STRUCTURE, METAMORPHISM, AND GEOCHRONOLOGY OF
THE NORTHERN WOLVERINE COMPLEX NEAR CHASE MOUNTAIN,
AIKEN LAKE MAP-AREA, BRITISH COLUMBIA

by

RANDALL RICHARDSON PARRISH
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to the required standard

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Department of Geological Sciences

The University of British Columbia
2075 Wesbrook Place
Vancouver, Canada
V6T 1W5

Date January 1, 1977
ABSTRACT

The Wolverine Complex (Armstrong, 1949; Roots, 1954) and similar rocks to the northwest are of Precambrian age, correlate with Windermere-type stratigraphy, and are poly-metamorphic and poly-deformational. Wolverine rocks near Chase Mountain have experienced two periods of tight to isoclinal folding (F1, F2) overprinted by one or more periods of northwest-trending large-scale open folding (F3), small-scale crenulation folding of various orientations, and minor faulting. The two earlier periods of folding were accompanied by metamorphism culminating in amphibolite facies at the close of F2. These early folds are recumbent to gently inclined, and the geometry of F2 is consistent with north-eastward transport of rocks in nappe-like fashion. An F3 large-scale fold deformed earlier foliations into an upright to steeply eastward inclined antiform which correlates with structures mapped to the north by Mansy (1972, 1974).

Geochronometric data strongly suggest that metamorphic culmination occurred in mid-Cretaceous or earlier time, and that many rocks in widespread areas south of 56°N have experienced resetting of K-Ar and Rb-Sr dates during the Eocene. A stock of biotite quartz monzonite, termed the Blackpine Lake granitic stock, has a Rb-Sr whole rock isochron age of 62±7 m.y. @ 0.7052±.0002 Sr 87/86 and a mineral isochron age of 44.7±2 m.y., and it intrudes the Wolverine Complex. In surrounding schists, gneisses, pegmatites, and muscovite-bearing granitic rocks related to the metamorphism, Rb-Sr mineral dates (muscovite, plagioclase, K-feldspar, whole rock) range from 52-84 m.y. and reflect partial to complete resetting, whereas K-Ar dates are entirely reset to 43-47 m.y. Rb-Sr mineral dates on
biotite from metamorphic rocks are anomalously younger than K-Ar dates, a problem which is not understood.

Though Eocene volcanic rocks and sediments indicative of rapid uplift occur within or flanking the Omineca Crystalline Belt, their spatial distribution bears no relationship to the areas of resetting of K-Ar and Rb-Sr dates. Though an entirely satisfactory explanation remains elusive, the resetting of dates must in part be due to the thermal effect of intrusions of granitic rock similar to the Blackpine Lake granitic stock which is shown to have disturbed dates in surrounding metamorphic rocks.
# Table of Contents

## PART I. Structure and Metamorphism

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Introduction</td>
<td>1</td>
</tr>
<tr>
<td>Previous Work</td>
<td>4</td>
</tr>
<tr>
<td>Geologic Setting</td>
<td>5</td>
</tr>
<tr>
<td>Stratigraphy</td>
<td>7</td>
</tr>
<tr>
<td>Structural Geology</td>
<td>13</td>
</tr>
<tr>
<td>Chase Mountain Area</td>
<td>13</td>
</tr>
<tr>
<td>Phase 1</td>
<td>13</td>
</tr>
<tr>
<td>Phase 2</td>
<td>16</td>
</tr>
<tr>
<td>Phase 3</td>
<td>22</td>
</tr>
<tr>
<td>Other Minor Structures</td>
<td>25</td>
</tr>
<tr>
<td>Stereographic Analysis</td>
<td>25</td>
</tr>
<tr>
<td>Domains 1, 2</td>
<td>29</td>
</tr>
<tr>
<td>Domains 3, 4, 5</td>
<td>32</td>
</tr>
<tr>
<td>Domain 6</td>
<td>34</td>
</tr>
<tr>
<td>Summary</td>
<td>34</td>
</tr>
<tr>
<td>Blackpine Lake Area</td>
<td>35</td>
</tr>
<tr>
<td>Phases 1, 2</td>
<td>35</td>
</tr>
<tr>
<td>Structural Relations of the Blackpine Lake Stock</td>
<td>38</td>
</tr>
<tr>
<td>Relations between Crystal Growth and Deformation</td>
<td>38</td>
</tr>
</tbody>
</table>

## Metamorphic Petrology

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Introduction</td>
<td>43</td>
</tr>
<tr>
<td>Chase Mountain Area</td>
<td>43</td>
</tr>
<tr>
<td>Blackpine Lake area</td>
<td>51</td>
</tr>
<tr>
<td>Summary of P-T-fluid conditions</td>
<td>58</td>
</tr>
<tr>
<td>Granitic Rocks</td>
<td>58</td>
</tr>
</tbody>
</table>

## PART II. Geochronology and the Eocene Resetting Event

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Introduction</td>
<td>61</td>
</tr>
<tr>
<td>Geologic Setting of the Wolverine Complex</td>
<td>63</td>
</tr>
<tr>
<td>Geochronology of the Wolverine Complex and Related Rocks</td>
<td>69</td>
</tr>
<tr>
<td>Review of Previous Work</td>
<td>69</td>
</tr>
<tr>
<td>Rb-Sr and K-Ar Data of this Study</td>
<td>69</td>
</tr>
<tr>
<td>Discussion and Problems of Interpretation</td>
<td>77</td>
</tr>
<tr>
<td>Early Tertiary Tectonic Setting</td>
<td>81</td>
</tr>
<tr>
<td>Summary</td>
<td>85</td>
</tr>
<tr>
<td>Bibliography</td>
<td>86</td>
</tr>
</tbody>
</table>
LIST OF FIGURES

2. Structural map of thesis area showing orientations of compositional layering.
4. Structural map of thesis area showing phase 1 data.
5. Photographs of phase 1 folds and related structural features.
6. Structural map of thesis area showing phase 2 data.
7. Photograph of phase 2 folds showing crenulated schistosity.
8. Photograph of phase 2 fold refolding phase 1 fold.
9. Photograph of phase 2 folds depicting general form.
10. Sketches of east-verging phase 2 folds.
11. Sketch of structural cross-section east of Chase Mountain.
12. Photograph of dislocation zone.
13. Schematic and somewhat conjectural tectonic profile near Chase Mountain.
14. Sketch of phase 3b folds.
15. Cross-sections of Mansy across Wrede and Russell Ranges.
16. Sketches of later, brittle minor structures.
17. Equal-area stereonet diagram of late minor structures.
18. Map of the Chase Mountain area showing domain boundaries.
19. Equal-area stereonets of structural features, Domain 1-5.
21. Structural map of S0, S1, and S2 orientations across fault east of Chase Mountain.
22. Photograph of refolded lineations.
23. Photograph of phase 1 folds, Blackpine Lake area.
24. Photograph and sketch of folds and related pegmatites, Blackpine Lake area.
25. Equal-area stereonet of structural data, Blackpine Lake area.
27. Photomicrographs of metamorphic textures in schists.
28. Photograph and photomicrograph of muscovite pseudomorphs of kyanite.
29. Diagram depicting interrelations of deformation, metamorphism, and intrusion.
30. Map of thesis area showing distribution of samples referred to in text.
31. Photomicrograph of muscovite pseudomorphs after kyanite.
32. PH2O-T diagram of general metamorphic conditions.
33. T-X diagrams of P_T=6 kbars for the system SiO2 - CaO - MgO - K2O - Al2O3 - CO2 - H2O
34. T-X diagram of P_T=6 kbars for samples 17, 18, and 271-2.
35. T-X diagram of P_T=6 kbars for sample 87-2.
36. T-X diagram of P_T=6 kbars for Fe-free, Na-free system.
37. T-X diagram of P_T=6 kbars for system containing biotite, plagioclase, and actinolitic amphibole.
38. T-X diagram of P_T=6 kbars for Blackpine Lake area.
39. Photomicrographs of Rock 304B showing relations of garnet, staurolite, sillimanite, and muscovite.
40. PH2O - T diagram of conditions of Sample 304b near Blackpine Lake.
41. Map of Wolverine Complex and related metamorphic rocks showing all available geocronologic data.
38. Rb-Sr evolution diagram for Schist of sample 134.
39. Rb-Sr evolution diagram for "Wolverine" Granitic rocks.
40. Rb-Sr evolution diagram for Blackpine Lake Granitic Stock.
41. Rb-Sr diagram showing all whole rock determinations.
42. Sketch map showing the tectonic setting during Eocene time in North-central British Columbia.
LIST OF TABLES

I. Representative assemblages from the Chase Mountain Area 46
IIa. K-Ar data from the Geological Survey of Canada. 70
IIb. K-Ar analytical data from this study. 71
III. Rb-Sr analytical data from this study. 74
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PART I. STRUCTURE, STRATIGRAPHY, AND METAMORPHISM OF THE CHASE MOUNTAIN-
BLACKPINE LAKE AREA

INTRODUCTION

The thesis area is situated at the northwestern end of the Wolverine Complex as defined by Armstrong (1949) and Roots (1954), as shown in figure 1. Rocks include metamorphosed clastic sediments, now mica schist, micaceous and feldspathic quartzite, biotite-muscovite gneiss, and rare, discontinuous layers of calc-silicate marble, amphibolite, tourmaline schist, and quartz-pebble conglomerate. Metamorphic grade is amphibolite facies, in some areas lower sillimanite grade. The rocks have been involved in polyphase deformation. Broad open folding and later faulting are superimposed on earlier syn-metamorphic near-isoclinal folding and flattening. The geochronologic evolution is complex and indicates a long and complicated thermal history.

This study was undertaken because it was felt that the structural, metamorphic, and geochronologic evolution of the Wolverine Complex was poorly known and warranted detailed study. The specific area near Chase Mountain was chosen because it is located at the northwestern end of the Complex where the dominant structural anticlinorium plunges northwesterly beneath lower grade and structurally higher rocks. It was hoped that metamorphic isograds and a change in structural style might be documented. The Blackpine Lake granitic stock was examined for deformational and geochronologic history in hopes that the absolute ages of structural events might be determined. The area was also selected because of the proximity of Eocene K-Ar dates, so typical of the Wolverine and Shuswap Complexes, and nearby Early Cretaceous dates on similar 'Wolverine' rock
types (Wanless et al, 1971). It was felt that a combined K-Ar and Rb-Sr approach to the 'Eocene resetting' problem on both metamorphic and igneous rocks might shed some insight into the thermal history.

The significance of the Wolverine Complex to the evolution of this part of the British Columbia cordillera is poorly known, and its relation to plate tectonics even more obscure; a regional synthesis is presented to provide a useful reference frame in which to view the accumulated data pertinent to these problems.
Figure 1. Geologic map of Wolverine Complex and adjacent areas of north-central British Columbia.
PREVIOUS WORK

Armstrong (1949) defined the Wolverine Complex as the metamorphic and igneous rocks constituting the Wolverine Range of the Fort St. James map-area. He also mapped portions of the Aiken Lake map-area with Roots (Armstrong and Roots, 1948) extending the northwestern end of the Complex. Roots (1954) defined and examined the uppermost Proterozoic and lowest Cambrian Tenakihi and Ingenika groups in the northeastern half of the Aiken Lake map-area in greater detail, briefly subdividing the stratigraphy and crudely documenting metamorphic zonations and major structures. Roots also mapped the remainder of the Aiken Lake sheet on a scale of 1" = 4 miles, in which mainly Paleozoic and Mesozoic rocks are found.

