# CRUSTAL EXTENSION AND SUBSEQUENT CRUSTAL THICKENING ALONG THE CORDILLERAN RIFTED MARGIN OF ANCESTRAL NORTH AMERICA, WESTERN PURCELL MOUNTAINS, SOUTHEASTERN BRITISH COLUMBIA

by

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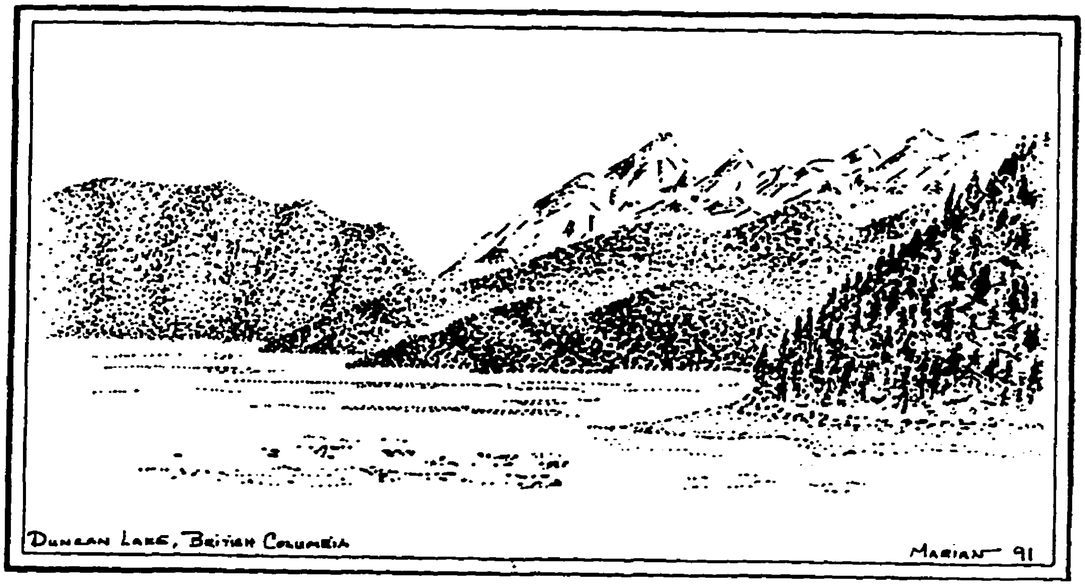


*In memory of*

**John Carter**

*of Nelson, B. C.,*

*whose hospitalitiy and assistance was precious,*

*and whose passion for the little-known places of the*

*Purcell and Selkirk Mountains was completely infectious*

# ABSTRACT

Geological mapping across a conspicuous zone of stratigraphic and structural contrasts in the west-central Purcell Mountains of southeastern British Columbia provides the basis for new insight into the tectonic evolution of a segment of the former North American rifted continental margin.

The Neoproterozoic Windermere Supergroup of the west-central Purcell Mountains is exposed in two contrasting stratigraphic domains that resulted from syn-depositional normal faulting: a thin southeastern succession of laterally continuous units, and a thick, laterally variable northwestern succcession. Strata in both domains comprise distinct lower and upper clastic sequences, separated by a regional carbonate-bearing marker unit and possibly significant unconformity at the base of the upper clastic sequence. The development of the Windermere basin and the deposition of the lower clastic sequence were controlled both locally and regionally by two sets of extensional faults (NNW- and NE-trending) that were active during an episode of significant crustal attenuation associated with intracontinental rifting. The upper clastic sequence was deposited during local normal faulting and perhaps regional uplift of the basin margins, in a tectonic setting that remains enigmatic.

The Neoproterozoic to Lower Cambrian Hamill Group is divisible into several distinct lithostratigraphic units. The uppermost shallow marine quartzite unit is continuous and rests unconformably on the lower fluvial or marine units, or directly on the Horsethief Creek Group. Lateral stratigraphic variations indicate that the distinct stratigraphic successions of lower Hamill Group in the Kootenay Arc and in the Purcell anticlinorium were deposited in separate half-grabens that were bounded to the east by normal faults. The unconformity beneath the upper Hamill Group in the west-central Purcell Mountains is a regional unconformity in the Hamill and Gog Groups of the southern Canadian Cordillera. It is the stratigraphic expression of the transition from continental rift to continental drift and thermal subsidence of a passive continental margin that occurred in Early Cambrian time between 549 Ma and before about 520 Ma.

Significant east-west changes in the thickness and mechanical properties of the Neoproterozoic to Early Paleozoic continental margin of Laurentia controlled mechanisms of crustal thickening during Mesozoic convergence. The Kootenay Arc boundary fault is a regionally significant thrust fault that separates rocks of the Purcell anticlinorium from rocks of the Kootenay Arc and coincides with a regional zone of structural divergence. It also coincides with a pre-exisitng Neoproterozoic to Early Paleozoic normal fault that marks perhaps the most significant "hinge zone" in the former rifted margin. Deformation of the supracrustal rocks that were deposited on the former North American continental margin can be divided into two distinct stages that reflect the nature of the lithosphere that underlay them on either side of this "hinge zone." Phase 1 involved short-lived development of a metamorphic hinterland, characterized by west-verging ductile deformation above an eastward-propagating tectonic wedge of North American strata, rapid and large horizontal shortening, and as much as 30 km of post-metamorphic exhumation in response to the abrupt increase in basal décollement gradient at the "hinge zone". The propagation of the Cordilleran basal décollement across the "hinge zone" is recorded by the initial development of the foreland basin, by uplift and cooling of the metamorphic hinterland and by a transition to thin-skinned east-verging deformation.

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# GENERAL INTRODUCTION

## PURPOSE AND PROBLEMS

Most of the world's Phanerozoic mountain belts are the result of plate convergence along former rifted continental margins. Deformed continental margins expose a record of thermal and mechanical processes from various depths below the earth's surface that were related to both divergent and convergent plate margins. Thus they provide insight into fundamental plate tectonic processes.

The purpose of this thesis is to understand and to illustrate the tectonic evolution of a segment of the outer edge of the former Cordilleran rifted margin of the North American continental plate. The west-central Purcell Mountains of southeastern British Columbia (Fig. 1-1) are well-suited to this purpose because they expose strata that record, over a long period of geologic time, both divergent and convergent phases of the evolution of this margin. Mesoproterozoic to Lower Paleozoic rocks were deposited on the western margin of Laurentia (the continental ancestor of the North American craton) during several phases of continental crustal attenuation and subsidence associated both with active rifting and with passive "drift" along this margin. These strata were subsequently deformed, regionally metamorphosed and intruded by granitoid plutons during Mesozoic accretion of terranes exotic to North America.

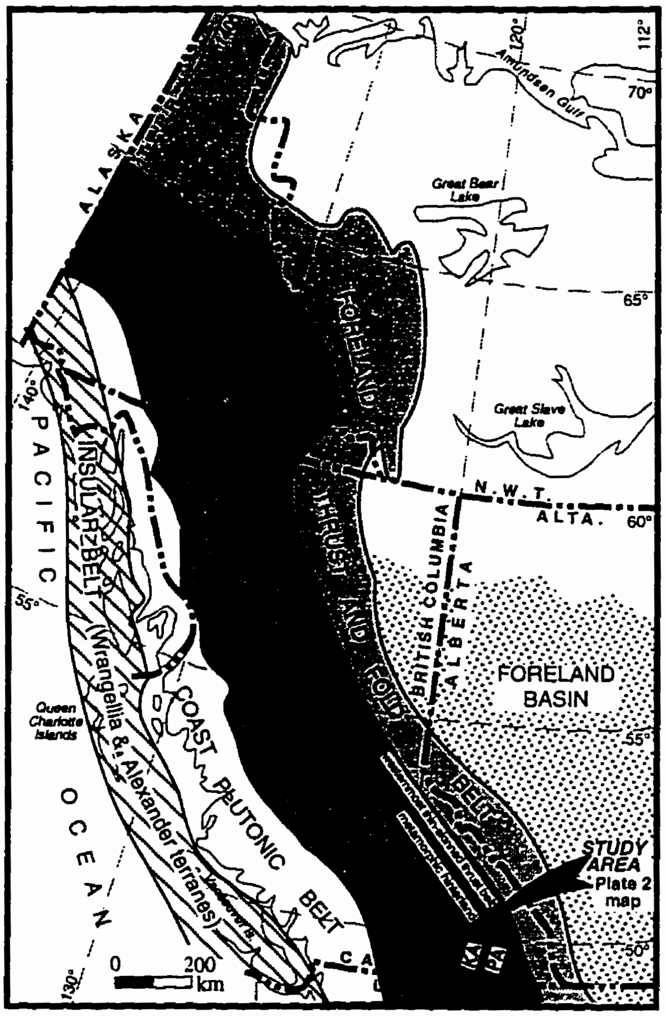


**Figure 1-1:** Morphogeological belts of the Canadian Cordillera (after Wheeler and McFeeley, 1991), showing the location of the study area. The Omineca belt comprises rocks of both ancestral North America and exotic terranes that were regionally metamorphosed during Mesozoic terrane accretion.

Four distinct sedimentary successions were deposited on this segment of the western margin of Laurentia that is now exposed in the west-central Purcell Mountains. These include:

1) The Mesoproterozoic Purcell Supergroup, which comprises fine-grained siliciclastic and carbonate rocks that accumulated in a deep intracontinental basin
2) The Neoproterozoic Windermere Supergroup, which comprises coarser, immature siliciclastic and carbonate rocks that accumulated during the continental rifting that initially defined the western margin of Laurentia
3) The uppermost Neoproterozoic to Lower Cambrian Hamill Group, which comprises coarse, quartz-rich siliciclastic rocks that were deposited at the base of a wedge of sediments that accumulated during subsidence of the young continental margin
4) Contrasting successions of pre-Mississippian Lower Paleozoic clastic and carbonate strata that, to the east, comprise predominantly shallow-water facies and are separated by unconformities, and, to the west, are immature, thick and deeply subsided

The deposition of the Mesoproterozoic Purcell Supergroup is not related to the birth and evolution of the Cordilleran rifted margin and thus it is not a focus of this thesis. The Windermere Supergroup and Hamill Group were related, at least in part, to two distinct episodes of continental extension that are recorded along the entire length of the North American Cordilleran margin. These two successions are the focus of the investigation of the extensional history of the margin in this thesis. The overlying Lower Paleozoic strata were deposited on the margin following the extensional episodes that initiated it, in a tectonic setting or settings that remain enigmatic and are not addressed directly in this thesis. Nonetheless, their distribution and the structures that controlled it were profoundly influenced by the previous configuration of the rifted margin, and these structures also played a significant role in the subsequent Mesozoic tectonic evolution of this margin.

These deformed Mesoproterozoic to Lower Paleozoic strata are exposed in the west-central Purcell Mountains in two distinct tectonostratigraphic domains, that both lie in the southern Omineca belt of the Canadian Cordillera (Fig. 1-1). The domains are the product both of Proterozoic and Paleozoic extension and subsidence, and of contrasting styles of Mesozoic compressional deformation. The eastern domain is the Purcell anticlinorium, which comprises the westernmost, weakly metamorphosed part of the foreland thrust and fold belt that is exposed primarily in the Canadian Rocky Mountains and Foothills to the east. The Purcell anticlinorium exposes thick Proterozoic strata and thin, discontinuous Lower Paleozoic strata. The western domain is the Kootenay Arc, which comprises the eastern part of the metamorphic "hinterland" of the deformed margin and contains the Mesozoic suture between North America and Intermontane Superterrane. It exposes more complexly deformed, and formerly hotter and more plastic rocks. The Kootenay Arc comprises much thicker, deeper-water Lower Paleozoic strata and virtually no Proterozoic strata.

Many aspects of the plate tectonic setting(s) and tectonic evolution of this deformed margin remain puzzling and controversial. This study in the west-central Purcell Mountains was undertaken in order to address the following fundamental questions about the evolution of the North American Cordilleran rifted margin:

1. What does the distribution of lithofacies within Neoproterozoic, uppermost Neoproterozoic to Lowermost Cambrian and Lower Paleozoic sedimentary strata deposited on this continental margin reveal about the structures that controlled the sedimentary basins in which they were deposited and about the nature of the underlying crust during each time period?

2. When did the transition from active continental extension (rifting) to passive, thermally-driven subsidence (dirfting) occur? How and where is this rift-to-drift transition recorded in the stratigraphic record?

3. What is the nature of the structural transition from refolded, west-verging ductile deformation of the metamorphic hinterland (Kootenay Arc) to east-verging, thin-skinned deformation of the western foreland thrust and fold belt (Purcell anticlinorium)?

4. What are the timing and conditions of deformation in both domains and how is their tectonic evolution linked?

5. How did the crustal configuration that was established during Neoproterozoic and Early Paleozoic extension affect mechanisms of crustal thickening during Mesozoic plate convergence?

## APPROACH

The foundation of this thesis is a new 1:75, 000 geological map (Warren, 1996b; Plates 1-3, in pocket) of the west-central Purcell Mountains of southeastern British Columbia. The map includes new 1:50,000 mapping of an approximately 1000 $km^2$ area, which was integrated in the field and on the map with previous geological mapping from adjacent areas to the east and the west. The scope of the interpretations presented in this thesis is two-fold: 1) new stratigraphic and structural relationships and interpretations from the west-central Purcell Mountains are supported primarily by field data from this map area; 2) the regional extent and implications of these relationships are examined by regional synthesis of these new and previously published field data from the southeastern Canadian Cordillera. Both local and regional interpretations are supported by new and published sedimentological, biostratigraphic, geochronological, thermobarometric and geochemical data. [Note: local place names in the text and figures are located on the geological map (Plate 1), and regional place names are located on Fig. 1-2, at the end of this introduction]

The geologic map provides the basis for three self-contained chapters that describe three distinct episodes during the tectonic evolution of this segment of the Cordilleran rifted margin. Each forms the basis of a manuscript to be condensed and submitted for journal publication.

- Chapter 2 describes the stratigraphy of the Neoproterozoic Windermere Supergroup (the basal Toby Formation and the overlying Horsethief Creek Group), with specific focus on rapid lateral thickness and lithofacies variations within it and within the underlying strata on which it rests. These stratigraphic variations form the basis of a discussion of the local and regional configuration and evolution of the fault-bounded intracontinental basin in which the Windermere Supergroup was deposited.

- Chapter 3 describes the stratigraphy of the uppermost Neoproterozoic to lowermost Cambrian Hamill Group, with specific focus on the identification of a regionally significant unconformity within the Hamill Group. A regional comparison of the distribution of strata above and below the unconformity is the basis of a discussion of the configuration and distribution of rift basins in which the lower Hamill Group and equivalent Gog Group strata in the Rocky Mountains were deposited, and of the subsequent transition to a continental shelf setting on a thermally-subsiding passive continental margin.

- Chapter 4 describes the structural, metamorphic and igneous relationships that are a product of crustal thickening during Mesozoic terrane accretion to this segment of the North American rifted margin. The specific focus is on elucidation of the kinematic link between the structure and evolution of the metamorphic hinterland and the structure and evolution of the thin-skinned foreland thrust and fold belt to the east. This chapter also incorporates the conclusions of the previous two chapters and examines the influence of the previous Proterozoic and Paleozoic tectonic history on the Mesozoic tectonic evolution. The pre-

Mesozoic configuration and Mesozoic tectonic evolution of this area are summarized in a restored regional cross-section presented in Plate 4 (in pocket).

Chapter 5 summarizes the major conclusions presented in each of the preceding chapters. Appendices 1 through 3 present suplementary new $^{40}Ar/^{39}Ar$ geochronological, thermobarometric, and geochemical data that are discussed in the body of the thesis. These appendices also describe the methods used to collect and analyse these data. Appendix 4 presents the methods and assumptions used in palinspastic restoration of the regional cross section. Appendix 5 comprises digital photographic plates that accompany descriptions in Chapters 2-4. Reviews of literature relevant to the problems addressed in this thesis are presented in the introductions and discussions of each of the chapters and are not presented here, in order to avoid repetition.

## ORIGINAL CONTRIBUTIONS

- ***Geological mapping:*** New 1:50,000 scale geological mapping of a 1000 km$^2$ section of the western Purcell anticlinorium and easternmost Kootenay Arc constitutes a substantial contribution to field data in the southeastern Canadian Cordilllera. Significant contributions from the mapping include: stratigraphic subdivision of the previously undivided Horsethief Creek Group into 16 lithostratigraphic map units, subdivision of the previously undivided Hamill Group into 5 lithostratigraphic units, and documentation of a regionally significant fault zone (the Kootenay Arc boundary fault) at the sharp structural transition between this segment of the Kootenay Arc and the Purcell anticlinorium. An up-to-date compilation map (Warren, 1996) from this work and from previous mapping to the east and west illustrates in detail the nature of the stratigraphic and structural contrasts between the eastern Purcell anticlinorium, the western Purcell anticlnorium and the Kootenay Arc. The map and the regional cross-section constructed from it represent the first well-constrained transect through the entire Purcell anticlinorium and Kootenay Arc.

- ***Windermere rift basin geometry:*** Lateral thickness and facies changes revealed by stratigraphic subdivision of the Horsethief Creek Group and compilation of previously published data detailing stratigraphic variations beneath the Toby Formation indicate that the distribution of the Windermere Supergroup lithofacies was controlled by two sets of syn-depositional normal faults that trend north-south and northeast-southwest.

- ***Hamill rift basin geometry and identification of a rift-to-drift transition:*** Stratigraphic subdivision of the Hamill Group reveals a regionally significant unconformity within quartz-rich sedimentary strata of the Hamill Group. The unconformity can be identified throughout the southeastern Canadian Cordillera. This unconformity represents the latest Neoproterozoic to Earliest Cambrian rift-to-drift transition. The Neoproterozoic to Lowermost Cambrian clastic and volcanic strata below it record sedimentation in north-trending, discontinuous, fault-bounded extensional basins. The Lower Cambrian clastic and carbonate strata above the unconformity record more uniform subsidence driven by thermal contraction on a newly-formed passive continental margin.

- ***Palinspastically restored regional paleogeographic maps:*** Compilation and re-interpretation of a large body of previously published data from the southeastern Canadian Cordillera show that the configurations and tectonic evolution of the Windermere Supergroup and Hamill Group basins that have been documented in the west-central Purcell Mountains are also valid on a much larger, more regional scale. New and previous data were used to reconstruct the three-dimensional configurations of these basins on palinspastically restored, pre-Mesozoic maps of the southeastern Canadian Cordillera. The maps show the distribution and thicknesses of the Windermere and Hamill rift basins and illustrate the paleogeography preceding and immediately following the rift-to-drift transition. These maps are a significant contribution because: 1) they illustrate for the first time fundamental differences in basin configuration and probable upper crustal behavior between Windermere time and Hamill time; 2) they are the first isopach maps of Lower Cambrian strata that show strata above and below the rift-drift unconformity separately, and 3) their palinspastically restored bases assure a more realistic conceptualisation of the rift basins, the early passive margin and the crust that underlay them.

- ***New geothermobarometric and $^{40}Ar/^{39}Ar$ geochronological data:*** These analytical data provide new constraints on the thermal history and tectonic burial history of this segment of the western Purcell anticlinorium and eastern Kootenay Arc. These data are compatible with previous data collected along strike to the south and to the north and thus link and strengthen previous interpretations of the syn-metamorphic burial, uplift and structural history of this segment of the continental margin. Geobarometric data also contribute entirely new insight into the evolution of the structural contrasts between the Kootenay Arc and the Purcell anticlinorium across the Kootenay Arc boundary fault.

- ***New insight into the controversial age of deformation in the Kootenay Arc:*** Field relationships, re-interpretation of previously published structural data and recently published geochronologic data indicate that the west-verging structures in the Kootenay Arc are Middle Jurassic and not mid-Paleozoic. This interpretation has profound implications for discussions of the mechanisms of terrane accretion and crustal thickening addressed in this thesis, as well as for discussions of the mid-Paleozoic tectonic setting of the Cordilleran margin.

- ***First palinspastic restoration of the Kootenay Arc:*** A regional cross section that is restored in several steps shows the Mesozoic structural evolution of the Kootenay Arc and the Purcell anticlinorium. It provides new insight into the evolution of the complex, west-verging structures in the Kootenay Arc and into their link with east-verging structures of the Purcell anticlinorium. It provides the first constraints on amounts and rates of shortening for this segment of the Cordillera, which can be compared with previously published data from the foreland thrust and fold belt to the east. The restoration also shows that this technique need not be confined to thrust and fold belts and that it can be applied meaningfully, in conjunction with other boundary conditions supplied by geochronology and thermobarometry, to rocks that have undergone previous extension and poly-phase, ductile deformation. Major uncertainties and assumptions are involved in palinspastically restoring these hinterland strata. However, the result is still a better-constrained and more realistic interpretation of the tectonic evolution than those presented in tectonic "cartoons" that are not drawn to scale or that ignore critical field relationships.

- ***New tectonic model:*** The tectonic model presented in Chapter 4 incorporates the new contributions outlined above and builds upon previously published interpretations and models. The model contributes a new illustration of how Proterozoic and Paleozoic crustal extension affected the mechanisms of Mesozoic horizontal crustal shortening and vertical crustal thickening.

Many other orogenic belts that are developed on previously rifted continental margins also display a similar, sharp transition and reversal in structural vergence from a metamorphic hinterland to a thin-skinned foreland thrust and fold belt. The original contributions of this thesis not only provide significant insight into the evolution of this segment of the Canadian Cordillera but also provide the basis for comparison with the extensional and compressional evolution of other deformed continental margins.

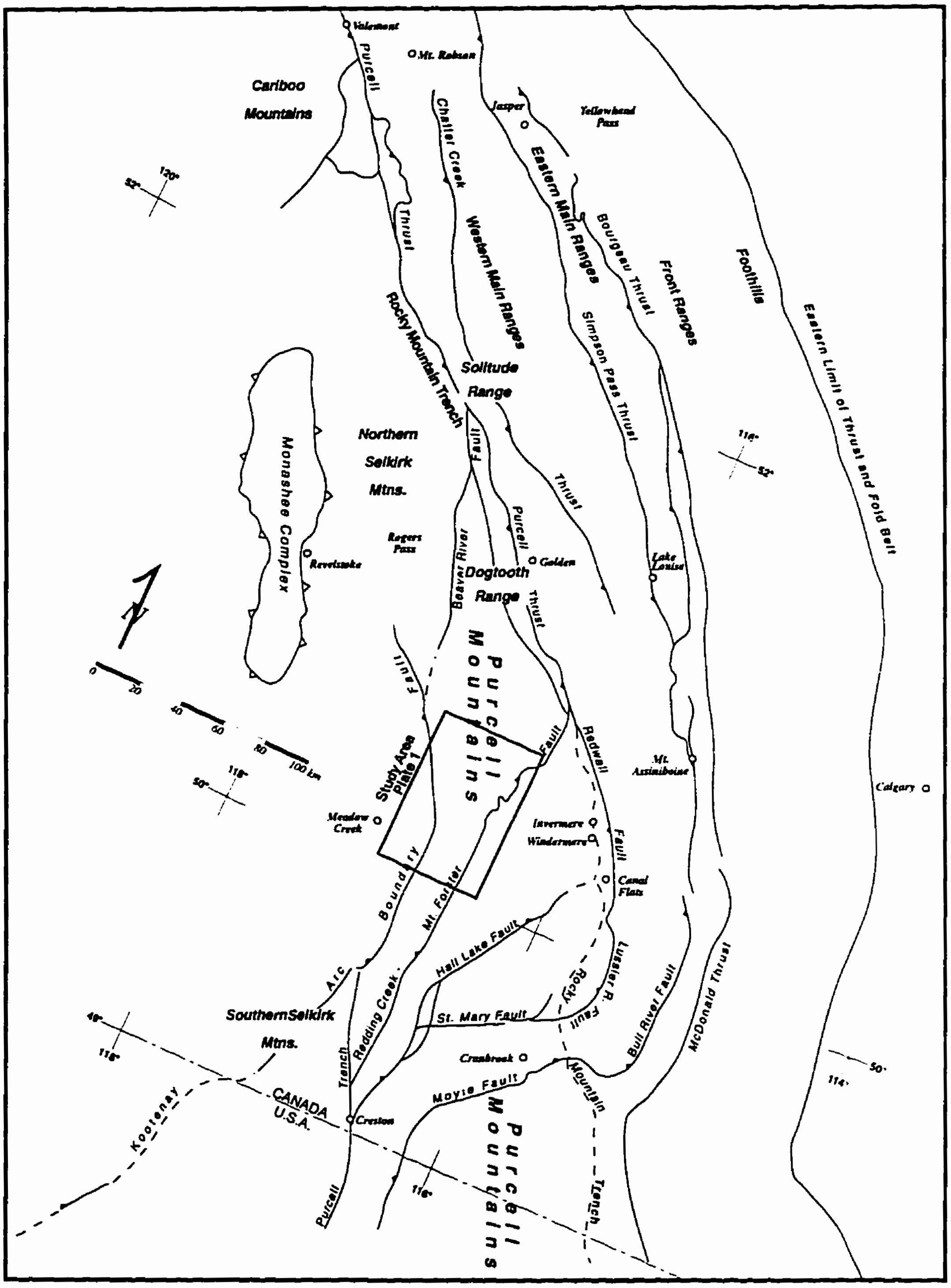

**Figure 1-2:** Simplified map of the southern Canadian Cordillera, showing major structures and regional geograhpic names given in Figs. 2-2 and 3-2, and throughout the text. Faults modified after Wheeler and McFeeley (1991). See also Fig. 2-7 and Plate 1 for place names in the central Purcell Mountains

# Chapter 2

# THE BIRTH OF THE SOUTHERN CANADIAN CORDILLERA: CONFIGURATION AND EVOLUTION OF THE WINDERMERE BASIN DURING NEOPROTEROZOIC RIFTING

## ABSTRACT

The distribution of sediment in the Neoproterozoic Windermere Supergroup of the Purcell anticlinorium, southeastern British Columbia, reflects both local and regional structural controls on the evolution of the basin that was the foundation of the western margin of Laurentia, and, furthermore, it records two distinct episodes of regional uplift of the basin margins.

The Windermere Supergroup of the northern Purcell anticlinorium comprises the glaciomarine Toby Formation, overlain by distinct lower and upper submarine fan clastic sequences of the Horsethief Creek Group. The upper clastic sequence unconformably overlies a finer siliciclastic and carbonate marker unit at the top of the lower clastic sequence. Conspicuous thickness and lithofacies changes in both the Toby Formation and overlying Horsethief Creek Group, particularly in the lower clastic sequence, correspond to syn-sedimentary normal faults that cut the underlying Purcell Supergroup. These faults were parallel to larger-scale faults that defined both the north-northwest-trending margin of the basin to the east and the more abrupt, northeast-trending margin of the basin to the south. The configuration of the basin indicates an abrupt transition from thick to highly attenuated underlying crust, particularly from southeast to northwest.

Two distinct episodes of normal faulting and mafic volcanism in the basin reflect distinct pulses of tectonic uplift and relative sea level fall on the basin margins. The first episode is

recorded by the Toby Formation and the lower clastic sequence and is related to significant crustal attenuation during the initiation of intracontinental rifting. The emergence of uplifted cratonic basement to the east and uplifted Mesoproterozoic Purcell Supergroup along major faults to the southeast is recorded by differences in the provenance of sediment in the Horsethief Creek Group that is exposed in the northern and southwestern Purcell anticlinorium, respectively. The sediment was transported to the basin and trapped by local, fault-controlled topography.

The second episode of faulting immediately followed a regional relative highstand event (the carbonate/fine clastic marker unit) and was recorded by the deposition of coarse, craton-derived sediment of the upper clastic sequence and by normal faults that cut the lower clastic sequence. The regional tectonic setting of the basin at this time is uncertain, but the very large volume (up to several km thick and 300 km restored width from 50° to at least 62°N latitude) of continuous, coarse, immature sediment that characterizes much of the upper part of the Windermere Supergroup is more compatible with relative sea level fall due to regional uplift on a margin of uncertain tectonic setting, rather than eustatic sea level fall only.

Recently published geochronologic, biostratigraphic and chemostratigraphic data from the Windermere Supergroup indicate that this apparently rapidly-deposited succession was deposited over a span of 200 million years. Thus the possibility of at least one significant, previously unrecognized hiatus within the Windermere Supergroup must be considered.

## INTRODUCTION AND PREVIOUS WORK

The Neoproterozoic Windermere Supergroup represents a critical interval in the geologic record of the Canadian Cordillera because it records the birth of the western margin of Laurentia. The sedimentary basin that was established during the deposition of the Windermere Supergroup profoundly influenced not only the sedimentary history of the ensuing continental passive margin throughout Paleozoic time but also its structural evolution during Mesozoic time. However, little previous work has addressed the initial configuration of the Windermere rift basin or basins and the thermal and mechanical properties of the underlying lithosphere that are recorded by this configuration. Insights into these properties are critical not only to further understanding of the birth of continental margins in general but also toward resolving recent controversy concerning the plate tectonic setting of the Windermere Supergroup. Thus the purpose of this paper is to use new and previously published stratigraphic data to reconstruct and to discuss the tectonic significance of the three-dimensional configuration of a well-exposed part of the Windermere basin in the southern Canadian Cordillera.

Neoproterozoic rocks correlative with the Windermere Supergroup extend from Alaska to Mexico (e.g. Stewart, 1972). Numerous previous workers had established that the Windermere Supergroup, up to 10 km thick, was deposited in a rift basin or basins during an episode of crustal extension that began at about 780-750 Ma (Armstrong et al., 1982; Roots, 1983; Devlin et al., 1985; Jefferson and Parrish, 1989) and resulted in subsequent Neoproterozoic or Early Cambrian birth of the Cordilleran Laurentian margin (e.g. Aitken, 1969; Stewart, 1972; Lis and Price, 1976; Eisbacher, 1981; Pell and Simony, 1987). However, the timing of the transition from rifting and continental extension to separation and continental drift on the western margin of Laurentia remains controversial (Bond and Kominz, 1984; Devlin and Bond, 1988; Ross and Murphy, 1988; Ross, 1991b; Dalrymple and Narbonne, 1996). Several recent models and plate tectonic reconstructions show that within the Neoproterozoic supercontinent "Rodinia," the rifted counterpart to the western margin of Laurentia was the eastern margin of Australia and adjacent Antarctica (Dalziel, 1991, 1992; Hoffman, 1991; Moores, 1991). More recent reconstructions include or cite paleomagnetic, lithostratigraphic and biostratigraphic evidence for separation of these continental masses during the Neoproterozoic breakup of "Rodinia" between about 750 and 720 Ma (Powell et al., 1993; Powell et al., 1994; Storey, 1993; Pelechaty et al., 1996; Torsvik et al., 1996). However, late Neoproterozoic to Early Cambrian normal faulting and volcanism, a widespread sub-Cambrian unconformity or unconformities and interpretations of the Early Paleozoic subsidence history of the western margin of Laurentia imply that continental separation, the initiation of seafloor spreading and thermal subsidence did not occur until latest

Neoproterozoic to Early Cambrian time (Aitken, 1969; Bond and Kominz, 1984; Bond et al., 1984; Devlin, 1988; Devlin and Bond, 1989; Lickorish and Simony, 1995; Warren; 1996; see Chapter 3). Some authors have proposed a reconciliation of these apparently conflicting sets of data by suggesting that Neoproterozoic to Lower Cambrian strata record two distinct and complete cycles of rift and drift (e. g. Ross, 1991b), but a rifted counterpart for the Early Cambrian event has not been identified, and the evidence for Neoproterozoic passive margin remains equivocal, particularly in the southern Canadian Cordillera.

The Windermere Supergroup is widely exposed in the southern Canadian Cordillera in the Cariboo Mountains, the northern and southern Selkirk Mountains, the Main Ranges of the Rocky Mountains and the Purcell Mountains, where it was established by Walker (1926) near Windermere, B. C. (Fig. 2-1). It primarily comprises immature siliciclastic rocks but contains minor carbonate strata. It is distinguished in the southern Canadian Cordillera from underlying and overlying siliciclastic strata by its thick intervals of grit[1], pebble conglomerate and coarse arkosic wacke. Several regional lithostratigraphic correlations that are the basis for tectonic interpretations have been proposed for the southern Canadian Cordillera (e.g. Brown et al., 1978; Mansy and Gabrielse, 1978; Pell and Simony, 1987), but exposure in separate fault panels, differences in metamorphic grade, lateral stratigraphic variations and lack of biostratigraphic data render these correlations unreliable.

Significant steps have been made toward regional correlation of these laterally discontinuous and largely unfossiliferous strata, as well toward elucidating the stratigraphic and structural relationships that record the initiation of the Windermere Supergroup basin. Ross and Murphy (1988) made use of a regionally extensive, fine clastic and carbonate marker unit (Old

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[1] "Grit" is an informal lithological name commonly used in the literature of the Canadian Cordillera and, particularly, of the Windermere Supergroup. This term refers to immature granule or small pebble conglomerates that contain abundant feldspar and are typically poorly sorted with a pelitic or wacke matrix. In parts of this thesis I distinguish qualitatively between arkosic grit (equivalent to "grit" as defined above), quartzose grit (more abundant clast and matrix quartz relative to feldspar and pelite) and calcareous grit (primarily quartz and carbonate grains in a calcareous matrix).

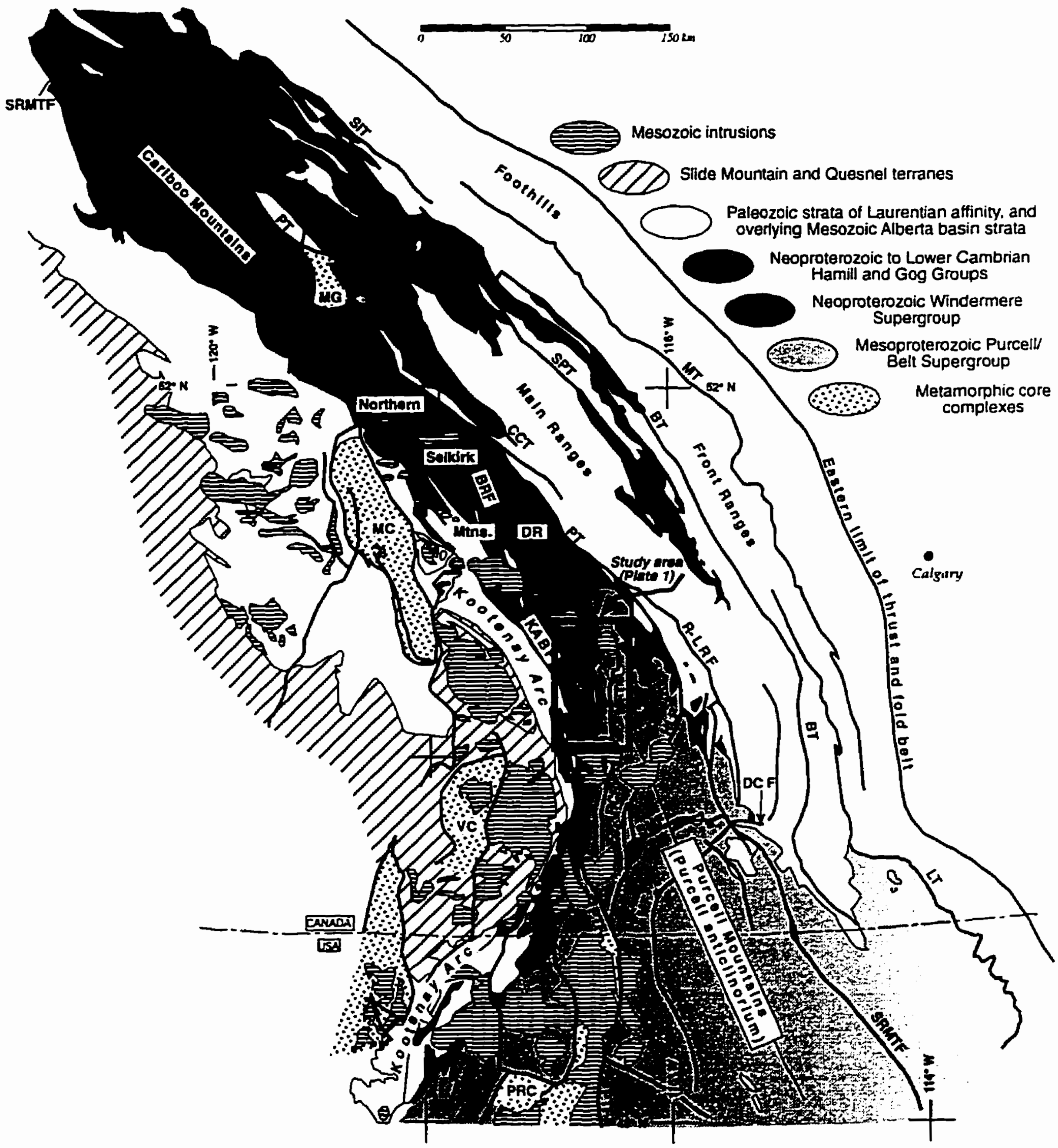



**Figure 2-1:** Simplified geological map of the southeastern Canadian Cordillera, showing regional distribution of exposures of the Windermere Supergroup, and major structures and physiographic provinces discussed in the text. After Wheeler and McFeeley (1991). Abbreviations: **MT** = McConnell thrust; **BT** = Bourgeau thrust; **LT** = Lewis thrust; **SIT** = Snake Indian thrust; **SPT** = Simpson Pass thrust; **CCT** = Chatter Creek thrust; **PT** = Purcell thrust; **R-LRF** = Redwall-Lussier River fault; **DCF** = Dibble Creek fault; **MF** = Moyie fault; **SMF** = St. Mary fault; **HLF** = Hall Lake fault; **RCF** = Redding Creek fault; **MFT** = Mount Forster thrust; **SRMTF** = Southern Rocky Mountain Trench fault; **PTF** = Purcell Trench fault; **BRF** = Beaver River fault; **KABF** = Kootenay Arc boundary fault; **PRC** = Priest River Complex; **VC** = Valhalla Complex; **MC** = Monashee Complex; **MG** = Malton gneiss; **DR** = Dogtooth Range.

