, 1e rusty and/or grey marble, 1f calc-silicate, 1g feldspathic grits, 1h conglomerate, 1i semipelite, 1j quartzite.
2f quartzite, 2g psammite.

GRIT MEMBER: 1a psammite, 1b pelite, 1c interbedded pelite; psammite and semipelite.
1d interbedded psammite and calc-silicate, rusty and/or grey marble.
1f calc-silicate, feldspathic grits, 1h conglomerate, 1i semipelite.
1j quartzite.

TURAL CROSS SECTIONS, NORTHERN BIG BEND

15 OF/DE 15
N. DOGTOOTH MOUNTAINS
ROGERS PASS
21Km. N. OF ROGERS PASS
SONATA MOUNTAIN
NORTHERN BIG

SIMONY AND WIND (1970)
Poulton AND SIMONY (1980)
Poulton AND SIMONY (1980)
Brown, Tippett AND LANE (1978)

BASE
UPPER SLATE UNIT
CARBONATE UNIT
MIDDLE SLATE UNIT
FELDSPATIC GRIT UNIT

OF
UPPER CLASTIC DIVISION
SLATE DIVISION
CARBONATE DIVISION
GRIT DIVISION

HAMILL
UPPER PELITIC MEMBER
MIDDLE MARBLE MEMBER

GROUP
STRUCTURE SYMBOLS

Axial Surface Trace of Second Phase Antiform / Synform
Axial Surface Trace of Third Phase Antiform / Synform
Normal Fault, mapped, assumed
Reverse Fault, mapped, assumed
Stratigraphic Contact, mapped, assumed
Bedding (So)
Second Phase Cleavage (S2), direction and amount of plunge shown by arrow
Second Phase Folds, symbol indicating sense of vergence viewed downplunge
Third Phase Cleavage (S3), direction and amount of plunge shown by arrow
Third Phase Folds, symbol indicating sense of vergence viewed downplunge
Fourth Phase Cleavage/Folds symbols as above
poles to bedding, cleavage, axial surface etc
Inclination direction - hinge lines, bedding/cleavage intersections

Domain Boundaries
Sedimentary tops
Nepheline syenite
Permanent ice or snowfield

Scale 1:50,000

Miles
Kilometres
STRUCTURE SYMBOLS

- Second Phase Antiform
- Synform
- First Phase Antiform
- Synform
- Assumed
- Assumed
- Mapped
- Assumed
- Bedding [So]
- S2, direction and amount of plunge shown by arrow
- Symbol indicating sense of vergence viewed downplunge
- S3, direction and amount of plunge shown by arrow
- Symbol indicating sense of vergence viewed downplunge
- Folds: symbols as above
- Ge, axial surface etc
- Lines, bedding/cleavage intersections
- Sedimentary tops
- Nepheline syenite
- Permanent ice or snowfield

Scale 1:50,000
13: STRUCTURAL GEOLOGY OF THE NORTHERN BIG BEND

14

BIGMOUTH CREEK
STOCK

BIGMOUTH CREEK

ARGONAUT

FRENCH GLACIER

DOMAIN B
FIGURE 23 SAMPLE LOCATION MAP: SOUTH OF THE SELKIRK FAN AXIS
FIGURE 28: DISTRIBUTION OF FABRIC ELEMENTS FROM GOLDSTREAM RIVER TO COLUMBIA REACH.
3 OF/DE

ZONE 3
### ZONE 2

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FOR LIST OF SYMBOLS AND
THIN SECTION NUMBERS
REFER TO APPENDIX I

ZONE BOUNDARIES
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10 OF/DE