More recent work in the Aiken Lake map-area has been done by the Geological Survey of Canada. Mansy (1971, 1972, 1974, 1976) has mapped the northern half in greater detail and has delineated detailed stratigraphy and structure. Paleozoic and Mesozoic sedimentary and igneous rocks to the west have been recently studied by Irvine (1975, 1976), Monger (1974, 1976), Woodsworth (1976), Garnett (1974), Richards (1976), and earlier by Lord (1948). Wanless et al (1967, 1971, 1973, 1974) in association with workers from the Geological Survey of Canada have been the principle creators of K-Ar data on plutonic rocks of the Hogem composite batholith. Eibach (1974a) has studied the Sustut and Sifton Basins, outliers of which occur in the Aiken Lake map-area. A regional tectonic synthesis has been prepared by Gabrielse (1967) for the northern Canadian Cordillera.
The area studied (figure 1) lies within the extensive uppermost Pre-
cambrian metamorphosed clastic sequence broadly correlative with the Kaza-
Miette (Winderemere) assemblage of rocks farther south in the McBride map-
area (Campbell et al, 1973; Gabrielse, 1972) and with the Missinchinka
Group to the northeast across the Rocky Mountain Trench (Irish, 1970). This
group of rocks forms the backbone of the Omineca Crystalline Belt and
comprises a deformed, metamorphosed, uplifted block nearly 60 km. wide
that is fault-bounded on the east adjacent to the Rocky Mountain Trench
and Rocky Mountains and on the west against the northern extension of the Quesnel Trough.

The Quesnel Trough can be divided into an eastern group of fairly
continuous Mississippian - Permian (±Upper Triassic ?) clastic sediments
rocks with conglomerate and a western group of Upper Triassic volcanic,
the Takla Group. The eastern upper Paleozoic rocks are intruded by gabbro
sills and alpine and zoned ultramafic rocks, and are associated with abun-
dant basalts. The western Takla rocks are intruded by various phases of
the Hogem composite batholith of Early Jurassic (175 my.) and Early Cre-
taceous (120 my.) ages (Eadie, 1976; Garnett, 1974; Woodsworth, 1976).
The contact between these two terranes is generally faulted, although
depositional contacts have been shown in some places.

The Hogem batholith is composite and consists of early mafic phases
followed by more calc-alkaline and syenitic plutonic rocks. Much of the
older Hogem terrane is deformed, especially near its western margin with
the Pinch fault system, which separates it from an assemblage of upper
Paleozoic and Upper Triassic eugeoclinal sedimentary, mafic and ultra-
mafic plutonic, and mafic volcanic rocks farther west including both the
Stuart Lake belt of Monger and Patterson (1974) and the Asitka Group of
Lord (1948) and Richards (1976). The Hazleton Group, consisting of vol-
canics and sediments of Early Jurassic age, rests upon these upper Paleozoic and Upper Triassic rocks, and they in turn are faulted and folded during later Mesozoic and Cenozoic deformation (Richards, 1976).

The boundary between the Precambrian rocks and the upper Paleozoic rocks to the west is a fault. Roots (1954) concluded that it was a steep normal fault, whereas Mansy (1974) and Gabrielse et al (1976) provide evidence that the contact is in places either a steep thrust or a reverse fault. Several periods of movement are probable.

To the east, normal faults separate the Rocky Mountain Trench floor from the highlands on either side. Across the Trench to the east, east-verging foreland thrust-fold belt geometry in general prevails in late Precambrian - Mesozoic miogeoclinal rocks (Irish, 1970), although to the north, the style of folding becomes more reminiscent of the Mackenzie Mountains where older Proterozoic strata are present. These older Proterozoic rocks (Purcell - Belt equivalent strata) are not exposed beneath the Ingenika - Tenakihi Group rocks in the Aiken Lake map-area, and they may be absent altogether.

Within the Rocky Mountain Trench region (Sifton Basin), detrital sediments of Upper Cretaceous to Eocene age are present and record high-relief sedimentary and metamorphic source terrane (Eisbacher, 1974a). K-Ar dates from detrital mica and granitic boulders indicate uplift and cooling of metamorphic core zone rocks by mid-Cretaceous time, although the erosional history is no doubt more complex.

The geology of this general area is dominated by the effects of Mesozoic and Cenozoic plutonic and deformational events, thereby obscuring possible Paleozoic activity, and the geological evolution is in many ways analogous to that of the Shuswap complex of southern British Columbia where Triassic and Paleozoic rocks are juxtaposed against higher grade metamorphic rocks of complex history.
Stratigraphic relations within the study area are considerably disrupted because of obvious local repetition and structural complexity. Rock types that are most abundant consist of thin-layered coarse quartz-mica schists and micaceous, feldspathic, and schistose quartzites with complete gradation between these lithologies. Many rocks show porphyroblastic garnet and plagioclase, and mineral segregations are well developed in places. Quartzo-feldspathic mica gneisses which appear to be feldspar-rich equivalents of the schists and quartzites are also abundant. Calc-silicate marbles, amphibolites, tourmaline schists, and quartz pebble conglomerates are also present but are quite rare and thin; consequently, they were difficult to map for any great distance. Nearly all conspicuous marker layers were found to pinch out into the surrounding schists within several hundred meters. With the exception of a few thick quartzites, the abundant schistose lithology proved to be discouraging during mapping because of its discontinuous and gradational nature caused by boudinage, repeated folding, and possibly discontinuous inherited depositional characteristics.

Because of these stratigraphic complexities, interpretation of the structural evolution relies upon minor structures; the large scale effect of early syn-metamorphic deformation is largely unknown.

The monotonous nature of the lithology was occasionally interrupted by pegmatite sills and bodies, both concordant and discordant. These igneous rocks are minor constituents in the area near Chase Mountain and south of Ravenal Creek (figure 2a), but become very numerous in the high grade rocks near Blackpine Lake (figure 2b). These rocks will be dealt with in a later section. Recrystallized mylonitic rocks are present south and east of Ravenal Creek; they occur in a small area as a 10 meter thick layer, dipping 20°-30° to the southwest. This mylonitic zone separates rocks of similar
Figure 2a. Structural and geologic map of thesis area near Chase Mountain showing attitude of compositional layering.
Figure 2b. Geologic and structural map of thesis area near Blackpine Lake.
schistose lithology and its relation to the structural sequence is unclear.

The grain size of the mica schists increases in a very general way to the southeast and correlates with an increase in quantity of pegmatites. There is probably a corresponding increase in the metamorphic grade although this is difficult to prove due to scarcity of aluminous schists or other grade indicators.

Original bedding in most cases has been transposed into the axial plane of the earliest folds; however, recognizable load structures, graded bedding, and slightly deformed cross-bedding were observed in a few places. The original sediments may have been distal parts of turbidite sequences. With increasing deformation and flattening, these structures become unrecognizable. Despite the rare deviations from the dominant lithology, the lasting impression of these rocks is one of a uniform, monotonous sequence of metamorphosed dirty sandstones, siltstones, and quartz-rich semi-pelitic sediments. No estimate of stratigraphic thickness can be made because of structural complications that render the stratigraphic sections of Roots (1954) for Tenakihi Group rocks of dubious value.

The lithology of the Tenakihi Group mapped in this study in many ways resembles a more highly metamorphosed stratigraphic correlative of the Middle Miette and Kaza Groups as described in the McBride map-area by Campbell and others (1973) and Sutherland Brown (1963). There, the Middle Miette consists of coarse sandstone, pebbly sandstone 'grit', quartz siltstone and argillite, the coarser units being very immature both texturally and chemically, in places containing up to 25% feldspar grains. Overlying the Middle Miette, the Upper Miette group consists of mudstones, silty argillites, a zone of coarse clastic rocks, and a thick limestone correlative with the Cunningham Formation which rests on the Kaza Group west of the Rocky Mountain Trench. The Middle Miette is correlative with the Kaza Group west of the
Trench which lies below a thick, recessive argillaceous unit known as the Isaac Formation, which in turn forms the base of the Cariboo Group.

These same rocks (Tenakihi-Kaza-Middle Miette) are probably also broadly correlative with the lower part of the Missinchinka Group described by Irish (1970) in Halfway River map-area, and correspond in general to the lower portion of the Windermere Series exposed through the entire length of the British Columbia Cordillera (Gabrielse, 1972). Figure 3 shows the relationships between these rock groups.

Similar lithologic correlatives to the Upper Miette group are present in the Aiken Lake area; these rocks have been recently described and correlated by Mansy (1972). Briefly, from oldest to youngest, they are as follows:

1) gritty clastic unit: shale, siltstone, quartzo-feldspathic sandstone and conglomerate (Tenakihi and lower Ingenika groups).

2) thin-bedded calcareous phyllite, equivalent to the Isaac Formation.

3) resistant, well-bedded limestone, equivalent to Cunningham Formation.

4) diverse unit of limestone, phyllite, impure quartzite, sandy limestone, dark green shales, and impure sandstone, equivalent to the Yankee Belle Formation.

5) pure white quartzite, equivalent to the Yanks Peak Formation.

6) impure quartzite interbedded with shale, equivalent to the Midas Formation.

7) blue-gray limestone with siltstone in its middle part, equivalent to the Mural Formation of lowest Cambrian age.

The above units 3 and 4 correlate with the Good Hope Group and units 5 to 7 are equivalent to the Atan Group in the Cassiar Mountains (Mansy, 1972).

In general, the distinction made by Roots (1954) between the Tenakihi Group and the lower Ingenika Group has been found difficult to apply, both in the study area where the strata at Blackpine Lake and near Chase Mountain are identical (but belong to Ingenika and Tenakihi Groups, respec-
<table>
<thead>
<tr>
<th>age</th>
<th>Cassiar Mountains</th>
<th>Omineca Mountains</th>
<th>Cariboo Mountains</th>
<th>Rocky Mountains</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.6E</td>
<td>Atan Group</td>
<td></td>
<td></td>
<td>Base not exposed</td>
</tr>
<tr>
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Figure 3. Correlation chart of uppermost Precambrian and lowermost Cambrian rocks in north and east-central British Columbia.

...tively) and elsewhere as noted by Mansy (1976). In many places, the apparent lithologic contrast may be a function of metamorphic grade and structural style rather than rock type. The rocks in the study area near Chase Mountain are located near the 'apparent' base of the exposed section, and they lie several thousands of feet structurally below the thicker quartzites and limestones in the Finlay and Swannell Ranges near Pelly Lake where the structural history appears much simpler (Mansey, 1972, 1974).
Figure 4. Structural map of thesis area showing phase 1 data.
There are at least four phases of deformation recognized within the Chase Mountain area (Parrish, 1976). Phases 1 and 2 (F1, F2) appear to be related to ductile rock conditions associated with metamorphism; they are characterized by N to NW-trending, tight to isoclinal, gently inclined to recumbent folds with abundant mineral recrystallization and development of axial plane foliation. Later phases 3a and 3b trend NW and N respectively, are open upright folds, and appear to be post-metamorphic. These later phases have produced an elongate antiformal dome or culmination in the earlier foliations and are probably related to the large scale structures in the Ingenika Group rocks to the north in Wrede Range and near Pelly Lake in the Russell Range.

The study area has been divided into six domains possessing moderately homogeneous geometry for stereographic structural analysis.

Phase 1 (Figure 4)

F1 folds are approximately similar isoclinal folds trending from N-S to NW-SE, and are characterized by a gently dipping axial plane schistosity. Amplitudes vary from a meter or so to several tens of meters, and the folds are most commonly rootless and extremely flattened (figure 5). F1 folding has resulted in transposition of the bedding into the axial plane, so that the two are subparallel. The planar and linear elements of this event have been refolded both by F2 and F3 folds; F1 orientations therefore vary across the major antiform in a systematic way (figure 4). The extreme flattening of F1 hinges is no doubt due to renewed closure during F2 folding so that the original form of F1 folds prior to F2 is not known. Features seen in the field associated with F1 are isolated fold hinges (figure 5a), spectacular boudins (figure 5b), discontinuous and transposed layering, and an extremely flattened, flaggy appearance of compositional layering. When
Figure 5a. $F_1$ fold hinges within a quartzo-feldspathic layer, looking northwest.

Figure 5b. Isoclinal $F_1$ fold with axial plane schistosity in quartz-mica schist, looking NW.

Figure 5c. Photograph of hinge of $F_1$ fold, looking NW, showing the development of axial plane schistosity.

Figure 5d. Photograph of large boudin of quartzite in flaggy schist, looking NW.
quartz pebble conglomerate is seen, a marked flattening and an elongation parallel to fold axes is particularly evident with aspect ratios of the pebbles typically 1:4:15. It is not known whether this strain is regionally representative or not.