Fort Point Formation, Baird Brook division; see Fig. 2-2) in the Cariboo Mountains, Western Main Ranges of the Rocky Mountains and northernmost Purcell Mountains (Charlesworth et al., 1967; Aitken, 1969; Carey and Simony, 1985; McDonough and Simony, 1986; Kubli, 1986). They concluded that this marker unit represents a eustatic highstand event that interrupted submarine fan grit deposition and is thus a timeline and powerful chronostratigraphic correlation tool throughout the basin. However, the marker unit was identified only in successions in which the base of the Windermere Supergroup is not exposed.

Stratigraphic and structural relationships at the base of the Windermere Supergroup were the focus of several detailed studies in the Purcell and southern Selkirk Mountains (Aalto, 1971; Atkinson, 1975; Lis and Price, 1976; Bennett, 1986; Root, 1987; Pope, 1989 and 1990). These studies established that the distribution of lithofacies at the base of the Windermere Supergroup was controlled locally by syn-depositional normal faults. However, the upper part of the Windermere Supergroup is not exposed in these areas due to erosion and bevelling beneath several Early Paleozoic unconformities (e.g. Reesor, 1973; Root, 1987) and the marker unit discussed by Ross and Murphy (1988) was not recognized. Kubli (1990) mapped the regional marker unit in the Dogtooth Range at the northern termination of the Purcell Mountains (Fig. 2-1) and also located it to the south and east in the area mapped by Root (1987), thus establishing a more firm stratigraphic link between the base of the Windermere Supergroup and the rest of the succession to the north. However, the stratigraphic successions beneath the marker unit differed conspicuously, and since the two successions are exposed in separate fault panels, and the base of the northern succession is not exposed, they could not be correlated directly.

Previously published data and interpretations have thus far not provided sound regional correlations and illustrations of the lithofacies and thickness changes that occur between the base of the Windermere Supergroup at its type locality in the eastern Purcell Mountains and the thicker successions to the north and west. Uncertainties in regional stratigraphic correlations render it difficult to reconstruct the regional configuration and structural evolution of the rift basin and to distinguish eustatic from local tectonic controls on sedimentation. These uncertainties thus

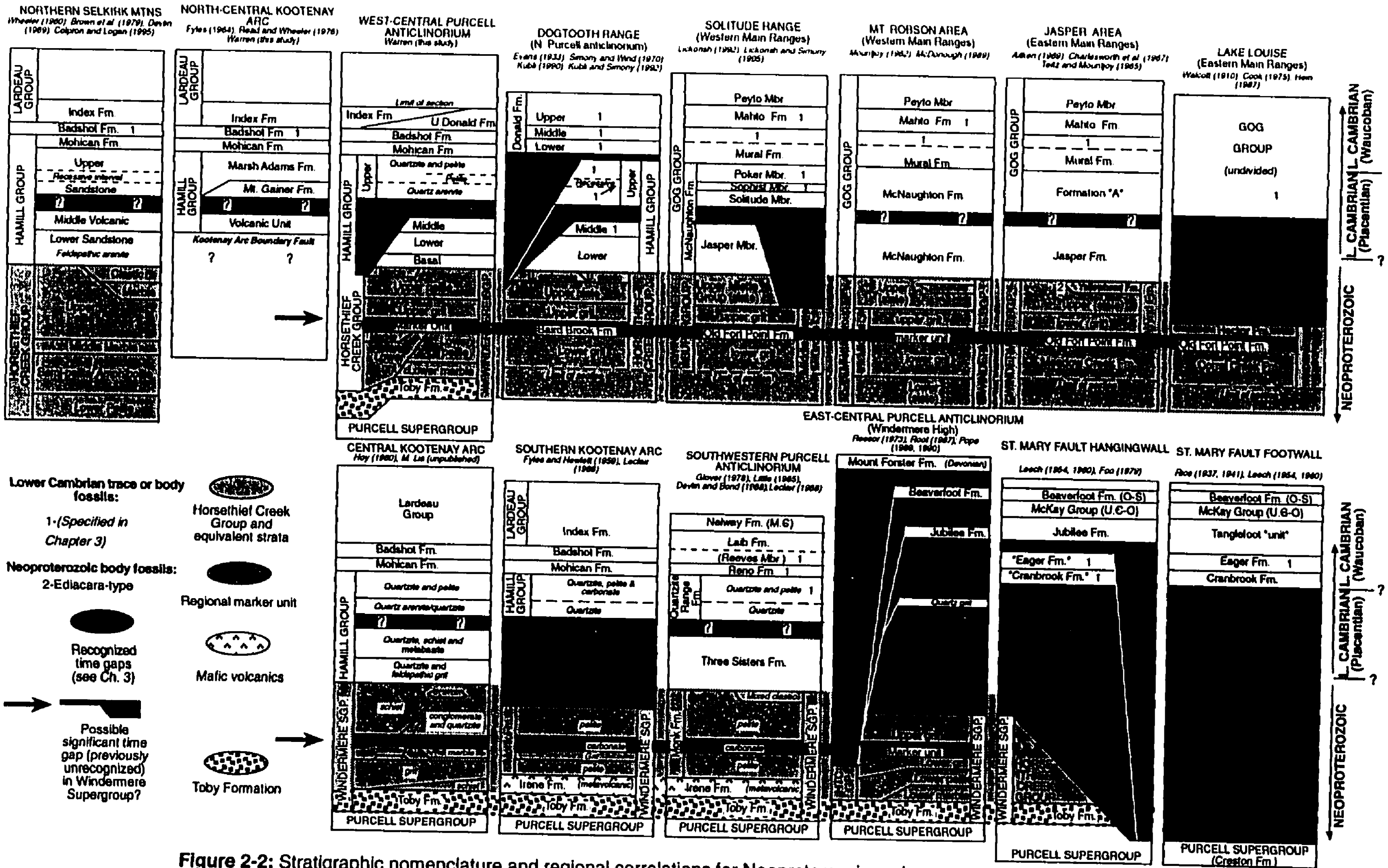



**Figure 2-2:** Stratigraphic nomenclature and regional correlations for Neoproterozoic and younger strata, southeastern Canadian Cordillera. Vertical axis not to scale. Fossil data are discussed in more detail in Chapter 3. "Waucoban" is used in the sense of Fritz et al. (1991). Place names located on Fig. 1-2.

contribute to the continuing debate about the nature and precise timing of the transition from continental rifting to continental separation and post-rift thermal subsidence on the western margin of Laurentia (Bond and Kominz, 1984; Ross and Murphy, 1988; Ross, 1991b; Devlin and Bond, 1989).

The Purcell Mountains is an excellent area for addressing these problems because:

1) the base of the succession and its underlying "basement" (the Purcell Supergroup) are well exposed
2) there is relatively continuous exposure of the Windermere Supergroup and underlying strata across strike as well as along strike in the northern Purcell Mountains, thus providing the opportunity to reconstruct the basin in three dimensions
3) there is a significant body of previous detailed mapping and stratigraphic/ sedimentological data from the Windermere Supergroup with which new data can be integrated.

This paper describes newly-established stratigraphic relationships within the previously undivided Horsethief Creek Group in the west-central Purcell Mountains, east of Duncan Lake, British Columbia. These relationships provide a critical "missing link" for stratigraphic correlation between previously described successions to the north, east and south within the Purcell and Selkirk Mountains and the Western Main Ranges of the Rocky Mountains. They also illustrate the nature of the lithofacies changes that occur between these successions. A synthesis of new and previous stratigraphic and structural data illustrates the regional configuration and tectonic evolution of the Windermere Supergroup basin and its underlying continental lithosphere in the southern Canadian Cordillera. This paper will demonstrate that the initiation of the basin and deposition of a large volume of grit in submarine fans was controlled by two sets of syn-sedimentary normal faults. The underlying continental crust was thinned abruptly from southeast to northwest and less abruptly from east to west. This episode of crustal attenuation was marked by mafic volcanism, deep tectonic subsidence to the west and north and significant topographic uplift to the east and south. Deposition adjacent to normal fault scarps locally obscured the record of a Neoproterozoic post-glacial, eustatic sea level rise. This episode of extension had probably ceased prior to a significant eustatic highstand event that is recorded regionally in the Windermere basin (Old Fort Point Formation and regionally correlative marker units of Ross and Murphy, 1988; Ross, 1991b). A second episode of syn-sedimentary normal faulting post-dated the highstand, but it apparently did not affect underlying "basement" strata, and it is difficult to distinguish in the stratigraphic record from eustatic processes. Nonetheless, sediment transport and deposition in the basin during this second episode of faulting were strongly influenced by the basin configuration and topography that was established during the first episode.

## REGIONAL SETTING

The Purcell anticlinorium (Fig. 2-1) is a broad, gently north-plunging structure in which rocks of the Mesoproterozoic Purcell Supergroup and the unconformably overlying Neoproterozoic Windermere Supergroup are extensively exposed. The Purcell anticlinorium comprises the westernmost part of the foreland fold and thrust belt of the southern Canadian Cordillera (Price, 1981), and it is exposed primarily in the Purcell Mountains of southern Canada and the adjacent northern United States. The anticlinorium is offset on its southeastern flank by normal faults marking the southern Rocky Mountain Trench. Its southwestern flank crosses Kootenay Lake and the Purcell trench fault into the southern Selkirk Mountains, and it is bounded to the west by a fault contact with the structures of the Kootenay arc (Höy, 1974 and 1980; Leclair, 1988; Warren and Price, 1995; see Chapter 4), a narrow belt of complexly deformed supracrustal rocks which contains the Jurassic "suture" between North America and Intermontane Superterrane (Monger and others, 1982). The Purcell anticlinorium plunges out to the north in the Dogtooth Range, and it terminates to the south of the International Border along the northwest-trending Lewis and Clark Line (Sears, 1994).

The Purcell anticlinorium at the latitude of Duncan Lake/ Windermere, B. C. is characterized by upright or east-verging folds and a regional axial planar cleavage. Most of the Purcell and Windermere Supergroup rocks in the Purcell anticlinorium are in the lowermost greenschist facies of regional metamorphism, but upper greenschist to amphibolite facies metamorphism and more severe penetrative deformation affected the westernmost Purcell anticlinorium and adjacent Kootenay arc (Fyles, 1964; Reesor, 1973; Chapter 4). Most of the regional metamorphism, deformation and plutonism is considered to be the result of Mesozoic terrane accretion (Monger and Price, 1979; Price, 1981; Archibald and others, 1983, 1984; Chapter 4).

Several regionally significant, tranverse faults cut the Purcell anticlinorium and its regional, north-trending cleavage. The faults are oblique, northeast-trending, right-lateral reverse faults where they cut across the anticlinorium, but southward and northward they are thrust faults that are parallel to the regional strike on the flanks of the anticlinorium (Fig. 2-1). The transverse faults include the Mount Forster thrust east of Duncan Lake, and the Hall Lake, St. Mary and Moyie faults to the south of the study area. Contrasting Proterozoic and/ or Lower Paleozoic stratigraphic successions in the hangingwalls and footwalls of the Mount Forster, Hall Lake, St. Mary and Moyie faults indicate that they are in part followed older normal faults that were reactivated as oblique right-hand reverse faults during Mesozoic compression (Lis and Price, 1976; Foo, 1979; Price, 1981; Root, 1987; Höy, 1993).

The Windermere Supergroup is exposed primarily in the northern and western parts of the Purcell anticlinorium (Fig. 2-1, and Plate 1). It unconformably overlies a thick sequence of carbonate and fine siliciclastic rocks of the Mesoproterozoic Purcell Supergroup (Belt Supergroup in the United States) that comprises the core of the Purcell anticlinorium. The Windermere Supergroup in the northern Purcell anticlinorium includes, at its base, the Toby Formation (Walker, 1926), a heterogeneous unit up to 500 m thick that is characterized by diamictite. The Toby Formation is overlain by several kilometres of immature siliciclastic and carbonate rocks of the Horsethief Creek Group (Walker, 1926). Much greater thicknesses of Horsethief Creek Group strata are exposed to the west and north than to the east and south (Reesor, 1973). On the western flank of the Purcell anticlinorium, near the International Border, the Toby Formation is up to 1500 m thick and is overlain by the Irene Volcanic Formation (Daly, 1912), a succession of mafic volcanic rocks between 1500 and 3000 m thick (Daly, 1912; Park and Cannon, 1943). The Irene Volcanic Formation is overlain by the Monk Formation (Daly, 1912), which is considered mostly equivalent to the Horsethief Creek Group (Little, 1960). South of the International Border the equivalents of the Toby Formation and Irene Volcanic Formation are the Shed Roof conglomerate and the Huckleberry greenstone or Leola volcanics, respectively (Park and Cannon, 1943; Little, 1960).

The Windermere Supergroup is overlain by different uppermost Neoproterozoic to Lower Paleozoic strata in different parts of the Purcell anticlinorium (Fig. 2-2). Locally, Cambrian, Ordovician or Devonian strata rest with angular unconformity on incomplete successions of the Windermere Supergroup. Several angular unconformities beneath these Lower Paleozoic strata are associated with intermittent Early Paleozoic uplift and tilting of several high-standing crustal blocks, which include the "Dogtooth High" at the northern termination of the Purcell anticlinorium (Kubli and Simony, 1992), the "Windermere High" in the footwall of the Mount Forster thrust (Reesor, 1973; Root, 1987; see Plate 4), the footwall of the St. Mary fault (Rice, 1937 and 1941) and "Montania" in the footwall of the Moyie fault (Deiss, 1941). In the blocks to the south of the St. Mary and Moyie faults, Lower Cambrian and Upper Devonian strata, respectively, rest directly on the Purcell Supergroup (Rice, 1937 and 1941).

Unfossiliferous Neoproterozoic to Lower Cambrian strata of the lower part of the Hamill Group and the equivalent Three Sisters Formation (Chapter 3), wherever they are preserved, rest on nearly the same stratigraphic level at the top of the Horsethief Creek Group and Monk Formation. The contact is gradational to locally disconformable (see Chapter 3).

# NEW LITHOSTRATIGRAPHIC MAPPING OF THE WINDERMERE SUPERGROUP IN THE WEST-CENTRAL PURCELL ANTICLINORIUM

## INTRODUCTION

The west-central Purcell anticlinorium comprises that part of the anticlinorium in the hangingwall of the Mount Forster thrust between the Fry Creek batholith, to the south, the Bugaboo batholith, to the north, and the Horsethief Creek batholith, to the east (Plate 1). The Windermere Supergroup in the west-central Purcell anticlinorium comprises the Toby Formation (Walker, 1926), at the base, and the overlying Horsethief Creek Group (Horsethief Formation of Walker, 1926).

The Toby Formation contains laterally variable diamictite, grit, argillite and carbonate. It unconformably overlies the Dutch Creek Formation and the overlying Mount Nelson Formation, which are the uppermost formations in the Purcell Supergroup (Walker, 1926; Reesor, 1973). Clasts in the diamictite are derived primarily from these two formations.

The Horsethief Creek Group in this region is characterized by immature, commonly coarse (grit and pebble or cobble conglomerate) clastic sedimentary rocks and minor resedimented carbonate rocks. The coarse clastic rocks contain detritus derived from a granitic basement source. The Horsethief Creek Group is overlain by more quartz-rich strata of the Neoproterozoic to Lower Cambrian Hamill Group (Walker and Bancroft, 1929).

A regionally significant stratigraphic discontinuity occurs within the overlying Hamill Group, so that locally, the upper part of the Hamill Group rests directly on the Horsethief Creek Group (Chapter 3). The nature of the contact between the lower part of the Hamill Group and Horsethief Creek Group is more difficult to assess, but a possible interpretation is that the contact is conformable in some parts of the basin and unconformable in others. These relationships and their regional tectonic significance are discussed in subsequent sections of this chapter and in Chapter 3. However, it is important to stress here that the sedimentary and tectonic history of the base of the Hamill Group may be intimately linked with, and locally continuous with, that of the upper part of the Horsethief Creek Group.

The Horsethief Creek Group in the west-central Purcell anticlinorium can be divided into several mappable lithostratigraphic units (Plates 1 - 3)[2]. Lower and upper siliciclastic sequences are separated throughout much of the area by a distinct carbonate-bearing marker unit that caps the lower sequence. Both the lower and the upper clastic sequence display abrupt southeast-to-northwest lithofacies and thickness changes, that result in conspicuous stratigraphic contrasts between a relatively thin, laterally continuous succession exposed in the Jumbo Creek, upper Toby Creek and Glacier Creek watersheds to the south and east, and laterally variable successions more than twice as thick, that are exposed in the Howser Creek, Tea Creek, upper Horsethief Creek and Stockdale Creek watersheds to the north and west (Plates 1 and 2). However, the base of the upper clastic sequence and the carbonate-bearing marker unit, as well as some other contacts, can be traced across the zones of stratigraphic contrast, so that it is possible to correlate map units from northwest to southeast. Thickness and lithofacies changes in the Horsethief Creek Group correspond closely with thickness and lithofacies changes in the underlying Toby Formation and apparently with changes in the stratigraphic level within the Purcell Supergroup on which the Toby Formation rests.

The vertical and lateral variations within the Windermere Supergroup and overlying strata of the western Purcell Range are illustrated by a series of stratigraphic columns (Figs. 2-3 - 2-5). Thicknesses of map units shown in the columns were estimated during 1:20,000 and 1:50,000 mapping, with the aid of 1:15,000 aerial photographs. Thicknesses of distinct lithofacies (e.g. the marker unit and the complete Horsethief Creek Group succession shown in Fig. 2-6) were estimated by pacing throught he succession and correcting the thickness for bedding attitude and topography. Most sections are located on the limbs of map-scale folds within homoclinally-dipping sequences. The section in lower Howser Creek (Fig. 2-3; Section 8), however, may contain

[2] Lithologic designators that appear in parentheses after map unit names in the text correspond to lithic designators for these units on Plates 1-3.

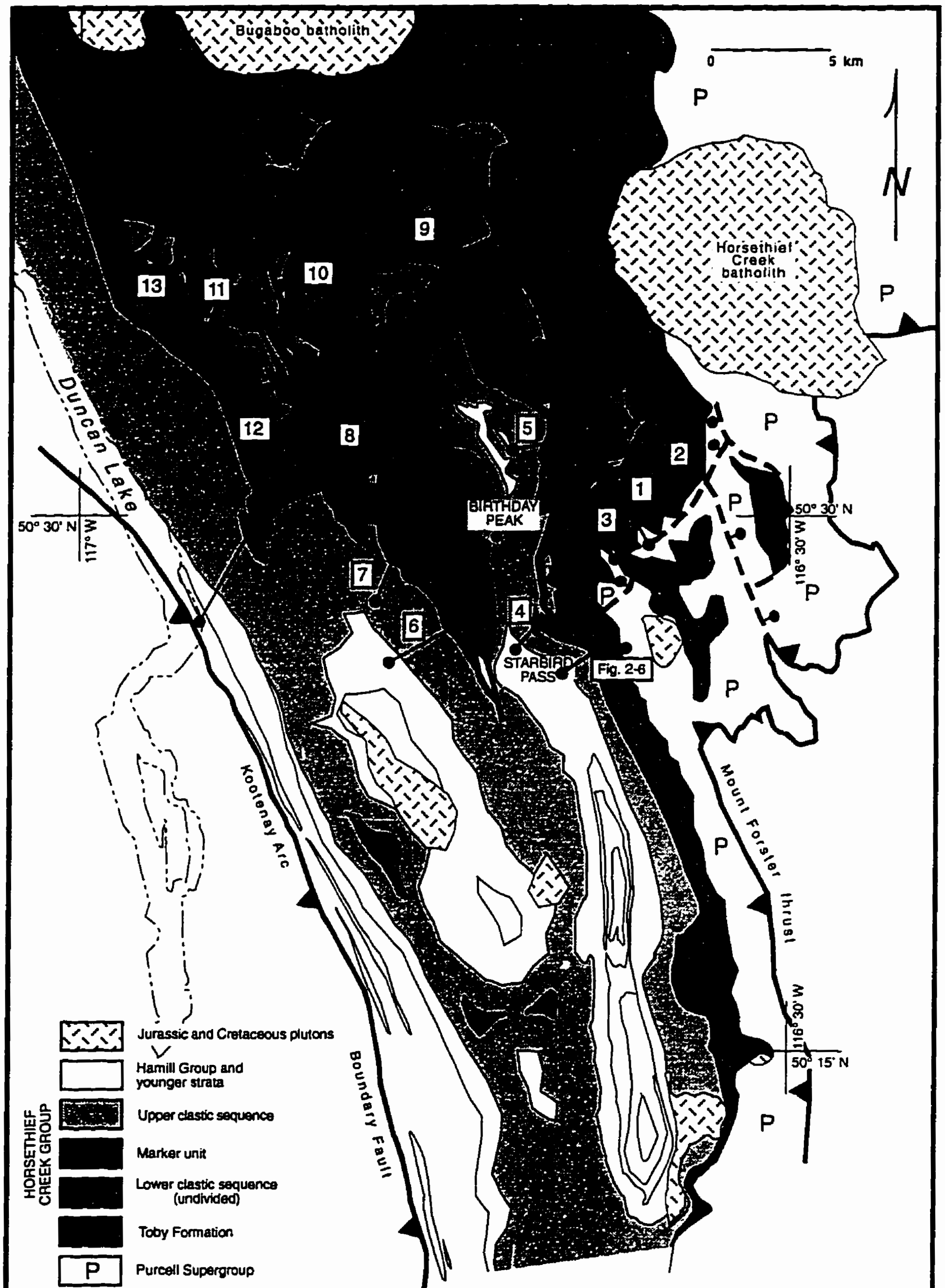



**Figure 2-3:** Simplified geological map of the west-central Purcell anticlinorium, showing distribution of the Windermere Supergroup and locations of the stratigraphic columns shown in Figures 2-4, 2-5 and 2-6. Refer to Plates 1-2 for details of geology.

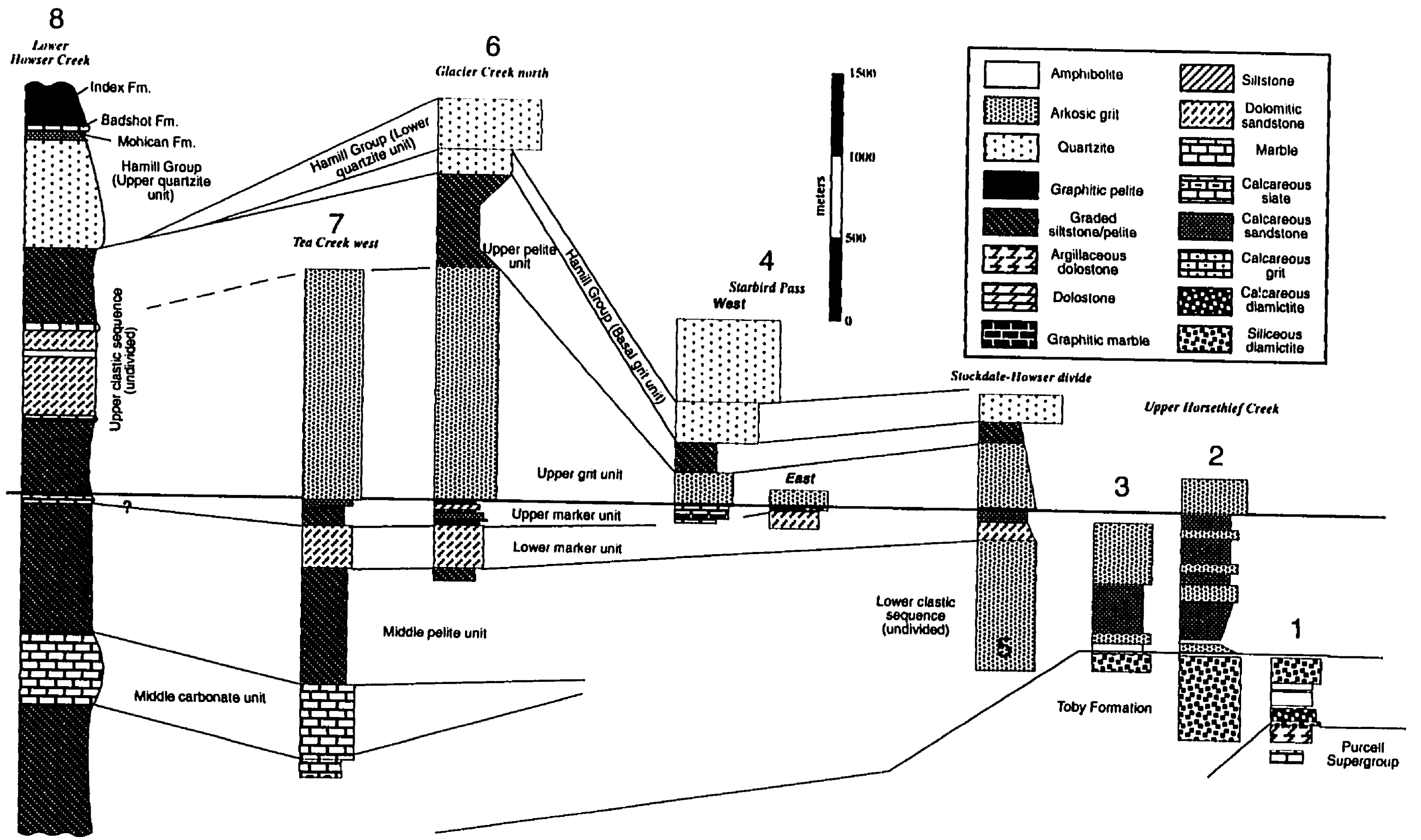



**Figure 2-4:** Stratigraphic columns from the west-central Purcell anticlinorium, illustrating lateral variations in the Windermere Supergroup from southeast to northwest. Horizontal datum is base of upper clastic sequence.

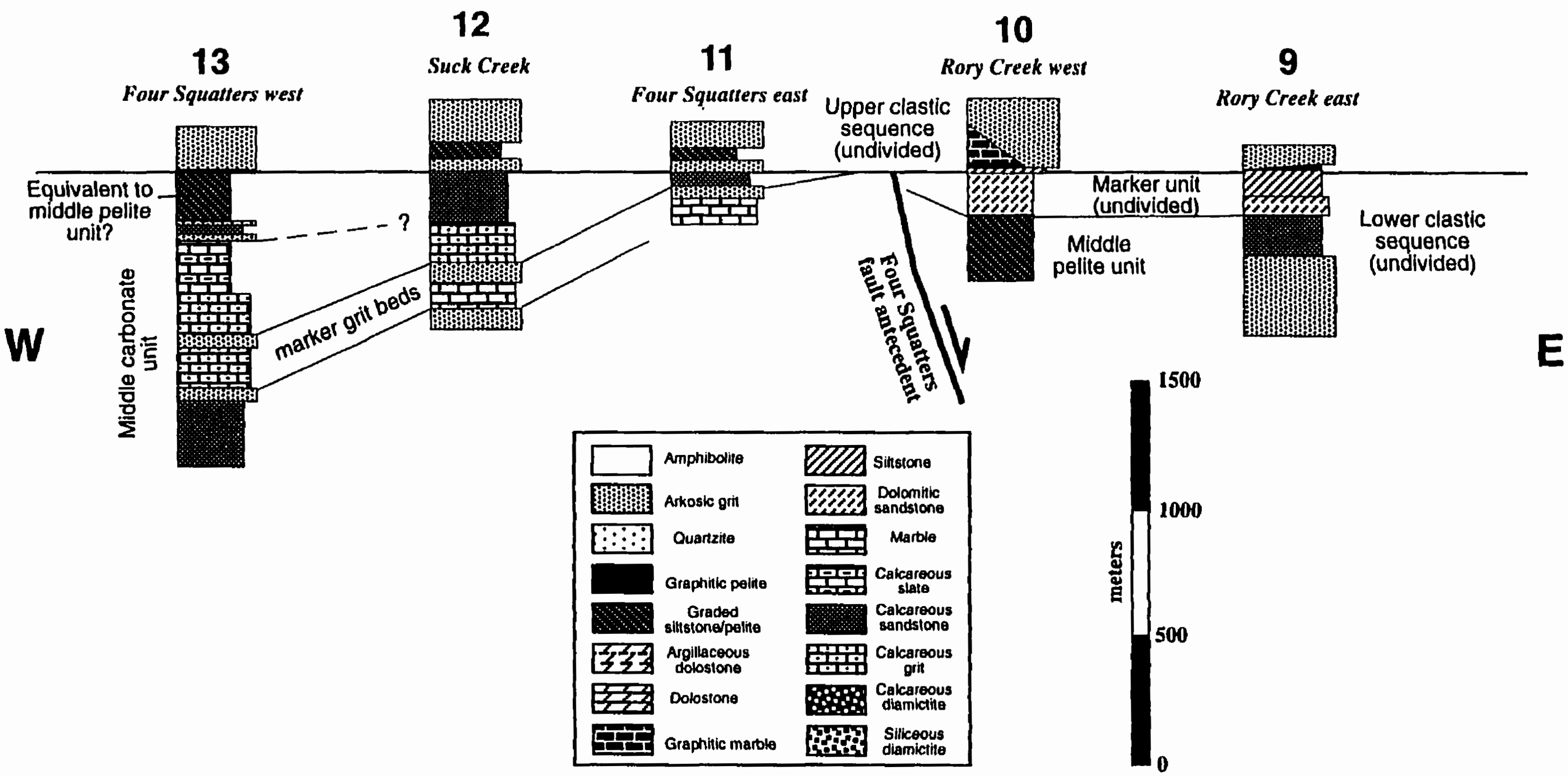



**Figure 2-5:** Stratigraphic columns from alpine exposures of the Horsethief Creek Group, south and west of the Four Squatters Glacier, illustrating relationships across the Four Squatters fault. Horizontal datum is base of upper clastic sequence.

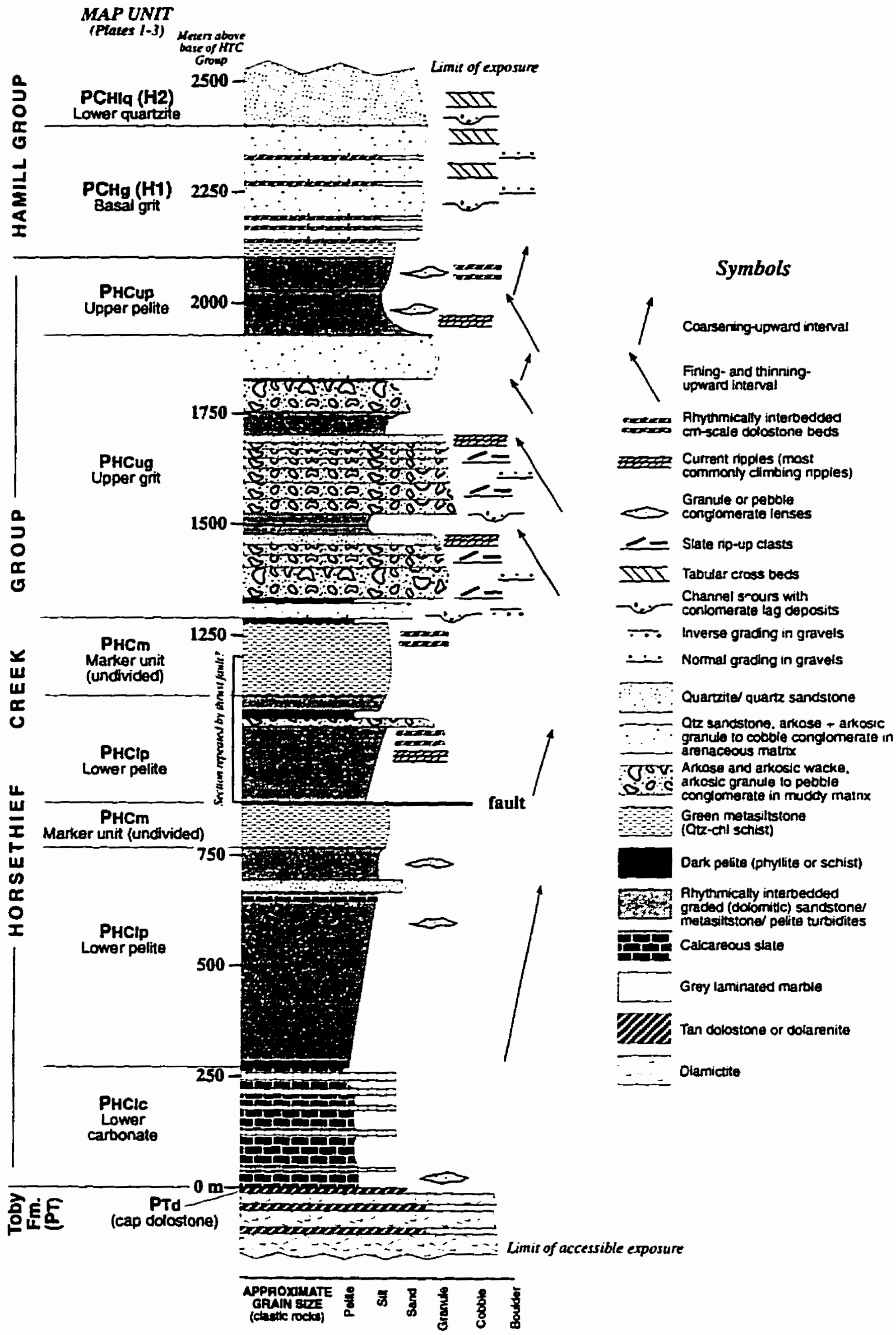



**Figure 2-6:** Complete stratigraphic section through the Horsethief Creek Group, from alpine exposure at the headwaters of Jumbo Creek (north fork). Section also shown on Photo Plate 1. Section is representative of lithostratigraphic units and their sedimentological features in the thin succession of Windermere Supergroup strata that is exposed in the southeastern part of the study area.

significant unidentified reversals in facing direction. The effect of cleavage development on original stratigraphic thicknesses could not be evaluated, but it is probably significant in argillaceous rocks. A detailed stratigraphic section (Fig. 2-6; Photo Plate 1) presents sedimentological data and lithostratigraphic subdivisions that are representatitve of much of the Horsethief Creek Group in the study area.

## THE TOBY FORMATION (PT)

### *Description*

The Toby Formation (Walker, 1926) is exposed extensively in the eastern Purcell anticlinorium in both the footwall and the hangingwall of the Mount Forster thrust, primarily to the east of the area that was mapped during this study (Fig. 2-7 and Plate 1). Since relationships within and beneath the Toby Formation are critical to understanding the evolution of the Windermere Supergroup basin in this region, a large body of previous work from the footwall of the Mount Forster thurst (Atkinson, 1975; Root, 1987; Pope, 1989, 1990), and from the hangingwall east of the Horsethief Creek batholith (Bennett, 1985, 1986), as well as from the present study area, is summarized and included in this discussion.

The Toby Formation unconformably overlies the Dutch Creek Formation and overlying Mount Nelson Formation in both the hangingwall and footwall of the Mount Forster thrust (Walker, 1926; Reesor, 1973; Root, 1987). It is gradationally to abruptly overlain by the Horsethief Creek Group (Pope, 1989; this study). It varies from zero to 600 m in thickness.

The lithofacies which distinguish the Toby Formation are polymict, poorly-sorted, matrix-supported pebble to boulder conglomerates and breccias (diamictite). The clasts are quartzite, carbonate and, less commonly, slate that clearly are locally derived from the underlying Dutch Creek and Mount Nelson Formations. Very rare granitic cobbles occur near Canal Flats (Fig. 2-7), east and south of the study area (Leech, 1956; Reesor, 1973).

Composition, texture and clast size in diamictite of the Toby Formation vary abruptly, commonly at the scale of a single outcrop. The matrix of Toby Formation conglomerate is most

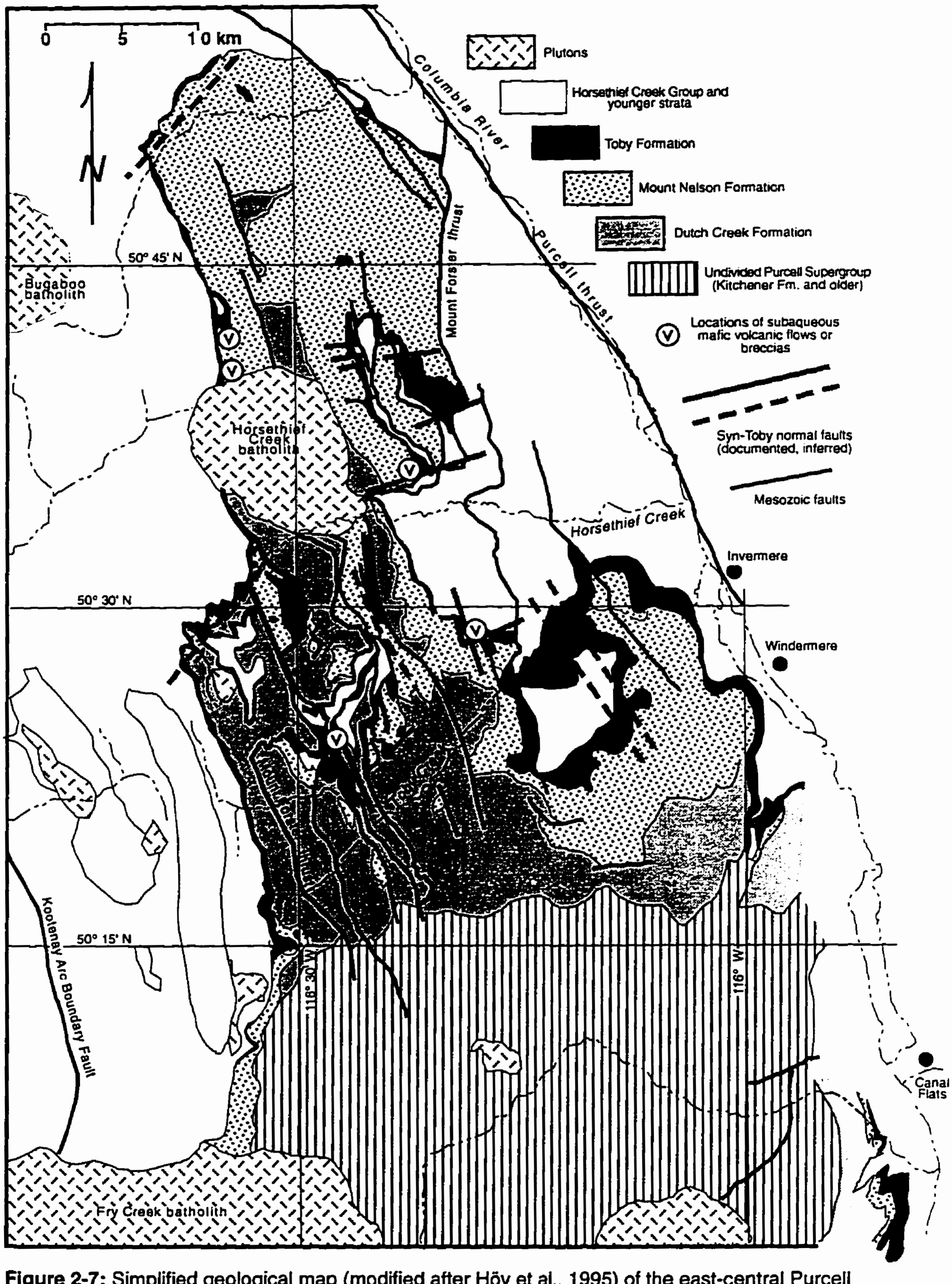



**Figure 2-7:** Simplified geological map (modified after Höy et al., 1995) of the east-central Purcell anticlinorium, showing exposures of the Neoproterozoic Toby Formation. Locations of subaqueous volcanic flows in the Toby Formation and syn-Toby normal faults (enlarged beyond exposure for emphasis) discussed by previous workers or in this study are also shown (see text for references). Refer to Plate 1 or to Root (1987) and Pope (1990) for details of the geology in the Toby Creek area.

commonly brown/ red-weathering argillite or sandy argillite. Locally the matrix is dolomitic or calcareous and, less commonly, comprises quartz arenite, wacke or grit. Conglomerate and breccia of the Toby Formation are interbedded locally with thick sequences of argillite, graded grit and pebble conglomerate beds, and minor grey and buff carbonates (see Pope, 1989, for detailed lithofacies descriptions).

A previously unrecognized, distinct, homogeneous, structureless buff dolostone interval (P Td), 10-50 m thick, occurs at the top of the Toby Formation in the homoclinal panel that is exposed in the hangingwall of the Mount Forster thrust (Warren and Price, 1993; Plate 1 and Photo Plate 1). The dolostone interval appears to be continuous for a strike length of 20 km. It is distinct from and abruptly overlain by dark calcareous slate and marble of the Horsethief Creek Group.

Volcanic rocks are intercalated locally with clastic rocks in the Toby Formation (Fig. 2-7). Intermediate to mafic subaqueous volcanic flows near the base of the Toby Formation are up to several hundred meters thick, and they are associated with syn-sedimentary normal faults (Reesor, 1973; Bennett, 1985, 1986; Root, 1987; Pope, 1989, 1990). Clasts of mafic volcanic rocks occur near some of the flow localities, implying erosion and local transport of the flows during deposition of the Toby Formation (Reesor, 1973; Bennett, 1985, this study). At other localities, locally-derived clasts of Purcell Supergroup strata are incorporated in an andesitic to mafic matrix (Reesor, 1973; this study). Diabase dikes intrude the upper Purcell Supergroup strata in the footwall of the Mount Forster thrust. They are interpreted as feeder dikes that intruded during deposition of the Toby Formation, and they are apparently spatially associated with Neoproterozoic normal faults that cut the Purcell Supergroup (Pope, 1989).

Stratigraphic variations within the Toby Formation reflect its relationships to the underlying Purcell Supergroup and to the overlying Horsethief Creek Group. Where the base of the Toby Formation follows a single stratigraphic unit within the Purcell Supergroup over a distance of several kilometres, lithofacies within the Toby Formation are laterally continuous, and the contact with the overlying Horsethief Creek Group is abrupt. In contrast, where the Toby

Formation cuts abruptly through Purcell Supergroup strata, or rests with obvious angular unconformity on underlying Purcell Supergroup strata, it is generally characterized by abrupt lateral variations in lithofacies and by marked thickness changes. The contact with the overlying Horsethief Creek Group is most commonly gradational, defined by a transition from diamictite in an arenaceous matrix to overlying pebble conglomerate and grit.

Abrupt facies changes in the Toby Formation are more common in the footwall of the Mount Forster thrust, where they are associated with numerous syn-sedimentary normal faults that cut the underlying Purcell Supergroup. In the hangingwall of the Mount Forster thrust, south of Horsethief Creek, the Toby Formation lies directly on the thin, uppermost map unit of the Dutch Creek Formation (Root, 1987), and it is internally layered and capped by buff dolostone (P Td) over an interval of nearly twenty kilometres along strike (Plate 1). However, a notable and abrupt change in stratigraphic relationships within, beneath and above the Toby Formation occurs where the strike of the base of the Windermere Supergroup is abruptly deflected from N-S to NE-SW in upper Horsethief Creek (Plate 1). The Toby Formation in this watershed rests on various subdivisions of the Dutch Creek and Mount Nelson Formations, its thickness varies abruptly, and it lacks internal lateral continuity. The buff dolostones appear as discontinuous lenses at several stratigraphic levels in the Toby Formation, and they contain locally abundant quartz sand grains. Discontinuous intervals of graded grit are common. The Toby Formation appears to thicken abruptly where it overlies the Mount Nelson Formation, reaching a maximum thickness of 500-600 m in upper Horsethief Creek (Fig. 2-4). The contact with the overlying Horsethief Creek Group is gradational rather than abrupt.

*Contact relationships beneath the Toby Formation and evidence for normal faulting during deposition of the Toby Formation*

The Toby Formation rests unconformably on many stratigraphic levels in the Mount Nelson and Dutch Creek Fomations within numerous different thrust fault panels in the eastern Purcell anticlinorium (Bennett, 1986; Root, 1987; Pope, 1990). At some localities the base of the

Toby Formation cuts abruptly across Purcell Supergroup strata within a single thrust fault panel. At other localities the Toby Formation rests on different stratigraphic units across thrust faults. Thickness and lithofacies changes in the Toby Formation and in the lower Horsethief Creek Group across many of these faults indicate that they must have originated as normal faults prior to or during deposition of the Toby Formation and/ or the lowermost Horsethief Creek Group (Atkinson, 1975; Bennett, 1986; Root, 1987; Pope, 1989). Most these structures are roughly north-south trending (Plate 1) and have less than 200 meters of pre-Mesozoic stratigraphic separation (Root, 1987; Pope, 1989). They most commonly show west-side-down displacement, but some show east-side-down displacement. One clearly observable fault of pre- to syn-Toby age (Atkinson, 1975; Fig. 2-7, "Paradise" locality), and one inferred fault of pre-Toby age (Root, 1987) are northeast-trending. The northwest sides moved down, with stratigraphic separations of greater than 250 metres and about 1000 metres, respectively. Mafic volcanic rocks are commonly associated with the downthrown sides of both the north- and northeast-trending normal faults.

Several relationships suggest that syn-tectonic deposition also preceeded the deposition of the Toby Formation. The uppermost unit in the Mount Nelson Formation is locally preserved in blocks that were down-faulted immediately prior to deposition of the Toby Formation (Atkinson, 1975; Root, 1987). Root (1987) interpreted the Toby Formation as locally conformable above this unit. Bennett (1986) separated a distinct interval from the top of the Mount Nelson Formation and the base of the Toby Formation in the hangingwall of the Mount Forster thrust, to the east of the Horsethief Creek batholith. (Frances Creek formation; informal nomenclature of Bennett, 1986). This unit is cut by NE-trending normal faults (Fig. 2-7) that are truncated at the base of the Toby Formation (as re-defined by Bennett, 1986). Facies relationships show that at least part of the Frances Creek formation was deposited while these faults were active (Bennett, 1985).

*Depositional enviroment, paleogeography and tectonic setting of the Toby Formation*

The Toby Formation was deposited in a subaqueous environment during regional extension, normal faulting and subaqueous mafic volcanism (Bennett, 1985; Root, 1987; Pope, 1989). The Toby Formation that is now exposed in the eastern Purcell anticlinorium accumulated in down-dropped area of numerous small-scale, north-trending grabens and half-grabens, where it primarily overlies the Mount Nelson Formation. The Toby Formation that is exposed in the Jumbo and Toby Creek valleys to the west accumulated on a relatively high-standing block, on which the Mount Nelson Formation was not preserved or was never deposited. Northeast-trending, northwest-dipping normal faults also controlled locally the deposition of the Toby Formation (Atkinson, 1975; Bennett, 1986; Root, 1987). The relationships beneath the Toby Formation that are shown on Reesor's (1973) map imply that a larger-scale northeast-trending, northwest-dipping fault or fault zone must occur in the Horsethief Creek valley. This inferred fault is interpreted as a significant structure that controlled the abrupt thickness and facies changes, and change in strike in the Toby Formation between upper Jumbo and upper Horsethief Creeks (Plates 1 and 2).

Striated clasts and dropstones in the Toby Formation (Eisbacher, 1981; G. Narbonne, pers. comm.) support the glaciomarine origin for the Toby Formation proposed by Aalto (1971). Evidence for further subaqueous, gravity-driven reworking of the glaciomarine deposits, perhaps on submarine slopes proximal to fault scarps, includes abundant turbidites throughout the Toby Formation (Root, 1987; Pope, 1989; this study), particularly toward the top, slumping in argillaceous facies (Atkinson, 1975; Pope, 1989), debris flow facies (Bennett, 1985; Root, 1987), volcanic clasts in diamictite or conglomerate facies, commonly proximal to flows (Aalto, 1971; Reesor, 1973; Bennett, 1985; Root, 1987), and rounded intraformational pebble conglomerates (Atkinson, 1975; Root, 1987). Turbidites in the upper part of the Toby Formation that contain blue quartz and feldspar indicate that sediment derived from a crystalline basement source, as well as from the underlying Purcell Supergroup, was reaching the basin by the end of Toby time. A U-Pb zircon concordia date of 1781+/-14 Ma (Loveridge et al., 1981) from one of the rare granitic clasts in the Toby Formation indicate that this basement source was most likely the belt of Hudsonian basement that lies to the east of the Purcell Supergroup north of 50°N (Ross, 1991a; Ross and Parrish, 1991). This source could have been exposed as little as a few kilometers to the east of the granitic clast locality near Canal Flats (southeastern limit of Fig. 2-7). Work in more recent and modern glaciogenic deposits indicates that most glaciomarine deposits are accumulated in the waning stages of glaciation, when large volumes of meltwater are supplied to the margin (e.g. Eyles and Eyles, 1992). Thus the upward increase of basement-derived sediment in the Toby Formation might record melting of glaciers that had previously blocked the transport of more distal sediment to the basin. Alternatively, it might record eastward stepping of the basin margin normal faults with time.

The buff dolostone at the top of the Toby Formation is similar to "cap dolostones" that are associated with Neoproterozoic glaciogenic deposits in the Adelaide geosyncline of southeastern Australia and other Neoproterozoic successions worldwide (Ross et al., 1995; G. Ross and G. Narbonne, pers. comm.). These "cap dolostones" are interpreted as dolostone that was very rapidly chemically precipitated in shallow water as a result of upwelling of $CO_2$-rich deep-ocean waters toward the end of glaciation (Kennedy, 1996).

Glaciogenic deposits similar to the Toby Formation occur at the base of the Windermere Supergroup in the northern Canadian Cordillera, and they also accumulated in fault-bounded rift basins (Eisbacher, 1981; Yeo, 1981; Aitken, 1991).

**HORSETHIEF CREEK GROUP (P HC)**

Two contrasting stratigraphic domains of Horsethief Creek Group strata are exposed in the west-central Purcell anticlinorium: a southern/eastern domain characterized by a thin stratigraphic succession and by continuous, clearly defined mappable units, and a northern/western domain characterized by a stratigraphic succession that is at least twice as thick and by laterally variable and discontinuous units (Figs. 2-3 - 2-5, and Plate 3; schematic stratigraphic relationships). The boundary between these two domains is a narrow, northeast-trending zone of abrupt thickness and lithofacies changes that coincides with upper Horsethief Creek and continues to the southwest through Starbird Pass toward the bend in Duncan Lake. Lithostratigraphic map units in the Toby Formation and Horsethief Creek Group in the upper

Jumbo Creek and Horsethief Creek valleys watersheds define a map pattern that provides a down-plunge view of the abrupt thickness and facies changes (Plate 1 and Plate 2, Sections X and Y).

The successions in both domains are divisible into continuous lower and upper clastic sequences, separated by a carbonate-bearing marker unit which links the two stratigraphic domains. The upper clastic sequence unconformably overlies the marker unit or older strata.

Only stratigraphic relationships within newly-subdivided Horsethief Creek Group strata in the western Purcell anticlinorium are discussed in this section. Relationships with strata in the eastern Purcell anticlinorium, in the footwall of the Mount Forster thrust, are included in a later discussion of regional relationships.

*The Horsethief Creek Group of the upper Jumbo, Toby and Glacier Creek watersheds*

The homoclinal panel of Horsethief Creek Group strata that is exposed in the upper Jumbo and Toby Creek watersheds (Plate 1) can be divided into five continuous mappable units:

- an upper grey pelite/ siltstone rhythmite unit, overlain by quartzose grits and quartzite of the Hamill Group (P HCup)
- an upper grit, conglomerate and black pelite unit (P HCug)
- a green argillite, dolomitic siltstone and black marble marker unit (P HCm)
- a lower rusty pelite/ siltstone rhythmite unit (P HClp)
- a lower dark marble and calcareous slate unit, that abruptly overlies the "cap dolostone" of the Toby Formation (P HClc)

The total thickness of the Horsethief Creek Group in this belt varies from about 2000 m to less than 200 m at the southern limit of the study area, where part of the Horsethief Creek Group is cut out by an Early Paleozoic normal fault (Chapter 3). These map units define two distinct clastic sequences, separated by a sharp, locally erosional contact at the base of the upper grit unit.

A similar succession of lithostratigraphic units can be traced westward into the Glacier Creek and Hamill Creek watersheds (Figs. 2-4 and 2-5). However, the base of the Horsethief Creek Group is not exposed west of Jumbo Creek, and the original features of these rocks are obscured by increased metamorphism and penetrative deformation. Stratigraphic correlations and sedimentological interpretations become increasingly difficult farther to the west. Therefore, descriptions and discussions of these lithostratigraphic units are based primarily on observations from the sections in the upper Jumbo and Toby Creek watersheds.

## Lower clastic sequence

### *Lower marble and calcareous slate unit (P HClc)*

The basal unit of the Horsethief Creek Group in this domain comprises medium grey to black, laminated limestone or marble and homogeneous or thinly bedded dark calcareous slate. It rests abruptly on the buff dolostone that caps the Toby Formation along this belt. The thickness of this unit varies from about 100 to 250 metres. Marble is most abundant near the base, where intervals several metres thick are common. The carbonate is locally pisolitic, or oolitic (Root, 1987, and this study). Rare discontinuous beds of intraformational pebble conglomerate occur near the base, and Root (1987) observed beds of angular brown dolostone pebbles in a black limestone matrix at the base of the section in southern Jumbo Creek. Lenses of quartz pebble conglomerate, grit and graded beds of siltstone/slate are common near the top. Individual beds thicken upward toward the gradational contact with the overlying rusty pelite/siltstone unit.

### *Lower pelite unit (P HClp)*

The lower pelitic unit (Photo Plate 2A) comprises thinly bedded, rusty-weathering, medium to light grey pyritiferous slate, brown siltstone/grey slate (Tde) and graded sandstone/slate (Tae or Tabe) classic turbidites (Walker, 1984; terminology of Bouma, 1962). Ripple cross-lamination (Tc) is rare. The lower contact with the underlying calcareous unit is gradational and has been placed where pelitic material is more abundant than carbonate. Beds generally vary in thickness from 1 to 10 cm. Sandy beds and discontinuous lenses of quartz grit and pebble conglomerate occur throughout the sequence, but are more abundant toward the top, particularly toward the north. Siltstone and sandstone beds also are more commonly dolomitic toward the top of the unit and to the north. Meter-thick beds of brown dolostone occur near the top of the unit at the north end of Jumbo Creek. The lower pelite unit is about 500 metres thick, except at the southern limit of the study area, where its upper part is truncated by a fault. It is otherwise conformably overlain by the green argillite unit.

### *Carbonate and fine clastic marker unit (P HCm)*

The marker unit gradationally overlies the lower pelite unit. The marker unit comprises distinct lower and upper intervals that can be mapped as separate units in the northwestern stratigraphic domain (Photo Plate 2D) but not in this southeastern stratigraphic domain. The lower interval comprises conspicuous, continuous homogeneous blue-green or dark green argillite. The green argillite is more resistant than the lower pelite due to a higher overall quartz content, particularly to the north, where thick intervals can be described as argillaceous quartzite. Sharply-bounded beds of brown or buff dolomitic sandstone, 2 to 10 cm thick, are common near the top. No graded bedding or other sedimentary structures were observed in this unit. Total thickness of the green argillite is about 200 metres.

The upper interval is not a continuous and mappable unit in this domain, except at the headwaters of the north fork of Glacier Creek (Plate 1 and Fig. 2-5). There it comprises about 100 m of interbedded channelled dolomitic sandstone, laminated black marble, dolostone and marble conglomerate, and dark slate or phyllite. At the headwaters of the south fork of Jumbo Creek, the green argillite is overlain by a thin interval of horizontally laminated reddish dolomitic sandstone, locally channelled, interbedded with green sandy slate. In the northern Jumbo Creek section (Fig. 2-6), the green argillite unit is capped by a recessive dark pelite about 10 metres thick, locally cut by channel scours filled by the overlying grits.

Upper clastic sequence

*Upper grit unit (P HCug)*

The upper grit unit is characterized by fining- and thinning-upward sequences of immature, thickly bedded pebble conglomerate, grit and coarse sandstone, commonly capped by dark grey pelite. It marks an abrupt vertical change in the character of the Horsethief Creek Group sediments. It rests sharply, and locally erosionally, on the underlying green argillite marker unit. It is gradationally overlain by the upper pelite/ rhythmite unit.

Individual beds of coarse sediment range from one to several metres thick, and are rarely laterally continuous for more than about 200 metres. Thick sequences of stacked or amalgamated beds, with no intervening pelite, are more common to the north in this unit than to the south (Photo Plate 2E). Individual grit or conglomerate beds are generally structureless, but may display inverse or inverse-to-normal grading. Sandy beds are commonly graded, and some contain rare planar laminations or current ripples. Flame structures, load casts and scour channels (Photo Plate 2E), up to 3 metres deep, are common where grit or sandy beds overlie pelite.

The grits are compositionally and texturally immature. Pebbles and small cobbles in the grit and conglomerate include, in order of decreasing abundance: white quartz, feldspar (up to 25%), blue quartz, and less common grey quartz, chert, dolostone and red quartz. Rip-up clasts of underlying dark pelite, up to 50 cm long, are common at the bases of coarse beds. Pebbles of granite, intraformational grit and metabasite also occur but are very rare. The matrix is generally poorly sorted arkose or wacke and is locally dolomitic, but mud-supported grit beds are abundant in the upper Toby Creek section. The grits are poorly sorted. Quartz is well-rounded to angular; feldspar is sub-rounded to angular, but most commonly angular. Textural maturity tends to be similar within a single bed but to vary between beds.

The grit unit is 600 to 800 m thick in Jumbo Creek. To the south, the lower part becomes thinner or is progressively cut out by a fault. It is absent at the southern edge of the study area (Fig. 2-4). In northern Jumbo Creek, beds in the lowermost 100 meters coarsen and thicken upward, but the remainder of the unit is characterized by beds which primarily fine and thin upwards toward the abrupt but gradational contact with graded classic turbidites of the base of the upper pelitic unit.

*Upper pelite (P HCup)*

The upper pelite unit comprises subtly, thinly bedded grey and grey-green slate or phyllite, and sandstone/siltstone/pelite classic turbidites. Pyrite is abundant in the slate and phyllite. The amount of quartz varies significantly along strike, and locally, in the Glacier Creek drainage, this unit is a homogeneous grey argillaceous quartzite. Discontinuous lenses of coarse quartz sandstone and arkosic grit in a pelitic matrix are common, and are probably channel deposits. Sandy turbidite beds are commonly dolomitic toward the top of this unit

Beds thin and fine upward abruptly above the gradational contact with the underlying grit, but the upper pelite unit generally become more quartz-rich upward toward the contact with the overlying Hamill Group. The upper contact with overlying quartzose grits of the lower Hamill Group generally appears conformable and gradational over a few tens of meters, but at a few localities the contact is very sharp, and could be unconformable. The thickness of the upper pelite varies from less than 50 m to nearly 500 m on the ridge between Glacier and Tea Creeks. It is generally about 200 m thick on the eastern limb of the Blockhead Mountain syncline in upper Jumbo and Toby Creeks, and it is about 400 m thick on the western flank of the Purcell anticlinorium.

*Basal grit unit of the Hamill Group (PC Hg)*

The "transition" from the Horsethief Creek Group to the overlying Hamill Group is marked by a distinct unit that comprises interbedded quartzose and arkosic grit, cross-bedded quartz arenite, chloritic schist or phyllite and several distinct tan dolostone beds (see Chapter 3). Its upper contact is everywhere gradational into the cross-bedded quartz arenite of the lower Hamill Group, except possibly in the lower Glacier Creek watershed. The lower contact, as described above, is commonly gradational, but may be locally unconformable. This "transitional" unit contains primarily shallow-water facies that are similar to much of the overlying Hamill Group, and it has proven to be easier to map as part of the Hamill Group. Accordingly, this unit is included in the Hamill Group, although it has been mapped as part of both the Horsethief Creek Group and the Hamill Group at different stages in this study (Warren and Price, 1992 and 1993).

*The Horsethief Creek Group of the Howser, Stockdale and upper Horsethief Creek watersheds*

The succession of Horsethief Creek Group strata exposed in the Howser Creek and upper Stockdale Creek and Horsethief Creek (Plates 1 and 2) drainages differs markedly from the succession exposed to the south and east (Figs. 2-4, 2-5). The critical differences in the northern/western succession are: greater thickness by a factor of at least two, greater lateral variability, and a conspicuously coarser, more heterogeneous, structureless lower clastic sequence.

The base of the northern/western succession of Horsethief Creek Group strata is exposed only in the upper Horsethief Creek watershed, where the Toby Formation grades upward into a thick sequence of heterogeneous, turbiditic grit, conglomerate and resedimented carbonate rocks. This lower clastic sequence grades laterally westward and northward into a sequence of finer and more laterally continuous turbiditic clastic and resedimented carbonate rocks (Plates 2 and 3). The lower clastic sequence is capped by a thin but distinctive marker unit of fine-grained clastic and carbonate rocks in much of the Howser Creek drainage.

The upper clastic sequence of the Horsethief Creek Group is characterized by immature coarse clastic rocks (grit, sandstone and minor conglomerate) which are easily traced northward from the grit unit in northern Jumbo and Glacier Creeks. Grit and conglomerate grade westward and possibly upward into a well-layered sequence of siliceous carbonates, calcareous slates, quartzites and minor grits. The contact with underlying strata is unconformable. The contact with the overlying Hamill Group is only exposed in a shallow syncline in the upper part of the Howser Creek watershed near Stockdale Glacier, and in a steeply-dipping homoclinal belt immediately to the east of northern Duncan Lake (Plate 1). In upper Howser Creek, the lower part of the Hamill Group gradationally overlies slate of the Horsethief Creek Group that can be correlated southward with the upper pelite unit described in the previous section. East of Duncan Lake, the upper part of the Hamill Group unconformably overlies undivided graded quartzite and pelite of the upper clastic sequence. The thickness of the entire northwestern succession of Horsethief Creek Group is difficult to estimate, but must exceed 4000 m (Plate 2 and Fig. 2-5).

Lower clastic sequence

The lower clastic sequence of the Horsethief Creek Group in the northwestern stratigraphic domain comprises a basal coarse, calcareous unit that overlies the Toby Formation Photo Plate 2B). The basal unit is overlain by a coarse, lower clastic unit. A middle carbonate unit and overlying middle pelite unit are mappable, western facies equivalents to the upper part of the lower clastic unit. The sequence is capped by a fine, carbonate-bearing marker that is equivalent to the marker unit to the south and that is divisible into distinct lower and upper units. The lower clastic sequence is unconformably overlain by the upper clastic sequence. The entire lower clastic sequence thickens westward and northward from about 800 m to at least 2000 m (Plate 2 and Figs. 2-4 and 2-5).

*Basal calcareous grit and carbonate conglomerate unit (P HClcg)*

The lower part of the lower clastic sequence exposed between upper Horsethief and Stockdale Creeks (Plate 1) is a heterogeneous calcareous unit. It comprises dark siliceous marble and calcareous slate, sandstone and conglomerate, interbedded with discontinuous intervals of thickly-bedded arkosic grit. The lower contact with the underlying Toby Formation is gradational. Diamictite containing locally-derived boulders and cobbles grades upward into calcareous grit and conglomerate containing progressively more feldspar and blue quartz. This basal grit, generally 50 to 100 m thick, in turn grades upward into heterogeneous, turbiditic calcareous marble, sandstone, grit and conglomerate (Photo Plate 2B). The basal calcareous unit is overlain by the lower clastic unit. The thickness of the basal calcareous unit varies from about 200 to 500 m.

The calcareous strata contain abundant subangular to well-rounded, coarse blue and white quartz sand grains, black pisolitic limestone or marble, and pebble to cobble conglomerate of dark limestone, buff dolostone, quartz and feldspar in a matrix of dark siliceous marble. Quartz sandstone, arkosic grit and rusty-weathering slate are also commonly interbedded with calcareous strata. Individual beds are sharply defined at both the base and the top. Graded beds occur in the arkosic grits, but they are rare within the carbonate intervals.

*Lower clastic unit (P HCga)*

The lower clastic unit comprises discontinuous intervals, 100 to 500 m thick, of grit, pelite and calcareous rocks that cannot be mapped individually. It is best exposed in the alpine terrain between upper Horsethief Creek and Stockdale Creek. The lower clastic unit overlies the basal calcareous unit in upper Horsethief Creek and to the north of the Horsethief Creek batholith. Its base is not exposed to the west and north. Its upper part grades westward into several continuous and mappable units that are discussed below (Plates 2 and 3), and it is overlain by either the marker unit or the upper clastic sequence of the Horsethief Creek Group. The lower clastic unit is less than 600 m thick on the divide between Horsethief and Stockdale Creeks, and it thickens westward.

The lower clastic unit contains conspicuous and abrupt lateral and vertical lithofacies variations. At least three "cycles" of alternating tan or white arkosic grit and grey calcareous grit and slate occur in this unit between upper Horsethief and Stockdale Creeks. The grit intervals are sharply bounded, less than 100 m thick, and they coarsen upward abruptly at the bases. They are otherwise apparently structureless. They contain abundant blue quartz as well as feldspar. Recessive intervals between them comprise grey pelite and calcareous slate, blue and white quartz sandstone and grit. These heterogeneous calcareous intervals are similar to the basal calcareous unit but could not be mapped separately at 1:50,000. The lower clastic unit comprises a single thick interval of tan grit a few kilometers to the west on the same divide. On the divide between upper Stockdale Creek and the Howser Creek watershed, the lower clastic unit comprises a thicker sequence of primarily green grit and conglomerate, with minor interbedded green pelite and siliceous marble. In the Howser Creek watershed the ratio of pelite to grit in the lower clastic unit is greater.

*Middle carbonate unit (P HCmc)*

The middle carbonate unit comprises a well-layered succession of coarsely crystalline grey marble, calcareous grit and minor interbedded dark pelite or calcareous slate. It is exposed in the Tea Creek and lower Howser Creek watersheds and in alpine exposures adjacent to the Four Squatters Glacier. It is a continuous and mappable lateral equivalent to carbonate strata in the upper part of the lower clastic unit. It therefore conformably overlies the rest of the lower clastic unit. It is overlain conformably by the middle pelite unit, also equivalent to the upper part of the lower clastic unit, in the Tea Creek and lower Rory Creek watersheds. It is overlain unconformably by the upper clastic unit to the west and north. The middle carbonate unit is about 500 m thick in Tea Creek, and it thickens westward and northward. The lowermost 500 m everywhere comprises resistant, light-weathering marble, siliceous marble and calcareous grit that contains blue and white quartz sand grains. The upper part comprises recessive, darker siliceous marble, calcareous slate and minor light-colored arkosic grit or sandy dolostone.

*Middle pelite unit (P HCmp)*

The middle pelite primarily contains rusty- or brown-weathering pelite (Te?) and graded sandstone (Ta?) and sandy pelite that contains lenses of grit and quartz pebble conglomerate (channel deposits?). Individual beds are generally 5 to 50 cm thick. This unit is probably turbiditic, although individual Bouma divisions are much more difficult to recognize than in the pelites in the Horsethief Creek Group to the south.The middle pelite unit conformably overlies the middle carbonate and it is conformably overlain by the green argillite of the lower marker unit. It is a finer, lateral facies equivalent of the uppermost part of the lower clastic unit that is exposed in the upper Horsethief Creek and Stockdale Creek watersheds. The middle pelite is exposed primarily on the alpine slopes on either side of Tea Creek, but also on one ridge to the west of lower Rory Creek. It is not recognized elsewhere. It is about 700 m thick in the Tea Creek watershed.

*Lower marker unit: Green argillite/quartzite (P HClm)*

A distinctive, resistant, green argillite or argillaceous fine-grained quartzite unit abruptly but conformably overlies the lower clastic unit or the middle pelite unit in tributary watersheds to the south of Howser Creek. Dolomitic siltstone to medium quartz sandstone beds occur throughout the unit but are most common toward the top. Sandstone beds are commonly horizontally laminated and locally contain current ripples (Tb and Tc). Detrital grains of feldspar or blue quartz are absent. Rhythmic, sharply-bounded intervals of distinctive light dolostone or light grey limestone, 2 to 10 cm thick, are also common near the top. This unit grades into the basal dolomitic lithofacies of the overlying upper marker unit in the Tea Creek watershed. The green argillite/quartzite is about 100 m thick. This unit can be traced southward into the lower part of the undivided marker unit that is exposed in the upper Jumbo Creek drainage. In the Rory Creek watershed, to the north, it cannot be distinguished as a mappable unit from abundant green-weathering argillaceous rocks in the lower clastic unit. It is not known whether the marker unit is everywhere present on the divide between Horsethief and Stockdale Creeks.

*Upper marker unit: carbonate lithofacies (P HCum)*

The upper marker unit (Photo Plate 2D) comprises five distinct lithofacies, of which only the lowermost appears to be laterally continuous. These lithofacies include: 1) basal rhythmically-bedded tan or orange-weathering dolomitic siltstone and sandstone, and interbedded rusty slate and dolostone; 2) poorly sorted pebble to boulder carbonate conglomerate (Photo Plate 2C); 3) interbedded horizontally laminated calcareous sandstone and black limestone; 4) chanelled dolomitic sandstone; and 5) black, locally graphitic, pyritiferous pelite. The carbonate conglomerates commonly contain clasts of the underlying black limestone and/ or siliceous dolostone. Clasts in the conglomerates include light dolostone, grey dolomitic marble, white sandy dolostone and dark grey, laminated limestone (Photo Plate 2C). The matrix is a fine- to medium-grained dolomitic quartz sandstone. Detrital feldspar and blue quartz were not observed within the upper marker unit.

The upper marker unit conformably overlies the green argillite/ quartzite of the lower marker unit.The marker unit is abruptly overlain by grits of the upper clastic unit, and it is partially or completely cut out at several localities beneath the overlying grit of the upper clastic unit. The upper marker unit is 140 metres thick on the divide between Glacier and Tea Creek (Fig. 2-4, 2-5). At most other localities it is significantly thinner.

Upper clastic sequence

The upper clastic sequence of Horsethief Creek Group strata in the northwestern stratigraphic domain comprises an upper grit unit and overlying upper pelite unit that can be mapped separately in the tributary watersheds to the south of Howser Creek, and an undivided upper clastic unit to the west and north. The grit and pelite units are continuous with and equivalent to the upper grit and upper pelite units, respectively, in the Jumbo and Glacier Creek watersheds, although they are both thicker and coarser to the north. Metamorphic and sedimentary facies changes, coupled with increased strain, render the upper clastic sequence indivisible to the west and north (Plates 1 and 2). Overall, the entire upper clastic sequence is finer and it contains more carbonate strata toward the north and west. The upper clastic sequence unconformably overlies the marker unit or underlying strata. It appears to rest with angular unconformity on the middle carbonate unit to the west of the Four Squatters fault (Plate 1) and to rest disconformably on the upper marker unit to the east of this fault. The undivided upper clastic unit is unconformably overlain by the upper part of the Hamill Group along the eastern shore of Duncan Lake. The upper pelite unit is gradationally overlain by the base of the Hamill Group in the Eyebrow syncline (Plate 1).The entire upper clastic sequence has a maximum thickness of about 2000 m, immediately to the north and west of Starbird Pass (Fig. 2-5).

*Upper grit unit (P HCug)*

The upper grit unit comprises resistant, thickly-bedded intervals of grit, conglomerate and coarse sandstone with interbedded dark pelite. Individual beds are generally a few meters thick, and they are amalgamated in intervals that are tens of meters thick and contain little interbedded pelite. Individual beds generally become finer-grained and thinner upward in each of these intervals. Sedimentary structures in the grit beds include flame structures, load casts and channel scours which cut the underlying beds at high angles. However, many of the grit beds are massive, structureless and laterally continuous for up to hundreds of meters. Grains and pebbles include abundant white quartz, feldspar (as much as 40% of clasts), blue quartz, chert, granite and chloritic lithic fragments, in order of decreasing abundance. Slate rip-up clasts, up to 0.5 m long, are common at the bases of beds. Most clasts and grains are subrounded to subangular. The matrix of the grits is more commonly arenaceous than pelitic, and many of the grits are clast-supported.

The upper grit unit is thickest, coarsest and most laterally variable immediately to the north of Starbird Pass, where the entire upper clastic sequence thickens abruptly to the west and north. On the divide between Stockdale and Howser Creeks, texturally and compositionally immature, thickly-bedded, laterally continuous grit beds are intercalated with less common, thick (>10 m), discontinuous beds of well-rounded, clast-supported white quartz pebble to cobble conglmerate and with intervals of thinly-bedded (1-10 cm) dark slate and sandy slate that are up to several meters thick. The conglomerates contain only rare clasts of subrounded feldspar and angular sandy calcareous slate. Some of the quartz pebbles and cobbles are weakly imbricated and indicate transport from the east. The conglomerate lenses are interpreted as submarine channel deposits and the pelitic or silty intervals as levee deposits. In the west face of Birthday Peak in the upper Tea Creek drainage (Photo Plate 2F), stacked channels of light-coloured coarse material, tens of meters thick, are thickest at this locality and become abruptly thinner or are absent to the north and south, suggesting a classic "steershead" channel geometry. To the north, south and west, cliff-forming grits of the same stratigraphic interval are more homogeneous, laterally continuous and thinner.

*Upper pelite (P HCup)*

The upper pelite unit comprises rhythymically bedded sandy turbidites and lenses of grit. It gradationally overlies the upper grit unit and it is also coarser and thicker than the equivalent upper pelite unit to the south in the Glacier and Jumbo Creek watersheds. On the north ridge of Birthday Peak, the upper pelite unit comprises brown-weathering graded coarse sandstone/ pelite and grit beds, and lenses of pebble conglomerate. The sandy beds are commonly dolomitic and the pelite contains abundant pyrite. The upper pelite unit is distinguishable from the underlying upper grit because it is more recessive, weathers darker and is more thinly bedded, but the contact is difficult to place. A few kilometers to the north, the contact is more abrupt and the upper pelite unit is much finer, darker and more pyritiferous. In this locality the upper pelite unit becomes more quartz-rich and dolomitic upward toward the gradational contact with the overlying Hamill Group. The uppper pelite unit is about 300 m thick there, and at least twice as thick on Birthday and Eyebrow Peaks.

*Undivided upper clastic unit (P HCuc)*

The upper clastic unit in the northwestern stratigraphic domain comprises basal grit and conglomerate, calcareous grit, abundant graded quartz arenite or quartz wacke, siliceous or sandy carbonate, pelite and minor metabasite.The upper clastic unit generally becomes finer-grained upward, and except for coarse grits exposed at the base of the unit across the entire area, it becomes finer-grained laterally to the west and to the north. The grit intervals in the upper clastic unit are similar to those in the correlative upper grit unit to the south, but they contain less conglomerate. The amount of calcareous strata in the upper clastic unit increases conspicuously to the west and to the north (mappable but discontinuous marbles are designated P Hcc, of variable stratigraphic position). Thick intervals of mature, graded quartz sandstone are common in the lower Howser Creek valley and to the west of the Four Squatters Glacier. The abundance of quartz is noteworthy. Sedimentary structures distinguish these intervals from the overlying upper part of the Hamill Group. The upper Horsethief Creek Group quartzites are characterized by 1-10 cm graded beds with semi-pelitic tops; clean quartzite of the overlying upper Hamill Group is best identified by conspicuous cross beds. The thickness of the upper clastic unit is estimated at about 1500 m near the mouth of Howser Creek.

Lenses and locally continuous belts of greenstone and amphibolite (P HCa), up to a few hundred meters thick, are common in the upper clastic unit on the steep west flank of the Purcell anticlinorium. The metabasites are in sharp contact with the metasedimentary rocks, and their contacts are apparently parallel to the bedding or compositional layering. It is not known whether the metabasic rocks occur at or near a particular stratigraphic level, or are distributed throughout the western part of the upper clastic unit. It is not clear whether the metabasites represent flows, sills or dikes within the sequence. However, the occurrence of rare mafic pebbles elsewhere in this map unit suggests that at least some of the metabasites were syn-sedimentary intrusions. Furthermore, a suite of several greenstone and amphibolite samples from this unit has a trace element geochemical signature which is distinct from the geochemical signatures of metabasic rocks of the Hamill Group and Lower Paleozoic Lardeau Group exposed east of Duncan Lake (Appendix 4), implying that the igneous rocks in the Horsethief Creek Group were not feeder dikes to metabasites in overlying strata.