Phase 2 (Figure 6)

F2 folds are distinguished from F1 folds by the presence of a folded and crenulated schistosity in the cores of F2 folds (figure 7), refolding of F1 hinges (figure 8), local divergence of axial plane orientation (figure 8), and by their tight (as opposed to sub-isoclinal) form with rounded hinges (figure 9). A near horizontal to moderately dipping axial plane foliation is often associated with these folds, but it is only rarely developed as penetratively as F1 schistosity, and relics of a folded earlier fabric are generally present. Many exposures (figure 10 and Parrish, 1976) illustrate these folds 'cascading' or verging to the east-northeast, and these vergences have been shown in the stereographic diagrams. No significant reversal of vergence was recognized throughout the area, and it is not known whether there is a corresponding lower limb of a very large structure in which vergences would be reversed.

Several large exposures were observed where there is a progressive transposition of the compositional layering into the F2 axial plane (figure 11a). This transposition is accomplished by increased flattening and tightening of hinges, accompanied by high strain and/or slip along discreet zones of structural dislocation resulting in the shearing off and loss of one limb of a fold. Mylonitic rocks are sometimes associated. This structural style (figure 11a) is recognizable on both small and large (over 100 meters) scales.

Many dislocation or fault zones are present throughout the study area (figure 11b), and these are likely F2 structures; their displacements are
Figure 7. Photograph of folded schistosity in hinge region of $F_2$ fold, looking S.

Figure 8. Photograph of refolded $F_1$ hinge on the limb of later $F_2$ fold, looking S. The $F_1$ hinge was tightened during $F_2$ deformation.
Figure 6. Structural map of Chase Mountain area showing phase 2 fold orientation.
Figure 9a. Photograph of $F_2$ fold in interlayered quartzites and schists.

Figure 9b. Photograph of pegmatite deformed during phase 2 deformation.
Figure 10. Sketches of east-verging phase 2 folds. All of the locations exhibit very strong development of rodding or lineation parallel to the fold axes.
Figure 11a. Detailed sketch of the increasing development of flattening and transposition of layering during phase 2 folding, NW view.

Figure 11b. Photograph of dislocation zone in mica schists, looking W.
not known but are not thought to be great as no exotic rocks were found.

In spite of the lack of mappable stratigraphic units, an attempt was made to construct a tectonic profile across the area showing gross geometry (figure 12). The section indicates that existing data are consistent with eastward transport of rocks in nappe-like fashion where the lower limbs of large recumbent folds are either sheared off or not exposed. Lack of stratigraphic control renders the section somewhat conjectural.

Phase 3

Two additional distinct sets of folds are present within the area, but their relative ages are indeterminate. Phase 3a is the major northwest trending antiformal fold which culminates in the thesis area and deforms earlier F1 and F2 fabrics (figures 4, 6). This is a very large feature, being an antiform at least 15 km. in width. It is an open upright fold, the limb dips being 30°–40°, and it does not appear to be associated with minor structures which reflect its geometry and orientation. This structure causes the variable orientation of F1 and F2 planar and linear structures as well as the regional changes in attitude of the compositional layering. The antiform plunges both NW and SE away from the Chase Mountain area, and
the axial trace is not offset by later faulting or folding within the study area. In addition, a minor culmination - depression pair is present 3 km. southeast of the Summit of Chase Mountain and near Ravenal Creek, respectively (figure 2). Metamorphism does not appear to have accompanied this folding.

Phase 3b is an areally restricted set of folds of variable but considerable size and wavelength that is present in a northwest trending zone near the crest of the F3a antiform southeast of Ravenal Creek. The folds themselves are north-trending, upright to steeply inclined to the east, and have variable form. They possess flexural slip to flexural flow geometry, and their form varies from open at higher structural levels to tight at depth (figure 13). These structures deform both F1 and F2 folds and are en echelon near the F3a antiformal crest. These folds are colinear (but not coplanar) with F1 and F2 folds found in the immediate vicinity, and it appears likely that the older structural fabric has been reactivated upon compression in the core of the F3a antiform. This explanation implies that they result from a 'room' problem in the core of the larger structure. This interpretation is supported by the increasing closure of the fold limbs at depth as well as their en echelon and spatially restricted nature.

The large antiform is most probably related to the large folds and associated thrusts mapped by Roots (1954) and Mansy (1972, 1974) in the central Swannell Ranges underlain by the Ingenika Group rocks, especially near the Wrede Range, the major structure of which is shown in figure 14. The geometry of the large antiforms are reasonably compatible, although in the ranges that are at low metamorphic grade (chlorite), there appears to be a lack of the earlier deformations present near Chase Mountain. More work is obviously needed to substantiate or reject this view. Recent work by Mansy (1976) in greenschist and lower amphibolite facies rocks in central Swannell Ranges indicate the presence of earlier isoclinal folds and a more complex history than previously recognized, in accord with the
Figure 13. Sketches of \( F_3 \) folds, looking S. These folds deform \( F_1 \) folds, as shown in (c), as well as \( F_2 \) structures.

Figure 14. Structure sections across Wrede and Russell Ranges from J. L. Mansy.

[Diagram of structural sections across Wrede and Russell Ranges with labels and geological sections.]
data presented in this paper. However, the structure in the very low grade Ingenika Group in Russell Range near Pelly Lake is fairly simple (Mansy, 1972, 1974); an earlier, isoclinal folding is missing there. It is clear that differences in pressure and temperature conditions affect the way that rocks behave under stress, and caution must be used when correlating structures in rocks of variable lithology and different metamorphic grade. Indeed, structures developed at higher pressure and temperature conditions may not be present in rocks that are structurally higher and at different grade. This has been shown by Campbell (1970) in the Cariboo Mountains.

F3a structures in the study area and those correlative structures to occurred during the north in the Wrede Range V, either non- or only low-grade metamorphism, and the author believes that these structures fold metamorphic isograds. If so, the metamorphic conditions during F3a folding near Chase Mountain and during analogous folding in Wrede Range were probably not significantly different, and the structural correlation is reasonably justified.

Other Minor Structures

Later folds of more brittle deformational style are locally present throughout the area. These include small kink and crenulation structures and associated lineations, flexural slip folds of small amplitude (figure 15a), faults, and joints. No convincingly consistent orientations were found in these late structures, (figure 15b) and their significance is not known.

Stereographic Analysis

The orientation of F1, F2, F3a, and F3b structures is summarized in stereographic fashion in figures 16 and 17. Domains 1 and 2 define the northwest plunging region of the F3a antiform near Chase mountain, and domains 3, 4, and 5 define the southeast plunging hinge southeast of Ravenal
Figure 15a. Sketches of late structures, viewed towards direction in parentheses.
Figure 15b. Equal-area stereonet plot of orientations of late structures; solid circle represents lineation or fold axis, open circle denotes axial plane or cleavage of late structure, triangle denotes attitude of dike.

- Open circle: axial plane of late fold
- Open triangle: dike
- Solid circle: fold axis of late structure
Figure 16. Structural map of Chase Mountain area showing domain boundaries.
Creek. Domain 6 is overwhelmed by F3b folds and neatly depicts F3b geometry (figure 17b).

Domain 1, 2 (figure 17a-1,2)

On a statistical basis, F1 and F2 fold axes (L1, L2) near Chase Mountain are coaxial, trending 324°±10° and plunging less than 20°. F1 axial planes (S1) in domain 1 cluster about 105°/24° NNE at the hinge but are very scattered along a partial great circle girdle roughly perpendicular to the concentration of L1. The concentration of S1 coincides with poles to compositional layering, confirming the near-isoclinal nature of F1 folds.

When S1 and the compositional layering, S_o, are compared for domains 1 and 2, it is clear that S1 forms a scattered array of points that lie along a partial small circle girdle approximately 20°-30° about a non-cylindroidal vertical fold axis. This indicates the presence of a dome-shaped culmination defined by the planar elements S_o and S1. In domain 2, fold axes of F1 and F2 vary from 145°/0° to 182°/35°, most points being from the southward-plunging part of this dome. In general, S1 and S_o dip gently to the south and southwest in domain 2, and to the north in domain 1.

Phase 2 structures are very common in domain 1, but less so in domain 2. Despite the fact that great care was taken to appropriately identify F1 and F2 folds, it is possible that some folds were mistakenly identified in the field due to their similar appearance and lack of other diagnostic criteria. In domain 1, fold axes L2 trend from 291°/10° to 340°/10° with most data clustering near 325°±10°/13°±11°; they are characterized by a strong sense of asymmetry or vergence toward the northeast as indicated in figure 17a. S2 data are scattered along a great circle girdle with a calculated fold axis of 322°/19°, which defines the F3a fold and verifies that F3a de-
Figure 17a. Equal-area stereonets of structural features, domains 1-5.

- compositional layering
- F1 axial planes
- F2 axial planes
- F1 fold axis
- F2 fold axis
- unspecified fold axis or lineation

sense of vergence
Figure 17b. Equal-area stereonet of orientations of structures in domain 6 which is dominated by phase 3b folds.

- Composition: layering
- $\perp F_2$ axial planes
- $\perp F_{3b}$ axial planes
- Mineral lineation
- $F_2$ fold axes
- $F_{3b}$ fold axes
forms and is also colinear with F1 and F2 on a statistical basis. The fact that S2 planes do not fall on a small circle locus is due to non-random distribution of data and outcrop. Fold F3a, the regional large scale antiform, is a post-F2 event that deforms earlier structures.

In domain 1 below an important zone of dislocation or faulting, F2 structural data are markedly distinct from F2 orientations above this fault, as shown in figure 18. Below the fault, S2 strikes SSE and dips 20°-40° southwest. The rocks below are characterized by extreme flattening and strong transposition of foliation into the F2 axial plane. This fault is thought to be analogous or continuous to that shown in figure 6 and it is probably of late F2 age.

Domains 3, 4, and 5 (figure 17a - 3,4,5)

Domains 3, 4, and 5 respectively occupy the SE-plunging hinge region and adjacent limbs of the major F3a antiform (figure 16). In domain 5, the dominant foliation strikes northwest and the dips steepen northeasterly from 18°, 500 meters east of the crest, to about 50°, at a distance of 3 km. from the crest. Axes of F2 folds plunge to the north near the crest but plunge southwest farther away. Phase 1 folds where seen are isoclinal and trend NW-SE.

Domains 3 and 4 were studied in greater detail and reveal the dominance of F2 structures that overwhelm F1 folds. Generally, L1 and L2 are roughly colinear and plunge 10°-25° to 176°± 5°. Because most data lie on the south to southwesterly dipping limb of the F3a antiform, poles to S1, S0, and S2 do not fall on a well-defined girdle. However, combined with poles to data from domain 5, their distribution can be considered consistent with a near-great circle distribution about the F3a fold axis, 120°/10°. Due to considerable structural complexity, the scatter is great.
Figure 18. Simplified structural map of change in orientations of $S_0$, $S_1$, and $S_2$ surfaces across major fault, domain 1.

Figure 19. Photograph of refolded lineations. The deformed linear structure is probably related to $F_2$, which is in turn refolded by a later $F_{3b}$ (?) fold.
Where seen, F2 folds retain the strong east-northeast vergence recorded farther to the northwest in domains 1 and 2.

Domain 6 (figure 17b)

Domain 6 contains F3b folds that are numerous and that overwhelm earlier F1 and F2 structures. Poles to S0, the compositional layering, and S1 are scattered about a well defined great circle whose calculated pole or fold axis is due north and horizontal (±5°). This orientation is subparallel to minor F3b folds seen in the field as well as subparallel to older F2 fold axes and regional lineation. Refolded folds and refolded lineations are locally present here (figure 19). F3b axial planes, S3b, cluster about 176°/60°E and bisect a nearly vertical limb and a moderately eastward dipping limb. Inclination of these folds is therefore westward. As stated previously, these folds are tightest at lowest structural levels and more open at higher levels and are interpreted as folds developed en echelon as a 'room' problem in the core region of the large F3a antiform.