*Summary of relationships between the southeastern and northwestern stratigraphic domains*

The two contrasting stratigraphic domains of the Horsethief Creek Group in the west-central Purcell Mountains can be correlated confidently by using the Toby Formation, the marker unit, the upper grit and the upper pelite unit, all of which are continuous between the two stratigraphic domains (Figs. 2-4, 2-5). The correlation shows clearly that the Horsethief Creek

Group is divisible into two distinct clastic sequences, and that both sequences display abrupt increases in thickness and changes in lithofacies to the north and west of Starbird Pass (Plate 3). Lithofacies in both sequences are abruptly coarser and less continuous northwestward from Starbird Pass and are gradually finer-grained and more laterally continuous toward the northwestern flank of the Purcell anticlinorium.

*Evidence for normal faulting during the deposition of the lower clastic sequence*

The abrupt changes in lithofacies and thicknesses in map units of the lower clastic sequence of the Horsethief Creek Group imply that there was an abrupt increase in the supply of coarse clastic sediment and in subsidence of the basin from southeast to northwest, at the latitude of Starbird Pass and Horsethief Creek. Relationships beneath the Toby Formation in the Horsethief Creek and Farnham Creek watersheds (Plate 1) indicate that the Toby Formation was deposited during an episode of normal faulting, as discussed above, and that the northwest side of Horsethief Creek represents a down-dropped block of underlying Purcell Supergroup. Several indirect arguments suggest that this episode of normal faulting continued during the deposition of the lower part of the Horsethief Creek Group, and that the thickness and lithofacies changes in the lower Horsethief Creek Group are not merely a result of basin topography that was established during the deposition of the Toby Formation. Root (1987) mapped several small-scale growth faults in the lowermost part of the Horsethief Creek Group in the footwall of Mount Forster thrust, demonstrating that there was active normal faulting in the basin following the deposition of the Toby Formation. The magnitude change in thickness of the Horsethief Creek Group in the western Purcell anticlinorium (at least 2 km) and the abundance of coarse, calcareous sediment in the lower clastic sequence to the northwest implies that there were local, uplifted sediment sources (either Purcell Supergroup carbonate rocks or intrabasinal Horsethief Creek Group carbonate rocks) as well as granitic or gneissic basement sources of sediment throughout the deposition of the lower clastic sequence. The abrupt boundaries between massive grit intervals and intercalated slate intervals has been interpreted to record very abrupt channel abandonment in submarine fans, a hallmark of intermittent faulting and uplift of the basin margins (Heritier et al. 1979; Kubli, 1990). Thus, the abrupt thickness and facies changes within the Toby Formation and Lower clastic sequence are interpreted as evidence for a significant, northeast-trending, northwest-side-down normal fault system that was active during the deposition of both the Toby Formation and the lower Horsethief Creek Group.

*Evidence for normal faulting during deposition of the upper clastic sequence*

The base of the upper clastic sequence is interpreted as an unconformity throughout the study area. In the southeastern stratigraphic domain, the upper part of the undivided marker unit commonly is partially or totally cut out by erosion beneath the overlying grit unit. To the north of lower Howser Creek, on the west side of the Four Squatters fault, the lower and upper marker units are completely absent, and the upper clastic unit rests directly on carbonate and grit of the lower clastic unit (Plate 1). In this area, the lower clastic unit contains several laterally continuous grit and carbonate marker intervals which can be followed above treeline. The relationships between these intervals and the base of the upper clastic unit strongly suggest an angular unconformity beneath the upper clastic unit (Fig. 2-5). At least 500 metres of the lower clastic unit appear to be bevelled from west to east beneath this contact. All of this stratigraphic truncation occurs to the east of the Four Squatters fault, which shows little thrust displacement of the upper clastic unit (Plates 1 and 2). However, the marker unit and the uppermost part of the lower clastic unit are present to the east of this fault. Therefore, the Four Squatters fault must have existed prior to the deposition of the upper clastic sequence. Strata to the west of the fault were uplifted, tilted and bevelled by erosion prior to the deposition of the upper clastic sequence. The fault was subsequently reactivated with little displacement during Mesozoic deformation. Faults with little apparent displacement near Birthday Peak coincide with an abrupt east-to-west increase in thickness of the upper clastic sequence, and they may have originated as west-side-down normal faults prior to or during its deposition.

*Depositional environments of the Horsethief Creek Group in the west-central Purcell anticlinorium*

The strata that comprise the Horsethief Creek Group in the west-central Purcell anticlinorium were deposited below storm wave base, primarily by turbidity currents. Because the lower clastic sequence, the marker unit that caps it, and the upper clastic sequence record distinct episodes of sediment accumulation in this basin, their depositional environments are discussed separately below.

Lower clastic sequence

The sharp base of the fine, turbiditic lower clastic sequence exposed in the Jumbo and upper Toby Creek watersheds records an abrupt change in depositional environment following deposition of the dolostone that caps the Toby Formation. The top of the Toby Formation marks the end of a widespread episode of Neoproterozoic glaciation, and thus it most likely represents a eustatic sea level rise. The overlying fine, turbiditic sediments of the lower clastic sequence (marble and calcareous slate unit, lower pelite unit and base of the marker unit) coarsen upward gradually. This coarsening-upward succession of turbidites has been interpreted previously (Root, 1987) as an outer submarine fan (lower pelite unit) that prograded gradually onto resedimented or hemipelagic calcareous sediments (lower marble and slate unit). Resedimented carbonates could have been derived from lower Horsethief Creek Group marbles and dolostones that are exposed to the east (Root, 1987) or from Purcell Supergroup carbonate rocks. The ooids in the lower clastic sequence indicate that at least some of the calcareous strata are resedimented. They are significantly larger than ooids that occur in the Purcell Supergroup, indicating that intrabasinal, shallow-water Horsethief Creek Group strata that lay to the east were a more likely source. The minor amounts of feldspar in the lower pelite unit also indicate some contribution from a granitic basement source.

In contrast, the lithofacies, provenance and lateral stratigraphic variations in the northwestern lower clastic sequence suggest that tectonic uplift of both basement and local Purcell or Windermere Supergroup sources to the south and east directly resulted in rapid northwestward transport and distribution of sediment, and in deposition of coarse, more "proximal" facies. The base of the lower clastic sequence in the upper Horsethief Creek watershed records an influx of coarse clastic sediment that began during the deposition of glaciomarine sediments of the Toby Formation. Abundant pisoids indicate that at least some of the coarse, resedimented carbonate was probably derived from Horsethief Creek Group strata within the basin. The carbonate was mixed with coarse, terrigenous, basement-derived sediment. The northwestern part of the basin was more deeply subsided than the southern part, and a thicker, coarser succession prograded very rapidly over the top of the Toby Formation. The coarse sediment was perhaps deposited by high-density turbidity currents (Lowe, 1982) and/ or other sediment gravity flows, but the precise mechanisms of sediment transport are difficult to determine in the thickly intercalated grit, pelite and carbonate of the northwestern lower clastic sequence because diagnostic sedimentary structures or facies associations are lacking. The succession that is exposed in the upper Horsethief Creek valley could record a channelled inner fan environment (e.g. Walker, 1984) or an apron of sediment deposited primarily by sediment gravity flows adjacent to a steep slope. The fact that grit intervals in the lower clastic sequence are discontinuous and coarsest to the south and east (in contrast with continuous grit intervals in the upper clastic sequence) suggests that deposition of the lower clastic sequence was controlled by local, rather than basinwide processes. If the grit intervals are channel deposits, then their sharply-bounded, cyclical nature and lack of fining- and thinning-upward intervals suggest that the channels were abruptly abandonned, rather than gradually filled in and abandonned (Heritier et al., 1979). Kubli (1990) described a similar succcession in the lower part of the Horsethief Creek Group in the Dogtooth Range, and suggested that abrupt channel abandonment might indicate intermittent pulses of source uplift and rapid sedimentation due to local syn-depositional normal faulting in the basin.

The rocks of the lower clastic sequence in the northwestern stratigraphic domain show a westward and northward increase in lateral continuity, and overall decrease in grain size. These relationships are interpreted to indicate increasing distance northwestward from the paleo-slope or submarine fan axis down which the sediments of the lower clastic sequence were transported.

### Marker unit

Most of the carbonate and fine-grained clastic marker unit represents an abrupt and marked drop in the supply of coarse clastic sediment to the basin, in accordance with previous interpretations of this interval by Ross and Murphy (1988) and Kubli (1990). The silty green argillite/dolomitic sandstone unit (lower marker unit) indicates that more reworked as well as finer sediment was being supplied to basin. The carbonate-siltite rhythmites indicate that turbidity currents continued to be the mechanism of transport. The black, sulfidic pelite lithofacies of the upper marker unit indicates locally anoxic, perhaps deep conditions. The deposition of the marker unit is consistent with a rapid rise in relative sea level, which would trap coarse sediment on the shelf, and/or consistent with cessation of normal faulting along basin margins and retreat of the source area (Ross and Murphy, 1988; Kubli, 1990). The coarse, discontinuous lithofacies in the upper marker unit (channelled dolomitic sandstone, carbonate clast conglomerate) could indicate a short-lived sea level fall that resulted in down-cutting of channels into an established carbonate shelf to the east or into exposed Purcell Supergroup strata. Kubli (1990) also inferred a relative sea level fall during an overall highstand recorded by the correlative Baird Brook diivision in the Horsethief Creek Group in the Dogtooth Range.

### Upper clastic sequence

The upper grit unit at the base of the upper clastic sequence reflects an abrupt influx of immature sediment derived from a granitic basement source, as well as resedimented carbonate probably derived from the underlying marker unit and/or a Windermere carbonate shelf. This unit has been interpreted previously to record a significant relative sea level fall (Ross and Murphy, 1988; Kubli, 1990) which resulted in rapid progradation of terrigenous sediment into the basin. Textures and sedimentary structures indicate that most of the grit beds throughout the upper clastic unit were deposited by high-density turbidity currents (Lowe, 1982), implying rapid sedimentation and a constant supply of coarse sediment for all but the uppermost part of this sequence. However, the precise depositional environment of this regionally widespread grit unit is difficult to determine. This unit could record rapid progradation (coarsening- and thickening-up at base) of a widespread channelled and leveed portion of a submarine fan (Walker, 1984, 1992). Abundant thinning- and fining-upward sequences with clearly erosional bases imply channel infilling and abandonment upward in the sequence (Walker, 1984). Interbedded fine turbidites (Tde) and dark pelites could be interpreted as levee deposits. Alternatively, Ross et al. (1995) interpreted the upper grit intervals in the Windermere Supergroup in the southern Canadian Cordillera as the record of a regionally significant, deep-sea longitudinal sediment transport system that moved sediment parallel to the basin margin, westward of slope turbidites that are not exposed in the southern Canadian Cordillera. This interpretation has important implications for the plate tectonic setting of the Windermere Supergroup, discussed below.

However, the three dimensional geometry of channels in the Birthday Peak region, the lateral thickness and lithofacies variations in the entire unit across the region and sparse paleocurrent indicators (pebble and cobble imbrication, rare ripples) suggest a sediment transport system in which sediment was moved downslope primarily from southeast to northwest or from east to west. The "steershead" channels and interbedded pelite in the anomalously thick succession at Birthday Peak might represent a submarine fan axis or major lobe axis in this region. Thick lenses of well-rounded, clast-supported quartz cobble conglomerate near this locality are interpreted as lag deposits in major distributory channels. Lateral transition to more "sheetlike" (parallel-bedded, more laterally continuous) composite or amalgamated grit and sandstone beds to the north and southwest may reflect transition to channelled distributory lobes. The abundant graded quartz arenite, calcareous sandstone and schist of the lower Howser Creek watershed to the northwest may represent an unchannelled, sand-rich lower fan (Walker, 1992). If a submarine fan model such as those outlined by Walker (1984, 1992) does not apply to the sand and grit of the upper Horsethief Creek Group, alternative models, such as the deep-sea sediment dispersal system outlined by Ross et al. (1995), must still account for the southeast-to-northwest facies and thickness changes and transport of sediment in the upper part of the Horsethief Creek Group in this area.

The fining-up succession from the underlying grit unit indicates that the contact between the two units records increased subsidence of the basin relative to rate of sedimentation, corresponding to retreat of the source area or a rise in relative sea level. The turbidites of the upper pelite in the southeastern stratigraphic domain of the Horsethief Creek are interpreted as outer fan deposits. The coarsening-up sequence toward the base of the locally conformably overlying shallow-water deposits of the Hamill Group indicates infilling of the basin and a subsequent fall in relative sea-level. The upward increase in carbonate in the distal turbidites of the upper pelite may indicate that a carbonate platform had been established elsewhere in the basin by the end of Horsethief Creek time. Successions that have been interpreted as carbonate platforms occur in the uppermost part of the Windermere Supergroup in the Main Ranges of the Rocky Mountains and the Cariboo Mountains to the north and east (Yellowhead and Byng platforms) and perhaps in the Dogtooth Range to the north (see Fig. 2-2).

**Correlation of the Horsethief Creek Group in the Purcell anticlinorium**

The newly-subdivided Horsethief Creek Group strata discussed above provide a critical link for correlating Windermere Supergroup strata throughout the entire Purcell anticlinorium (Fig. 2-2). Contributions from this study in the west-central Purcell anticlinorium link previously studied Windermere Supergroup strata in its type locality in the footwall of the Mount Forster thrust with equivalent but different successions in the northern and western Purcell anticlinorium. These new data also reveal the nature of the lithofacies and thickness changes that cause many of the differences in the Horsethief Creek Group succession throughout the Purcell anticlinorium. The correlations in turn strengthen or clarify previously proposed correlations between the Windermere Supergroup in the Purcell anticlinorium and the Windermere Supergroup in the northern Selkirk Mountains, Cariboo Mountains and Main Ranges of the Rocky Mountains.

Regional documentation of the nature of lithofacies and thickness changes in the Windermere Supergroup provides the basis for the restored regional paleogeographic interpretation (Fig. 2-8) discussed in the next section.

*Dogtooth Range (northern Purcell anticlinorium)*

The Horsethief Creek Group in the northern Purcell Range and Dogtooth Range has been subdivided previously, from bottom to top, into grit, slate, carbonate and upper clastic units (Evans, 1933; Poulton and Simony, 1980). This sequence is unconformably overlain by the Hamill Group, and its base is not exposed. The total exposed thickness is estimated at nearly 3000 m (Kubli, 1990).

Kubli (1990) recognized a thin, interbedded slate, limestone and dolomitic siltstone interval within the grit division (informal Baird Brook division), which separates sedimentologically distinct lower and upper grit sequences. The Baird Brook division contains an assemblage of lithofacies that is virtually identical to the assemblage of lithofacies that comprise the upper and lower marker units of this study. The underlying lower grit sequence is characterized by thickly interbedded and sharply bounded grit and pelite intervals; the upper grit sequence is characterized by abundant grit-dominated fining- and thinning-upward sequences. This succession is considered equivalent to the lower clastic unit, marker unit and upper grit unit in the Howser, Stockdale and upper Horsethief Creek watersheds of this study area. The gradational contact and fining-upward trend between the grit division and the overlying slate division (Kubli, 1990) is similar to the contact between the upper grit and upper pelite units in the Jumbo Creek, Glacier Creek and upper Tea Creek drainages. The slate division of the Dogtooth Range comprises turbiditic pelite, carbonate and calcareous sandstone similar to much of the upper pelite unit and undivided upper clastic unit in the lower Howser Creek drainage. The turbiditic carbonate division of the Dogtooth Range is probably equivalent to the carbonate-rich lithofacies of the upper clastic unit that are exposed to the north and west of the Four Squatters Glacier. The upper clastic division is poorly preserved in the Dogtooth Range, and it is considered either a

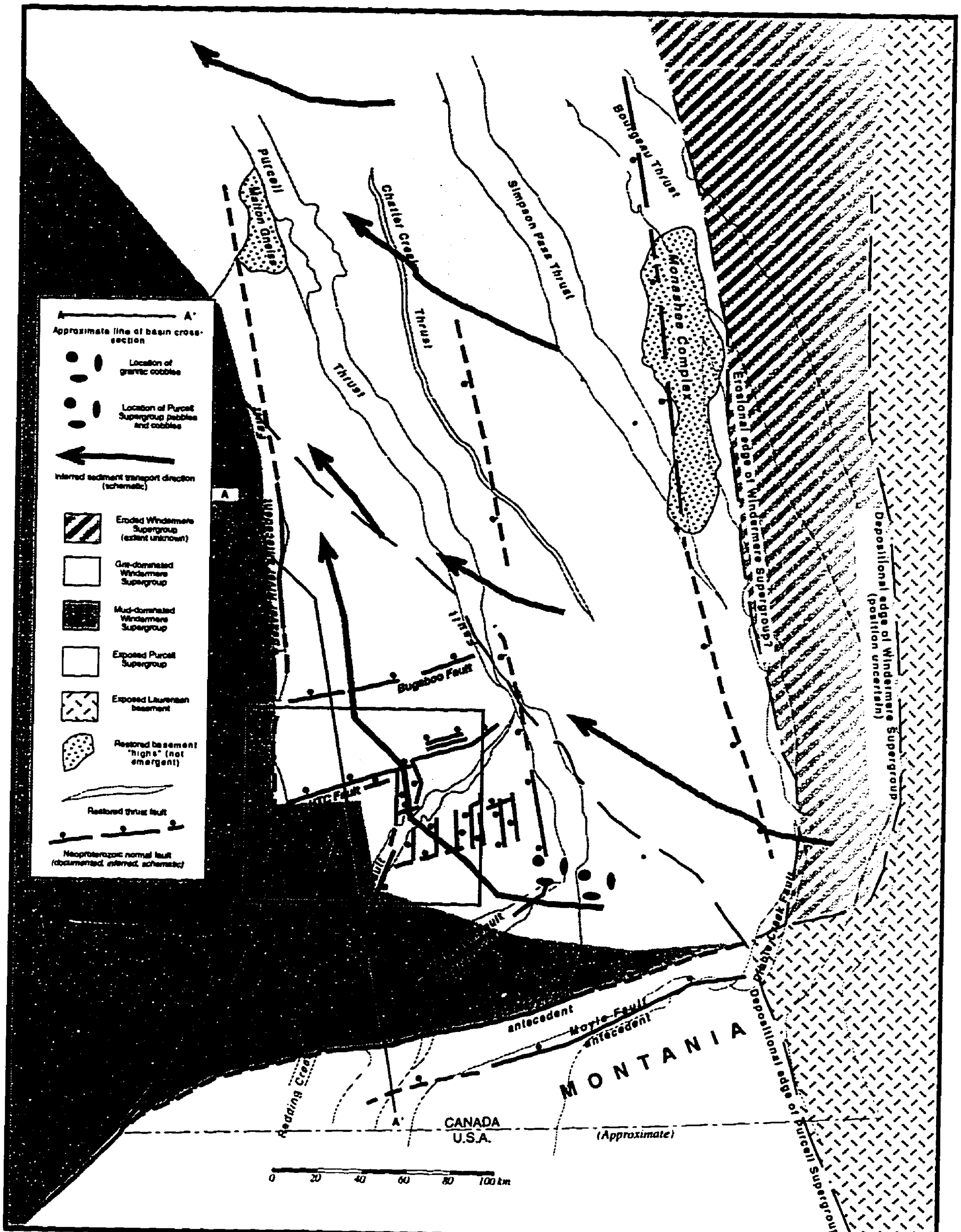



**Figure 2-8:** Paleogeographic interpretation for the southern Canadian Cordillera in lower Windermere (pre marker unit) time, shown on palinspastically restored base. Palinspastic map was constructed by restoring major thrust faults with respect to a "pin line" at the eastern limit of Cordilleran deformation, using restored cross sections of Price and Mountjoy (1970) and Price and Fermor (1985) for the Rocky Mountains, and restored cross sections and maps of Kubli and Simony (1992, 1995) and Warren (this study, Plate 4) for the Purcell anticlinorium (see Fig. 3-8). The Monashee Complex is assumed to be autochthonous. See text for sources and discussion of stratigraphic data. Square area represents approximate area of Plate 1. Basin cross section is shown in Figure 2-9.

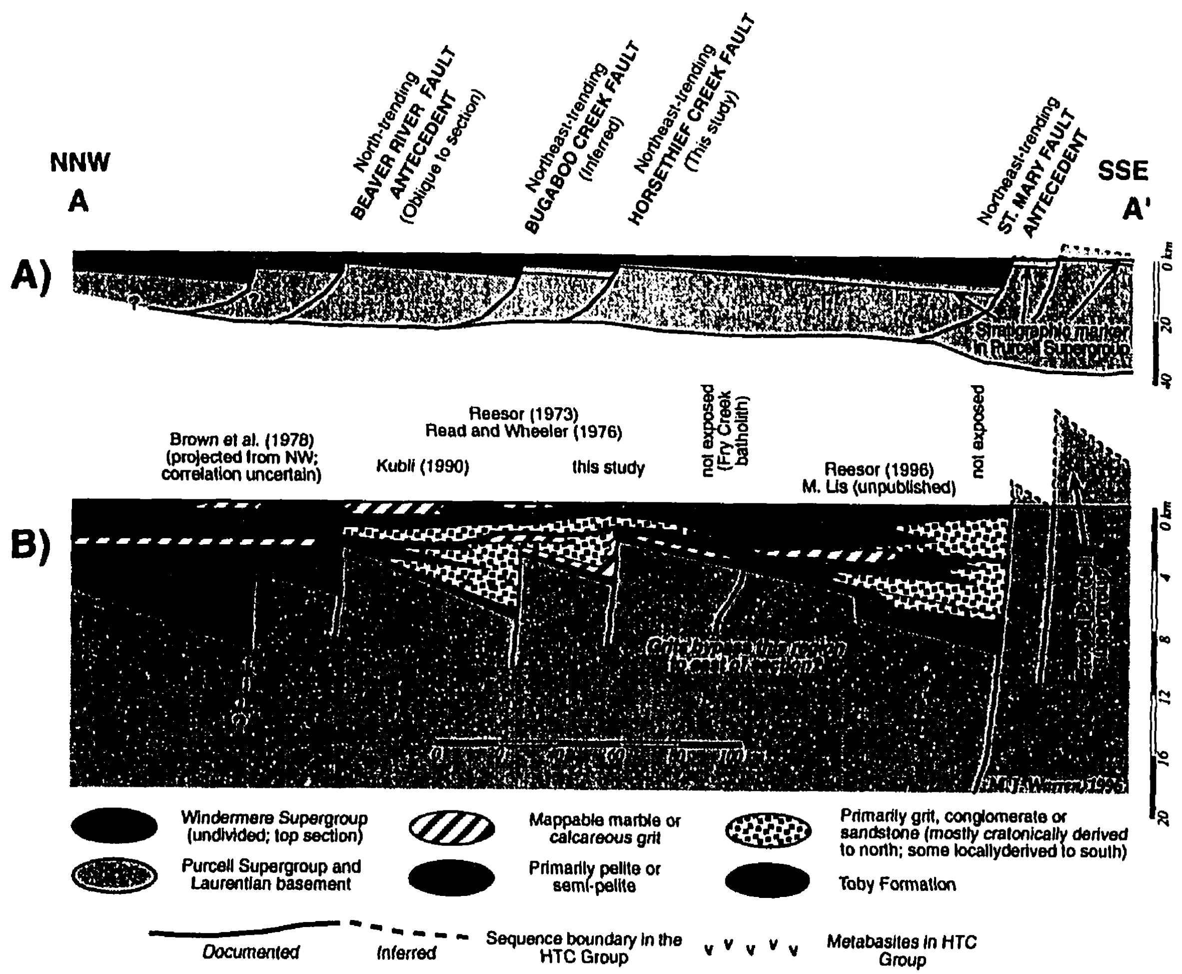



**Figure 2-9:** Cross section through the restored Windermere basin (see Fig. 2-8). A) No vertical exaggeration, showing inferred continental crustal attenuation from southeast to northwest. B) 5X vertical exaggeration, showing gross lithofacies changes. Coarse sediment derived from the craton to the east of the section is trapped by locall fault-controlled topography and mixed with sediment derived locally from the underlying Purcell Supergroup, particularly toward the south.

lateral facies equivalent of part of the carbonate division or, alternatively, a lateral facies equivalent of the basal unit of the Hamill Group (Kubli, 1990). The succession in the Dogtooth Range is therefore considered equivalent to the succession in the northwestern stratigraphic domain of this study.

Stratigraphic descriptions provided by other previous workers (Reesor, 1973; Read and Wheeler, 1976) suggest that the subdivisions of the Horsethief Creek Group are continuous between the study area and the Dogtooth Range, and strengthen the proposed correlations between the two areas. In the western Purcell anticlinorium to the north of Duncan Lake, the Horsethief Creek Group has been divided into a lower clastic unit of coarse grit and interbedded marble, and a finer upper clastic unit, separated by a marble marker (Read and Wheeler, 1976). The marble unit could be equivalent to the marker unit of this study and the Baird Brook division (Kubli, 1990), or to the marble that occurs at the top of the lower clastic sequence to the north of Howser Creek. To the north of the Bugaboo batholith, the northernmost exposures of the Toby Formation (Fig. 2-7) are overlain by dark limestone, in turn overlain by several hundered meters of interbedded grit and slate (Reesor, 1973). Helicopter reconnaisance suggests that the basal calcareous unit and lower clastic unit of this study area continue north of the Horsethief Creek batholith to this locality, and Evans' (1933) map shows that the grit division of the Dogtooth Range extends southward to where it directly overlies the Toby Formation in the immediate hangingwall of the Mount Forster fault. Bennett (1986) showed that the Horsethief Creek Group that is exposed in the Mount Forster syncline to the east of the Horsethief Creek batholith (Fig. 2-7) also comprises a basal calcareous interval, overlain by thickly interbedded grit and slate.

*Footwall of the Mount Forster fault*

The lower part of the Horsethief Creek Group that is exposed in the western Purcell anticlinorium in the hangingwall of the Mount Forster thrust can be correlated with the incomplete succession of Horsethief Creek Group strata that is exposed in the footwall of the Mount Forster thrust. The basal part of the Horsethief Creek Group in the eastern Purcell anticlinorium comprises either distinct dark limestone/marble and interbedded pelite that abruptly overlies the Toby Formation (Reesor, 1973; Atkinson, 1975; Root, 1987; Pope, 1989; Pope, 1990) or calcareous and arkosic grit that grades upward from the Toby Formation into calcareous sandstone, slate and marble (Pope, 1989). The calcareous succession is overlain by interbedded grit and slate.

Thinly bedded limestone, siltstone and red and green slate occur above this succession of intercalated grit and slate between lower Toby Creek and lower Horsethief Creek (Reesor, 1973; Root, 1987). This interval has been examined by Kubli (1990) and correlated with the Baird Brook division (marker unit) in the Dogtooth Range. Similar thinly bedded limestones have been reported by Pope (1989, 1990) near the upper limit of the section exposed on a ridge along strike to the south. The interval described by Reesor (1973) rests at nearly the same stratigraphic level (about 800 m) above the Toby Formation as does the marker unit in the Jumbo and upper Toby Creek watersheds.

Most of the Horsethief Creek Group exposed in the footwall of the Mount Forster thrust is thus considered to be equivalent to the lower clastic sequence exposed in the upper Jumbo Creek and upper Toby Creek watersheds, although the footwall sequence contains more grit and conglomerate. The uppermost part of the section described by Reesor (1973) and Root (1987) contains beds of grit and pebble conglomerate that coarsen and thicken upward abruptly above the "marker unit." It is thus reasonable to propose that the top of this section is equivalent to the base of the upper clastic sequence.

Many of the abrupt, small-scale changes in lithofacies and thickness in the lower part of the Horsethief Creek Group in the footwall of the Mount Forster thrust coincide with abrupt changes in the underlying Toby Formation and are related to syn-sedimentary faults that were active during the deposition of the Toby Formation and the lower Horsethief Creek Group (Root, 1987; Pope, 1989). The Horsethief Creek Group appears to contain more grit where it and the Toby Formation overlie down-dropped blocks of Mount Nelson Formation, rather than uplifted blocks of Dutch Creek Formation. None of these lithofacies or thickness changes is as conspicuous as those that occur in the Horsethief Creek Group to the north and west of Starbird Pass.

*Southwestern Purcell anticlinorium*

The belt of Windermere Supergroup strata that is exposed on the western flank of the Purcell anticlinorium (this study) is intruded by the Cretaceous Fry Creek batholith (see Plate 1 and Fig. 4-2), but continues southward beyond the batholith to Kootenay Lake as a west-dipping, homoclinal belt (Fig. 4-2). At Kootenay Lake it is offset by faults associated with the Purcell trench (e.g. Reesor, 1996). Within this belt, he Toby Formation, generally less than 200 m thick, overlies the Mount Nelson Formation, and it is overlain by 4000 to 8000 m of Horsethief Creek Group strata, which are in turn overlain by the basal grit of the Hamill Group (Reesor, 1996; M. Lis, unpublished field data). The Horsethief Creek Group in the northern part of this belt is nearly twice as thick as it is immediately to the north of the Fry Creek batholith. It also doubles in thickness over a distance of about 10 km southward from the batholith (M. Lis, unpublished field data).

The Horsethief Creek Group strata in the southernmost part of this belt contrasts conspicuosly with the Horsethief Creek Group in other parts of the Purcell anticlinorium, because it contains much more pelitic rock and fine-grained sandstone relative to grit and pebble conglomerate. The succession in the northern part of the belt comprises intercalated schist, grit, mature quartzite and minor marble. Mappable marble units occur locally near the bottom of the succession (Reesor, 1996), and a thin but continuous belt of grit and pebble conglomerate in the upper part of the succession could be equivalent to the base of the upper clastic sequence of this study. The thicker succession near Kootenay Lake contains more conspicuous grit intervals and at least two distinct carbonate cobble conglomerate intervals, both beneath and above a thick (1500 m) marble unit that appears to pinch out or become discontinuous to the north (Reesor, 1996; M. Lis, unpublished field data). The cobble conglomerate intervals contain clasts of Purcell

Supergroup strata that are exposed beneath Lower Cambrian strata in the footwall of the St. Mary fault to the southeast (Lis and Price, 1976).

The base of the Horsethief Creek Group in this area comprises a distinct succession of laterally continuous schist and overlying white quartzite that apparently coarsens upward and northward (Reesor, 1996; M. Lis, unpublished field data). This succession is truncated northward by a steep fault, and it is absent to the north of the fault (Reesor, 1996). The Toby Formation and the underlying Mount Nelson Formation are offset by this fault. The Toby Formation is thicker to the south of the fault, and it rests unconformably on a lower stratigraphic level within the Mount Nelson Formation to the north of the fault. Thus the fault was originally a south-side-down normal fault that was active during the deposition of the Toby Formation and the lower part of the Horsethief Creek Group. Reesor's (1996) map indicates that this fault probably had no more than a few hundred meters of stratigraphic separation.

According to Lis and Price (1976), the anomalously thick, laterally variable succession of Horsethief Creek Group strata that is exposed along the southeastern of Kootenay Lake was deposited on the down-dropped side of a major northeast-trending Neoproterozoic normal fault, an antecedent to the St. Mary fault, that was unconformably overlapped by Lower Cambrian quartz arenites (see Chapter 3). Up to 7 km of Purcell Supergroup strata were eroded from the uplifted footwall of this normal fault, and up to 9 km of Windermere Supergroup strata were deposited on the down-dropped side prior to deposition of Lower Cambrian strata. The Horsethief Creek Group northwest of the St. Mary fault includes intervals of conglomerate, both in the middle and near the top of the succession, that contain cobbles of Purcell Supergroup strata. The conglomerates indicate that the Purcell Supergroup was, at least intermittently, a source of sediment for the Windermere Supergroup in this area throughout the deposition of the Horsethief Creek Group, as well as during the deposition of the Toby Formation. The relative abundance of pelite, quartz schist and clean quartzite to arkosic grit in parts of this succession might also indicate fine-grained siliciclastic rocks of the Purcell Supergroup as a significant component of the sediment source.

The relationships described above suggest that the Horsethief Creek Group thins gradually northward and westward from the St. Mary fault to upper Horsethief Creek and Starbird Pass above a fault-bounded crustal block that was tilted to the southeast (Fig. 2-9). The syn-sedimentary Horsethief Creek-Starbird Pass normal fault of this study (Plates 1 and 2) was subparallel to the antecedent of the St. Mary fault and had the same sense of displacement (northwest-side-down). The smaller, south-side-down normal fault shown by Reesor (1996) to the north of the St. Mary fault is at a high angle to it and may have been one of the north-northwest-trending faults that were also active during the deposition of the Windermere Supergroup (Root, 1987; Pope, 1989; M. Lis, unpublished field data; and this study).

## DISCUSSION

### INITIAL CONFIGURATION OF THE WINDERMERE BASIN

A growing body of evidence from the Purcell Mountains shows that subsidence in this part of the Windermere basin was controlled by two sets of normal faults (Fig. 2-9): 1) a northeast-trending, northeast-dipping set (Atkinson, 1975; Lis and Price, 1976; Bennett, 1986; Root, 1987; Pope, 1989, 1990; and this study); and 2) a north-trending set of both west- and east-dipping faults. Faults of the northeast-trending set commonly were reactivated during Mesozoic compressive deformation as right-hand reverse faults. They are oblique to the dominant structural strike, and thus their effect on the distribution of the Windermere Supergroup and on lithofacies and thickness changes within it is apparent. The north-trending set are more nearly parallel to the regional tectonic strike and were reactivated more commonly during Early Paleozoic normal faulting (Chapter 3), as well as during Mesozoic thrust faulting. Thus they are more obscured by younger deformation, but they can be inferred on the basis of differences in the stratigraphic successions within and beneath the Windermere Supergroup in adjacent thrust panels (Root, 1987; Pope, 1989; this study).

The regional-scale map pattern defined by the contact between the Purcell Supergroup and the Windermere Supergroup in the Purcell anticlinorium (Figs. 2-2 and 2-7) thus reflects both sets of Neoproterozoic normal faults as well as Mesozoic north-trending folds that are parasitic to the Purcell anticlinorium. This contact "steps" northeastward across the Mesozoic axis of the Purcell anticlinorium, and the abrupt northeastward deflections correspond to thickness and facies changes in the Windermere Supergroup, and to changes in the level of erosion in the underlying Purcell Supergroup (St. Mary fault antecedent of Lis and Price, 1976 and "Horsethief Creek fault" of this study). I have inferred the presence of another northeast-trending normal fault at the latitude of Bugaboo Creek, immediately to the north of the study area ("Bugaboo fault" on Figs. 2-8 and 2-9), where the base of the Windermere Supergroup abruptly changes strike from north to northeast and is characterizied by notable coarse breccias and abrupt variations in lithofacies and thickness (Reesor, 1973, and pers. comm., 1993)

These relationships indicate that the Purcell Supergroup was attenuated by north- and northwest-dipping normal faults, and that a northwestward-thickening wedge of Windermere Supergroup was deposited on top of it. This interpretation is supported by the fact that no known exposures of the Purcell Supergroup or its underlying crystalline basement occur to the north of Bugaboo Creek or to the west of the Purcell anticlinorium (restored positions; see Fig. 2-9). This interpretation is also supported by the fact that provenance of Windermere Supergroup sediment in the Purcell anticlinorium records uplift and erosion of basin margins both to the south and to the east (Fig. 2-9). The Windermere Supergroup in the Purcell anticlinorium is underlain by the Purcell Supergroup wherever the base of the Windermere Supergroup is exposed, but many of the facies changes across syn-sedimentary faults are defined by thick accumulations of coarse, arkosic sediment that clearly was derived from the Laurentian craton to the east of the Purcell Supergroup. This sediment was presumably eroded and rapidly transported to the basin, where it became locally trapped by fault-related topography.

However, the Windermere Supergroup succession in the hangingwall of the St. Mary fault contains cobbles and small boulders at the base, middle and top of the succession (Lis and Price,

1976) that clearly were locally derived from the underlying Purcell Supergroup. The Purcell Supergroup was uplifted and bevelled in the footwall of the St. Mary fault as much as 7 km, prior to deposition of Upper Cambrian strata (Lis and Price, 1976; M. Lis, unpublished field data). This succession of Windermere Supergroup may be anomalous and difficult to correlate regionally because it was derived primarily from fine-grained siliciclastic and carbonate strata, and thus contains more quartz sand and mud, and because recurring uplift and erosion in the footwall of this major fault may have obscured the regional highstand event recorded elsewhere by the regional marker unit.

Stratigraphic evidence from the northern Canadian Cordillera (e.g. Eisbacher, 1981; Cecile et al., 1997) also shows that NE-trending normal faults controlled sedimentation at the base of the Windermere Supergroup and in younger strata along much of the strike length of the Canadian Cordillera. Eisbacher (1982), and other workers in the southern Canadian Cordillera (e.g. Price, 1981; Ross, 1991a) have concluded that transverse normal faults in the Mackenzie Mountains and in the Purcell Mountains, respectively, have been inherited from antecedent structures in the underlying Hudsonian basement, which has a northeast-trending structural grain (Ross, 1991a). These previous anisotropies were thus favorably oriented for reactivation during Neoproterozoic and younger extension.

## TECTONIC SETTING AND HISTORY OF THE WINDERMERE SUPERGROUP IN THE CENTRAL PURCELL MOUNTAINS

The Horsethief Creek Group in the Purcell Mountains records two pulses of submarine fan progradation (the lower and upper clastic sequences) in a regional, deeply-subsided basin. These two pulses of thick sediment accumulation are separated by a starved interval (the marker unit), interpreted as a basin-wide relative sea-level highstand. This interpretation is in accordance with previous interpretations of the regional depositional setting of the Horsethief Creek Group and equivalent strata in the southern Canadian Cordillera (Poulton and Simony, 1970; Pell and Simony, 1987; Root, 1987; Ross and Murphy, 1988; Pope, 1989; Kubli, 1990).