Summary

The following observations are pertinent:

1) Structural overprinting has caused considerable scatter in the data. The distribution of poles to S0, S1, and in most cases S2 clearly defines the geometry of F3a and F3b folds which post-date these earlier events.

2) It is clear that previously developed structural fabric and anisotropy provide continued influence on the orientation of later structural features. The colinearity of folds in any region or domain attests to this.

3) The major F3a antiform is divisible into two sub-culminations, centered 3 km. and 8 km. respectively southeast of the crest of Chase Mountain, as defined by the orientations of S0, S1, and S2 planar and L1 and L2 linear structures.

4) In at least one area east of Chase Mountain, a gently dipping zone of structural dislocation separates rocks having markedly different fabric orientations and styles, but the regional importance of this type of feature is not clear.
5) F2 deformation has consistent northeastward vergence, and no reversal of this vergence was recognized. It is possible that the entire sequence observed lies on the upper limb of a large recumbent east-facing antiform that may have suffered detachment along its lower synclinal hinge, in a nappe-like structure. This interpretation, although consistent with mesoscopic geometry, is entirely conjectural and cannot be proved without knowledge of structures at depth.

Blackpine Lake Area (Figures 20-22)

Phases 1, 2

The structural evolution of the Blackpine Lake area is similar in its earlier stages to that of the Chase Mountain area. F1 folds are present which are sub-isoclinal and possess axial plane schistosity (figure 20); F2 folds are present that fold this schistosity and related transposition foliation. Both generations of folds are tight to sub-isoclinal, and both seem to be syn-metamorphic and formed in a ductile rock regime. Pegmatitic bodies and sills are intruded before, during, and after folding and bear corresponding deformational fabric (figure 21); many synkinematic granitic sills are intruded parallel to the axial planes of F2 folds. These early folds cluster about northwest trending (323°±5°), gently plunging (10°-20°) axis orientations, and F1 and F2 axial planes dip at moderate angles (30°-50°) to the north and northwest (figure 22). In many cases, it was not possible to classify all early folds as either F1 or F2, and consequently, the structural history is less clear. The grade of metamorphism is higher here and may account for increased development of F2 axial plane schistosity, making it more difficult to distinguish these fold sets. Poles to S0, however, are scattered about a fairly well defined great circle whose calculated fold axis is 340°/0°-30°, and this may reflect later open folding correlative with Phase 3 near Chase Mountain.
Figure 20. Photograph of F, fold in the Blackpine Lake area, looking NW. The fold is cut by a small pegmatite.

Figure 21. Photograph of folded and deformed granitic sills in the Blackpine Lake area, looking S. Black lines outline folds.
Figure 22. Equal-area stereonet plot of orientations of structural data in the Blackpine Lake area.

- • compositional layering
- ○ F1 axial planes
- □ F2 axial planes
- □ foliation in Blackpine Lake stock
- × lineation or fold axis of unknown generation
- ○ F1 fold axis
- □ F2 fold axis
Structural Relations of the Blackpine Lake Granitic Stock

The Blackpine Lake granitic stock is foliated, and in thin section appears strongly deformed, possessing mortar fabric (figure 23a). The surrounding country rocks possess a similar texture that can be interpreted to have been caused by flattening approximately normal to the foliation (figure 23b, c, d). Roots (1954) stated that foliation in the granitic stock was crudely parallel with the margin of the intrusion, and he attributed this to igneous flow processes. He also considered this body to be part of the Wolverine Complex and related to the 'granitizing' activity. However, undulatory extinction and strong recrystallization in quartz indicate significant post-crystallization deformation, at least near the margins. The outcrop pattern and domal foliation (Roots, 1954) are interpreted in this study to be due to forceful, possibly diapiric emplacement with corresponding deformation of surrounding country rock. These processes clearly post-date early folding as well as the peak of regional metamorphism.

Mortar fabrics are best developed in the quartz-rich country rock surrounding the stock; however, undulose extinction and sutured quartz grain boundaries are present in many rocks throughout the Chase Mountain-Blackpine Lake area and indicate that deformation has continued (or resumed) after most recrystallization has ceased.

Relations between Crystal Growth and Deformation

In general, the metasediments have acquired their mineralogy and texture by synkinematic recrystallization. The schistosity defined by the preferred orientation of biotite and muscovite was originally formed during F1 folding. F2 folding deformed this foliation and in many rocks, a new generation of micas has grown parallel to S2 (figure 24a).
Figure 23a. Photomicrograph of Blackpine granitic stock showing mortar fabric and recrystallized quartz grains. X-nichols, 25X.

Figure 23b. Photomicrograph of pegmatite of country rock showing kink bands in mica and recrystallized quartz. X-nichols, 25X.

Figure 23c. Photomicrograph of pegmatite showing kinked micas and mylonitic texture of ribbon quartz. X-nichols, 25X.

Figure 23d. Photomicrograph of pegmatite showing broken feldspar grain with quartz fracture filling and mylonitic fabric. X-nichols, 25X.
Figure 24a. Photomicrograph of relict F_2 fold with some growth of new micas. Plane-polarized light, 25X.

Figure 24b. Photomicrograph of relict fold with well-developed polygonization of micas indicating that recrystallization outlasted deformation. Plane-polarized light, 25X.

Figure 24c. Photomicrograph of sericite + chlorite pseudomorph after synkinematic garnet. X-nichols, 25X.

Figure 24d. Photomicrograph of chlorite + muscovite + quartz pseudomorph after synkinematic garnet. Plane light, 25X.
During this F2 folding, micas were strongly kinked, bent, and crenulated in the cores of F2 folds (figure 24b), indicating a pre-F2 metamorphism producing the schistosity. Where unchloritized garnets retain their form, it is clear that the foliation wraps around porphyroblasts of garnet as well as plagioclase (figure 24c, d); 'spiral' or 'snowball' garnets are rare to absent and this observation supports a flattening as opposed to shear origin for the foliation. Following F2 deformation, post-kinematic polygonization of micas and recrystallization of quartz have given rise to present fabrics. The metamorphism accompanying F2 probably outlasted deformation in most areas. Late brittle folds, kinks, and fractures that have deformed the rocks are probably related to the development of undulose extinction and somewhat sutured grain boundaries in quartz throughout the area. The late (post F2) development of cataclastic and mortar fabric surrounding the Blackpine stock may be an entirely separate and local event. Retrograde alteration of garnet to chlorite and kyanite to sericite is widespread (figure 25), and this is most likely a post-F2 low grade metamorphic event which developed later than the polygonization of micas. Relationships are illustrated in figure 26.
Figure 25a,b. Photomicrographs of sericite replacement of kyanite. 25X.

Figure 26. Schematic diagram showing relations between deformation, metamorphism, granitic intrusion, geochronologic disturbance, and time.
METAMORPHIC PETROLOGY

Introduction

The Chase Mountain and Blackpine Lake areas have been subjected to both amphibolite facies metamorphism (medium grade of Winkler, 1974) and later retrogression. This statement is supported by the following data:

1) presence of staurolite, kyanite, and/or sillimanite in aluminous schists;

2) absence of the equilibrium assemblage chlorite + muscovite + quartz throughout the study area;

3) stability of muscovite + quartz into the sillimanite-fibrolite field;

4) lack of in situ melting of pelitic compositions;

5) alteration of kyanite, staurolite, and garnet to muscovite + chlorite.

Because of the low alumina content of the metasediments, diagnostic mineral assemblages are only rarely developed in pelitic rocks. Calc-silicate marbles, however, host a variety of mineral associations, some of which provide data on temperature and composition of the fluid phase. Rocks that do provide critical P-T-X\textsubscript{CO}_2 information are scattered, and specific isograds could not be mapped (figure 27).

Chase Mountain Area

Because detailed compositional data on coexisting mineral phases was not obtained, and because aluminous schists were so rare, the pressure of metamorphism is not well determined. Muscovite pseudomorphs after kyanite were found at several localities (figure 28), and the assumption will be made that kyanite represented the stable aluminosilicate polymorph during the principle metamorphic episode. Using the aluminosilicate triple point of Richardson, Gilbert, and Bell (1969) of 5.5 kbars at 622° C, the pressure of metamorphism will be assumed to have been about 6 kbars.
Figure 27. Map of area showing distribution of samples referred to in text.

Near Blackpine Lake, staurolite breakdown occurs in the sillimanite field, which places an upper limit of pressure at about 7–8 kbars.*

In the Chase Mountain area, no staurolite or fibrolite were found, but the absence of the equilibrium assemblage chlorite + muscovite + quartz indicates stability in the amphibolite facies or medium grade of Winkler (1974). Retrograde metamorphism has produced chlorite ± quartz ± sericite pseudomorphs after garnet (figure 24c) and muscovite pseudomorphs after kyanite (figure 28). Pelites and carbonates are characterized by the relevant assemblages listed in table I.

* The author is aware of the experimental and thermodynamic problems with the determination of kyanite-sillimanite-andalusite equilibria, but the data of Richardson, Gilbert, and Bell is thermodynamically consistent and seems as good as data from other workers.
Figure 28. Photograph of muscovite pseudomorphs after kyanite. (See figure 25 for photomicrograph.)

Figure 29. General pressure-temperature conditions of metamorphism.
TABLE I. Representative assemblages from the Chase Mountain area.

Pelites and semi-pelites:*

\[
\begin{align*}
\text{qtz} + \text{mus} + \text{biot} + \text{ox} & \pm \text{plag} \ (\text{oligoclase}) \pm \text{gnt} \pm \text{cc} \\
\text{qtz} + \text{mus} + \text{biot} + \text{plag} + \text{cc} + \text{czo} & \pm \text{sph} \pm \text{apa} \pm \text{ox} \\
\text{qtz} + \text{mus} + \text{ky} \ (\text{pseudomorphs}) & + \text{biot} \pm \text{gnt} \pm \text{plag} \pm \text{ox} \\
\text{qtz} + \text{plag} + \text{kf} + \text{biot} + \text{mus} + \text{ox} & \\
\text{qtz} + \text{biot} + \text{gnt} + \text{Mg-chl} & \\
\text{qtz} + \text{tour} + \text{ep} + \text{mus} & \\
\end{align*}
\]

Amphibole-bearing rock:

\[
\begin{align*}
\text{qtz} + \text{hbl} & \pm \text{gnt} + \text{plag} \pm \text{czo} \pm \text{sph} \pm \text{biot} \\
\text{hbl} + \text{gnt} + \text{plag} + \text{biot} + \text{qtz} + \text{cc} + \text{kf} & \\
\text{hbl} + \text{gnt} + \text{plag} + \text{mus} + \text{biot} + \text{czo} + \text{sph} & \\
\end{align*}
\]

Calc-silicates:

\[
\begin{align*}
\text{czo} + \text{act} + \text{biot} + \text{qtz} + \text{plag} + \text{kf} & \pm \text{mus} \pm \text{Mg-chl} \pm \text{sph} \\
\text{cc} + \text{trem} + \text{qtz} + \text{plag} + \text{kf} + \text{biot} + \text{sph} & \\
\text{cc} + \text{czo} + \text{qtz} + \text{diop} + \text{Mg-chl} + \text{kf} + \text{trem} & \\
\text{qtz} + \text{cc} + \text{mus} \pm \text{biot} \pm \text{czo} \pm \text{sph} \pm \text{chl} & \\
\end{align*}
\]

* Retrograde chlorite and muscovite are present in many of these rocks, but have not been included in equilibrium assemblages.

Mineral abbreviations

\[
\begin{align*}
\text{qtz} - \text{quartz}, \text{mus} - \text{muscovite}, \text{biot} - \text{biotite}, \text{plag} - \text{plagioclase}, \\
\text{gnt} - \text{garnet}, \text{chl} - \text{chlorite}, \text{ky} - \text{kyanite}, \text{kf} - \text{K-feldspar}, \text{ox} - \text{oxides}, \text{ep} - \text{epidote}, \text{czo} - \text{clinozoisite}, \text{cc} - \text{calcite}, \text{hbl} - \text{hornblende}, \\
\text{trem} - \text{tremolite}, \text{act} - \text{actinolite}, \text{diop} - \text{diopside}, \text{sph} - \text{sphene}, \text{apa} - \text{apatite}, \text{tour} - \text{tourmaline}, \text{ctd} - \text{chlorotoid}, \text{alm} - \text{almandine}.
\end{align*}
\]
All pelites appear to have equilibrated above the stability of quartz + muscovite + Fe-chlorite and within the stability field of quartz + muscovite + plagioclase due to the lack of primary chlorite (except Mg-rich variety) and the lack of Al₂SiO₅ + K-feldspar and in situ migmatitic rocks that have experienced melting according to the reaction, muscovite + quartz + albite + H₂O = melt + Al₂SiO₅ (Winkler, 1974). Assuming P = 6 kbars, these limits restrict the P-T environment to the shaded region of figure 29 assuming P_{total} = P_{H₂O}.

The composition of the fluid phase in all rock types was H₂O-rich because of the ubiquitous presence of clinozoisite (or zoisite) in calc-silicates and most amphibolites. Grossular garnet is not present in any calc-silicates and clinozoisite (zoisite) + quartz remains a stable assemblage. At 6 kbars, this demands that X_{CO₂} be less than 0.1 (Storre, 1973; Hewitt, 1975). This is consistent with observations by Ghent and Devries (1972) in similar types of rocks. The stability of zoisite + quartz is limited by the reaction,

zoisite + quartz = grossular + anorthite + H₂O,

and the stability of zoisite alone by the reaction,

zoisite + CO₂ = calcite + anorthite + H₂O,

as shown in figure 30a.

Dilution of the anorthite content of plagioclase shifts the isobaric, univariant curve toward lower temperatures and more H₂O-rich fluid compositions as depicted schematically in figure 30b. Hewitt (1973) has calculated this displacement for the reactions.

zoisite + CO₂ = calcite + anorthite + H₂O, and

muscovite + calcite + quartz = anorthite + K-feldspar + vapor

at 6 kbars fluid pressure, as shown in figure 30a. Plagioclase compositions in rocks of this study are less than An_{60}, generally about An_{50}, which limit X_{CO₂} to less than .05 or so, varying slightly with temperature.
Figure 30a,b. T-X diagrams for SiO₂-CaO-Al₂O₃-K₂O-MgO-H₂O-CO₂:
a) calibrated diagram  b) schematic diagram showing displacement.
Rocks 17, 18, and 271-2 contain the assemblage,

\[ \text{quartz} + \text{muscovite} + \text{calcite} + \text{clinozoisite} + \text{plagioclase}, \]

which restricts them to the region below the stability of \( \text{An}_{40} + \text{K-feldspar} \) and on the reaction of zoisite breakdown (figure 31). Using the displaced curves of Hewitt (1973), conditions for these rocks with oligoclase-andesine plagioclase are approximately \( T \leq 510^\circ, \ X_{\text{CO}_2} = .04, \) and \( P_{\text{fluid}} = 6 \text{ kbars}. \) The fluid composition of the rocks that contain zoisite + calcite + plagioclase is being buffered by the reaction, and the rock is constrained to remain on this reaction until one of the components is used up.

Sample 87-2 contains the assemblage,

\[ \text{kt} + \text{trem} + \text{diop} + \text{cc} + \text{czo} + \text{Mg-chl} + \text{qtz}, \]

which lies on the reaction \( \text{trem} + \text{cc} + \text{qtz} = \text{diop} + \text{vapor}, \) on the \( \text{H}_2\text{O-rich} \) side of zoisite + \( \text{CO}_2 = \text{cc} + \text{anorthite} + \text{H}_2\text{O}, \) below zoisite + \( \text{cc} + \text{qtz} = \text{grossular} + \text{vapor}, \) and above biot + \( \text{cc} + \text{qtz} = \text{amphibole} + \text{kt} + \text{H}_2\text{O} \) (figure 32). The fluid composition is less than .08 and probably less than .05 \( X_{\text{CO}_2}, \) and without any further constraints, the temperature at 6 kbars may vary greatly (less than 550°) while retaining the same mineralogy. The stability of phlogopite + \( \text{cc} + \text{quartz} \) is enhanced with increasing substitution of FeO, TiO\(_2\), and F into phlogopite and with a realistic composition, this curve probably would restrict the temperature to greater than 500° or so. Unless \( X_{\text{CO}_2} \) were less than .02, the temperature must be greater than about 510°.

Sample 273 contains the assemblage,

\[ \text{kt} + \text{biot} + \text{actinolite} + \text{cc} + \text{czo} + \text{An}_{42} + \text{qtz}. \]

This assemblage and its stability has been discussed by Hewitt (1975); it is usually found in amphibolite grade rocks in regional metamorphic terrane. This assemblage must lie on the curves zoisite + \( \text{CO}_2 = \text{cc} + \text{anorthite (\text{An}_{42})} + \]

\* Curve for \( \text{trem} + \text{cc} + \text{qtz} = \text{diop} + \text{vapor} \) is extrapolated from 5kb curve of Slaughter, Kerrick and Wall (1975).
Figure 31. T-X diagram showing conditions of metamorphism for samples 17, 18, and 271-2 assuming a pressure of 6 kb. $P_{\text{fluid}}$.

Figure 32. T-X diagram showing conditions of metamorphism for sample 87-2 for the same pressure conditions.
CO$_2$ and biot + cc + qtz = kf + amphibole + vapor. It must have already crossed the reaction mus + cc + qtz = anorthite + kf + vapor to produce a bulk composition suitable to place it on the biot + cc + qtz reaction. This condition presents problems if phases are pure as illustrated in figure 33a. However, as noted by Hewitt (1975), FeO, TiO$_2$, and F substitution in phlogopite enhances the thermal stability of this reaction. Na, however, reduces the stability of the mus + cc + qtz reaction due to substitution in plagioclase; these two effects must bring these curves together so a reasonable topology is produced to account for the natural occurrence of this assemblage (figure 33b). There is insufficient data to determine the quantitative effect of such solid solution upon stability relations, so that the phlogopite + cc + qtz reaction is of little value. However, the presence of An$_{42}$ and the absence of diopside indicate conditions approximating $X_{CO_2} \leq 0.04$ and $T = 510^\circ$ at 6 kbars $P_{fluid}$, as shown in figure 33b.

These calc-silicates are distributed over a large area near Chase Mountain and their mineralogies are all compatible with H$_2$O-rich fluid conditions and temperatures at 6 kbars fluid pressure of about 510$^\circ$-530$^\circ$, or lower amphibolite facies. Given the limited compositional range of rocks studied, it is not possible to argue that metamorphic conditions vary in any systematic way across the field area.

Blackpine Lake Area

The Blackpine Lake area appears to have been metamorphosed to higher temperatures than the Chase Mountain area; this is concluded by the following data:

1) both fibrolite and coarse sillimanite have been observed in addition to muscovite pseudomorphs after kyanite;

2) the reaction trem + cc + qtz = diop + vapor has occurred in calc-silicate marbles;
Figure 33a. T-X diagram of Fe and Na-free system at 6 kb.

Figure 33b. Schematic T-X diagram showing a possible topology in a system with Na and Fe that could explain the mineralogy of sample 273.
3) staurolite bearing pelites also contain sillimanite + garnet + biotite;
4) muscovite + garnet bearing granitic rocks and pegmatites are very common;
5) muscovite + quartz remains stable.

In pelites, the following assemblages have been observed:

- \( \text{qtz} + \text{mus} + \text{biot} + \text{fib} + \text{plag} \) (An\(^{27}\)) + op
- \( \text{qtz} + \text{mus} + \text{biot} + \text{gnt} + \text{staur} + \text{fib} + \text{plag} + \text{op} + \text{chl} \) (retrograde)
- \( \text{qtz} + \text{mus} + \text{biot} + \text{gnt} + \text{plag} + \text{op} + \text{ky} \) (pseudomorph).

And, of special significance, sill + gnt + biot and staur + gnt form inclusions in a mus + biot + qtz + plag + fib + staur + and schist. In calc-silicates, the assemblage \( \text{cc} + \text{diop} + \text{trem} + \text{czo} + \text{plag} \) (An\(^{30}\)) + sph + apa is common.

Calc-silicates probably remain on the curve \( \text{zo} + \text{CO}_2 = \text{plag} + \text{cc} + \text{H}_2\text{O} \) and above the stability of trem + cc + qtz. Assuming \( P_{\text{fluid}} = 6 \text{ kbars} \) once again, \( X_{\text{CO}_2} = .05 \) and \( T = 530^\circ \), but less than \( 640^\circ \) because \( \text{zo} + \text{cc} + \text{qtz} \) remains stable (figure 34). Presence of fibrolite and sillimanite in adjacent rocks indicates \( T \geq 622^\circ \) based on the Richardson, Gilbert, and Bell triple point.

Pelites contain fibrolite (300, 304, 305) that is usually associated with muscovite or less commonly biotite, but the reaction(s) leading to its formation is not clear. Probably Na-muscovite + qtz = fib + albite + K-richer musc + H\(_2\)O is occurring, forming slightly more sodic rims on plagioclase, analogous to that observed by Guidotti (1970). When staurolite is present, it is usually surrounded by a sheath of coarse and fine-grained muscovite without fibrolite (figure 35a), and no obvious reaction relations can be texturally ascertained. However, in 304b, the presence of both sill + biot + mus (figure 35c) inclusions in garnet seems significant and points towards either more than one period of metamorphic equilibration, disequilibrium, or toward a buffered reaction. In the matrix of this rock staurolite
Figure 34. T-X diagram of conditions of metamorphism for calc-silicate marbles near Blackpine Lake. Hatched line indicates possible conditions. Because sillimanite is found in pelites, temperature are nearer the high temperature limit.
Figure 35a. Photomicrograph of staurolite with halo of sericite near larger garnet. Staurolite + quartz are not in contact. Plane light, 25X.

Figure 35b. Photomicrograph of sillimanite + biotite + muscovite inclusion in garnet. X-nichols, 160X.

Figure 35c. Photomicrograph of staurolite + sericite (alteration) inclusion in garnet. Note absence of quartz with staurolite. X-nichols, 40X.

Figure 35d. Photomicrograph of andalusite + muscovite + biotite + fibrolitic sillimanite in matrix of rock. X-nichols, 32X.
is surrounded by pseudomorphous coarse-and fine-grained muscovite and andalusite + mus + biot (figure 35d) and may be involved in reaction relations. Fibrolite grows in and appears to be replacing muscovite and andalusite of the matrix, and andalusite porphyroblasts are poikilitic about muscovite, biotite, quartz, oxides, and plagioclase, but not staurolite. Andalusite is itself being replaced by fibrolite, muscovite, and chlorite, and may have been late synkinematic.

Reaction relations of staurolite and especially muscovite pseudomorphs after staurolite have been dealt with by Albee (1972) and Guidotti (1970). Where pseudomorphs are of coarse muscovite without obvious retrogressive textures and where chemical equilibrium between phases can be shown, the pseudomorphs are likely prograde (Guidotti, 1970). Staurolite crystals in this study are not large and are surrounded by coarse and fine-grained muscovite generally without fibrolite. There is evidence for retrogression of garnets and kyanite, for instance, and most likely there is prograde and retrograde muscovite growth after staurolite, and perhaps more than one generation of staurolite. The following reactions are probably involved:

\[ \text{Na-mus + qtz = fib + albite + K-richer mus + H}_2\text{O} \]
\[ \text{staur + Na-mus + qtz = fib + biot + K-richer mus + albite + H}_2\text{O ± gnt} \]
\[ \text{low Zn-staur + mus + qtz = fib + biot + Zn-richer staur + H}_2\text{O ± gnt}. \]

Rock 304b is interpreted to have formed staurolite and upon further heating, to have crossed into sillimanite field and probably crossed the reaction staur + mus + qtz = gnt + sill + biot + H\text{\textsubscript{2}}O (figure 36). The fugacity of water was probably buffered by this reaction while performing solid solution substitution in the manner described in the above reactions. Upon lowering of pressure, probably related to contact metamorphism by the granitic stock, andalusite porphyroblasts crystallized and were partially replaced by fibrolite and later retrograde muscovite and chlorite. Ambi-
Figure 36. Generalized pressure-temperature diagram of path of sample 304b in the Blackpine Lake region, assuming an H₂O pressure of 6000 bars prior to the formation of post-kinematic andalusite.
guities resulting from alteration and lack of compositional data for coexisting phases precludes making more definite statements concerning equilibrium conditions.