The Toby Formation and the overlying lower clastic sequence of the Horsethief Creek Group clearly were deposited in an active, intracontinental rift setting, as demonstrated previously by many other workers. Extension and mafic volcanism began immediately prior to the deposition of the Toby Formation (Bennett, 1985; Root, 1987; Pope, 1989). During the depostion of the Toby Formation, the basin began to subside on a regional scale to the north of the St. Mary fault and to the west of a depositional edge that is no longer preserved (see Fig. 2-9), but that must have been abrupt. Rifting was accompanied by submarine volcanism, and deposition and reworking of primarily locally-derived glaciomarine sediments on submarine slopes. Normal faulting continued during deposition of the lowermost part of the Horsethief Creek Group, which marks the first regional influx of detritus derived from the craton to the east. This observation perhaps indicates that the faults that defined the uplifted basin margin stepped eastward with time or, alternatively, that the transport of sediment from the craton to the basin had been blocked during the deposition of the Toby Formation.

The green argillite unit and the overlying carbonate marker unit record a marked decrease in the amount of coarse clastic sediment supplied to the basin, interpreted as evidence for a basin-wide sea-level high stand (Ross and Murphy, 1988; Kubli, 1990). Several workers attribute this event to melting of globally widespread Neoproterozoic ice sheets (Ross et al., 1995, and references therein). Two glaciations are recorded by diamictites in the Windermere Supergroup of the Mackenzie Mountains in the northern Canadian Cordillera (Aitken, 1991). The Rapitan Formation, at the base of the Windermere Supergroup, is probably equivalent to the Toby Formation (Ross et al., 1995). The Ice Brook Formation, which is stratigraphically higher, records a glaciation that could have been a precursor to the highstand event recorded by the marker unit in the southern Canadian Cordillera. The Vreeland Formation in the southern Rocky Mountains is a diamictite that occurs at a stratigraphic position close to that of the Old Fort Point Formation in the Miette Group (McMechan, 1996), a proposed equivalent of the regional marker unit in the Purcell Moutains (Ross and Murphy, 1988). A third glaciation, for which there is no direct lithostratigraphic evidence, is revealed by chemostratigraphic data from the upper part of the Windermere Supergroup in the Mackenzie Mountains (Kaufman et al., 1997).

The regional highstand was followed by a relative sea-level fall and an abrupt influx of coarse cratonic sediment (the upper clastic sequence). The tectonic setting of the upper clastic sequence is more controversial. Pell and Simony (1987) argued that the upper clastic sequence recorded a second episode of crustal extension. More recently, Ross (1991) has argued that the upper Windermere Supergroup was deposited on a passive continental margin following a major eustatic sea level fall, and that the local unconformity or sequence boundary beneath the upper clastic sequence could represent the rift-to-drift transition (G. Ross, pers. comm., 1994). Certainly, the more significant episode of crustal attenuation, which defined the regional Windermere basin, occurred during the deposition of the Toby Formation and at least come of the lower clastic sequence.

However, several observations cast doubt on a passive margin setting for the upper part of the Windermere Supergroup, although they are by no means conclusive:

1) The deposition of the upper clastic sequence in the west-central Purcell Moutains was accompanied by local normal faulting that caused local tilting and bevelling of the lower clastic sequence beneath the upper clastic sequence (Fig. 2-5), and perhaps by mafic volcanism. Stratigraphic evidence from the southwestern Purcell Mountains (Lis and Price, 1976) and from the upper part of the Middle Miette Group (Fig. 2-2) in the Western Main Ranges of the Rocky Mountains (Carey and Simony, 1985) for syn-sedimentary normal-faulting supports the argument that the upper part of the Windermere Supergroup was not deposited in a tectonically quiescent setting.

2) The grits in the upper part of the Windermere Supergroup are more than a kilometer thick and were deposited as a sheet of coarse, immature sediment that extends northward from 50°N to at least 62°N latitude, and westward from the Rocky Mountains to a restored width of at least 300km (my estimate). The magnitude of the eustatic sea level fall required for these submarine fan facies to prograde over that much of the basin, and for the basin to accumulate so

much coarse sediment must be very great. Perhaps one of the Neoproterozoic glaciations could account for so significant a sea-level fall. However, if the correlations discussed above are valid, then the third glacial event identified by Kaufman et al. (1997) does not occur until apparently higher(?) in the Windermere succession. Furthermore, the gravels on the northern United States Atlantic passive margin that are associated with the Pleistocene glacial lowstand are only a maximum of about 150 m thick (Riggs and Belknap, 1988), an order of magnitude less than much of the upper Windermere grit.

3) There is no well-preserved Neoproterozoic passive margin "miogeoclinal" succession similar to the Lower Paleozoic succession of the Canadian Cordillera or the Mesozoic succession of the Atlantic Ocean. It could be argued that a significant part of the Windermere Supergroup has been removed beneath a significant sub-Cambrian unconformity, and Ross (1991) argued that the shelf facies of the Windermere passive margin were deposited largely to the east of the present exposires of the Windermere Supergoup in the southern Canadian Cordillera. However, constraints on the timing of deposition of the Windermere Supergroup and overlying strata, discussed below and in Chapter 3, imply that the time gap between the Windermere Supergroup and younger strata is not everywhere large, and there is evidence locally that sedimentation may have been uninterrupted.

4) The sedimentological model of Ross et al. (1995) for the Windermere grits involves a deep-sea longitudinal sediment transport system, that developed seaward of a shelf and slope that are not exposed in the southern Canadian Cordillera. If the deep-water transport system is a valid depositional model, I would nonetheless argue that the Horsethief Creek Group grits in the Purcell Mountains were deposited on significantly attenuated continental crust, rather than oceanic crust, since the continental "basement" (the Purcell Supergroup) that floored the deep-water environment is exposed in this area. Furthermore, the Windermere Supergroup in the Purcell, Rocky, Selkirk and Cariboo mountains is overlain by shallow-water siliciclastic and carbonate shelf facies that would not have been deposited regionally over oceanic crust. Thus if

there was oceanic crust at the latitude of the douthern Canadian Cordillera during Windermere time, it lay westward of the restored position of the presently exposed Windermere Supergroup.

**DURATION OF DEPOSITION OF THE WINDERMERE SUPERGROUP**

It is important to point out that the deposition of the apparently rapidly-deposited Windermere Supergroup must have spanned more than 200 million years, the tectonically enigmatic time between the well-documented Neoproterozoic initiation of intracontinental rifting and the well-documented Early Cambrian initiation of thermal subsidence. The possibility of previously unrecognized, significant time gaps within the Windermere Supergroup should perhaps be considered and taken into account in regional correlations.

The onset of Neoproterozoic rifting was marked by mafic to intermediate volcanism and associated intrusions at the base of the Windermere Supergroup. These rocks have yielded Rb-Sr and U-Pb zircon ages of 780-770 Ma in the northern Canadian Cordillera (Armstrong et al., 1982; Roots, 1983; Jefferson and Parrish, 1989) and a Sm-Nd whole rock age of 762+/-44 Ma in northern Washington state (Devlin et al., 1985). Three distinct assemblages of Neoproterozoic Ediacaran fauna occur in the upper 3 km of the Windermere Supergroup succession in the Mackenzie Mountains of the northern Canadian Cordillera (Narbonne et al., 1994; Narbonne and Aitken, 1995), and Ediacaran fauna also occur at the top of the succession in the Cariboo and Rocky Mountains (Fig. 2-2) of the southern Canadian Cordillera (Hofmann et al., 1985; Ferguson and Simony, 1991). New geochronological and chemostratigraphic constraints on the age of the Precambrian-Cambrian boundary and on the duration of Ediacara (Grotzinger et al., 1995) indicate that most Ediacaran fauna worldwide (equivalent to the upper two assemblages in the Mackenzie Mountains; Narbonne et al., 1994) flourished after a maximum age of 550 Ma, just six million years before the beginning of Cambrian time. It is not certain whether the Ediacaran fauna that occur in the southern Canadian Cordillera are included in the pre- or post-550 Ma assemblages (G. Narbonne, pers comm., 1997). Thus the deposition of the Windermere Supergroup succession in the northern Canadian Cordillera must have continued to at least 550

Ma and perhaps to within a few million years of the end of Neoproterozoic time, and it might have continued for as long in the southern Canadian Cordillera.

**POSSIBLE PLATE TECTONIC IMPLICATIONS**

Previously published paleomagnetic, sedimentological and stratigraphic data from the northern Canadian Cordillera and from the Australian craton imply that continental separation between the northwestern margin of Laurentia and the eastern margin of Australia had occurred by a maximum of 750-720 Ma and by a minimum of about 600 Ma (Powell et al., 1993; Powell et al., 1994; Storey, 1993; Dalrymple and Narbonne, 1996; Pelechaty et al., 1996; Torsvik et al., 1996). Other stratigraphic and sedimentological data have been cited as evidence for earliest Cambrian continental separation (Bond and Kominz, 1984; Bond et al., 1985; Devlin and Bond, 1988). Ross (1991) has argued that continental separation and development of a passive margin in Neoproterozoic time could have been followed by a second episode of continental extension and separation of another continental fragment or fragments in earliest Cambrian time. The new data presented above do not necessarily conflict with the evidence for continental separation in Neoproterzoic time, but they do suggest that local extension, normal faulting and volcanism continued during upper Windermere time in the southern Canadian Cordillera.

I propose the following hypothesis as a possible reconciliation of new and previously published data and conclusions: birth of the western margin of Laurentia and perhaps the initiation of seafloor spreading occurred diachronously from north to south along the northwestern Laurentian margin. It is probably overly simplistic to visualize the entire length of the Cordilleran margin as having formed simultaneously by a single rifting event between only two continental blocks. Mesozoic opening of the northern Atlantic Ocean in eastern Canada occurred diachronously from south to north and spanned more than 100 Ma (Keen et al., 1990), although the pre-drift extension did not last for more than about 20 Ma in any one locality. Separation occurred between Africa and North America in Early to Middle Jurassic time, to the south, but separation between North America and Iberia did not occur until mid-Cretaceous time or later, to the north. An active transform fault separated the Africa-North America rift axis from the Iberia-North America rift axis.

Perhaps a similar pattern of tectonic evolution occurred on the western margin of Laurentia in Neoproterozoic to Early Cambrian time: intracontinental rfiting began along the entire Canadian Cordilleran margin at about 780-750 Ma, followed by separation of a continental block from the northern Cordilleran margin at about 750-720 Ma. Subsequent open ocean circulation along the northern Cordilleran margin is recorded by contourites in the upper Windermere Supergroup of the Mackenzie Mountains (Dalrymple and Narbonne, 1996). However, intracontinental extension continued intermittently in the southern Canadian Cordillera (upper clastic sequence), and did not culminate in continental separation there until Earliest Cambrian time, when a second continental block was detached from Laurentia. Some recent plate reconstructions for Neoproterozoic time show that the South China continental block, as well as Australia, might have been connected to northwestern North America, and that it might have separated at a different time than Australia from Laurentia (Li et al., 1996; Pelechaty, 1996).

The Liard Line is a major northeast-trending transverse lineament that divides the southern Canadian Cordillera from the northern Canadian Cordillera and across which there are fundamental differences in the configuration of the Neoproterozoic to Lower Paleozoic margin (Cecile et al., 1997). This structure has been interpreted as a previous transfer fault zone across which there was a reversal in the asymmetry of the rifted margin (Cecile et al., 1997). Perhaps the Liard Line, or an equivalent transverse structure, accommodated differential plate motions between Laurentia and crustal blocks to the west during diachronous rifting.

## CONCLUSIONS

The Windermere Supergroup of the west-central Purcell Mountains comprises two contrasting stratigraphic domains: a southern/eastern domain characterized by a thinner succession of homogeneous and laterally continuous units, and a northern/western domain characterized by a thicker succession of more heterogeneous and laterally variable units. Stratigraphic relationships show that these two domains are linked across a zone of abrupt thickness and facies changes, controlled by a northeast-trending syn-depositional fault zone. In both domains Horsethief Creek Group strata comprise two distinct clastic sequences, separated by a regional carbonate-bearing marker unit and an unconformity at the base of the upper clastic sequence, which possibly represents a significant time gap.

These relationships provide a stratigraphic link, and the basis for confident correlation, between the base of the Windermere succession exposed in the vicinity of its type locality, in the footwall of the Mount Forster fault (Walker, 1926; Reesor, 1973; Root, 1987; Pope, 1989, 1990), and the Horsethief Creek Group exposed in the northwestern Purcell Mountains and Dogtooth Range, in the hangingwall of the Mount Forster thrust (Evans, 1933; Reesor, 1973; Read and Wheeler, 1976; Kubli, 1990). These relationships also strengthen previously proposed correlations between the Horsethief Creek Group of the Purcell Mountains and the Miette Group of the Rocky Mountains (e.g. Pell and Simony, 1987; Ross and Murphy, 1988).

Two pulses of tectonic activity controlled the sedimentation of the Windermere Supergroup in the Purcell Mountains. The first began immediately prior to the deposition of the Toby Formation (Atkinson, 1975; Bennett 1986; Root, 1987; Pope, 1989), and continued during deposition of at least the lowermost part of the lower clastic sequence of the Horsethief Creek Group (Root, 1990, and this study). This initial subsidence of the Windermere basin was controlled by two sets of extensional faults: a N-trending, primarily west-side-down set (Root, 1987; Pope, 1989), and a northeast-trending, northwest-side-down set (Atkinson, 1975; Bennett, 1986; Root, 1987; this study). The northeast-trending set is clearly expressed in present map-patterns and in the distribution of Windermere Supergroup rocks at a variety of scales throughout the central Purcell anticlinorium, and it accommodated significant upper crustal stretching along deep, listric normal faults, from southeast to northwest.

The second period of tectonism was recorded locally by block-faulting, minor volcanism and rapid influx of a large volume of coarse, immature sediment into the basin. I have argued that a eustatic sea level fall alone cannot account for such a large thickness of grit over such a large area of the southern Canadian Cordillera, and that the base of the upper clastic sequence records renewed regional tectonic uplift of the basin margins.

Thus, in this interpretation, the upper part of the Windermere Supergroup in the southern Canadian Cordillera does not simply record sedimentation on a passive continental margin following continental separation in Neoproterozoic time, although sedimentological, stratigraphic and paleomagnetic data from the northern Canadian Cordillera and the Australian craton imply that continental separation between the northwestern margin of Laurentia and the eastern margin of Australia had occurred between 750-720 Ma and about 600 Ma. I propose that seafloor spreading might have occurred diachronously from north to south, and involved separation of more than one continental block from north to south, as it did from south to north in the northern Atlantic Ocean. Or, alternatively, continental separation by 720 Ma along the entire length of the Canadian Cordillera was followed by rifting and separation of a second continental fragment in upper Windermere time in the southern, but not the northern, Canadian Cordillera. Further data from the Cordilleran rifted margin and investigation of analogs in recent or modern rift systems are required to test these ideas.

# Chapter 3

# A REGIONAL "BREAK-UP" UNCONFORMITY WITHIN THE NEOPROTEROZOIC(?) TO LOWER CAMBRIAN HAMILL GROUP, S.E. CANADIAN CORDILLERA, AND IMPLICATIONS FOR SYN- AND POST-RIFT BASIN CONFIGURATION AND REGIONAL PALEOGEOGRAPHY

## ABSTRACT

A regionally significant unconformity within the Neoproterozoic to Lower Cambrian Hamill and Gog Groups of southeastern British Columbia records the transition from active rifting of an extending continental margin to thermally-driven subsidence of a passive margin. This unconformity has been identified in the west-central Purcell anticlinorium and in the adjacent Kootenay Arc structural domain in southeastern British Columbia. The uppermost, shallow marine quartzarenite/quartzite unit of the Hamill Group rests unconformably either on underlying fluvial and/or shallow marine units of the lower Hamill Group, which are sedimentologically distinct from it, or else it directly overlies the Neoproterozoic Windermere Supergroup. The lower units of the Hamill Group, beneath the unconformity, become thin and pinch out to the west in the Purcell anticlinorium; to the east they thicken and coarsen and are abruptly truncated by a Neoproterozoic to Cambrian fault that was a west-side-down normal fault but has been reactivated as a Mesozoic thrust fault. To the west, in the Kootenay Arc, the upper quartzite unit rests on a distinctly different sequence of lower Hamill Group strata that contains mafic volcanic rocks. These relationships and other sedimentological data indicate that the lower part of HamilllGroup was deposited in isolated basins that were bounded to the east by syn-depositional, west-dipping normal faults.

The unconformity within the Hamill and Gog Groups and equivalent strata is a regionally significant feature. Strata above this unconformity are more regionally continuous and contain primarily Waucoban Lower Cambrian faunas. Below this unconformity stratigraphic units are discontinuous, contain less mature sediment and locally include mafic to intermediate volcanic rocks. These generally unfossiliferous strata are no younger than Lowermost Cambrian (Placentian), and they could be primarily Neoproterozoic. Palinspastically restored thickness data from the lower parts of the Hamill/Gog Groups show that these strata were deposited in at least three shallow, north-trending basins that were controlled primarily by west-dipping normal faults and were superposed on much thicker accumulations of the Neoproterozoic Winderemere Supergroup. This episode of extension and normal faulting was followed by deposition of a more continuous, westward-thickening passive margin sequence (upper Hamill/Gog and overlying Lower Cambrian strata) on a subsiding continental margin. The regional unconformity within the Hamill/Gog Groups is thus interpreted as the stratigraphic expression of the rift-to-drift transition. Its age is bracketed between 549 and about 520 Ma, which is consistent with the latest Proterozoic to Earliest Cambrian continental separation and initiation of seafloor spreading adjacent to the western margin of ancestral North America proposed in several recent models.

## INTRODUCTION

The nature and timing of the continental extension and subsequent transition to the seafloor spreading and continental drift that initiated the Early Paleozoic passive margin in the southern Canadian Cordillera remain critical and controversial questions. Arguments for continental separation along the western margin of Laurentia during both Neoproterozoic and Earliest Cambrian time are supported by geophysical, geochronologic, lithostratigraphic, biostratigraphic and chemostratigrahic data. Paleomagnetic, geochronologic and stratigraphic data have been cited to show that Laurentia, Australia and Antarctica were linked prior to 750 Ma and that continental separation between the northwestern margin of Laurentia and Australia/Antartica occurred by about 750-720 Ma (e.g. Powell et al., 1993, 1994). Most workers agree that the deposition of at least the lower part of the Neoproterozoic Windermere Supergroup occurred during an episode of intracontinental rifting that began at about 780-750 Ma in the Canadian Cordillera (e.g. Stewart, 1972; Bond and Kominz, 1984; Devlin and Bond, 1988, Ross, 1991b, and references therein). The tectonic setting of the upper part of the Windermere Supergroup remains enigmatic, particularly in the southern Canadian Cordillera (see Chapter 2). It has been interpreted to record a second episode of regional uplift related to intracontinental rifting (e.g. Pell and Simony, 1987), but more recently it has been interpreted to record deposition on the shelf and slope of a poorly-preserved Neoproterozoic passive margin following a major eustatic sea-level fall (e.g. Ross and Murphy, 1988; Ross, 1991b; Dalrymple and Narbonne, 1996). I have argued in Chapter 2 of this thesis that the upper part of the Windermere Supergroup, at least in the southern Canadian Cordillera, does not simply record deposition on a newly separated passive continental margin. I have suggested that perhaps separation of one or more continental fragments occurred in Neoproterozoic time in the southern Canadian Cordillera, following the initial continental separation and opening of the proto-Pacific ocean, or, alternatively, that final continental separation occured diachronously along the Cordilleran margin as it did along the Mesozoic Atlantic margin.

The strongest arguments for an episode of continental separation in latest Proterozoic to Earliest Cambrian time are based on analysis of tectonic subsidence rates recorded by the accumulation of westward-thickening Early Paleozoic sedimentary rocks of the southern Canadian Cordilleran "miogeocline." Tectonic subsidence curves constructed by "backstripping" the accumulation of these strata over time (Bond and Kominz, 1984; Bond et al, 1985) show an exponential decrease in tectonic subsidence with time and can be extrapolated to indicate that a rift-to-drift transition to margin-wide, thermally-driven subsidence (McKenzie, 1978) occurred in latest Proterozoic to Early Cambrian time.

Tectonic models of continental rifting predict that this rift-to-drift transition should also be recorded in the stratigraphic record as a regional "breakup" unconformity (Falvey, 1974; McKenzie, 1978). "Breakup" unconformities separate generally immature clastic and volcanic rocks that were deposited in fault-bounded basins from overlying mature, laterally continuous clastic strata that were deposited across these basins and across upthrown basement blocks after normal faulting had ceased. Break-up unconformities have been documented on a regional scale in Mesozoic sediments of the modern Atlantic margins (e. g. Balkwill, 1987; Mauffret and Montadert, 1987; Klitgord et al., 1988), as well as on ancient deformed continental margins such as the pre-Appalachian Neoproterozoic/Early Cambrian margin of eastern Laurentia (Hiscott and Williams, 1987; Simpson and Eriksson, 1989).

However, prior to this study, an unambiguous, regionally extensive "breakup" unconformity had not been identified in the latest Neoproterozoic to Lower Cambrian stratigraphic record of the southern Canadian Cordillera. A regional, abrupt lithological change between deep-water, immature and commonly coarse siliciclastic rocks of the Neoproterozoic Windermere Supergroup and overlying shallow-water, more mature quartzose sedimentary rocks of the Neoproterozoic to Lower Cambrian Hamill and Gog Groups marks the base of the "miogeoclinal" succession (Fig. 3-1), but the nature and significance of this stratigraphic

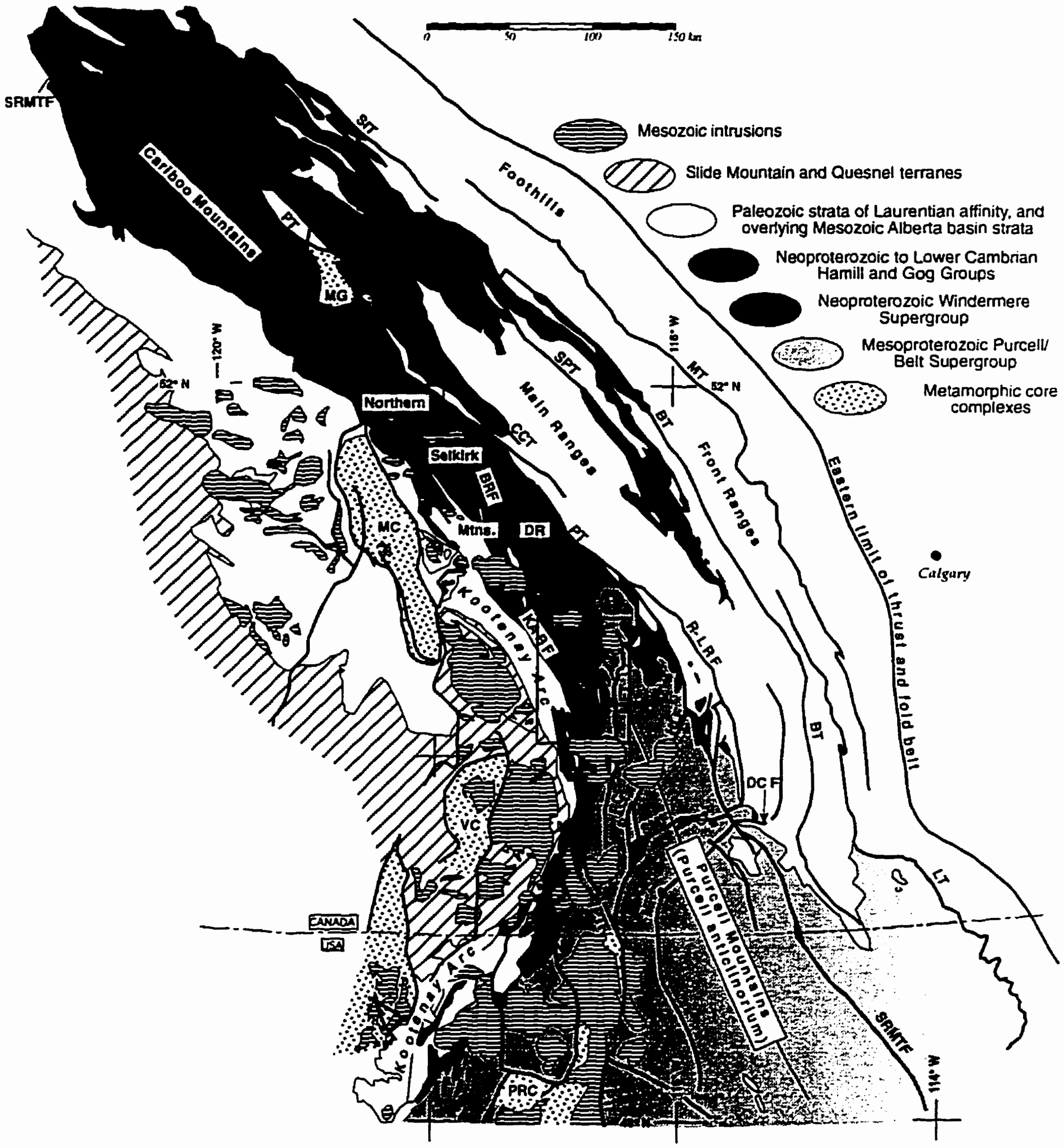



**Figure 3-1:** Simplified geological map of the southeastern Canadian Cordillera, showing regional distribution of exposures of the Hamill Group and equivalent strata, and major structures and physiographic provinces discussed in the text. After Wheeler and McFeeley (1991). Small rectangle = study area (Plate 2). Abbreviations: **MT** = McConnell thrust; **BT** = Bourgeau thrust; **LT** = Lewis thrust; **SIT** = Snake Indian thrust; **SPT** = Simpson Pass thrust; **CCT** = Chatter Creek thrust; **PT** = Purcell thrust; **R-LRF** = Redwall-Lussier River fault; **DCF** = Dibble Creek fault; **MF** = Moyie fault; **SMF** = St. Mary fault; **HLF** = Hall Lake fault; **RCF** = Redding Creek fault; **MFT** = Mount Forster thrust; **SRMTF** = Southern Rocky Mountain Trench fault; **PTF** = Purcell Trench fault; **BRF** = Beaver River fault; **KABF** = Kootenay Arc boundary fault; **PRC** = Priest River Complex; **VC** = Valhalla Complex; **MC** = Monashee Complex; **MG** = Malton gneiss; **DR** = Dogtooth Range.

boundary have been controversial. Many previous workers in the southern Canadian Cordillera (e.g. Mountjoy, 1962; Aitken, 1969) have concluded that the contact is a significant regional angular unconformity. However, many other workers have described both vertical and lateral gradations between the Windermere Supergroup and overlying rocks (e.g. Little, 1960; Hein, 1982; Devlin, 1986; Warren and Price, 1992, 1993; Urlwin and Ross, 1996). Furthermore, regional stratigraphic correlations suggest that the Hamill/Gog Groups rest on the same stratigraphic level within the Windermere Supergroup (pelite of the upper Horsethief Creek Group and slate of the Upper Miette Group) over much of their exposed extent (Fig. 3-2) and, finally, mafic to intermediate volcanic rocks lie above, not below, this lithological boundary (Wheeler, 1963; Devlin and Bond, 1988; Devlin, 1989; Lickorish and Simony, 1995; Warren and Price, 1993; Logan and Colpron, 1995).

At least two previous studies implied that the rift-to-drift transition is recorded slightly higher in the stratigraphic record. Devlin (1989) cited sedimentological evidence to argue that the rift-to-drift transition lies within the quartzites of the Hamill Group, but did not map a corresponding "breakup" unconformity. Lickorish and Simony (1995) demonstrated that there is an unconformity within the quartzose strata of the Gog Group in the Western Main Ranges of the Rocky Mountains, and that it overlaps lower Gog Group strata that were deposited in a fault-bounded basin, as well as uplifted and bevelled blocks of Windermere Supergroup. However, the regional extent of this unconformity and its tectonic significance were uncertain.

The question of the rift-to-drift transition is further complicated by the fact that the outboard portion of the Laurentian margin in the southern Canadian Cordillera records continued intermittent normal faulting, local uplift and erosion on several crustal "highs" (e. g. Deiss, 1941; Reesor, 1973; Kubli and Simony, 1992), intermittent volcanism, and deep subsidence and accumulation of coarse clastic sediments to the west (Lower Paleozoic Lardeau Group) throughout Lower Paleozoic time. These events not only have obscured some of the previous stratigraphic record, but also raise questions about the tectonic setting or settings of the outboard strata that are equivalent to the Lower Paleozoic "miogeocline," or passive margin

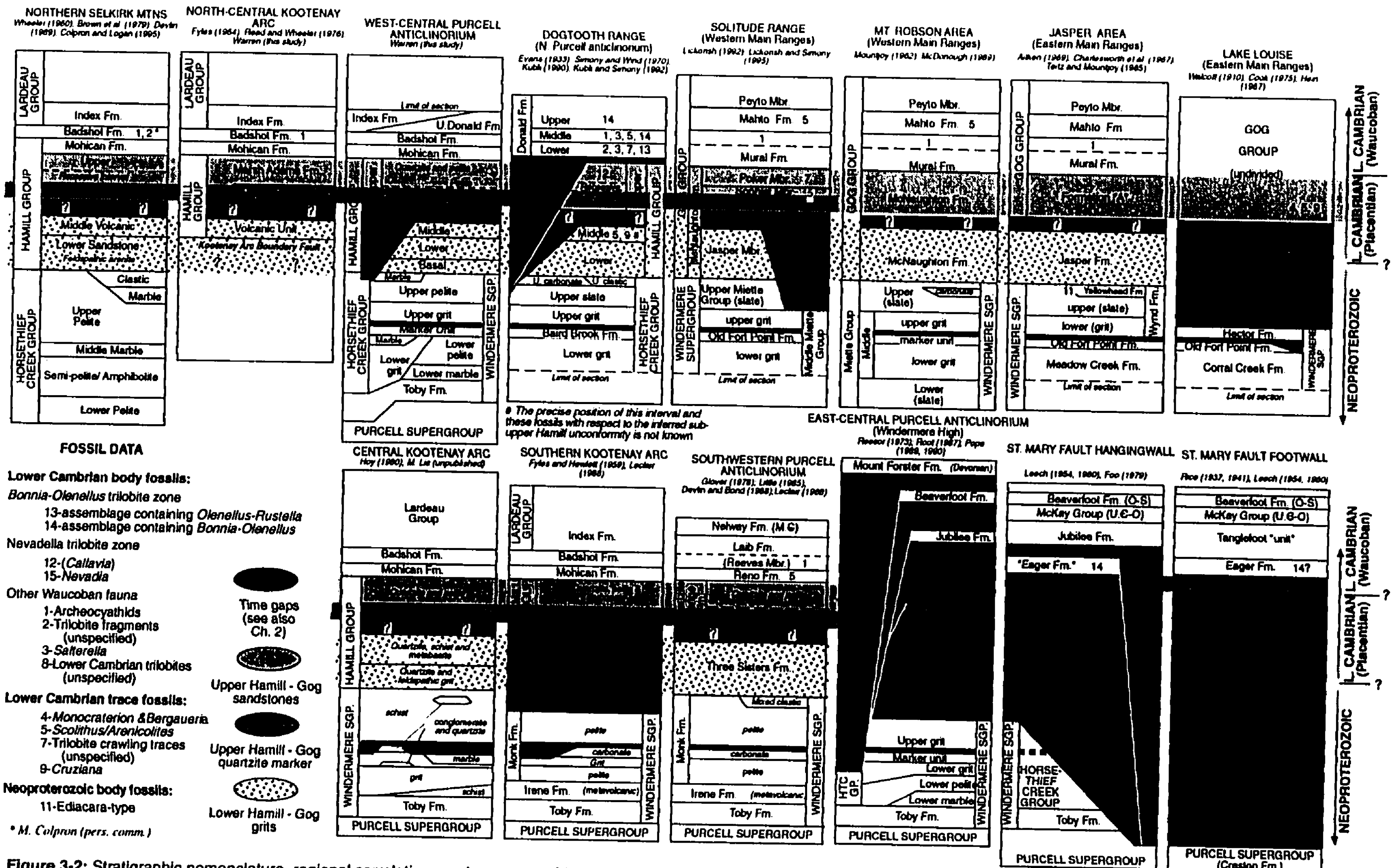



**Figure 3-2:** Stratigraphic nomenclature, regional correlations and summary of fossil data for Neoproterozoic and Lower Cambrian strata, southeastern Canadian Cordillera. Vertical axis not to scale. Many of the fossil data are discussed in Fritz et al. (1991), as well as in references above. The fossil *Callavia* is now known to be a proto-Atlantic fauna, and thus these specimens were misidentified in the southern Cordillera, but are nonetheless *Nevadella*-zone fauna (G. Narbonne, C. Barnes, pers. comm., 1996). "Waucoban" is used in the sense of Fritz et al. (1991). Place names are located on Fig. 1-2.

succession discussed by Bond and Kominz (1984). The stratigraphic record in the western part of the "miogeocline" is not consistent with the predicted behavior of a thermally-subsiding passive margin as recorded by equivalent rocks in the eastern part.

This chapter will address the question of the stratigraphic position and tectonic significance of the "breakup" unconformity and rift-to-drift transition in two parts. In Part I, I will present stratigraphic and sedimentological evidence from the west-central Purcell Mountains of southeastern British Columbia that shows that there is a significant unconformity within, rather than beneath, the previously undivided Hamill Group. I will demonstrate that the lower part of the Hamill Group was deposited in distinct half grabens that were bounded by normal faults that have since been reactivated as thrust faults. I will argue that the upper part of the Hamill Group and overlying Lower Cambrian strata were deposited continuously across these basins after normal faulting had ceased and before it was renewed later in Cambrian time.

In Part II of this chapter, I will use these new data from the western Purcell Mountains to provide a critical link to other tectonostratigraphic domains of the southern Canadian Cordillera and to demonstrate the regional extent and significance of this unconformity. I will argue that the unconformity within the Hamill Group is distinct on a regional scale from the local unconformity at the base of the Hamill Group and that it is a clear stratigraphic expression of a rift-to-drift transition as inferred by Bond and Kominz (1984) and Bond et al. (1985). The unconformities beneath and within the Hamill Group record time gaps that may be no more than a few million years and could be substantially less than that in some parts of the basin. These unconformities are interpreted to mark the initiation and cessation, respectively, of a brief period of regional extension and normal faulting that differed markedly from that during which the underlying Windermere Supergroup was deposited and that culminated in final continental separation and the initiation of thermal subsidence on this part of the western margin of Laurentia. Subsequent Cambrian normal faulting, local uplift and erosion, and deep subsidence to the west are interpreted to record differential thermal subsidence of underlying crustal blocks of varying thickness, and reactivation of faults that were active during earlier (syn-Windermere and syn-lower Hamill) crustal attenuation.

The new data from the west-central Purcell Mountains thus provide a basis for re-examination and re-interpretation of relationships at the Neoproterozoic-Paleozoic boundary in the southern Canadian Cordillera, as described by many other previous workers, and for the elucidation of the paleogeographic and tectonic setting of these rocks on a more regional scale.

## REGIONAL SETTING

The western margin of Laurentia is overlain by Neoproterozoic and Lower Paleozoic strata that were detached from their underlying basement and transported eastward onto the craton during Mesozoic terrane accretion (Monger and Price, 1979; Price, 1981). These rocks unconformably overlie Paleoproterozoic crystalline basement rocks in the undeformed Western Canada Sedimentary basin and parts of the southern Canadian Cordillera (Ross, 1991a; Parrish, 1991) and, in the northermost United States and southernmost Canada, they overlie Mesoproterozoic sedimentary strata of the Belt-Purcell Supergroup. They comprise two main stratigraphic assemblages: 1) coarse, immature clastic and minor carbonate rocks of the Neoproterozoic Windermere Supergroup; and 2) mature siliciclastic and carbonate shelf and slope deposits of the Lower Cambrian through Middle Jurassic miogeocline. The Neoproterozoic Windermere Supergroup and overlying quartzose Lower Cambrian(?) strata are closely associated spatially and are exposed in thrust sheets in the western Main Ranges of the Rocky Mountains and in the eastern Omineca belt. From northern Canada to the northern United States, the eastern limit of exposure of the Windermere Supergroup coincides with the eastern margin of the Phanerozoic Cordilleran miogeocline (Price, 1994). In the southernmost Canadian Cordillera, in the vicinity of the Moyie and St. Mary faults (Fig. 3-1), there is an abrupt, norheast-trending right-hand shift of about 200 km in the eastern limit of both the Windermere Supergroup and of the miogeocline.

Neoproterozoic and Lower Paleozoic strata that were deposited on the ancestral North American margin are exposed in the southeastern Canadian Cordillera in several distinct tectonostratigraphic domains (see Fig. 3-1). The Neoproterozoic Windermere Supergroup and the overlying Gog Group are exposed in the Western Main Ranges of the Rocky Mountain thrust and fold belt expose, north of latitude 51°N. The Purcell anticlinorium, a broad north-plunging structure to the west of the southern Main Ranges, exposes the Mesoproterozoic Purcell Supergroup, unconformably overlain by the Neoproterozoic Windermere Supergroup and Neoproterozoic to Lower Cambrian Hamill Group and equivalent rocks on its limbs and locally in down-dropped blocks in its core (Fig. 3-1). The Hamill Group is exposed primarily on its western limb and in the Dogtooth Range at its northern termination, at about 51°N latitude. The Purcell anticlinorium is cut obliquely by several northeast-trending thrust faults that include, from south to north, the Moyie, St. Mary, Hall Lake and Redding Creek - Mount Forster thrusts. The Moyie fault coincides with the abrupt change in location of the eastern limit of the Windermere Supergroup and of the Paleozoic miogeoclinal succession. The thin Paleozoic succession exposed to the south and east of this structure was deposited on a high-standing platform known as "Montania" (Deiss, 1941). These transverse faults are notable for the contrasts between the stratigraphic successions in their hangingwalls and footwalls, which indicate that antecedents to these structures experienced normal motion during Neoproterozoic and Early Paleozoic time (Lis and Price, 1976; Price, 1981).