Summary

Though chemical analyses of mineral phases were not done, reasonable estimates can be obtained for the conditions of metamorphism. The fluid phase has been shown to be $H_2O$-rich with no more than 10% $CO_2$ and probably no $CH_4$ as graphite is absent. Fluid compositions may have been relatively constant throughout the area. Given these conditions, temperatures varied from $510^\circ-530^\circ$ in the Chase Mountain area to nearly $650^\circ$ in the Blackpine Lake area. If $P_{H_2O} = P_{total}$, then melting would have begun to occur in place at slightly higher temperatures. Staurolite relations indicate temperatures up to $680^\circ$ and this may indicate that $P_{total}$ was greater than $P_{H_2O}$, or that other elements (Zn?) enhance the stability of staurolite. Later lower pressure contact (?) metamorphism in the area near the Blackpine Lake granitic stock is probably responsible for the crystallization of andalusite in some rocks. Retrograde metamorphism(s) has variably affected most rocks studied.

Granitic Rocks

As noted previously, pegmatites and small granitic sills and bodies are present throughout the Chase Mountain - Blackpine Lake area, and the quantity of granitic material bears a direct relation to the grade of regional metamorphism. In the Chase Mountain area, pegmatites form less than 1% of the rock volume and increase in quantity with lower structural level. In the Blackpine Lake area where sillimanite occurs, granitic rocks are very common and constitute greater than 10% of the rock volume.
and in places much more. Field evidence suggests that these granitic rocks were not melted in place, but represent material injected from elsewhere. The mineralogy of these rocks is simple with the following assemblages:

\[ \text{qtz + plag + mus + op + kf + gnt + biot + zircon}. \]

Most tabular granitic rocks are folded or flattened and nearly all have a deformational foliation defined by the preferred orientation of mica and the mortar, or ribbon structure of quartz which occurs as elongate flattened and recrystallized grains (figure 23b). Some, however, are cross-cutting and undeformed, showing only undulatory extinction in quartz grains. Most granitic rocks were emplaced synkinematically with probably F1 folds (figure 21). These rocks probably crystallized at pressures greater than about 3 kbars due to the coexistence of muscovite + quartz.

The Blackpine Lake granitic stock, as shown in figure 2, is a biotite quartz monzonite intrusion that contains no muscovite or garnet and is probably post-kinematic with respect to F2. It is semi-discordant and dome-shaped (Roots, 1954), slightly foliate, and ranges in composition from granodiorite to quartz monzonite. It is, however, deformed and contains abundant evidence of recrystallization of quartz grains. Elongate quartz grains usually have mortar structure and are separated from smaller, polygonal, strain-free recrystallized grains by sutured grain boundaries. Biotite crystals are variably bent and rotated into the foliation plane of the intrusion.

Roots (1954) considered both the pegmatites and the Blackpine Lake granitic stock as part of the Wolverine Metamorphic Complex and consequently related to the metamorphic and 'granitizing' activity. He also considered
the stock of muscovite-biotite granodiorite 5 km. northeast of Blackpine Lake (figure 2) to be similar to the Blackpine stock, contrary to the conclusions of this study. The evidence presented here and in Part II suggests that the foliated muscovite + garnet-bearing granitic rocks and pegmatites are distinct in age, mineralogy, intrusive relations, and origin from the coarse-grained Blackpine Lake granitic stock. Furthermore, it is suspected that the pegmatites and muscovite-bearing rocks are partially a product of anatexis at greater depths and bear a genetic relationship to the regional metamorphism of the country rocks.
PART II. Geochronology and the Eocene Resetting Event

INTRODUCTION

Throughout extensive regions of southern British Columbia, adjacent northeastern Washington, and northern and central Idaho, K-Ar dates on pre-Eocene igneous and metamorphic rocks have been reset to 45-55 m.y. Many of the metamorphic rocks are part of or correlative with the Shuswap Metamorphic Complex of southern British Columbia. Most workers attribute this to an Eocene thermal event accompanied by intense volcanism, plutonism, and hydrothermal alteration (Ross, 1974; Armstrong, 1974a; Miller and Engels, 1975; Medford, 1975). In many places Eocene volcanic rocks lie upon an erosion surface above granitic and gneissic rocks that in turn give Eocene apparent ages. Mathews (1976) has shown that apparent ages of gneisses decrease with increasing depth below this unconformity, leveling off at about 45 m.y. An additional explanation of Eocene resetting is rapid uplift and unroofing, and this has been invoked to explain Eocene dates in some areas of Idaho (Ferguson, 1975).

In north-central British Columbia, extensive Eocene resetting of K-Ar dates occurs in medium to high grade metamorphic rocks of the Wolverine Complex between latitudes 54°N and about 56°N. The Wolverine Complex (Armstrong, 1949), similar to the Shuswap Complex in many ways, is a belt of latest Precambrian (Winderemere) sediments that were regionally metamorphosed by middle Cretaceous time and perhaps earlier in the Mesozoic or Paleozoic. The explanation for the resetting of dates in these rocks is less obvious because volcanic rocks of Eocene age, where present, do not overlie the metamorphic complex.

The Eocene resetting of K-Ar dates in the Omineca Crystalline Belt
of the British Columbia cordillera is one manifestation of a profound disturbance during the early Tertiary, and it coincides with the abrupt tectonic transition from compressional continental arc tectonics to relative quiescence in the Oligocene and Miocene (Coney, 1972; Wheeler et al., 1972; Souther, 1976). This paper presents new Rb-Sr and K-Ar data and discusses both the Wolverine Complex and the Eocene 'event' in relation to the early Tertiary tectonic setting.
GEOLOGIC SETTING OF THE WOLVERINE COMPLEX

The Wolverine Complex and related lower grade rocks to the north (Figure 1) consist of latest Precambrian Windermere-equivalent strata (Mansy, 1972; Gabrielse, 1972) metamorphosed in middle Cretaceous or earlier time. Higher grade rocks of the Wolverine Complex are schists, quartzites, gneisses, amphibolites, and calc-silicate marbles intruded by granitic sills, dykes, small stocks, and pegmatites (Armstrong, 1949; Roots, 1954; Parrish, 1976). These medium to high grade rocks are involved in polyphase deformation including two generations of early synmetamorphic northwest-trending isoclinal folds and the development of nearly flat lying metamorphic foliation. These structures are overprinted by open to tight northwest-trending large scale folding and thrusting, principally southwest-directed (Mansy, 1971, 1972, 1974; Parrish, 1976). Later structures include brittle folds, joints, or faults, some of which are probably related to Eocene(?) faulting and deformation in the Rocky Mountain Trench region (Eisbacher, 1972).

Medium to high grade metamorphism accompanied the early phase(s) of tight to isoclinal folding which led to the development of the metamorphic foliation (Parrish, 1976). Kyanite, staurolite, and sillimanite bearing rocks are now exposed mainly in anticlinoria, indicating that this earlier metamorphic foliation and related gently dipping isogradic surfaces are probably folded by the later large scale northwest-trending folds, many of which are southwest vergent. A later post-kinematic prograde metamorphism has been noted by Mansy (1976) in some areas, and a still later retrograde metamorphism is present throughout extensive regions, especially near Chase Mountain.

The age of the high grade metamorphism and intense deformation is not well determined. Because medium grade rocks near and to the north and
northwest of Chase Mountain yield K-Ar ages of 87-128 my., (figure 37), this metamorphic period must be middle Cretaceous or earlier. Roots (1954) felt that this period of metamorphism was accomplished in post-Lower Cambrian, pre-Mississippian time because 1) Lower Cambrian rocks are involved and 2) sedimentary and volcanic rocks of the Lay Range of Mississippian to Permian (possibly including Triassic) age do not appear to be affected by pre-Mesozoic metamorphism (though no unconformity was observed). Further justification for this is the presence of granite, quartzite, and schist-bearing micaceous conglomerate in the Lay Range assemblage 20 miles west of Chase Mountain (Roots, 1954). A K-Ar determination on detrital white mica by the Geological Survey of Canada yields an age of 246 my. or Permo-Triassic (based on the revised time scale of Armstrong, 1974b). This conglomerate is areally very restricted and apparently unfossiliferous, but to date in this region is the only substantial evidence for pre-Mesozoic metamorphic and plutonic activity. Penetrative deformation of rocks of the Omineca Crystalline Belt occurred in pre-middle Cretaceous time because granitic rocks of the Early Cretaceous Cassiar batholith cut penetrative structures and metamorphic isograds (Gabrielse and Reesor, 1974).

In summary, it is known that many of the rocks of the Omineca Crystalline Belt are polymetamorphic and most are polydeformational. Thus Paleozoic activity in this area is possible but not proved; much of the deformation however, is probably Jurassic or Cretaceous in age based on stratigraphic, radiometric, and structural arguments. No unconformities separating lower Mesozoic or late Paleozoic rocks from metamorphic "basement" have been discovered, and where metamorphic and nonmetamorphic rocks are juxtaposed, several alternative explanations are possible and in fact documented in this type of metamorphic terrane (see Campbell, 1970; Coney, 1974; Davis, 1975).
Figure 37. Map of Wolverine Complex and related metamorphic rocks showing the geochronologic data, including dates of this study.
The Wolverine Complex is flanked to the northeast by the northern Rocky Mountain Trench which contains Upper Cretaceous to Eocene clastic sedimentary rocks (Sifton Basin) overlain by Quaternary glacial and post-glacial deposits (Eisbacher, 1974a). The origin of the Rocky Mountain Trench and its sediments has been discussed by Leech (1966) and Eisbacher (1972), and it is fairly clear that fanglomerates derived from the east and related sediments of the Sifton formation are the result of normal faulting very near the present Trench and the Dall Lake lineament (H. Gabrielse, personal communication, 1976). These rocks of Late Cretaceous and early Tertiary age have been affected by a late stage (probably Eocene) deformation involving kink folding, faulting, and possibly right lateral strike-slip faulting (Eisbacher, 1972).

Volcanic rocks of probably Eocene age are present in the Rocky Mountain Trench north of 56°, in part of the Dall Lake lineament, and as numerous felsic dikes in the Pelly lineament (H. Gabrielse, personal communication, 1976). Because their extent does not coincide with the area of Eocene dates, the significance of these volcanic rocks for resolution of the resetting problem is not clear.

The Rocky Mountains lie to the northeast of the Rocky Mountain Trench, and their stratigraphy and structure have been described by many workers (Bally, Gordy, and Stewart, 1966; Irish, 1970; Price and Mountjoy, 1970). The principle deformational period in the Rockies is thought to be Late Cretaceous and early Tertiary, but may extend back into the Early Cretaceous. Late Precambrian rocks correlative with those west of the Trench are present in the western ranges of the Rocky Mountains where metamorphic grade is very low. In addition, a small wedge of granitic gneiss with a kyanite-bearing aureole is present on the east side of the Trench near Fort Grahame (Gabrielse, 1971).
To the west of the Wolverine Complex and related Precambrian rocks lies a complex sequence of low grade upper Paleozoic and lower Mesozoic volcanic and sedimentary rocks of both "oceanic" and "volcanic arc" affinity (Monger, 1975; Richards, 1976) that have been intruded by a variety of igneous rocks ranging from ultramafic, both zoned Alaskan type (Irvine, 1975, 1976) and alpine type (Monger and Patterson, 1974; Monger, 1975) to gabbroic (Irvine, 1975, 1976) to granitic (Garnett, 1974; Gabrielse and Reesor, 1974; Woodsworth, 1976). The bulk of the granitic rock is Early Jurassic and late-Early Cretaceous in age. The Wolverine Complex is faulted against and in extensive areas thrust westward over these younger rocks (Mansy, 1974). The large scale southwest-vergent structures in the metamorphic terrane are probably associated with this faulting, but its age is not well known though probably Jurassic or Cretaceous. Recurrent movement during later times is also probable.