The Kootenay Arc, a narrow arcuate belt of coaxially refolded rocks to the west of the Purcell anticlinorium, exposes the Hamill Group and overlying deeper-water Lower Paleozoic strata that are distinct from primarily shallow-water Lower Paleozoic strata exposed in the Purcell anticlinorium and Western Main Ranges of the Rocky Mountains. North of 51°N, the northern termintion of the Purcell anticlinorium, the Cariboo Mountains expose Windermere Supergroup and younger strata that can be correlated with the succession in the Western Main Ranges (e.g. Ross and Murphy, 1988), and the northern Selkirk Mountains expose a succession similar to that in the Kootenay Arc.

Stratigraphic relationships within and beneath the Hamill/Gog Groups and equivalent rocks vary markedly within and between these tectonostratigraphic domains. These regional stratigraphic variations will be the focus of this chapter and will provide the basis for discussion of the paleogeographic and tectonic setting of the Hamill/Gog Groups. Emphasis will be placed on new stratigraphic and sedimentological data from the Hamill Group from the western Purcell anticlinorium and the adjacent central Kootenay Arc. These data will be intergrated subsequently with previous data and interpretations from the other tectonostratigraphic domains. Stratigraphic relationships and nomenclature within each of these domains are summarized in Figure 3-2.

## PART I:
## NEOPROTEROZOIC AND LOWER CAMBRIAN STRATIGRAPHY OF THE WEST-CENTRAL PURCELL ANTICLINORIUM AND ADJACENT KOOTENAY ARC

### TECTONOSTRATIGRAPHIC DOMAIN: WEST-CENTRAL PURCELL ANTICLINORIUM

The west-central Purcell anticlinorium comprises the part of the Purcell anticlinorium that is carried in the hangingwall of the Mount Forster fault, and that is bounded by the Cretaceous Fry Creek batholith to the south and the Bugaboo batholith to the north (Fig. 3-1, and Plate 2). It is separated from the Kootenay Arc by a steep fault zone, the Kootenay Arc boundary fault (Chapter 4). The Neoproterozoic and Lower Cambrian stratigraphy changes markedly between the eastern Purcell anticlinorium, in the footwall of the Mount Forster fault, the west-central Purcell anticlinorium and the central Kootenay Arc (Fig. 3-2). In the western Purcell anticlinorium, a thick succession of the Hamill Group is underlain by the Windermere Supergroup, which in turn unconformably overlies the Purcell Supergroup that is carried in the immediate hangingwall of the Mount Forster fault. In the eastern Purcell anticlinorium, Lower Cambrian strata are thin or absent in the footwall of the Mount Forster thrust, and Middle Cambrian to Devonian shallow-water strata commonly rest unconformably on Windermere Supergroup strata (Walker, 1926; Reesor, 1973; Root, 1987; Pope, 1989). In the central Kootenay Arc only the upper part of the Hamill Group is exposed. The Hamill Group succession differs from that to the east, and it is overlain by primarily deep-water Lower Paleozoic rocks.

The rocks of the west-central Purcell anticlinorium were deformed, regionally metamorphosed and intruded by plutonic rocks during Mesozoic convergence between North America and exotic terranes (see Chapter 4).The structure of the west-central Purcell anticlinorium is characterized by steep folds and a steep regional axial planar cleavage or schistosity. Several steep thrust faults, with displacements of probably less than one kilometer, are congruent with the regional foliation and folds. Locally, younger crenulation cleavages overprint the regional foliation. The regional metamorphism ranges from chlorite grade in the east and north to staurolite-kyanite grade in the southwest, immediately adjacent to the Fry Creek batholith (Fyles, 1964; Reesor, 1973). Penetrative strain also increases from east to west and from north to south. The Hamill Group is well exposed (Photo Plates 1 and 3A, 3B) in two upright, doubly-plunging map-scale synclines and in a narrow, more complexly deformed synclinal belt on the western limb of the anticlinorium. These structures are, from east to west, the Blockhead Mountain, the Mount Cauldron and Birnam Creek synclines (Plate 2). The base of the Hamill Group is also poorly exposed in a shallow syncline to the north and east of the Blockhead Mountain syncline.

Thin sections of metasedimentary rocks show extensive recrystallization of nearly all matrix minerals across the entire domain, and of an increasing amount of clastic framework material toward the south and west. Detrital feldspar and lithic clasts of pebble size and greater are locally abundant, but most detrital feldspar and lithic fragments of sand size or smaller have been destroyed during regional metamorphism. This has been demonstrated by comparison of point-counted samples from the Hamill Group in the west-central Purcell anticlinorium with samples of similar coarse framework compositions from the less-metamorphosed Dogtooth Range (Devlin, 1986). Thus, quantitative sedimentary petrography is rendered largely invalid for most of the west-central Purcell anticlinorium, but texture and composition of coarser detrital material nevertheless provides important constraints on sedimentary processes and provenance. Primary sedimentary structures are well preserved in most of the Hamill Group and in medium- to coarse-grained rocks of the Horsethief Creek Group, except within the Birnam Creek syncline along the western flank of the Purcell anticlinorium. There, strongly deformed cross beds and rare channel scours are preserved only in the most competent strata.

## TECTONOSTRATIGRAPHIC DOMAIN: KOOTENAY ARC

The central Kootenay Arc is bounded to the east by the Kootenay Arc boundary fault, a steep thrust fault that juxtaposes it against the Purcell anticlinorium (Chapter 4). The central Kootenay Arc as discussed here extends northward from the mid-Cretaceous(?) Fry Creek batholith to the Battle Range batholith (Figs. 4-2 and 3-1).

The stratigraphic succession in the central Kootenay Arc is transitional between miogeoclinal strata of the North American shelf to the east and deep-water strata of oceanic or island arc affinity in Intermontane terrane to the west, and it is stratigraphically linked to both North America and Intermontane terrane (Klepacki, 1985; Colpron and Price, 1995). The succession comprises the Hamill Group and overlying Lower Cambrian shallow-water carbonate and fine siliciclastic strata. The Lower Cambrian strata are overlain by deep-water, immature clastic and mafic volcanic rocks of the Lower Paleozoic Lardeau Group. The Lower Paleozoic strata are unconformably overlain by fine-grained clastic rocks of the Mississippian Milford Group (Read and Wheeler, 1976), and the Milford Group is stratigraphically overlain (Klepacki, 1985) by mafic volcanic rocks of the Kaslo Group (Slide Mountain terrane?), which is in turn overlain by marine clastic rocks of the Slocan Group (Quesnel terrane).

Mesozoic deformation and metamorphism have partly obscured original stratigraphic and structural relationships in the central Kootenay Arc. The rocks have undergone upper greenschist to upper amphibolite facies metamorphism and two to three episodes of coaxial folding. Thicknesses of units are difficult to estimate in the Kootenay Arc due to complex, ductile deformation. Sedimentary structures are poorly preserved. Sedimentary facing indicators are commonly lacking in sections which probably contain reversals in facing direction. Thicknesses in the stratigraphic sections (Figs. 3-4, 3-5) shown for the central Kootenay Arc should, therefore, be considered less reliable than those shown for the west-central Purcell anticlinorium. Nonetheless, lithostratigraphic units are quite distinct in the central Kootenay Arc, and it is possible to link new subdivisions of the Hamill Group in this area with previous subdivisions to the north (Read and Wheeler, 1976; Devlin, 1989) and south (Höy, 1980) and with some correlative units in the Purcell anticlinorium (Fig. 3-1).

## STRATIGRAPHY

### *The Windermere Supergroup*

The Windermere Supergroup in the west-central Purcell anticlinorium comprises the Toby Formation and the overlying Horsethief Creek Group. (Walker, 1926; Evans, 1933; Walker, 1934; Reesor, 1973; see Chapter 2). The Windermere Supergroup unconformably overlies the Dutch Creek and Mount Nelson Formations of the Purcell Supergroup (Walker, 1926; Reesor, 1973; Root, 1987). It is in turn overlain by the Hamill Group (Photo Plate 1). The Windermere Supergroup can be divided into three distinct lithostratigraphic sequences: the Toby Formation, at the base, comprising glaciomarine and sub-glacial sediments that were deposited during extension, normal faulting and local mafic volcanism (Aalto, 1971; Root, 1987; Pope, 1989; Bennett, 1985) and two deep-water coarse clastic successions within the Horsethief Creek Group (lower and upper clastic sequences of Chapter 2). The Windermere Supergroup varies in thickness from less than 250 meters, at the southeastern limit of the domain, to more than 4000 meters in the north and west. New lithostratigraphic subdivisions within the Horsethief Creek Group, lateral variations in these rocks and their depositional environments and tectonic setting are discussed in detail in Chapter 2. The base of the overlying Hamill Group everywhere rests on the same lithostratigraphic unit of the upper clastic sequence (upper pelitic unit), except in the northwestern part of the area (Plate 2).

*The Hamill Group*

The Hamill Group (Walker and Bancroft, 1929) comprises a dominantly quartz sandstone/quartzite succession, about 600 to 2000 m thick, in the west-central Purcell anticlinorium and in the adjacent part of the Kootenay Arc to the west. It comprises two distinct stratigraphic successions that can be correlated between the two tectonostratigraphic domains. These two successions are discussed separately below. The Hamill Group overlies the Horsethief Creek Group in the west-central Purcell anticlinorium, its base is not exposed in the segment of the Kootenay Arc that lies to the west. The Hamill Group is conformably overlain by calcareous schist of the Mohican Formation (Fyles and Eastwood, 1962), which is in turn overlain by marble of the Badshot Formation (Walker and Bancroft, 1929), in both domains.

Strata assigned to the Mohican Formation were originally included in the Hamill Group at its type locality in the west-central Purcell anticlinorium (Walker and Bancroft, 1929), but were subsequently included as a map unit with the Badshot Formation in the central Kootenay Arc and westernmost Purcell anticlinorium (Fyles, 1964; Reesor, 1973). The Mohican Formation will be discussed in this paper as a mappable lithostratigraphic unit separate both from the underlying subdivisions of the Hamill Group and from the Badshot Formation.

**The Hamill Group in the west-central Purcell anticlinorium**

The Hamill Group in the west-central Purcell anticlinorium ranges in thickness from less than 750 m to about 2000 m. It can be divided into four mappable lithostratigraphic units (Fig. 3-3, Photo Plates 1, 3A, 3B, and Plates 1-3):

- Unit H4: upper quartzite and pelite (**CHuq** on Plates 1-3)
- Unit H3: middle pelite (**PCHp**)
- Unit H2: lower quartzarenite (**PCHlq**)
- Unit H1: basal feldspathic grit and arenite (**PCHg**)

The upper unit is locally divisible into two distinct subunits:

•Unit H4a: lower mature quartzarenite

•Unit H4b: interbedded quartzite and pelite

Basal unit H1 comprises interbedded coarse-grained feldspathic arenite and grit, pelite and carbonate that grades into unit H2, a more homogeneous thick coarse quartzarenite or quartzite unit. Unit H2 is more abruptly overlain by Unit H3, which comprises calcareous schist, pelite and minor quartizte and conglomerate. Units H1, H2 and H3 thin and pinch out toward the west. The upper Unit (H4) is a finer-grained, more thinly bedded and generally a more mature quartzarenite and pelite unit that unconformably overlies unit H3, unit H2 or the Horsethief Creek Group.

New subdivisions of the Hamill Group in the west-central Purcell anticlinorium and their lateral variations are shown in a series of stratigraphic sections located on the limbs of major structures (Figs. 3-4; 3-5). Incompetent strata in the Hamill Group such as the middle pelite unit H3 and the upper pelitic part of unit H4 commonly are deformed strongly because of the contrast in competency with underlying and overlying strata, but most of this deformation is clearly related to flexural slip or ductile attenuation during folding, and does not involve any significant offset within the sections. Faults of known significance are shown in the sections. Uncertainties in thicknesses and incompleteness in sedimentclogical data increase from east to west, because of the increase in ductility and severity of deformation, but the lithostratigraphic units clearly can be traced from east to west.

*Unit H1: Basal feldspathic arenite and grit*

The basal unit (H1) of the Hamill Group comprises a distinct, commmonly rusty weathering, succession of white and pale green or grey, thickly cross-bedded coarse feldspathic arenite, grit, and conglomerate, and interbedded dark green or grey pelite and tan-weathering thin carbonate or dolarenite. Unit H1 typically is about 250 meters thick, but it varies in thickness from 0 to 650 m. It generally overlies the fine-grained turbiditic upper pelite unit of the Horsethief Creek Group

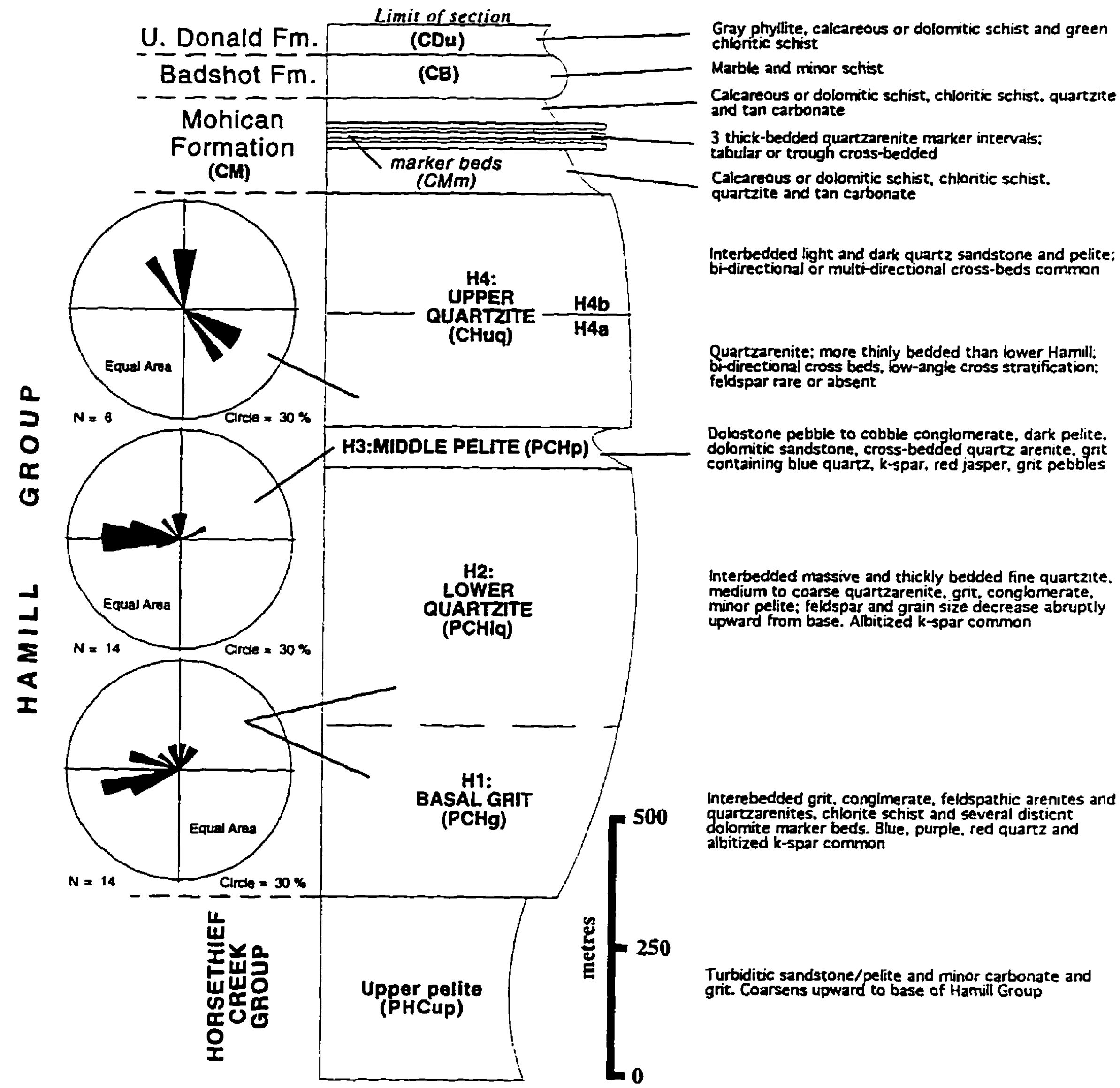



**Figure 3-3:** Composite stratigraphic column from the Blockhead Mountain syncline, showing lithostratigraphic subdivisions of the Hamill Group. Paleocurrent data (dip direction of foresets, corrected for bedding tilt) from one traverse east of Jumbo Pass, and from Devlin (1986), also from the eastern limb of the syncline. Thicknesses approximate.

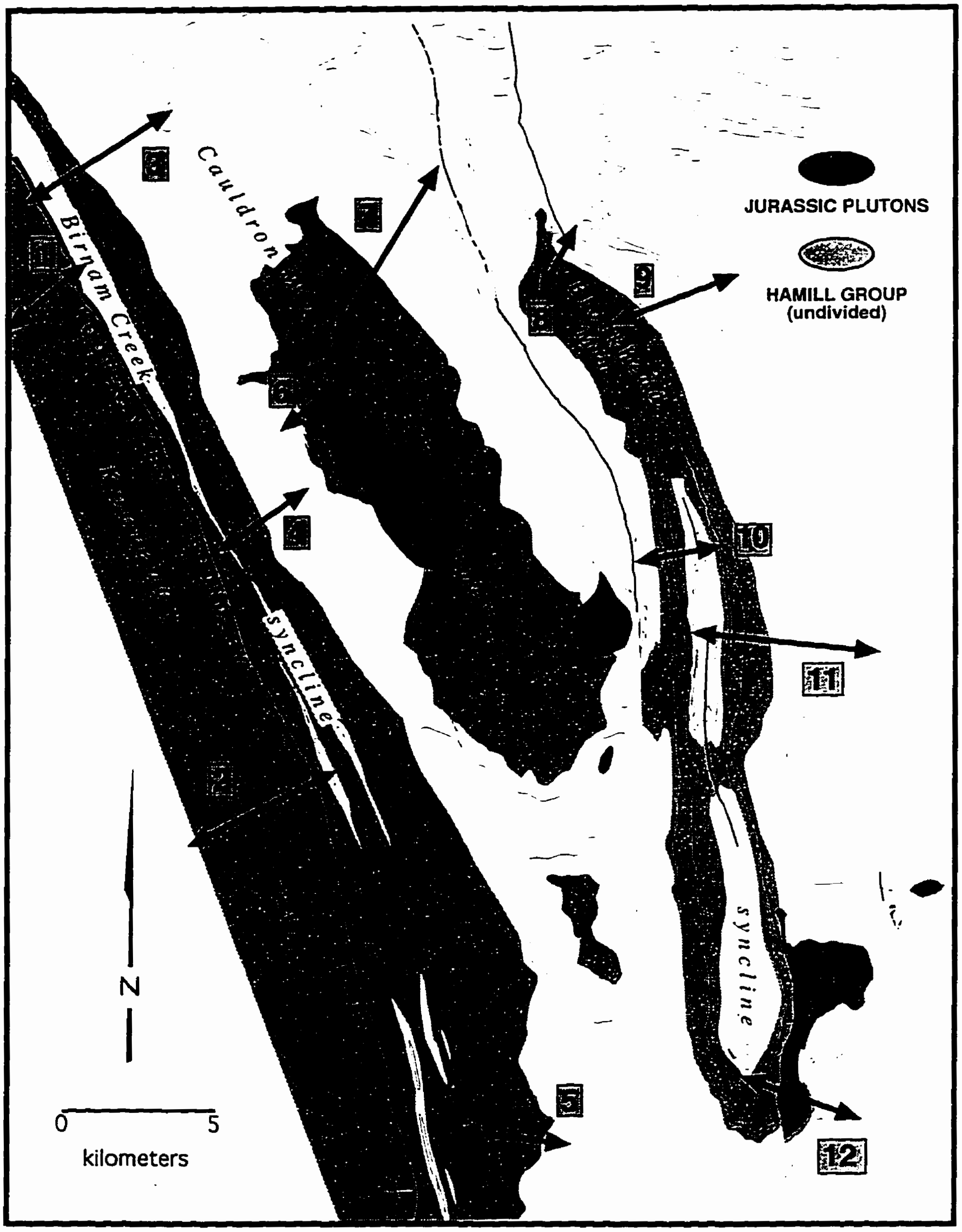



**Figure 3-4:** Simplified geological map of part of the study area, showing the locations of the stratigraphic columns presented in Figure 3-5.

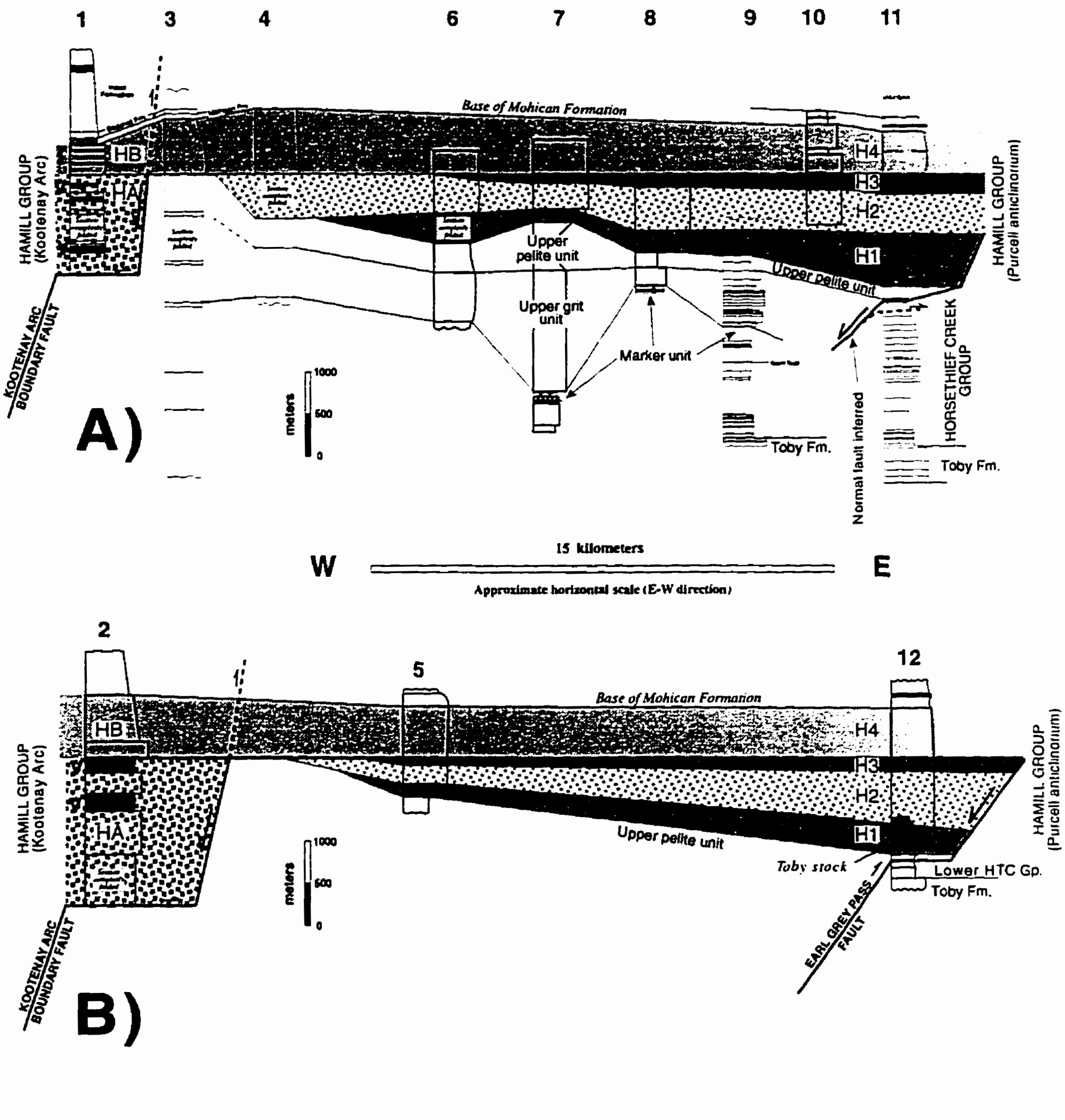



**Figure 3-5:** Stratigraphic columns through the Hamill Group and Horsethief Creek Group in the west-central Purcell anticlinorium and adjacent Kootenay Arc. Column locations given in Figure 3-4. **A)** E-W stratigraphic variations across northern exposires of the Hamill Group in the study area. **B)** E-W stratigraphic variations across southern exposires of the study area, showing relationships across the Earl Grey Pass fault that indicate that it is a reactivated normal fault. Horizontal distances not to scale. **V** = position of mafic volcanic rocks.

gradationally, but locally the contact is abrupt. The basal unit (H1) grades upward into more resistant cross-bedded massive or thickly-bedded quartzarenite and minor grit of the lower unit H2. Unit H1 pinches out to the north and west, and it is thickest at the eastern limit of its exposure. At some localities it is clearly absent, but elsewhere both the lower and upper contacts may be difficult to place, so that only intervals containing the distinct tan-weathering dolostone and green pelite beds can be identified as unit H1.

The feldspathic arenites and grits of the unit H1 are characterized by abundant tabular and less abundant trough cross beds, in sets 0.5 to 1.5 meters thick. The original dip of foresets from the Blockhead Mountain syncline was primarily to the west-northwest (Fig. 3-3). Shallow scoured channels with pebble to small cobble lag deposits are also common. Ungraded lenses of conglomerate, tens of centimeters thick, occur rarely. Channels typically are less than a meter wide and several tens of centimeters deep. Channels are most commonly cut into underlying pelitic beds. Foresets, reactivation surfaces and the bases of beds are also commonly defined by coarser sediment only one clast thick, and by darker matrix that contains heavy minerals. Single beds are typically tabular and laterally continuous for tens of meters. Beds are less commonly wedge-shaped and rarely lenticular.

Clasts in the quartzose rocks range from coarse sand to large cobbles (up to 25 cm). The abundance of grit- to pebble-size sediment and of feldspar in this unit, as compared to the rest of the Hamill Group, is notable (Devlin 1986, and this study). The clasts include, generally in order of decreasing abundance: white, grey and pale green quartzarenite (intraformational), white vein quartz, white poly-crystalline quartz, blue quartz, purple or "amethystine" quartz, red quartz or rare jasper, dolostone, microcline feldspar, plagioclase feldspar, dark grey or black chert and rip-up pelite clasts confined to the bases of beds. Feldspar makes up as much as 23% of the detrital grains in arenite or grit samples at the base of the Hamill Group in the Blockhead Mountain syncline (Devlin, 1986). Feldspars are subrounded to angular. They are rarely larger than one cm, and most are between two and five mm (as are most quartz grains). Microtextures in many plagioclase grains suggest that they were potassium feldspar grains that were albitized during regional metamorphism (Reesor, 1973, and this study). Quartz clasts are most commonly rounded, but several sections contain lenses or channels of angular, poorly sorted pebble to cobble conglomerate. Textural and compositional maturity of sediment varies markedly between beds, although the sediment is quite homogeneous and texturally mature in sections of tabular cross-bedded feldspathic arenite and grit. In general, grain size decreases and textural and compositional maturity increases upward through unit H1.

The quartzose rocks in the basal unit (H1) are interbedded with several distinct meter-thick, tan-weathering, homogeneous dolostone beds or beds of white quartzarenite/grit in a tan dolomitic matrix. Ankerite and limonite are abundant in these dolomitic beds. The arenaceous beds may be graded. Devlin (1986) observed cross-stratification in some of these beds at one locality. Several single beds were traced in alpine exposures for more than 1 km. On the west limb of the Cauldron Mountain syncline (Plate 2), the dolomitic intervals do not form continuous beds but occur as lenses of dolostone pebble to cobble conglomerate and quartz grit or conglomerate in a tan dolomitic matrix. Green or dark grey pelites, 0.5 m to several m thick, are also common in basal unit H1. The green pelites are fine-grained, homogeneous Fe-chlorite schist containing no discernible sedimentary structures, and they most commonly overlie the tan dolostone beds. The grey pelite beds are typically fine-grained quartz-muscovite schist or phyllite, but some contain rounded coarse sand grains or pebbles of grey and red quartz, as well as rare fragments of black chert and grey or green pelite. Rare sedimentary structures include planar laminations (<5mm), and cm-scale rhythmically bedded dark and light pelite near the base of the unit, similar to that in the upper part of the underlying Horsethief Creek Group.

Several lateral facies variations are notable in unit H1. Pebble and cobble conglomerates and obvious channels are more common in eastern exposures. Rare structureless beds, 1 to 2 m thick, of quartzite breccia in an arenaceous matrix occur at one locality on the eastern limb of the Blockhead Mountain syncline. However, in westernmost exposures of unit H1, adjacent to where it pinches out, there is also abundant grit and quartz-dolostone pebble conglomerate, in a semi-pelitic matrix. Individual beds are less continuous and there is a greater amount of interbedded pelite and semi-pelite relative to quartz or feldspathic arenite. Successions of cross-bedded arenite are less common in the west and north, and the contact with the underlying Horsethief Creek Group is more difficult to place. The northernmost exposures of unit H1, isolated in a shallow syncline to the north of the Blockhead Mountain syncline (Plate 2), differ markedly from those to the south. At this locality unit H1 comprises interbedded pelite, metasiltstone and less abundant quartz sandstone. The matrix is pelitic rather than arenaceous, and the unit is finer grained than to the south. It is distinguished from the underlying upper pelite unit of the Horsethief Creek Group only by an abrupt upward increase in quartz content and by the presence of the tan-weathering dolostone beds.

*Interpretation*

Unit H1 of the Hamill Group must have been deposited in an environment that received an abundant sediment supply and was dominated by strong, largely uni-directional traction currents that caused migration of 2-D and 3-D dunes (Walker, 1984). The occurrence of the sandy carbonate marker beds implies a marine environment. Coarsening-upward successions in the upper part of the Horsethief Creek Group and the gradational to locally unconformable contact between fine-grained turbiditic deposits of the upper Horsethief Creek Group and coarser traction deposits of unit H1 of the Hamill Group imply that a relative sea level fall accompanied initial deposition of the Hamill Group. Devlin (1986) interpreted the lower part of the Hamill Group in the Blockhead Mountain syncline as a shallow marine deposit, but he interpreted very similar successions of facies in the lower part of the Hamill Group and equivalent strata as little as 60 km to the south as alluvial braidplain deposits on the basis of abundant and well-developed channel systems. The similarities in facies, sediment composition and paleocurrent data (Devlin, 1986) between these localities suggest a strongly fluvially infuenced shallow marine environment in the west-central Purcell anticlinorium, perhaps close to the intersection of an alluvial braidplain with the shoreline. Alternatively, the cross-bedded facies could have been produced by strong ebb or flow tidal currents. The carbonate intervals could record sea-level fluctuations and/or fluctuations in the supply of terrigenous clastic sediment.

The westward and northward decrease in coarse, channeled and cross-bedded facies implies less strong currents and/or less available coarse sediment to the west and north. Conglomeratic facies of basal unit H1 are most notable at the extreme eastern and western limits of exposure, where this unit is thickest and thinnest, respectively. The easternmost exposures comprise a thick succession both of well-rounded conglomeratic channel or lag facies and of rare, angular structureless breccias, suggesting that easternmost exposures were either more proximal to a major sediment source, or to the location of major channel systems, or both.

The abundant blue quartz, angular feldspar and poly-crystalline and vein white quartz indicate that the sediment in unit H1 was, at least in part, derived from a high-grade basement source, but the abundant well-sorted and well-rounded quartz indicates that there was more extensive reworking of the detritus, or contribution from a different source, than in the feldspathic grits of the underlying Windermere Supergroup. The Windermere Supergroup grits could provide a local source of angular blue quartz and feldspar. The source of the red or purple quartz remains undetermined. More than one source is considered likely for unit H1. The Purcell Supergroup or the craton are possible sources for the more reworked component of the sediment; the craton is considered more likely because most of the quartzite in the Purcell Supergroup is finer-grained than that in the basal unit (H1) and the lower unit (H2) of the Hamill Group.

### *Unit H2: Lower quartzarenite*

The lower unit (H2) comprises abundant pale grey, pale green and white coarse quartzarenite and less abundant feldspathic arenite and grit. It conformably overlies the basal feldspathic grits of unit H1, and is overlain either by the recessive rocks of the middle unit (H3) or by quartzarenite of the upper unit (H4) of the Hamill Group. Unit H2 varies from zero to 600 m thick, but it is typically about 500 m thick. It thins and pinches out to the north and west.

Unit H2 resembles the underlying unit H1 but overall it is finer grained, more homogeneous, and more texturally and compositionally mature. Interbedded pelites are rarely more than a few cm thick and are most commonly less than one cm thick. The lower contact is gradational and is placed at the last occurrence of tan dolostone and/or thick-bedded chlorite schist or green phyllite. Devlin (1986) desribed this contact as coinciding with an abrupt decrease in the amount of feldspar but although this is true in the Blockhead Mountain syncline, the decrease in the amount of feldspar upward in the section is more subtle to the west in the Mount Cauldron syncline. On the western limb of the Blockhead Mountain syncline, abundant subangular feldspar (5-15%) was observed in cross-bedded strata throughout the top of unit H2, within only a few tens of meters from the contact with the overlying unit H3 of the Hamill Group.

Tabular and trough cross beds similar to those observed in unit H1 are common, but unit H2 is also characterized by intervals of massive or thickly-bedded, apparently structureless, vitreous pale green or white quartzite that are many tens of meters thick (Photo Plate 3C). They may be in part the result of recrystallization of finer-grained quartzarenites and the destruction of most primary sedimentary structures. Subtle cross-bed foresets are locally discernible in these intervals and are defined by slightly darker and coarser sand grains. Paleocurrent directions from cross beds in unit H2 from the Blockhead Mountain syncline are similar to those from unit H1 (Fig. 3-3). Channels or scoured surfaces are rare in unit H2 and carbonate strata are also rare except in the central Blockhead Mountain syncline, where disrupted carbonate beds occur throughout much of the Hamill Group. Unit H2 is notable for lateral continuity of individual beds and of successions of thickly bedded to massive quartzite. Cliff exposures on both limbs of the Cauldron Mountain syncline and on the west limb of the Blockhead Mountain syncline show two prominent breaks in slope within unit H2, that correspond to two thin but laterally continuous recessive pelitic intervals. These breaks are visible for 20 km along strike in the Cauldron Mountain syncline.

*Interpretation*

The sedimentary structures and paleocurrent directions in unit H2 are similar to those of the underlying unit H1, suggesting that a high-energy depositional environment characterized by dominantly uni-directional migration of 2-D and 3-D dunes persisted, and that the difference between the two units was one of more prolonged reworking of the sediment. The decrease in the amount of pelite, the absence of the carbonate marker beds, and less abundant or obvious erosion surfaces in unit H2 suggest that the sediment supply was more constant and/or sea-level fluctuations played a less important role than for unit H1.

In the easternmost exposure of unit H2 in the Blockhead Mountain syncline, there is an unusual thick interval of interbedded quartzite breccia, blue quartz and feldspar grit and pebble conglomerate, resedimented carbonate and mature quartzarenite. The angular quartz and feldspar fragments presumably could not have been transported far, and because these facies are restricted to easternmost exposures of unit H2, a local source of coarse detritus immediately to the east is proposed in addition to a more distant source of fine quartz.

*Unit H3: Middle pelite*

The middle pelitic unit (H3) of the Hamill Group comprises locally calcareous red-weathering pelite, dolostone-clast conglomerates, quartzose grit and cross-bedded quartzite or quartzarenite, and mud-matrix-supported granule or pebble conglomerates. It abruptly overlies the lower unit and marks a conspicuous break in the deposition of dominantly quartzose sediments in the Hamill Group. It is abruptly overlain by mature quartzarenite of unit H4. The upper contact is more abrupt toward the west. Unit H3 varies from 0 to 125 m in thickness. It is thickest to the east and pinches out to the west, where in one exposure (Mt. Banquo) it appears to be truncated beneath unit H4 (Plates 2 and 3). Unit H3 contains less coarse-grained material to the west. In the Mt. Cauldron syncline unit H3 comprises dominantly pelite and metasiltstone. On the western limb of the Blockhead Mountain syncline, it comprises two distinct recessive pelitic, calcareous and conglomeratic intervals, separated by 60 m of cross-bedded quartzarenite and grit.

Sedimentary structures are poorly preserved in unit H3 due to its overall fine grain size and to greater internal strain relative to the more competent units below and above. However, cross beds are well preserved in quartzarenites in the middle part of the unit in the Blockhead Mountain syncline. The cross beds generally differ from those typical of both units H1 and H2 but are similar to those in the overlying unit H4. They are closely spaced, occur in sets rarely thicker than 10 cm, and foresets show mutidirectional paleoflow (Fig. 3-3). Low-angle cross-stratification and planar bedding are also common in the quartzarenites. Overall, the quartzarenites in this unit are more thinly bedded than in the underlying units. However, intervals with thicker sets (0.5 m) of tabular or trough cross beds similar to those in the underlying units also occur in the middle unit in the Blockhead Mountain syncline. They are interbedded with intervals of smaller-scale cross bed sets. Pelite is more commonly interbedded with the smaller-scale cross-bedded quartzarenite. Rare ripple cross lamination (Devlin, 1986) and planar lamination are preserved in the pelites.

The conglomerates and grits that occur in unit H3 in the Blockhead Mountain syncline are markedly less mature than the quartzose grits of unit H2. The lower part of unit H3 contains several intervals, 1-5 m thick, comprising angular dolostone cobble conglomerate in a calcareous, quartzose arenaceous matrix. Granule to pebble conglomerate in a pelitic matrix is also common and is distinct because it contains abundant blue, purple and red quartz in addition to white quartz. Rare red jasper pebbles also occur in the conglomerate intervals. Quartzose grits interbedded with cross-bedded quartzarenite contain abundant blue quartz and rarer clasts of grit or laminated sandstone and angular single grains of albitized potassium feldspar.

Western exposures of unit H3 contain only a cross-bedded quartzarenite facies that is similar to unit H2, and they are very abruptly overlain by unit H4. Unit H3 pinches out or is truncated beneath the unit H4 to the west. However, eastern exposures contain intervals of interbedded cross-bedded quartzarenite facies that are characteristic of both the underlying unit H2 and the overlying unit H4 of the Hamill Group. These relationships may imply that the upper contact represents a transition from an unconformable to a conformable and gradational contact, from west to east.

*Interpretation*

Unit H3 was deposited in a shallow marine environment that received less sand than the underlying units, due either to a decreased sediment supply from the source or to a relative sea level rise. Rare wave ripples indicate that unit H3 was deposited at least in part above fair weather wave base. Weakly bi-directional paleocurrents suggest a tidally influenced setting, although the deposition of sediment now exposed to the east in the Blockhead Mountain syncline was partly controlled by strong uni-directional currents in a setting that must have been similar to that of units H1 and H2. The observation that large-scale trough and tabular cross-bedded facies and conglomeratic facies are restricted to eastern exposures suggests that the strong currents that transported the sand and coarser sediment into this environment were confined to the east.

The K-feldspar, red jasper and blue quartz imply a relatively proximal source similar to that which supplied the basal unit H1. The abrupt appearance of these clasts in a commonly pelitic matrix implies rejuvenation of that source despite an overall decrease in sandy sediment. Thus it is incompatible either with waning erosion of the main source area or with a relative sea level rise, unless a local source of immature sediment was tectonically rejuvenated at the same time.

*Unit H4: upper quartzarenite and pelite*

The upper unit (H4) of the Hamill Group comprises two members: a lower homogeneous and mature quartzarenite member H4a and an upper member H4b of thickly to thinly interbedded white, green and pink quartzarenite, argillaceous quartz sandstone and dark pelite. The contact between the two members is gradational and they are not everywhere distinct and mappable separately. On the east limb of the Blockhead Mountain syncline, they are separated by a distinct rusty-weathering recessive interval about 50-m thick. Unit H4 is approximately 500 m thick; the lower member H4a is about 200 m thick and the upper member H4b is about 300 m thick. Unit H4 of the Hamill Group overlies, from east to west, unit H3, unit H2 and the Horsethief Creek Group. Unit H4 grades upward into the more calcareous and schistose Mohican Formation.

Member H4a is the most texturally and compositionally mature mappable unit of the entire Hamill Group. On the east limb of the Blockhead Mountain syncline, where metamorphism is least severe, it contains abundant well-rounded and well-sorted medium- to coarse-grained quartzarenite. Homogeneous recrystallized quartzite intervals or beds are interpreted as finer-grained sandstone. Beds of quartz grit or pebble conglomerate, commonly one clast thick are found throughout member H4a but are rare. Feldspar is virtually absent, except in the Birnam Creek syncline where isolated angular pebbles, primarily of albite (formerly k-feldspar?), occur in lenses within a quartzite matrix. Individual beds in the member H4a range in thickness from 5 to 15 cm. Cross beds are common in coarser sediment. Sets of closely-spaced foresets are interbedded with beds containing low-angle cross-stratification. Foreset orientations are clearly more multidirectional than in the basal, lower and middle units, although quantitative paleocurrent data are sparse (Fig. 3-3; Photo Plate 3D).

The upper member H4b comprises laterally continuous intervals of thinly to thickly-bedded quartzarenite and argillaceous quartz sandstone, with interbedded pelite. Individual beds become thinner upward in this member, and the ratio of pelite to quartz sand increases. The proportion of pelite also increases westward (Reesor, 1973). Below the contact with the Mohican Formation, which is defined as the first appearance of calcareous or dolomitic beds, the unit H4 is a succession of distinct green, pink and white quartzite and interbedded dark grey to black pelite. The coloured quartzites are more conspicuous to the west than to the east, but this may reflect differences in metamorphic grade. Cross beds similar to those in the lower member (H4a) are found throughout the upper member (H4b). Horizontally laminated bedding is also common. Wave(?) ripple marks are rare and poorly preserved in the upper part of member H4b.

*Interpretation*

The sedimentary structures and paleocurrent data of unit H4 of the Hamill Group suggest that it was deposited in a tidally-influenced shelf setting, in which sediment was extensively reworked. This environment must have been different from that in which units H1 and H2 of the Hamill Group were deposited. The lateral and vertical variations in bed thickness and proportion of pelite suggest that western exposures of this unit represent more distal deposits, and that shelf subsidence and/or relative sea level rise outpaced coarse clastic sediment input. The transition to the Mohican Formation implies a decrease in clastic sediment supplied to the margin, which thus favoured carbonate production.

**The Hamill Group/Horsethief Creek Group contact in the west-central Purcell anticlinorium**

The contact between the Horsethief Creek and Hamill Groups is difficult to map in detail, despite a marked lithological contrast between the rhythmic turbidites of the upper part of the Horsethief Creek Group and the quartzose conglomerate, cross-bedded grit and arenite of the basal unit (H1) of the Hamill Group (Photo Plate 3A). Difficulties in mapping precisely the stratigraphic contact are compounded by deformation and metamorphism in this region. I concluded that this contact is apparently gradational over much of the west-central Purcell antclinorium, but that it is also locally unconformable. Evidence for a local unconformity is found toward the west, on the west limb of the Blockhead Mountain syncline and in the Cauldron Mountain syncline.

At most localities in the western Purcell Mountains, the upper pelite unit of the Horsethief Creek Group fines upward from the underlying grit unit, and coarsens upward into the overlying unit H1 of the Hamill Group. At a few localities the contact is abrupt, but generally it is gradational over tens of meters. Grit beds become more laterally continuous and abundant, and they contain a greater proportion of arenaceous, as opposed to pelitic, matrix upward through this transition. The Hamill Group almost everywhere rests on the same unit of the Horsethief Creek Group, except perhaps toward the northwest, where the uppermost unit (H4) of the Hamill Group may rest unconformably on a lower stratigraphic level in the undivided upper clastic sequence of the Horsethief Creek Group.

This contact was examined at the northern closure of the Blockhead Mountain syncline (Plate 2). Flexural slip is not severe, and cleavage is at a high angle to bedding. There, several thick, continuous intervals of grit or conglomerate, consisting of rounded quartz and feldspar pebbles in an arenaceous matrix, are interbedded with turbiditic pelite/sandstone of the upper Horsethief Creek Group. A laterally continuous sequence about 20 m thick of rhythmically-bedded grey pelite, similar to the upper Horsethief Creek Group, also occurs above the lowest beds of cross-bedded feldspathic arenite and tan-weathering dolostone and chlorite schist which mark the base of the Hamill Group. The contact is therefore considered gradational at this locality, perhaps controlled by small-scale fluctuations in sea-level superposed on an overall shallowing trend.

The abrupt nature of this contact at many localities may be due to local shearing and faulting between the incompetent upper Horsethief Creek Group and the competent Hamill Group. Bed-parallel shear along the contact, the result of flexural slip during folding, is common on the limbs of the Cauldron Mountain syncline and particularly the tight Blockhead Mountain syncline. Devlin (1986) and Root (1987) examined the abrupt contact on the east limb of the Blockhead Mountain syncline (Plate 2) where a lens of grit similar to that of the basal unit of the Hamill Group (H1) occurs several tens of meters below the contact, but they were unable to determine whether the lens of grit and quartzite in the upper part of the Horsethief Creek Group is a sedimentary lens or a tectonic sliver. The base of the Hamill Group is clearly folded and sheared at this locality (Devlin, 1986; and this study). However, continuous grit and arenite beds of unit H1 contain an increasing amount of interbedded pelitic material toward the base, and the upper part of the Horsethief Creek Group contains 5-m thick, discontinuous lenses of feldspathic quartz grit. It is important to note that the uppermost quartzose schist of the Horsethief Creek Group contains rare purple quartz pebbles similar to those which are characteristic of the overlying unit H1 of the

Hamill Group and thus had begun to receive sediment from the same source as the Hamill Group. Therefore, despite significant deformation, the contact is considered gradational in this locality.

However, on the western limb of the Blockhead Mountain syncline, the contact appears both gradational, to the south, and erosional, to the north. Quartz pebble conglomerate of the base of the Hamill Group is in very sharp contact with phyllite of the upper Horsethief Creek Group, and locally it fills shallow channels cut into the underlying pelite. The conglomerate is angular and poorly sorted at the base and becomes finer upward over 1-2 m. About 5 km to the south, matrix-supported quartz grit and pebble conglomerate of the Horsethief Creek Group fines upward into an interval of thinly-bedded pelite that abruptly but gradationally coarsens upward into clast-supported grit and conglomerate of unit H1 of the Hamill Group.

In alpine exposures near the northern closure of the Cauldron Mountain syncline, the contact is gradational for several hundred meters along strike, but it becomes more abrupt toward the north, until quartz pebble conglomerate and grit at the base of the Hamill Group sharply overlie pelite and sandy schist of the Horsethief Creek Group and appear to cut gently downward toward a dolomitic interval in the upper Horsethief Creek Group. On the western limb of this syncline, disrupted, discontinuous beds of grit and dolostone and dolostone cobble conglomerate at the base of unit H1 overlie strongly deformed gritty pelite and matrix-supported grit that is thought to be toward the base of the upper pelite unit of the Horsethief Creek Group. Unit H1 pinches out immediately to the west.

These relationships show that the Horsethief Creek Group is gradational into the basal unit H1 of the Hamill Group over some of the west-central Purcell anticlinorium, but that locally it rests with erosional unconformity on the Horsethief Creek Group, particularly toward the west. The succession that includes the upper pelitic unit of the Horsethief Creek Group and the basal grit unit H1 of the Hamill Group evidently records a shallowing-upward from below storm wave base to above storm wave base, perhaps accompanied by a change in sediment source. This does not preclude local breaks in sedimentation, particularly if the shallowing-upward trend records a sea-level fall and/or was accompanied by local tectonic uplift.

**The Hamill Group in the Kootenay Arc**

The Hamill Group in the central Kootenay Arc, east of Duncan Lake (Plate 2) comprises two lithostratigraphic units and is conformably overlain by the Mohican Formation. The maximum exposed thickness is about 1500-2000 m. The lower lithostratigraphic unit HA is a heterogeneous assemblage of schist, impure quartzite and minor conglomerate, carbonate and local intercalations of metabasite (Photo Plate 3E). Unit Ha is divided on Plates 1-3 into discontinuous intervals containing metabasite and dark quartzose schist (PCHv) and continuous intervals comprising primarily quartzite (PCHlqs). Unit HA is only exposed in the immediate hangingwall of the Kootenay Arc boundary fault, and its base is not exposed because it is truncated by this structure. The overlying upper unit HB (CHuq) is locally divisible into two members (not divided on Plates 1-3): a distinct but discontinuous clean quartzite unit (HBa) at the base, and an upper member of interbedded quartzite and pelitic schist (HBb) which grades into calcareous and dolomitic schist of the overlying Mohican Formation. The clean quartzite unit (HBa) and the interbedded quartzite/pelite unit (H2b) can be traced to the north along strike across Duncan Lake into the Mount Gainer and Marsh Adams Formations, respectively, of Read and Wheeler (1976). Fyles (1964) mapped the entire Hamill Group in the Duncan Lake area as undivided Marsh Adams Formation but interpreted discontinuous clean quartzite at the base of several sections of the Hamill Group as equivalent to Mount Gainer Formation. He did not map beneath these quartzites and thus did not recognize any metabasite in the lower part of the Hamill Group. Read and Wheeler (1976) mapped a thin metabasite unit or units at the base of the Mount Gainer Formation to the west of Duncan Lake, along strike from the stratigraphically highest metabasites exposed to the east of Duncan Lake.

*Unit HA: schist, impure quartzite and metabasite*

Unit HA is a heterogeneous assemblage of primarily clastic metasedimentary rocks. It comprises abundant Fe-, Mg- and Ca-rich quartzose schist, less abundant pelitic schist, impure quartzite, metabasite, and minor grit, conglomerate and dolostone or dolomitic schist. It is recognized only adjacent to the Kootenay Arc boundary fault, and its base is not exposed. Unit HA is generally overlain abruptly by massive or platy clean white quartzite of Unit HBa. The maximum exposed thickness of Unit HA is about 1500 m but is difficult to estimate because of complex folding adjacent to the Kootenay Arc boundary fault.

Unit HA contains abundant chlorite-biotite-epidote-plagioclase-quartz-(garnet) schist, which locally contains lenses of conglomerate or grit containing rounded quartz pebble. The conglomerates also contain rare plagioclase pebbles and mafic schist fragments. Graded bedding is commonly preserved in the coarse-grained rocks. Unit HA also contains abundant interbedded green, grey and white quartzite and interbedded quartz-muscovite-biotite-(garnet)-(chlorite) schist. Graded quartzite-pelite couplets, several cm thick, are locally preserved. Discontinuous lenses of metabasite and thin beds of calc-silicate schist or gneiss also occur. Lateral variations are abrupt throughout this unit, even at outcrop scale.

The metabasites (Photo Plate 3E) are up to tens of meters thick. Most are parallel to bedding and appear to be sills or flows, but some cut bedding at a low angle. No primary structures were observed in these metabasites. Light grey marble or tan dolostone, several meters thick, is commonly associated with the metabasites. The carbonates appear to underlie the metabasites, but sedimentary facing indicators are ambiguous in the localities where the carbonates were observed. The metabasites appear to occur at three distinct stratigraphic levels within unit HA, but one of these intervals may be a structural repetition due to folding in the immediate hangingwall of the Kootenay Arc boundary fault. The metabasites are associated with intervals of chlorite-biotite-quartz-epidote schist, which are interpreted as volcanogenic metasedimentary rocks. The presence of rare mafic fragments in pebble conglomerates in unit HA strengthens the argument that at least some of the metabasites are syn-sedimentary. However, some may cut the base of the overlying clean quartzite unit HBa.

*Interpretation*

The metabasites and mafic clastic rocks imply a tectonically active setting for unit HA. The graded quartzite-pelite couplets are interpreted as turbidites that would imply deposition in a deeper-water setting than those of the Hamill Group in the adjacent Purcell anticlinorium. Alternatively, these beds could be graded storm deposits. A similar , less metamorphosed succession of intercalated graded quartzite-pelite beds, discontinuous, immature coarse clastic rocks and pillowed mafic volcanic rocks is exposed along strike in the lower part of the Hamill Group in the northern Selkirk Mountains. These rocks have been interpreted by Devlin (1986, 1989) as turbidites and other sediment gravity flows that were deposited on or adjacent to a paleoslope that was proximal to the sediment source.

*Unit HB: quartzite and pelitic schist*

Unit HB comprises a discontinuous lower member of clean quartzite (HBa) and an upper member (HBb) of interbedded pelitic schist, impure quartzite and cleaner dark and light quartzite. It rests very abruptly on different rock types of Unit HA. The lower member, equivalent to the continuous Mount Gainer Formation to the north (Read and Wheeler, 1976) is a distinctive, pure, massive or thickly bedded white or very pale green, blocky-weathering quartzite. Sedimentary structures are not preserved in the recrystallized clean quartzite. The quartzite is typically a few tens of meters thick, but it is not everywhere recognized. The greatest observed thickness, on Lavina Ridge (Plate 2), is about 100 m. It grades upward and laterally into less pure quartzite and pelitic schist of unit HBb.

Unit HBb (the Marsh Adams Formation of Fyles, 1964) comprises between 150 and 250 meters of interbedded light to dark grey quartzite and quartz-muscovite-biotite-(garnet) schist. Beds of white quartzite, one to two m thick, are common near the base, and cm-scale beds of pink and green quartzite are common at the top. Strongly deformed cross beds in sets 5-10 cm thick are common in coarse-grained quartzite beds in the upper part of the unit. Unit HBb becomes more calcareous upward and grades into the overlying Mohican Formation.

The contact between units HA and HB is not well exposed, but several observations suggest that it is unconformable. The transition from immature metasedimentary and metavolcanic rocks to clean quartzite is everywhere abrupt. The clean quartzite (HBa) rests on several different rock types in the underlying unit HA. At one locality, where the contact intersects Duncan Lake (Plate 2), a graded quartz pebble conglomerate occurs at the base of unit HBa and abruptly overlies metabasite and quartz-muscovite-chlorite schist of Unit HA. The conglomerate appears to truncate compositional layering in the underlying rocks, but the outcrop is structurally complex, and the contact may be faulted. However, the conglomerate does contain pebbles of schist similar to immediately underlying schist in unit HA. The conglomerate is 1 m thick and grades upward into coarse quartz grit and overlying recrystallized quartzite.

*Interpretation*

A marine shelf setting is proposed for unit HB. It was deposited in an environment in which coarse sediment was much more extensively reworked compared to that of unit HA. This environment received less sand relative to mud with time, and it eventually gave way to carbonate-producing shallow marine environments of the Mohican and Badshot Formations. It existed across a belt that was at least 90 km wide, the approximate restored width of the Marsh Adams Formation in the central Kootenay Arc (Chapter 4 and Plate 4).

**Stratigraphic correlation and summary of lateral stratigraphic variations in the Hamill Group between west-central Purcell anticlinorium and Kootenay Arc**

The contrasting stratigraphic successions in the Hamill Group and overlying rocks between the central Kootenay Arc and the west-central Purcell anticlinorium can be correlated (Fig. 3-2). The Mohican and Badshot Formations are common to both domains, and new information from the Birnam Creek syncline near Duncan Lake shows that the lower part of the Index Formation also extends eastward into the Purcell anticlinorium (Fig. 3-5; Column 3). Subdivision of the Hamill Group in both domains shows that the uppermost units (units H4 and

HB) can be correlated, but that the lower parts of the Hamill Group are distinctly different in the two domains.

The uppermost quartzite and pelite unit of the Hamill Group is stratigraphically equivalent in both domains (units H4 and HB). It can be divided locally into a discontinuous lower mature quartzarenite, overlain by interbedded quartzite and less mature sandstone or quartzose schist or and pelite that grades into the overlying Mohican Formation. It is characterized by more thinly-bedded and thinly cross-bedded strata and by a lack of feldspar in comparison to underlying rocks in the Hamill Group. In both domains, it rests sharply on several distinct underlying units or rock types. The upper quartzite unit of the Hamill Group overlies, from east to west, unit H3, unit H2, the Windermere Supergroup and unit HA of the Kootenay Arc. However, the succession of units H1, H2 and H3 of the Hamill Group in the Purcell anticlinorium is distinctly different from the metabasite-bearing lower unit HA in the Kootenay Arc.

These relationships imply that there is an important regional unconformity beneath the upper quartzite unit of the Hamill Group (Fig. 3-5). The upper quartzite unit of the Hamill Group, the Mohican Formation, the Badshot Formation and the lower part of the Index Formation overlie this unconformity and can be correlated from the western Purcell anticlinorium to the Kootenay Arc. The lower part of the Hamill Group that lies beneath this unconformity is discontinuous and non-correlative between the two domains. These relationships are important because they show that the Kootenay Arc boundary fault follows an older structure along which different stratigraphic successions are juxtaposed beneath the unconformity.

In the west-central Purcell anticlinorium, the succession of the Hamill Group beneath the unconformity generally thickens and coarsens from west to east (Fig. 3-5). Both unit H1 and unit H3 pinch out to the west. Conglomeratic facies of unit H1 are generally coarser in the Blockhead Mountain syncline. Angular quartzite cobble and boulder breccia was observed in units H1 and H2 only on the eastern limb of the Blockhead Mountain syncline. Conglomeratic facies in the lower and middle units H2 and H3 are rare to the west of the Blockhead Mountain syncline. In the Kootenay Arc, the lateral continuity of the lower unit (HA) cannot be determined because it and the base of the continuous upper quartzite unit are only exposed in the eastern part of the Kootenay Arc

Although there is an important regional unconformity at the base of the upper quartzite unit, locally it appears to be gradational into the underlying unit H3 of the Purcell anticlinorium. In the Blockhead Mountain syncline, unit H3 contains intercalated cross-bedded quartz sandstone facies that are similar to those in both units H2 and H4, and there is no clear evidence of a hiatus between units H3 and H4. These relationships suggest that sedimentation may have been continuous in the east. Westward from the Blockhead Mountain syncline, the distinctly different successions that comprise the lower part of the Hamill Group in the Purcell anticlinorium and in the Kootenay Arc are separated by bevelled Windermere Supergroup, against which units H1 and H2 in the Purcell anticlinorium appear to onlap, and toward which unit H3 pinches out beneath the unconformity.

**Truncation of eastern exposures of the Hamill Group by a pre-Mesozoic normal fault**

At the southeastern limit of its exposure in the Blockhead Mountain syncline (Fig. 3-6a and Plate 2), the Hamill Group rests on an anomalously thin (<500 m) succession of Windermere Supergroup. The stratigraphic relationships are partially obscured by intrusion of the Middle Jurassic Toby stock and by several Mesozoic faults and ductile shear zones, but it is clear that more than 1500 m of Horsethief Creek Group are cut out across an enigmatic fault that has been traced northward for about 15 km (Plate 2; Photo Plate 3F). The structural and stratigraphic relationships across this fault, which are discussed in detail below, imply that the omission of part of the Horsethief Creek Group strata is due to pre-Middle Jurassic, west-side-down normal displacement on this fault. However, rotated and truncated cleavage, cross-cutting relationships between the Toby stock and contact and regional metamorphic assemblages and, to the north of Earl Grey Pass, repeated stratigraphic section in the Horsethief Creek Group demonstrate that this fault was a west-dipping, east-verging thrust fault during emplacement of the Middle Jurassic Toby

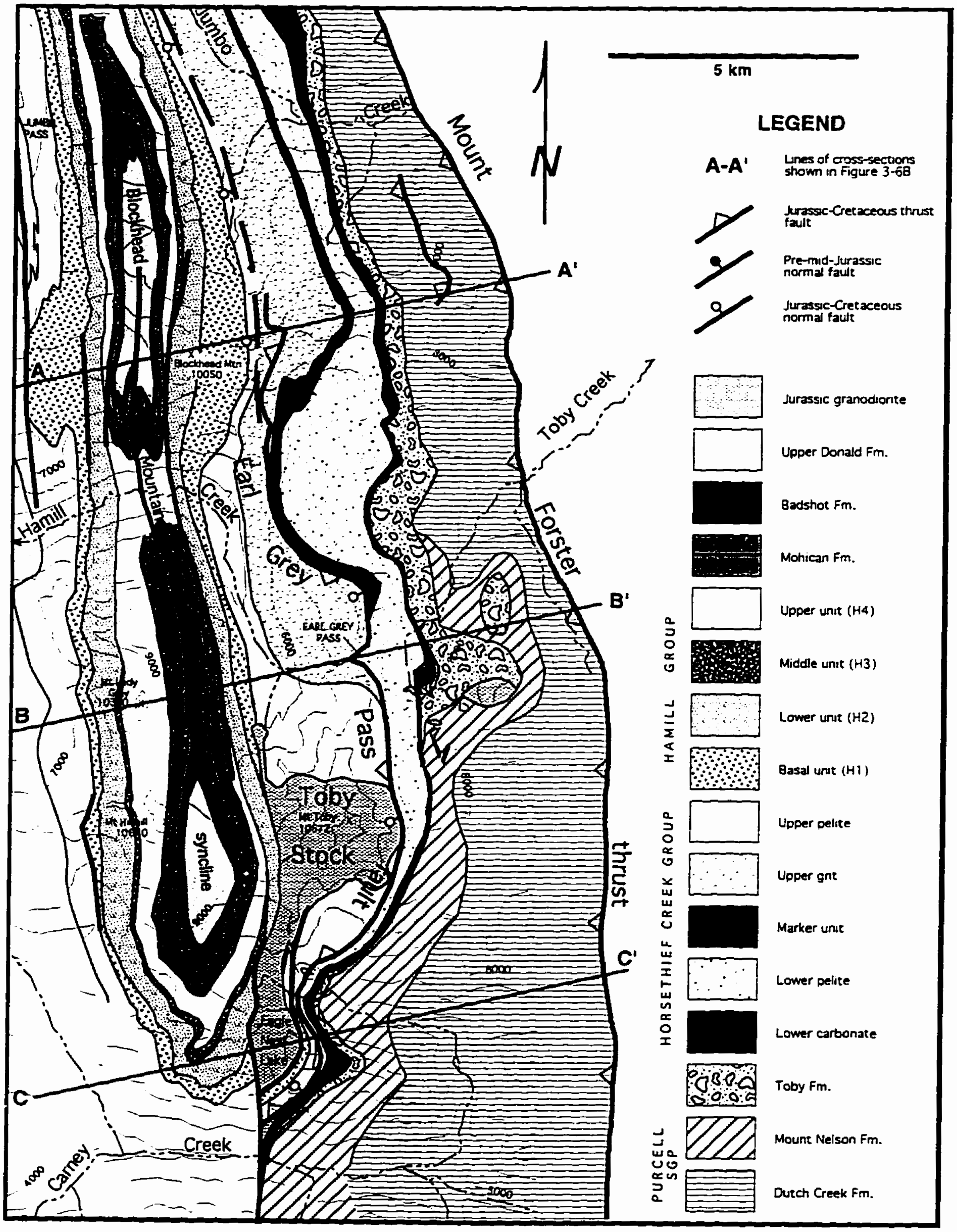



**Figure 3-6A:** Geological map of the southern part of the Blockhead Mountain syncline. The map illustrates stratigraphic realtionships across the Earl Grey Pass fault that indicate that it is a reactivated, post-Windermere normal fault. Cross sections sre shown in Figure 3-6B.

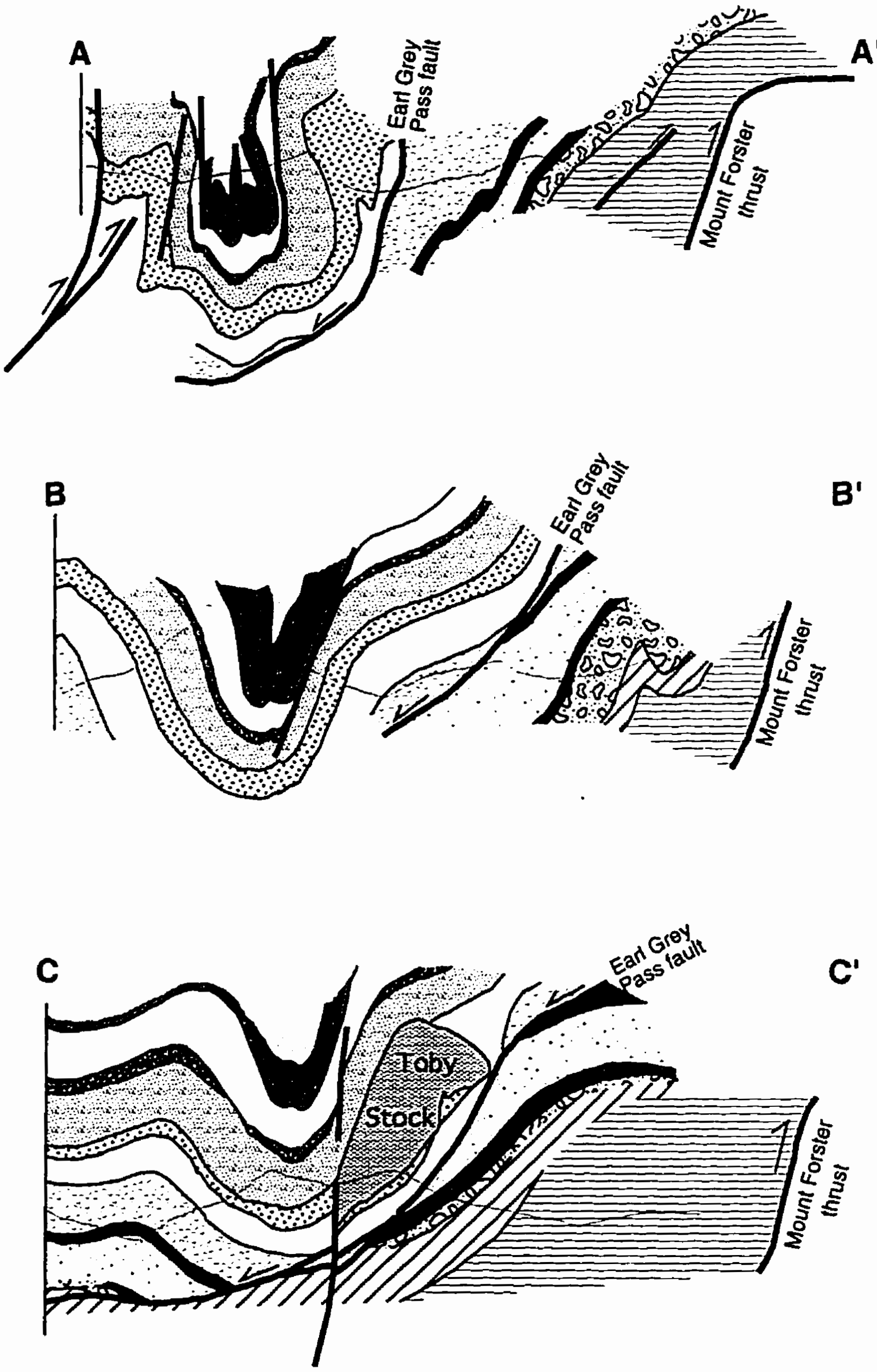



**Figure 3-6B**: Structural cross-sections to accompany Figure 3-6A, showing interpretation of reactivated Neoproterozoic to Early Paleozoic normal fault (Earl Grey Pass fault) at depth. See Figure 3-6A for legend.

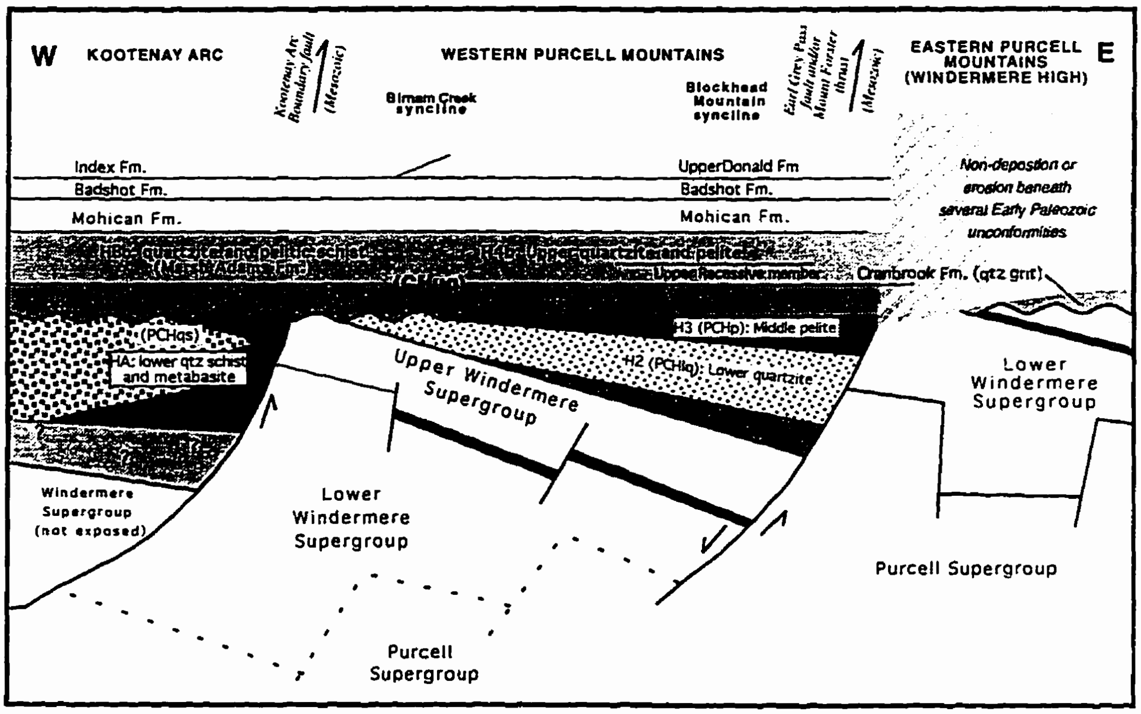



**Figure 3-7:** Summary of interpretation of stratigraphic relationships in and beneath the Hamill Group (shaded) in the Purcell anticlinorium and Kootenay Arc, illustrating the regional unconformity within the Hamill Group and its relationship to syn-depositional normal faulting. Thicknesses not to scale. Post-Badshot Formation normal faulting not depicted (see Fig. 4-3). Lithic designators in ( ) refer to map units on Plates 1-3.

stock, but after the formation of the cleavage that is axial planar to the Blockhead Mountain syncline (Chapter 4).

South of Earl Grey Pass, stratigraphic units in the Horsethief Creek Group are progressively cut out toward the south both in the hangingwall and in the footwall of this fault (Photo Plate 3F). East of the south end of the Toby stock, the upper pelite of the Horsethief Creek Group, which is overlain by unit H1 of the Hamill Group, is juxtaposed against the lower pelite of the Horsethief Creek Group, so that much of the Horsethief Creek Group is missing (Fig. 3-6a). South of this locality, and down dip on the fault, the lower pelite is cut out by the fault, and according to mapping by Reesor (1973), the Toby Formation is missing completely along this contact farther to the south of Carney Creek, so that the upper part of the Windermere Supergroup presumably lies directly on the Purcell Supergroup. These relationships indicate that an older normal fault has been reactivated, at least in part, as a thrust fault. The truncations observed in the footwall (east side) are interpreted as an oblique view of footwall ramp cutoffs on a normal fault. The corresponding hangingwall cutoffs in the Windermere Supergroup must lie down the dip of the fault beneath the Hamill Group strata that are exposed in the Blockhead Mountain syncline. Three structural cross sections through the southern part of the Blockhead Mountain syncline (Fig. 3-6b) illustrate an interpretation of the map relationships (Fig. 3-6a), and a "cartoon" of stratigraphic relationships (Fig. 3-7) illustrates an interpretation of the original normal fault.

This interpretation requires that this fault must continue beyond its observed trace to the north, although there is no direct evidence for it to the north of the divide between Hamill and Jumbo Creeks (Plate 2). Based on the interpretation presented below of the tectonic setting of the Hamill Group, this structure has been inferred to continue to the north within the upper pelite unit of the Horsethief Creek Group, and to die to zero displacement near the northern termination of the Blockhead Mountain syncline.

There are not enough data to constrain precisely the age of this structure, but two alternatives are considered. The hypothesized normal displacement could be either

Neoproterozoic to Early Paleozoic or it could be younger Paleozoic to Mesozoic, prior to peak metamorphism. Syn-orogenic Jurassic extension has not been documented elsewhere in this segment of the Cordillera, but it cannot be completely ruled out. Similarly, there is no regional evidence for late Devonian through Triassic extension. However, intermittent Neoproterozoic to Devonian uplift has been well documented in the eastern Purcell anticlinorium, associated with the "Windermere high" (Walker, 1926; Reesor, 1973; Root, 1987; Pope, 1989). Rocks as young as the Hamill Group are not exposed in the footwall of this fault, and on the "Windermere high" in the eastern Purcell Mountains Lower Paleozoic strata that may be equivalent to the upper quartzite unit of the Hamill Group are anomalously thin or absent due to Early Paleozoic erosion and/or non-deposition (Fig. 3-2). Therefore, there is better evidence, although indirect, to support the hypothesis that the normal fault mapped in the western Purcell anticlinorium was active during Neoproterozoic to Early Paleozoic uplift and extension.

**Tectonic setting of the Hamill Group in the west-central Purcell anticlinorium and Kootenay Arc**

A simple model for the depositional and tectonic setting of the Hamill Group in the west-central Purcell anticlinorium and adjacent Kootenay Arc (Fig. 3-7) can account for the lateral and vertical stratigraphic variations within the Hamill Group, including the local unconformites within and at the base of the Hamill Group, the sedimentary environments proposed for these strata and the observed and inferred Neoproterozoic to Early Paleozoic faults on the eastern limb of the Blockhead Mountain syncline and beneath the Kootenay Arc boundary fault, respectively. Units H1, H2 and H3 of the Hamill Group in the west-central Purcell anticlinorium were deposited in a syn-sedimentary half graben which deepened and was bounded to the east by a west-dipping normal fault. This fault may be represented by the reactivated normal fault on the eastern limb of the Blockhead Mountain syncline, or else it may be obscured by the Mount Forster thrust, a Mesozoic compressional structure. The inferred normal fault was active during the deposition of the basal, lower and middle units of the Hamill Group. The basal and lower units onlapped onto the emerging tilted block of Horsethief Creek Group to the west. The shallow marine basin received abundant sediment, probably supplied by rivers from the south, as well as some coarser, more locally-derived sediment from another uplifted block to the east. The supply of coarse and immature sediment decreased with time, although sedimentation kept pace with subsidence.

Immature clastic sediments and mafic igneous rocks of unit HA in the Kootenay Arc were deposited in a separate, deeper marine basin further to the west. Subsidence was more rapid in this basin. The western basin must have been separated from the uplifted and eroded block of Windermere Supergroup strata to the east by a west-dipping normal fault. Igneous activity and turbiditic deposition may have been localised along and adjacent to this structure. This fault appears to have been reactivated as the Kootenay Arc boundary fault, a Mesozoic compressional structure (Plates 1 and 2, and Fig. 3-7) that juxtaposes the western basin against the uplifted block.

The upper unit of the Hamill Group (units H4 and HB) in both the Purcell anticlinorium and Kootenay Arc was deposited after local normal faulting had ceased. It unconformably overlaps the underlying units of the Hamill Group and also the tilted and eroded edge of the block of Windermere Supergroup which separated the two basins. However, the apparently gradational nature of the contacts above and below the middle pelitic unit (H3) of the Hamill Group in the Blockhead Mountain syncline indicates that sedimentation in the down-dropped portion of the eastern basin may have been continuous during deposition of units H3 and H4. If so, little or no time elapsed between the cessation of normal faulting and the deposition of the upper quartzite unit (H4) of the Hamill Group across the tilted underlying strata. The upper quartzite unit and the overlying Mohican Formation record a change to a more open and regionally continuous shelf environment, in which clastic sedimentation was controlled by more margin-wide processes. The westward fining of the upper unit and westward thinning of the Mohican Formation suggest that the distal part of this shelf was to the west.