To the west of this late Paleozoic and early Mesozoic terrane lie clastic deposits of the Middle Jurassic to Early Cretaceous Bowser and Late Cretaceous to early Tertiary Sustut Basins. Both sequences have been folded and faulted and much of the deformation in the Sustut Basin involving compressional folding and thrusting is dated as early Tertiary (Eisbacher, 1974a).

The Wolverine Complex and related latest Precambrian rocks to the north (Ingenika group) thus form the westernmost exposures of Precambrian rocks believed to have been deposited along the cratonic continental margin (Gabrielse, 1972). These rocks are juxtaposed against younger upper Paleozoic and lower Mesozoic rocks to the west that record an entirely different set of structural and metamorphic conditions. The Wolverine Complex is part of the Cordilleran metamorphic core zone traceable southward into the Shuswap Complex, northeast Washington, parts of central and southeast Idaho, and northeast and eastern Nevada (Armstrong and Hansen, 1966; Snoke, 1975;
Coney, 1974). Though their histories are not identical, these metamorphic belts form a fundamental structural element of the Cordillera whose significance in plate tectonic terms is neither fully understood nor appreciated.
The Geological Survey of Canada has been the principle collector of K-Ar data in this region (Wanless et al, 1971, 1973, 1974), and these data are compiled in figure 37 and table IIa. These compilations clearly show the occurrence of Eocene dates clustering about 45-50 my. from 55° N to about 56° N, including some dates east of the Rocky Mountain Trench. Though data are incomplete, the evidence suggests that northwest from the Chase Mountain region, the apparent K-Ar ages abruptly increase from 45-55 my. to 87-128 my. Where mineral pairs have been analysed, the determinations are concordant with only a few exceptions differing by 10% or so. Eocene dates in the Wolverine Complex, however, are also relatively concordant but much younger. The coincidence of higher grade 'Wolverine' rocks with Eocene K-Ar ages at first would seem to suggest that the last important high grade metamorphism was during the Eocene; however, present evidence indicates that this is not the case. The key to the Eocene resetting problem may lie in the transition zone from young to older dates and in areas surrounding Eocene plutons.

Rb-Sr and K-Ar Data of this study

Near Chase Mountain a sample (no. 134) of muscovite-biotite schist yields a concordant K-Ar pair of 89 my. (biotite) and 92 my. (muscovite) as shown in table IIb. Less than 1 km. away a structurally discordant hornblende porphyry dike yields a hornblende K-Ar date of 42.6 my. This 42.6 my. date is identical to similar dated dikes in the Rocky Mountain Trench region (table IIa, samples GSC 73-55, 73-56). Rb-Sr analyses of muscovite, plagioclase, biotite, and whole rock from sample 134 yield comparable results (figure 38) and indicate that excess Ar retention is not a problem. Individual mineral isochron dates are discordant, and the biotite-plagioclase and biotite-
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<td>garnet</td>
<td>hornblende</td>
<td>64.8 ± 2</td>
<td>56°22'.1'</td>
<td>124°39'.1'</td>
</tr>
<tr>
<td>73-47</td>
<td>amphibolite</td>
<td>hornblende</td>
<td>53.8 ± 3</td>
<td>56°22'.1'</td>
<td>124°39'.1'</td>
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<tr>
<td>73-48</td>
<td>amphibolite</td>
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<td>40.5 ± 2</td>
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<td>124°39'.1'</td>
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<tr>
<td>73-49</td>
<td>amphibolite</td>
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<tr>
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<td>86.8 ± 3</td>
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<tr>
<td>73-52</td>
<td>granodiorite</td>
<td>muscovite</td>
<td>88.4 ± 4</td>
<td>58°00'.1'</td>
<td>126°48'.1'</td>
</tr>
<tr>
<td>73-53</td>
<td>qtz. monzonite</td>
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<td>94.9 ± 4</td>
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<td>125°47'.1'</td>
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<tr>
<td>73-54</td>
<td>qtz. monzonite</td>
<td>muscovite</td>
<td>104 ± 4</td>
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<td>49 ± 2</td>
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<tr>
<td>73-56</td>
<td>lamprophyre</td>
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<td>37 ± 2</td>
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### Table IIa (continued). Unpublished Geological Survey of Canada dates

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<th>age (m.y.)</th>
<th>latitude (N)</th>
<th>longitude (W)</th>
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<td>1</td>
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<td>47</td>
<td>56°32'</td>
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<td>schist</td>
<td>biotite</td>
<td>44</td>
<td>56°32'</td>
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<td>micaceous</td>
<td>muscovite</td>
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<td>biotite</td>
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### Table IIb. K-Ar data from this study

<table>
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<th>rock #</th>
<th>rock type</th>
<th>mineral</th>
<th>%K*</th>
<th>Ar (rad)/Ar (total)**</th>
<th>Ar 40 (rad)/K 40</th>
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</thead>
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<tr>
<td>134</td>
<td>schist</td>
<td>muscovite</td>
<td>8.01</td>
<td>0.915</td>
<td>5.528 x 10⁻³</td>
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<tr>
<td>134</td>
<td>schist</td>
<td>biotite</td>
<td>7.55</td>
<td>0.954</td>
<td>5.324 x 10⁻³</td>
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<tr>
<td>144</td>
<td>hornblende</td>
<td>hornblende</td>
<td>0.77</td>
<td>0.632</td>
<td>2.52 x 10⁻³</td>
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</tbody>
</table>

<table>
<thead>
<tr>
<th>rock #</th>
<th>apparent age (m.y.)</th>
<th>latitude (N)</th>
<th>longitude (W)</th>
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</thead>
<tbody>
<tr>
<td>134</td>
<td>muscovite 92.2 ± 3.3</td>
<td>56°32'</td>
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</tr>
<tr>
<td>134</td>
<td>biotite 88.9 ± 3.2</td>
<td>56°32'</td>
<td>125°18'</td>
</tr>
<tr>
<td>144</td>
<td>hornblende 42.6 ± 2</td>
<td>56°32'</td>
<td>125°19'</td>
</tr>
</tbody>
</table>

*K analyses performed by K. Scott of University of British Columbia

**Ar mass spectrometry performed by J. Harakal of University of British Columbia; \( \lambda K 40 \) to \( \lambda Ar 40 = 0.585 \times 10^{-10} \text{ yr}^{-1} \), \( \lambda \beta = 4.72 \times 10^{-10} \text{ yr}^{-1} \), \( \frac{\lambda K}{\lambda Ar} = 1.181 \times 10^{-4} \).
Figure 38. Rb-Sr evolution diagram for schist of sample 134.

Figure 39. Rb-Sr evolution diagram for "Wolverine" granitic rocks.
whole rock dates are anomalously younger (77 m.y.) than the K-Ar date (89 m.y.), and thus geologically unreasonable.* The muscovite-plagioclase-whole rock date is $106 \pm 6$ m.y. and confirms a Late Cretaceous or earlier age for the metamorphism.

The rocks near sample Vlie at the northwestern end of the Wolverine Complex, and they are metamorphosed to lower amphibolite facies but have suffered extensive retrogression of kyanite and garnet.

Farther south near Blackpine Lake (figure 37), K-Ar dates from two-mica granitic rocks are 47 and 43 m.y. respectively for a muscovite-biotite pair (table IIa; Wanless and others, 1971). Field investigation indicates that these muscovite + garnet-bearing granitic rocks are closely related to the regional metamorphism which in this specific area is sillimanite grade. Here, where the typical occurrence of Wolverine-type granitoid rocks interspersed with schists and gneisses is well-developed, Rb-Sr analysis from these muscovite-bearing granitic rocks yields a composite mineral isochron of $79 \pm 10$ m.y. and plagioclase-mineral isochrons of 52-84 m.y. (figure 39, table III) and are thus somewhat older than the K-Ar dates.

The area 10-15 km. south of Chase Mountain has been intruded by a small (8 sq. km.) stock of coarse-grained biotite quartz monzonite (figure 2), termed the Blackpine Lake granitic stock. This stock was originally mapped by Roots (1954) as part of the Wolverine Complex and yields a Rb-Sr whole rock isochron age of $62 \pm 7$ m.y. at $\text{Sr}^{87}_{86}$ of $0.7052 \pm 0.0002$ and a mineral isochron age of $44.7 \pm 2$ m.y. @$0.7058$ (figure 40, table III). This stock is deformed by past-crystallization deformation related to its diapiric (?) emplacement, but it is post-tectonic with respect to the earlier synmetamorphic

*In all 'Wolverine' Complex biotites analyzed in this study, this anomaly was encountered, and neither its cause nor its significance is understood. Similar results, however, have been encountered elsewhere in terrane of old metamorphic rock reset at a later date (Satir, 1974; R. L. Armstrong, personal communication, 1976).
### Table III. Rb-Sr analytical data

<table>
<thead>
<tr>
<th>sample #</th>
<th>rock/mineral</th>
<th>ppm Rb**</th>
<th>ppm Sr**</th>
<th>Rb 87/Sr 86</th>
<th>Sr 87* ± error</th>
</tr>
</thead>
<tbody>
<tr>
<td>134</td>
<td>schist</td>
<td>209</td>
<td>84</td>
<td>7.20</td>
<td>.7864</td>
</tr>
<tr>
<td>134</td>
<td>biotite</td>
<td>491</td>
<td>10.5</td>
<td>135.5</td>
<td>.9247</td>
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<tr>
<td>134</td>
<td>muscovite</td>
<td>252</td>
<td>85</td>
<td>8.58</td>
<td>.7897</td>
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<tr>
<td>134</td>
<td>plagioclase</td>
<td>34</td>
<td>233</td>
<td>0.42</td>
<td>.7767</td>
</tr>
<tr>
<td>300</td>
<td>schist</td>
<td>196</td>
<td>162</td>
<td>3.50</td>
<td>.7604</td>
</tr>
<tr>
<td>308</td>
<td>qtz. monzonite</td>
<td>140</td>
<td>254</td>
<td>1.60</td>
<td>.7113</td>
</tr>
<tr>
<td>308</td>
<td>muscovite</td>
<td>333</td>
<td>32</td>
<td>30.1</td>
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<tr>
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<td>biotite</td>
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<td>24.7</td>
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<td>308</td>
<td>orthoclase</td>
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<tr>
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<td>193</td>
<td>0.39</td>
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<td>muscovite</td>
<td>363</td>
<td>41</td>
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<tr>
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<td>granodiorite</td>
<td>123</td>
<td>446</td>
<td>0.80</td>
<td>.7098</td>
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<tr>
<td>324</td>
<td>aplite</td>
<td>254</td>
<td>63</td>
<td>11.7</td>
<td>.7367</td>
</tr>
</tbody>
</table>

**Blackpine Lake granitic stock**

|         | qtz. monzonite    | 204      | 307      | 1.92        | .7069          | .0001          |
|         | biotite           | 571      | 89       | 18.6        | .7180          | .0001          |
|         | orthoclase        | 552      | 402      | 3.97        | .7082          | .0001          |
|         | plagioclase       | 37       | 371      | 0.29        | .7060          | .0001          |
| 323     | qtz. monzonite    | 234      | 272      | 2.49        | .7073          | .0001          |
| 325     | granodiorite      | 148      | 317      | 1.35        | .7063          | .0001          |

*Sr 87/86 ratios have been adjusted to conform to the NBS standard SrCO₃ (987) actual value of .71022.

**1σ Rb and Sr concentrations are accurate to within 2% or 1 ppm., as determined by XRF.

\[ \lambda Rb \, 87 = 1.42 \times 10^{-11} \, \text{yr}^{-1} \]; \ Rb \, 87/Sr \, 86 = 2.894 \times \text{Rb/Sr}
**Figure 40.** Rb-Sr evolution diagram for Blackpine Lake granitic stock.

**Figure 41.** Rb-Sr diagram showing all whole rock determinations.
structures in the country rock (Parrish, 1976). Its foliation is not a primary igneous feature as thought by Roots (1954), and is best developed near contacts.

The whole rock Rb-Sr and $\frac{87}{86}$ Sr values of the syndeformational, syn-metamorphic muscovite-bearing granitic rocks do not define an isochron (figure 41), and it is not possible to date these igneous rocks with respect to the early tight to isoclinal folds that they are involved in; hence, the age of early structures is not known.

The Sr isotopic compositions of the country rock schists and gneisses were not systematically determined, but two determinations indicate that their ratios are much higher than the igneous rocks that intrude them (figure 41). This implies that the origin of the 'Wolverine' granitic rocks is both anatctic as well as magmatic and probably involves contamination of a less radiogenic source magma by highly radiogenic country rock.

The initial Sr $\frac{87}{86}$ ratio of the Blackpine Lake stock is .7052 ± .0002 and is much lower than both country rock schists (> .740) and muscovite-bearing 'Wolverine' granitic rocks (.710-.737). A different, less radiogenic magma is thus clearly distinguishable both in age, structural relations, and Sr isotopic composition from the synmetamorphic type.

Initial Sr isotopic compositions from the Hogem composite batholith to the west (figure 1) are very low, generally less than .704 (Eadie, 1976), and it is postulated that ancient Precambrian basement rock may extend beneath but essentially no farther west than the Wolverine Complex and related Precambrian metasediments. Obviously more information is needed to extend this boundary.
Discussion and Problems of Interpretation

Wolverine-type rocks are metasediments with granitic sills, dikes, and pegmatites present in abundance and with the metamorphic grade roughly corresponding to the sillimanite zone (see Part I). This rock association is now exposed in several anticlinoria throughout the entire length of the metamorphic belt from 54°N to 58°N where structural uplift is great. Though detailed structural and metamorphic data are lacking in most areas, regional studies (Mansy, 1976; Gabrielse, 1976) indicate that the high grade rocks throughout the belt have experienced a reasonably similar structural evolution to that near Chase Mountain, whereas in low grade areas 20 km. to the north of Chase Mountain, the structural history appears simpler and apparently does not involve early synmetamorphic isoclinal folding (Mansy, 1972, 1974).

These structural and metamorphic considerations demonstrate that similar rocks both north and south of about 56°N have experienced similar metamorphic and structural conditions, but they have very different geochronologic histories. The Eocene resetting event observed south of 56° cannot be responsible for the dominant metamorphic fabric of the higher grade rocks since rocks with the same fabric possess Early Cretaceous dates north of the reset region. The main period of metamorphism of these rocks must be late Early Cretaceous or older. This is supported by the presence of detrital mica and granite boulders yielding K-Ar dates of 95-117 m.y. in coarse clastic sediments found with the Sifton formation of Late Cretaceous to early Tertiary age in the Rocky Mountain Trench (Eisbacher, 1974a). When viewed in this structural-temporal context, the Wolverine Complex (Armstrong, 1949) is not unique because identical rocks are found in many anticlinoria where the structural uplift is sufficiently great (figure 1).
The problem of reset K-Ar dates is not easily explained. The explanations for reset ages in the Shuswap Complex of southern British Columbia (see Introduction) have serious shortcomings farther north. For instance, only a few scattered small Teriary intrusions into Wolverine-type rocks are known, and Eocene volcanics, though locally abundant in Rocky Mountain Trench, Dall Lake lineament, and Pelly lineament (H. Gabrielse, personal communication, 1976), are not widespread and bear no spatial correlation with K-Ar reset dates. These volcanics may have once been much more extensive, but this is not supported by present data.

Rapid block uplift cannot yet satisfactorily explain the observed dates because the sedimentary evidence of Eocene uplift and faulting present in fanglomerates in Rocky Mountain Trench and elsewhere does not bear a spatial relationship with the reset area. However, more data are needed to fully evaluate this statement.

Unusually high heat flow may also be appealed to; indeed, one might be tempted to relate such heat flow (with possible hydrothermal interactions) to the retrograde metamorphism observed throughout extensive areas. However, retrograde effects are present within both reset rocks and rocks preserving Cretaceous dates, and such high heat flow would have no other presently recognized manifestation except perhaps oxygen isotope variations. A further appeal can be made to the heat flux introduced upon the injection of various post-tectonic dikes and sills of feldspar and hornblende porphyry (Roots, 1954) and minette (Eisbacher, 1972) that are sporadically present in the metamorphic terrane as well as the Rocky Mountain Trench and Pelly lineament. But again, all dated dikes are within the 'older' terrane and indicate no appreciable affect of the dikes on the K-Ar ages of metamorphic rocks; and, in addition, nearly all determinations on dikes (table IIa) fall between 37 and 43 m.y. which is slightly younger and perhaps unrelated to Eocene resetting.
Several of these dikes were observed to occupy faults or joints which may be related to possible Eocene faulting in the Rocky Mountain Trench; similar relations have been observed by Eisbacher (1972) near Ware, British Columbia.

A workable explanation of the resetting phenomena in this region must take into account the following:

1) Pervasive resetting of K-Ar dates between 55°N and roughly 56°N during the Eocene;

2) Occurrence of young K-Ar dates in some metamorphic rocks directly east of the Rocky Mountain Trench as well;

3) Occurrence of older Rb-Sr mineral dates (52-84 m.y.) within the 45-50 m.y. K-Ar area near Blackpine Lake;

4) Absence of spatial relationship between Eocene volcanic rocks and reset metamorphic rocks;

5) Presence of sedimentary record of intense Eocene (?) uplift, but lack of spatial correlation with the reset area, given present data;

6) Occurrence of Early Cretaceous K-Ar dates in detrital rocks of the Sifton formation directly east of as well as farther north of the reset terrane;

7) Emplacement of dikes throughout the region in the Eocene, perhaps accompanied by faulting in the Rocky Mountain Trench;

8) Presence of a few small but important Eocene granitic stocks generally within the reset area (several more no doubt remain undiscovered).

Intrusion of small Eocene granitic plutons such as the Blackpine Lake stock accompanied by high heat flow, hydrothermal circulation, and perhaps oxygen isotope reequilibration is probably important in explaining the reset dates. The Blackpine Lake stock is likely responsible for the young K-Ar dates surrounding it and for the partial resetting of Rb-Sr dates. Faulting, doming, and rapid uplift may also be important factors, but the significance of these processes in explaining resetting of dates remains unclear and elusive. Based upon spatial arguments, block faulting, related (?) dike emplacement, and Eocene (?) volcanism are probably not related to the resetting event.
Rather, all of these events—deformation, uplift, erosion, volcanism, and intrusion may have been occurring simultaneously as part of an intense pulse of early Tertiary tectonism. The only clear correlation with the resetting of K-Ar and Rb-Sr mineral dates is the effect of small Eocene granitic intrusive rocks which have caused geochronologic disturbance in previously metamorphosed country rock.
Early Tertiary Tectonic Setting

In order to evaluate the significance of the Eocene K-Ar resetting event, it is necessary to view the event in terms of the early Tertiary tectonic setting in which volcanism, plutonism, deformation, and sedimentation are interwoven.

During the Eocene, volcanism was especially widespread and intense in British Columbia as shown in figure 42. In the Intermontane zone of central British Columbia, the Ootsa Lake volcanics occupy large regions from the Skeena Arch southward and are correlative with the Princeton and Marron volcanics of southern British Columbia and related volcanics in Washington and Idaho. These Eocene volcanics are related to small intrusive bodies of Eocene age in the Intermontane zone, the Katzberg intrusions (Richards, 1974), which extend without their volcanic equivalent into the eastern Bowser Basin and elsewhere including a few bodies in the metamorphic terrane. Eocene volcanic rocks are lacking in the Bowser Basin, but the Sloko volcanics lie farther to the north (Souther, 1972, 1976). Much larger bodies of intrusive rock are associated with the Sloko Group near the eastern margin of the Coast Plutonic Complex where many large discordant plutons yield Eocene radiometric dates. Though large regions of the central and eastern Coast Mountains have also been reset to young Eocene ages (Hutchinson, 1970), many of the ages on eastern border plutons are probably emplacement ages (Smith, 1975). During the intense volcanism and plutonism in the Intermontane zone and eastern Coast Plutonic Complex, metamorphic rocks of the Wolverine Complex were being reset.

Sedimentation during this igneous episode was confined mainly to the Sustut and Sifton basins where coarse clastic sediments were shed from high relief source terrane composed of granitic, metamorphic, sedimentary and
Figure 42. Sketch map illustrating tectonic events during the Eocene.
volcanic rocks. The present form of these basins is a result of both original basin extent as well as later erosional modification, and it is doubtful that the Sustut and Sifton basins were ever connected (Eisebacher, 1974a). It is likely that the Bowser and western part of the Sustut basins were source areas during the Eocene in addition to parts of the Omineca Crystalline Belt and the Rocky Mountains.

Deformation during the Eocene was considerable and extended across the Cordillera. The final stages of crustal shortening in the Rocky Mountains foothills took place in Paleocene and Eocene time (Bally, Gordy, and Stewart, 1966; Price and Mountjoy, 1970). Probable contemporaneous compressional deformation occurred in the Sifton and Sustut Basins deforming Upper Cretaceous, Paleocene(?), and Eocene strata by east-vergent folding and thrusting in the Sustut Basin and by kink folding and faulting in the Sifton Basin (Eisbacher, 1972, 1974a). Folds and thrusts present on the western part of the Upper Paleozoic and lower Mesozoic terrane that displace Permian, Upper Triassic, and Lower Jurassic rocks westward over Cretaceous and Tertiary sediments and older rocks (Richards, 1976) are probably of more than one age, and locally may be early Tertiary structures. Deformation of the Bowser and Sustut Basins continued from post-Lower Cretaceous to post-45 my. time controlling paleocurrent trends and sedimentation patterns (Eisbacher, 1974a). Early Tertiary uplift of the Coast Plutonic Complex has structurally impinged on the western Bowser Basin causing décollement style folding and thrusting (Eisbacher, 1974b).

Deformation, metamorphism, plutonism, and volcanism are widespread in the Mesozoic and early Tertiary history of the Cordillera but are especially intense in the Eocene. This Eocene episode was the concluding stage of an episodic, but consistently compressional regime extending from earlier in
the Mesozoic. This situation was followed by an abrupt transition to a quiescent tensional regime for much of the remaining Cenozoic (Wheeler et al., 1972), characterized by absence of acidic volcanism, plutonism, and folding, and by the presence of restricted block faulting, extrusion of basaltic plateau lavas, and intermittent sedimentation and peneplanation until the Pliocene rejuvenation.
Summary

In this study Rb-Sr whole rock isochron data show that a small Eocene plutonic body intrudes rocks grouped in the Wolverine Complex. Rb-Sr mineral isochrons on medium to high grade metamorphic and synmetamorphic granitoid rocks within the Eocene K-Ar reset terrane near the Blackpine Lake stock are partially reset to 52–84 my. from probable middle Cretaceous or older original dates. Metamorphic rocks to the north that are structurally continuous with the Wolverine Complex and with the Eocene reset province have K-Ar and Rb-Sr mineral ages of 89–106 my. indicating that the medium to high grade metamorphism and its accompanying structural disturbance are middle Cretaceous or older in age.

Although much work needs to be done to fully explain the phenomena, the Eocene resetting event is likely the result of both the intrusion of small bodies of granitic rock and the accompanying high heat flow. Faulting and uplift may be responsible for some of the pattern in K-Ar date distribution, but their importance is difficult to evaluate because of lack of detailed data in surrounding regions.

The Eocene event was accompanied by intense volcanism, plutonism, and deformation in other areas of northern and central British Columbia. It coincides with the cessation of compressional tectonics in the Cordillera and is abruptly followed by a tectonically quiescent regime until the Pliocene rejuvenation which has resulted in much of the present physiography.


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