*Lower Cambrian/Lower Paleozoic strata above the Hamill Group*

**The Mohican Formation**

The Mohican Formation (Fyles and Eastwood, 1962) comprises thinly interbedded calcareous schist, chloritic schist, quartz-muscovite schist, green phyllite and metasiltstone, and minor quartzite and dolostone or dolomitic marble. The Mohican Formation in the eastern part of the Kootenay Arc can be divided into a calcareous lower part that contains at least two distinct marble intervals, and a schistose upper part that is much less calcareous (Fyles, 1964). The Mohican Formation gradationally overlies unit H4 of the Hamill Group in the Purcell anticlinorium and unit HB in the Kootenay Arc. It is conformably overlain by the Badshot Formation. The Mohican Formation varies in thickness from 50 m to greater than 300 m in the Purcell anrticlinorium. It is as much as 300 m thick in the eastern part of the Kootenay Arc, and it thins to less than 10 m where it is exposed to the west of Duncan Lake (Fyles, 1964). Much of the variation in thickness is probably due to tectonic thinning and thickening of this incompetent unit.

Primary sedimentary structures are not discernible in the fine-grained strata of most of the Mohican Formation. The most noteworthy feature of the Mohican Formation is a marker unit approximately halfway between its lower and upper contacts that comprises three thickly-bedded quartzarenite intervals, separated by quartz-muscovite schist. These three intervals are laterally continuous along the entire length of exposure of the Mohican Formation in the Blockhead Mountain syncline, a distance of more than 20 km. The quartzarenite intervals are each several meters thick each and are characterized by abundant trough and tabular cross beds in sets 0.5 to 1.0 m thick. The same marker unit may occur in the adjacent Kootenay Arc at this latitude, and it also may occur as far south as the International Border (J. T. Fyles, pers. comm., 1994), which would imply that the sedimentary environment that produced the marker intervals persisted over a large area.

## The Badshot Formation

The Badshot Formation (Walker and Bancroft, 1929) is characterized by resistant, white to medium grey, commonly laminated marble or dolomitic marble and minor pelitic and dolomitic schist. In the Kootenay Arc it tends to contain more quartz and pelite toward the top. It abruptly but apparently conformably overlies the Mohican Formation. In the Purcell anticlinorium it is conformably overlain in the Blockhead Mountain syncline by schist or phyllite considered equivalent to the Lower Cambrian upper Donald Formation of the Dogtooth Range to the north (Fig. 3-1), and it is conformably overlain in the northern part of the Birnam Creek syncline by the lower Index Formation of the Lower Paleozoic Lardeau Group. Where it is overlain by the lower Index Formation in the Birnam Creek syncline it is about 100 m thick. Where it is overlain by pelite and calcareous schist in the Blockhead Mountain syncline, it is less than 20 m thick, but it is apparently tectonically attenuated along the eastern limb of the syncline at this locality. In the Kootenay Arc, the thickness of the Badshot Formation varies markedly along strike. At Lavina Ridge (Plate 2), it is about 100 m thick. It is absent to the north toward Duncan Lake, but reappears and thickens abruptly to nearly 500 m north and east of Duncan Lake (Fyles, 1964). Fyles (1964) implied that variations in thickness of the Badshot Formation were structural, but several similar abrupt variations in the thickness of the Badshot Formation in the Illecillewaet synclinorium in the northern Selkirk Mountains have been interpreted as stratigraphic (Colpron and Price, 1993). The Badshot Formation is unfossiliferous in the Purcell anticlinorium, but it contains archeocyathans and other Lower Cambrian fossils in the northern Selkirk Mountains to the northwest (Wheeler, 1963; Logan et al., 1996), where it is exposed in a belt that is continuous along strike from the central Kootenay Arc.

## The upper Donald Formation (Blockhead Mountain syncline)

Interbedded light grey, green and tan quartz-muscovite-chlorite phyllite or schist and calcareous schist conformably overlie the dolomitic marble of the Badshot Formation in the core of the Blockhead Mountain syncline. Root (1987) mapped these strata as the Index Formation of the

Lower Paleozoic Lardeau Group, which overlies the Badshot Formation in the Kootenay Arc (Fyles and Eastwood, 1962). These strata, however, is distinct from the lower Index Formation. The lower Index Formation phyllite or schist is characteristically black, commonly graphitic, and contains abundant black or graphitic marbles and dark quartzite above the contact with the Badshot Formation. The succession exposed in the Blockhead Mountain syncline is lithologically similar to the upper part of the Donald Formation in the Dogtooth Range to the north (Evans, 1933, Simony and Wind, 1970), which overlies an archeocyathid-bearing limestone (the middle Donald Formation) considered equivalent to the Badshot Formation (Wheeler, 1960; Simony and Wind, 1970). These strata therefore considered equivalent to the upper part of the Donald Formation.

**The lower Index Formation of the Lardeau Group**

The Lardeau Group (Fyles and Eastwood, 1962; Fyles, 1964) was defined and mapped in the northern Kootenay Arc, where it overlies the Badshot Formation. It comprises several thousand meters of complexly folded and penetratively deformed schist, metabasite, feldspathic grit, conglomerate and minor carbonate. In the central Kootenay Arc east of Duncan and Kootenay Lakes, only the lower part of the Lardeau Group, the Index Formation (Fyles and Eastwood, 1962), is exposed.

The Index Formation is divisible into a lower member comprising dark, commonly graphitic schist, phyllite, marble and quartzite and an upper member comprising green schist and phyllite, light grey marble, and lenses of metabasite, mafic-clast conglomerate and feldspathic grit (Fyles, 1964, and this study). Graphitic marble and schist at the base of the Index Formation abruptly but apparently conformably overlies light grey marble and schist of the upper Badshot Formation. The lower part of the Index Formation also includes graded, possibly turbiditic, interbedded black quartzite and pelite that grades upward into homogeneous black pelite. The black pelite passes upward into green and light grey schist and phyllite of the upper member. The lower member of the Index Formation also includes at least one altered dike or sill, perhaps isoclinally folded and thus repeated, that is subparallel to bedding and varies from mafic to ultramafic compositions over contiguous exposures. The intrusions are up to 10 m thick, and one can be traced for 3 km.

The Index Formation was previously unrecognized on the western limb of the Purcell anticlinorium. A succession of black phyllite with thin beds or laminations of dark or black quartzite overlies the Mohican and Badshot Formations in the Birnam Creek syncline, in the immediate footwall of the Kootenay Arc boundary fault where the fault intersects Duncan Lake (Plate 2). This interval is considered an equivalent to the lower member of the Index which is exposed about three km to the west in the hangingwall of the Kootenay Arc boundary fault. It is important to note that, in contrast, the Badshot Formation is conformably overlain to the east, in the Blockhead Mountain syncline, by the upper Donald Formation. Thus a facies change must occur above the Badshot Formation in the western Purcell anticlinorium between the Blockhead Mountain and Birnam Creek synclines.

The contact and stratigraphic relationships beneath and within the Index Formation in the Duncan Lake area support the conclusions, based on data from the northern Selkirk Mountains, that the Lardeau Group is an upright, stratigraphic sequence that depositionally overlies the Badshot Formation (Colpron and Price, 1995). These relationships and those described by Colpron and Price (1995) refute the hypothesis that the Lardeau Group is a sequence of units that is inverted because it had been internally imbricated and tectonically emplaced over the Badshot Formation (Smith and Gehrels, 1992b). Documentation of the nature of the stratigraphic contact between the Badshot and the Index Formations is critical to subsequent interpretations of the depositional and tectonic setting of the Lardeau Group and of its relationship to the Lower Paleozoic North American miogeoclinal succession that is exposed to the east.

*Depositional and tectonic settings of post-Hamill Lower Paleozoic strata in the west-central Purcell anticlinorium and eastern Kootenay Arc*

The Mohican Formation was probably deposited in a shallow marine, shelf setting that had been established during deposition of the upper part of the Hamill Group. The Mohican Formation records waning of coarse clastic input to the shelf, implying relative sea-level rise, and the initiation of conditions that locally favored carbonate production. The cross-bedded marker unit marks an influx of sand that was transported and reworked in a high-energy setting. The lateral extent of these three thin intervals suggests that this influx was controlled by a widespread relative sea-level fall.

The upper Donald Formation in the Blockhead Mountain syncline is lithologically similar to much of the Mohican Formation, and it probably was deposited in a similar shallow marine environment with moderate clastic input. The intervening Badshot Formation represents a hiatus in clastic input to this environment, during which carbonate production was favored. The upper Donald Formation near its type locality in the Dogtooth Range is lithologically similar to that in the Blockhead Mountain syncline, but it is less deformed and contains well-preserved sedimentary structures and Lower Cambrian fossils. Kubli (1990) interpreted the upper Donald Formation in the Dogtooth Range to record progradation of terrigenous clastic sediment onto an open, shallow marine shelf following a period of carbonate production on the shelf (the middle Donald or Badshot Formation).

The lower and upper members of the Index Formation of the Lardeau Group in the Kootenay Arc and the Birnam Creek syncline differ conspicuously both from the underlying, laterally continuous shallow marine strata in the Badshot and Mohican Formations and upper quartzite of the Hamill Group as well as from the laterally continuous, initially shallow marine Lower Paleozoic miogeoclinal rocks that are exposed in the eastern Purcell Mountains and in the Rocky Mountains to the east. The lower, black phyllite member of the Index Formation also differs markedly from the rest of the Lardeau Group, and it is important for subsequent discussions to emphasize that the depositional and tectonic setting of the lower Index Formation was probably different from that of the upper Index Formation and younger Lardeau Group strata.

The lower member of the Index Formation was deposited in a subsiding, anoxic marine basin that was superposed on the Badshot Formation. The contact between the Badshot and Index Formations represents the point at which shallow-water carbonate production could no longer keep pace with subsidence. The basin received fine, distal turbiditic carbonate and siliciclastic sediment, probably derived from the shelf environments to the east. The supply of sediment initially decreased as the basin deepened and relative sea level increased.

The upper member of the Index Formation records the initiation of submarine volcanism and rapid deposition, in more oxygenated conditions, of sediment eroded from uplifted feldspathic and carbonate sources and possibly from mafic volcanic sources. Local volcanism and deposition of coarse, immature sediment culminated in the extrusion of a thick, regionally extensive sequence of mafic volcanic rocks (the Jowett Formation) and in the deposition of a thick, regionally continuous but heterogeneous sequence of grit containing abundant potassium feldspar (Broadview Formation). These relationships imply that the the lower part of the Index Formation is a distal and deeper-water equivalent of the upper Donald Formation and was deposited in a setting that was relatively quiescent tectonically, and that the rest of the Lardeau Group was deposited in a down-dropped basin adjacent to uplifted source areas. The deep subsidence recorded by this succession, the stratigraphic contact with underlying Lower Cambrian strata that are correlative with Lower Cambrian strata in the miogeocline to the east, and trace-element geochemical data from mafic rocks of the Lardeau Group (Smith, 1990; Colpron and Logan, 1995; Warren, unpublished data, Appendix 3) together suggest that this basin was developed on significantly attenuated continental crust of the Laurentian margin.

# PART II:
# REGIONAL STRATIGRAPHIC RELATIONSHIPS BETWEEN NEOPROTEROZOIC TO LOWER CAMBRIAN STRATA IN THE SOUTHEASTERN CANADIAN CORDILLERA

The distinctive quartzarenite/quartzite marker unit that comprises the base of the upper part of the Hamill Group, and the unconformity that underlies it in the western Purcell anticlinorium and central Kootenay Arc, provide the basis for correlating and comparing the tectonic setting of the Hamill and Gog Groups northward and eastward into the northern Selkirk and Dogtooth Mountains and the Western Main Ranges of the Rocky Mountains, and also southward into the southern Purcell and Selkirk Mountains (Fig. 3-1). Throughout an extensive area of the southern Canadian Cordillera, a sheet of Lower Cambrian shallow marine quartzarenite/quartzite and interbedded pelite overlies a regional unconformity along which it overlaps older rock units, including the mainly unfossiliferous fluvial, shallow marine and deeper marine units of the Hamill and Gog Groups and the Three Sisters Formation, the underlying Windermere Supergroup and, southwest of the St. Mary fault, progressively older strata of the Purcell Supergroup (Fig. 3-2). The distribution of the older units of the Hamill and Gog Groups and pattern of erosional truncation of the underlying Windermere Supergroup indicate that they were deposited in a series of north-trending half grabens that were bounded mainly by syn-sedimentary, west-dipping normal faults (Kubli and Simony, 1992; Lickorish and Simony, 1995; and this study). Regional correlations and stratigraphic relationships that support this hypothesis are summarized and discussed below.

## THE DOGTOOTH RANGE

The stratigraphy of the Horsethief Creek and Hamill Groups in the Dogtooth Range, at the northern termination of the Purcell anticlinorium, is similar to that in the west-central Purcell anticlinorium (this study), and lithostratigraphic correlations between the two areas are straightforward (Fig. 3-2).The Hamill Group in the Dogtooth Range comprises three lithostratigraphic units. Evans (1933) originally correlated them with part of the Gog Group in the Rocky Mountains, but the precise correlations proposed subsequently were shown to be invalid, and subsequent workers have referred to the subdivisions informally as the lower, middle and upper units (Wheeler, 1963; and references cited in Kubli and Simony, 1992). These three units are considered equivalent to quartzose units H1 and H2, pelitic unit H3 and quartzose unit H4 of the Hamill Group in the west-central Purcell Mountains, respectively, primarily on the basis of sediment composition and sedimentary structures.

The lower unit of the Hamill Group in the Dogtooth Range comprises subarkose and quartzarenite that contains abundant blue and purple quartz. Tabular cross beds similar to those in units H1 and H2 in the west-central Purcell anticlinorium are common, and original foreset dips are to the southwest (Ellison, 1967; Kubli, 1990). The lower unit generally contains more feldspar toward the base (Devlin, 1986), although no distinct feldspathic basal unit has been described in the Dogtooth Range. However, Kubli (1990) suggested that some of the sandy upper clastic unit of the Horsethief Creek Group in the Dogtooth Range may be equivalent to the lowermost part of the Hamill Group. The lower unit of the Hamill Group in the Dogtooth Range is thus correlated with the lower quartzarenite unit H2 of the west-central Purcell anticlinorium, and it may also be equivalent to at least part of the basal feldspathic unit H1.

The middle unit comprises pelite and silty argillite interbedded with argillaceous sandstone, similar to unit H3 in the west-central Purcell anticlinorium. The upper unit comprises dominantly tabular cross-bedded white, pink and green quartzarenite, pebble conglomerates and minor interbedded pelite. Quartzarenites of the upper unit are subdivided by a pelitic or calcareous marker known as the upper recessive member (Wind, 1967). The proportion of interbedded pelite is greater above the upper recessive member (Evans, 1933). This stratigraphic succession is similar to that of unit H4 of the Hamill Group in the northern part of the Blockhead Mountain syncline (Fig. 3-3). The thicknesses of units H3 and H4 in the Dogtooth Range are similar to those of the middle and upper units of the west-central Purcell anticlinorium, but the lower unit in the Dogtooth Range is generally about 200 m thinner than units H1 and H2.

The arenites of the lower and upper units are distinctly different in the Dogtooth Range as they are in the west-central Purcell anticlinorium. The upper unit is markedly less feldspathic and more texturally mature than the lower unit (Ellison, 1967; Devlin and Bond, 1988). The upper unit contains Lower Cambrian trilobites (Evans, 1933) and trace fossils (*Skolithos*; Kubli, 1990), but the lower unit is unfossiliferous (Evans, 1933; Ellison, 1967). The upper unit has been interpreted as a shallow marine, tidally-dominated deposit (Ellison, 1967; Devlin and Bond, 1988), whereas the environment of the lower unit is controversial. Ellison (1967) interpreted it as shallow marine, and Devlin and Bond (1988) proposed a fluvial envornment. Kubli (1990) argued for a tidal setting, based on the abundance of tabular cross beds and the absence of trough cross beds. Nonetheless, it is sedimentologically distinct from the upper unit.

The Hamill Group thins markedly to the south and to the west in the Dogtooth Range, from about 1200 meters (Simony and Wind, 1970) to about 450 meters (Evans, 1933). Kubli and Simony (1992) concluded that the lower and middle units of the Hamill Group pinch out to the south and west, and that the upper unit rests directly and unconformably on the Horsethief Creek Group. In the western Dogtooth Mountains, the Hamill Group is missing entirely and the Lower Cambrian Donald Formation directly overlies the Horsethief Creek Group. However, immediately to the west, on the opposite side of the Beaver River fault (Fig. 3-1), 3000 meters of Hamill Group strata are conformably overlain by the Badshot Formation, which can be correlated on the basis of its *Archeocyathid* fauna with the middle Donald Formation (Wheeler, 1963).

Kubli and Simony (1992) interpreted these relationships as the result of onlap of the Hamill Group and the Donald Formation onto the the east flank of the "Dogtooth High," an emergent area produced by tilting of the Horsethief Creek Group above southwest-dipping listric normal faults that were active during deposition of the Hamill Group. The "Dogtooth High" is onlapped to the east by an eastward-thickening wedge of Hamill Group strata , and is bounded to the west by a zone of westward- or southwestward-dipping normal faulting.

It is not clear whether the absence of the upper unit of the Hamill Group on the "Dogtooth High" is due to non-deposition or to subsequent erosion associated with renewed normal faulting and block tilting, which is suggested by volcanism associated with the sub-Donald Formation unconformity (Kubli and Simony, 1992). Kubli and Simony (1992) implied that the lower, middle and upper units are a conformable sequence that onlapped progressively onto the emerging "Dogtooth High." An alternative interpretation of their data is that the upper unit unconformably overlies the middle unit and the Horsethief Creek Group and that it may have been continuous across the high prior to deposition of the Donald Formation. This hypothesis is supported by the occurrence of coarse breccias and channel fill that contain clean quartzite in the base of the Index Formation, which can be correlated with the upper Donald Formation on the basis of their positions immediately above the regional archeocyathan-bearing marble, in the northern Selkirk Mountains to the west of the "Dogtooth high" (Brown, 1991; Logan et al., 1996). The quartzite blocks and channel fill are interpreted as Hamill Group that was eroded from the western part of the "Dogtooth high" during renewed syn-Donald uplift and normal faulting. Whether the upper quartzite unit of the Hamill Group was deposited across the "Dogtooth high" or not, it can be argued in either case that deposition of at least the lower and middle units of the Hamill Group in the Dogtooth Range was controlled by the development of an eastward-deepening half-graben.

## THE NORTHERN SELKIRK MOUNTAINS AND KOOTENAY ARC

The northern Selkirk Mountains and much of the Kootenay Arc comprise a tectonostratigraphic domain that is distinct from the western and northern Purcell anticlinorium (Fig. 3-1). The upper quartzite unit of the Hamill Group, the Mohican Formation and the Badshot Formation are exposed in both the west-central Purcell anticlinorium and the Kootenay Arc, but they overlie markedly different stratigraphic successions of the lower part of the Hamill Group, and they are overlain by conspicuously different Lower Paleozoic successions. The succession of immature metasedimentary rocks and metabasites that underlies the upper quartzite of the Hamill Group in the central Kootenay Arc (this study) can be traced northward into the northern Selkirk Mountains, where it overlies the Horsethief Creek Group (Wheeler, 1963; Devlin and Bond, 1988; Colpron and Logan, 1995), and to the south of the Fry Creek batholith along the Kootenay Arc

(Höy, 1980). Throughout the Kootenay Arc and northern Selkirk Mountains, the clastic/metabasite unit is abruptly overlain by a distinct clean quartzite that appears to become thinner, finer and less continuous toward the north and west (Wheeler, 1963; Logan and Colpron, 1995).

The Hamill Group in the northern Selkirk Mountains comprises a lower feldspathic metasandstone unit, a middle unit comprising abundant subaqueous, mafic to intermediate metavolcanic rocks and turbiditic metasandstones, and an upper more mature quartz sandstone unit that grades upward into the Mohican Formation (Wheeler, 1963; Devlin , 1989; Logan and Colpron, 1995). In the eastern part of the northern Selkirk Mountains the middle unit of the Hamill Group contains abundant coarse, carbonate debris flow facies and less finer-grained turbiditic siliciclastic facies, is markedly and abruptly thinner than it is to the west, and lacks volcanic rocks (Devlin, 1989). Yet this sequence of Hamill Group is distinct from that exposed in the Dogtooth Range to the east and is separated from it by the tilted and bevelled Dogtooth high (Kubli and Simony, 1992). The lower unit of the Hamill Group west of the Beaver River fault gradationally overlies pelite of the upper Horsethief Creek Group (Devlin, 1986). Farther to the north, the stratigraphic contact between the Hamill Group and a sequence of mixed clastic and carbonate rocks of the upper part of the Windermere Supergroup is ambiguous (Logan and Colpron, 1995; M. Colpron, pers. comm., 1996).

In the Kootenay Arc to the south of the Fry Creek batholith, the Hamill Group comprises a thick succession of intercalated dark schist, amphibolite and quartzite, abruptly overlain by a distinctive clean quartzite that grades upward into interbedded quartzite and pelite (Fig. 3-2; Höy, 1980). The Hamill Group is unconformably overlain by the Mohican Formation. Its base is not exposed because it is truncated by the Kootenay Arc boundary fault (West Bernard fault of Höy, 1980), but in an adjacent structural panel in the immediate footwall of the Kootenay Arc boundary fault a similar thick succession of metabasites and immature clastic rocks is underlain by a unit of thickly cross-bedded quartzite and grit that is more feldspathic toward the base. The feldspathic grit apparently gradationally overlies pelite of the upper Horsethief Creek Group (Höy, 1980), but this contact also has been interpreted as unconformable (Lis and Price, 1976; M. Lis, unpublished data).

The lower feldspathic arenite unit of the Hamill Group in the immediate footwall of the Kootenay Arc boundary fault has been interpreted as fluvial (Höy ,1980; Devlin, 1986). The similar lower feldspathic unit that is exposed along strike in the northern Selkirk Mountains has been interpreted as shallow marine (Devlin, 1989). Devlin (1989) concluded that the conformably overlying immature clastic rocks and metabasites of the middle unit indicate an abrupt change to deeper marine sedimentation, with rapid subsidence and volcanism localized along a submerged normal fault near the eastern side of the basin. The proposed fault must have been located to the west of the normal fault that marked the western side of the "Dogtooth high". The two normal faults must merge toward the central Kootenay Arc (this study) and continue to the south (see Fig. 3-9). Devlin (1989) observed that the upper mature quartzite unit of the Hamill Group appears to be continuous across the inferred submerged normal fault in the northern Selkirk Mountains, and he suggested that the upper quartzarenite unit consists of reworked sediment that was deposited on a continental shelf following an episode of crustal stretching and normal faulting, and during the initial stages of thermal subsidence of a passive margin.

## SOUTHWESTERN PURCELL ANTICLNORIUM AND SOUTHERN KOOTENAY ARC

The Seeman Creek fault (Leclair, 1988), a southern equivalent of the Kootenay Arc boundary fault (Chapter 4) separates overturned strata primarily of the Hamill Group, Badshot Formation and Lardeau Group in the southern Kootenay Arc from a similar but distinct and much thicker succession on the southwestern flank of the Purcell anticlinorium (Fig. 3-2). A homoclinal succession of dominantly pelitic Windermere Supergroup strata, overlain by feldspathic arenite and grit of the Three Sisters Formation (Rice, 1941; Leclair, 1988), is exposed in the footwall of the Seeman Creek fault. The Three Sisters Formation is abruptly overlain by more mature quartzarenite, interbedded pelite and minor carbonate of the Quartzite Range Formation, the upper part of which contains Lower Cambrian trilobites (Rice, 1941) and trace fossils (Devlin, 1986). The Quartzite Range Formation is overlain by an archeocyathan-bearing carbonate unit that appears to be a lateral equivalent of the Badshot Formation. The Three Sisters Formation has been included by some previous workers in the Windermere Supergroup, based on its sedimentological immaturity and on an apparently gradational lower contact (Rice, 1941), but it also has been considered with the Quartzite Range Formation as equivalent to the Hamill Group, based on the abundance of quartz in it compared to grits of the underlying Windermere Supergroup (Little, 1960) and on its sedimentological similarities with the fluvial lower part of the Hamill Group to the north (Devlin, 1986; Devlin and Bond, 1988).

However, no unit similar to the Three Sisters Formation occurs in the Hamill Group that is exposed immediately to the west in the hangingwall of the Seeman Creek fault. In the immediate hangingwall of this fault, mature quartzite of the Hamill Group abruptly overlies a similar but thinner succession of Windermere Supergroup strata (Leclair, 1988). Mapping by Leclair (1988) shows that the Three Sisters Formation is abruptly truncated to the west against a normal fault of undetermined age in the immediate footwall of the Seeman Creek fault. A possible interpretation of this fault and of the contrasting stratigraphic sequences in the hangingwall and footwall of the Seeman Creek fault is that the Three Sisters Formation is truncated agianst a Neoproterozoic to Early Cambrian normal fault, and that quartzarenite and overlying quartzite and pelite of the Lower Cambrian Quartzite Range Formation and Hamill Group were deposited across the Three Sisters Formation, the normal fault and an uplifted block of Windermere Supergroup to the west after normal faulting had ceased.

## THE EASTERN PURCELL ANTICLINORIUM AND ADJACENT ROCKY MOUNTAIN TRENCH

The eastern Purcell anticlinorium and parts of the adjacent Rocky Mountain Trench and western Main Ranges in the hangingwall of the Redwall - Lussier River faults comprise four domains that are stratigraphically distinct from each other and from the Dogtooth Range and the western Purcell anticlinorium in the hangingwall of the Mt. Forster - Redding Creek fault (Fig. 3-1).

The domains are defined by changes in Lower Paleozoic stratigraphy across major faults. They include, from south to north, the block to the south and east of the St. Mary fault, the block between the St. Mary and Hall Lake faults, the block that is between the Mount Forster, Redwall - Lussier River and Hall Lake faults and the immediate hangingwall of the Mount Forster and Purcell faults in the Rocky Mountain trench. Lower Cambrian strata rest on progressively older rocks toward the south. They are markedly thinner than they are to the west, or else they are absent altogether. They do not include any feldspathic grit units similar to those in the lower part of the Hamill Group in the Dogtooth Range, western Purcell anticlinorium or northern Selkirk Mountains. The name "Cranbrook Formation" has been extended northward (Reesor, 1973; Foo, 1979; Bennett, 1986) from the type locality of that formation south of the St. Mary fault (Rice, 1937) to include quartzose strata that are presumably Lower Cambrian and that unconformably overlie the Windermere or Purcell Supergroups. However, the relationships between these quartzose strata north of the St. Mary fault, the Cranbrook Formation at its type locality and the Hamill Group to the west have remained uncertain.

The type Cranbrook Formation and the type overlying Eager Formation outcrop in the Purcell anticlinorium and in the Rocky Mountain Trench to the south of the St. Mary fault (Rice, 1937 and 1941). The Cranbrook Formation comprises cross-bedded quartzarenite and minor siltstone, carbonate and conglomerate. Its age and marine setting are confirmed by the presence of Lower Cambrian trilobites (Leech, 1954). The Cranbrook Formation is conformably overlain by argillite of the Lower Cambrian Eager Formation in the Cranbrook area (Rice, 1937) and in the Purcell Mountains to the west (Rice, 1941). The Cranbrook Formation rests unconformably on several formations within the Mesoproterozoic Purcell Supergroup (Rice, 1937), and cobbles and angular blocks of underlying Purcell Supergroup occur in the base of western exposures of the Cranbrook Formation (Rice, 1941). The Windermere Supergroup is entirely absent.

Strata that are perhaps equivalent to the Cranbrook and Eager formations outcrop in the Rocky Mountain Trench south of Canal Flats, B. C. and in the Hughes Range immediately to the east (Leech, 1954). In the Rocky Mountain Trench the "Cranbrook Formation" overlies a thinned succession of Horsethief Creek Group, and in the Hughes Range it overlies variable thicknesses of Toby Formation at the base of the Windermere Supergroup. The overlying "Eager Formation" is unconformably overlain by the Middle to Upper Cambrian Jubilee Formation in the Canal Flats area.

In the eastern Purcell anticlinorium south and east of the Mount Forster fault, north of the Hall Lake fault and west of the Redwall - Lussier fault, Lower Cambrian rocks are anomalously thin or entirely absent. Lower Cambrian strata thin toward the south and west from the Dogtooth Range, and toward the north from the Canal Flats area, at least in part due to erosional truncation beneath the Upper and(?) Middle Cambrian Jubilee Formation. This is one of a series of overstepping unconformities related to intermittent uplift and eastward tilting of the "Windermere High" in the footwall of the Mount Forster fault during the interval from Early Cambrian to Middle Devonian time (Walker, 1926; Reesor, 1973; Root, 1987; Pope, 1989).

A thin, discontinuous unit of unfossiliferous quartzite and quartz grit unconformably overlies the Horsethief Creek Group and is unconformably overlain by the Upper and(?) Middle Cambrian Jubilee Formation in the footwall of the Mount Forster fault in the eastern Purcell Mountains and in the Rocky Mountain Trench (Reesor, 1973; Bennett; 1986). This unit could therefore be Lower Cambrian, although it could be significantly younger than the Hamill Group and/or the Cranbrook Formation. The lower part of the Horsethief Creek Group, which is preserved beneath this unit, can be correlated with the lower part of the Horsethief Creek Group in the west-central Purcell anticlinorium and in the Dogtooth Range in the hangingwall of the Mount Forster fault (Root, 1987; Kubli, 1990; Pope, 1990; Chapter 2), but the uppermost part of the Horsethief Creek Group is missing, indicating that the quartz grit unit was deposited on strata that had been uplifted and eroded between Neoproterozoic and Middle Cambrian time.

A similar quartzarenite unit occurs unconformably above the Toby Formation and the Horsethief Creek Group and is unconformably overlain by the Jubilee Formation in the Rocky Mountain Trench in the immediate hangingwall of the Mount Forster fault (Reesor, 1973; Bennett, 1986). Evans' (1933) map shows that the Lower Cambrian upper quartzarenite unit of the Hamill

Group in the Dogtooth Range can be traced southward through the Rocky Mountain Trench to these exposures. Thus, the Rocky Mountain Trench in the hangingwall of the Mount Forster fault marks the upthrown eastern edge of the basin of Hamill Group strata that onlapped onto the "Dogtooth high" (Kubli and Simony, 1992; Kubli and Simony, 1994), against which the lower and middle units of the Hamill Group are truncated and across which the upper quartzarenite unit is continuous.

Exposures of the upper quartzarenite unit of the Hamill Group in the hangingwall of the Mount Forster fault occur within a few kilometers of exposures of the thin, unfossiliferous quartzarenite and grit unit in the footwall (Evans, 1933; Reesor, 1973; Bennett, 1986). Both units unconformably overlie the Windermere Supergroup and are unconformably overlain by the Jubilee Formation in this area. The displacement on the Mount Forster fault is probably less than 10 km, based on offset of the Toby Formation and of the axial trace of the Mount Forster syncline on Reesor's (1973) map. Therefore, it is more appropriate to correlate the unfossiliferous quartz grit unit with the upper quartzarenite unit of the Hamill Group, rather than with the Cranbrook Formation as proposed by Reesor (1973). This correlation would imply that this Lower Cambrian unit was probably originally continuous across the "Windermere high," and that it was removed by erosion on the western part of the high beneath the sub-Jubilee unconformity during subsequent eastward tilting of the "Windermere high." The west-dipping normal fault that the quartzarenite presumably overlapped is either obscured beneath the Mount Forster fault or is preserved as the reactivated normal fault on the eastern limb of the Blockhead Mountain syncline in the west-central Purcell anticlinorium.

The relationship between the Cranbrook Formation at its type locality and the Hamill and Gog Groups and Quartzite Ranges Formation to the north, east and west has been uncertain, due to absence of the overlying regional archeocyathan-bearing carbonate marker unit and to lithological differences between the Cranbrook - Eager and Hamill - Donald (Badshot) successions. However, other fauna indicate that the Cranbrook - Eager succession is at least in part correlative with the upper Hamill - Donald succession in the Dogtooth Range (Fig. 3-2). Both the upper part of the upper quartzarenite of the Hamill Group in the Dogtooth Range and the upper part of the "Cranbrook" Formation north of the St. Mary fault contain *Nevadella*-zone trilobites, and the Eager Formation both north and south of the St. Mary fault contains *Bonnia-Olenellus*-zone trilobites, as does the entire Donald Formation in the Dogtooth Range, including the archeocyathan carbonate (Evans, 1933; Rice, 1937; Leech, 1954; Fritz et al., 1991).

Stratigraphic relationships beneath the the Cranbrook Formation imply that it was deposited in a sedimentary environment and tectonic setting that was similar to that of the upper quartzarenite unit of the Hamill Group. The Cranbrook Formation, like the upper quartzarenite unit of the Hamill Group and the Quartzite Range Formation, is a shallow marine unit that was deposited unconformably on strata that were uplifted and eroded during Neoproterozoic to Early Cambrian time (Fig. 3-2), although the normal fault that was the antecedent to the St. Mary reverse fault experienced greater displacement, in part syn-Windermere (Lis and Price, 1976), than the post-Windermere faults that controlled the deposition of the lower Hamill Group and Three Sisters Formation (see Figs. 3-9 and 3-11, discussed below). The area in which the Cranbrook Formation is preserved subsequently experienced significant subsidence to produce the "Eager Trough" while Early Paleozoic uplift continued on "Montania" to the south and on the "Windermere high" to the north. Thus the Cranbrook and Eager Formations were isolated from exposures of the upper quartzarenite of the Hamill Group and younger strata in the northern Purcell anticlinorium and Kootenay Arc.

## THE MAIN RANGES OF THE ROCKY MOUNTAINS

Thick successions of dominantly quartz sandstones of the Gog Group and of dominantly grit and slate of the underlying Neoproterozoic Miette Group (Windermere Supergroup) occur in thrust sheets in the Main Ranges of the Rocky Mountains (Figs. 3-1 and 3-2). In the Jasper area the Gog Group contains the archeocyathan-bearing Mural Formation, which has long been considered a lateral equivalent of similar fossiliferous carbonates of the Donald, Badshot and Laib formations of the Dogtooth Range, and northern and southern Selkirk Mountains, respectively

(Little, 1960; Wheeler, 1963; Aitken, 1969) On the basis of this correlation, the part of the Gog Group that lies beneath the Mural Formation has been correlated with the Hamill Group. In the eastern Main Ranges near Lake Louise and Mount Assiniboine, the Gog Group is undivided and comprises the entire Lower Cambrian succession (Walcott, 1910; Deiss, 1939, 1940; Fritz et al., 1991).

Southward between Lake Louise and Jasper in the Eastern Main Ranges, the Gog Group rests unconformably on a markedly lower stratigraphic level in the Windermere Supergroup (on or above strata that have been interpreted as equivalent to the Old Fort Point Formation; Aitken, 1969), and it is also conspicuously thinner. (e.g. Mountjoy, 1962; Aitken, 1969). The base of the thicker Gog Group succession to the north comprises a distinct thick, feldspathic arenite and grit unit (the Jasper Formation or lower part of the McNaughton Formation; see Fig. 3-2) of intercalated fluvial and shallow marine deposits (Young, 1979). This unit is absent to the south, where the entire Gog Group comprises quartzose shallow marine strata (Walcott, 1910; Deiss, 1939, 1940).

A significant unconformity within the Gog Group has been documented in the Solitude Range of the Western Main Ranges (Lickorish and Simony, 1995). The coarse, feldspathic arenite and grit unit at the base of the Gog Group (lower part of the McNaughton Formation) is truncated abruptly toward the south and east against an inferred normal fault that is overlapped by a distinct shallow marine quartzarenite unit (Solitude member of the McNaughton Formation; nomenclature proposed by Lickorish and Simony, 1995). The feldpathic unit reappears in the footwall of the normal fault and gradually thickens in a direction perpendicular to the fault. The quartzarenite unit unconformably overlies both the discontinuous feldspathic basal unit and the upthrown, tilted and bevelled succession of Miette Group strata in the footwall of the fault. Lickorish and Simony (1995) thus concluded that this discontinuous feldspathic unit was deposited in the down-dropped segments of eastward-deepening half-grabens. On the basis of these relationships in the Western Main Ranges, these authors suggested that it is the unconformity within the Gog

Group, rather than that at the base of the Gog Group, that overlaps more deeply eroded Windermere Supergroup strata toward the south in the Eastern Main Ranges.

## DISCUSSION

The distribution and sedimentology of the Hamill/Gog Group strata and equivalent rocks that lie below the regional unconformity are conspicuously different from the distribution and sedimentology of the Hamill/Gog Group strata that lie above this unconformity. These differences indicate a change from an actively extending continental margin to a continental shelf setting during Hamill/Gog time. They are consistent with the initiation of post-rift thermal subsidence and with the establishment of a passive continental margin during latest Proterozoic to Early Cambrian time, as outlined by Bond and Kominz (1984) from analysis of tectonic susidence recorded by Lower Paleozoic miogeoclinal strata in the Rocky Mountains, and by Devlin and Bond (1988) from the sedimentology of the Hamill Group in the northern Selkirk Mountains. The palinspastic restorations of the lower and upper Hamill/Gog basins discussed below (Figs. 3-8 to 3-11) not only support these previous interpretations but also contribute a regional paleogeographic and tectonic framework for the the rift-to-drift transition and provide insight into the configuration and behavior of the upper continental crust during this phase of rifting and initial thermal subsidence. A compilation of fossil data from below and above the "break-up" unconformity (Fig. 3-2), coupled with updated absolute age constraints for the Precambrian-Cambrian boundary and Early Cambrian period (Grotzinger et al., 1995) provide improved precision in elucidating the timing of the rift-to-drift transition.

### REGIONAL PALEOGEOGRAPHY AND TECTONIC SETTING DURING DEPOSITION OF THE LOWER PART OF THE HAMILL GROUP

Several north-trending, fault-bounded shallow sedimentary basins are outlined by the distribution of the lower parts of the Hamill/Gog Groups and equivalent rocks (Fig. 3-9). The basin

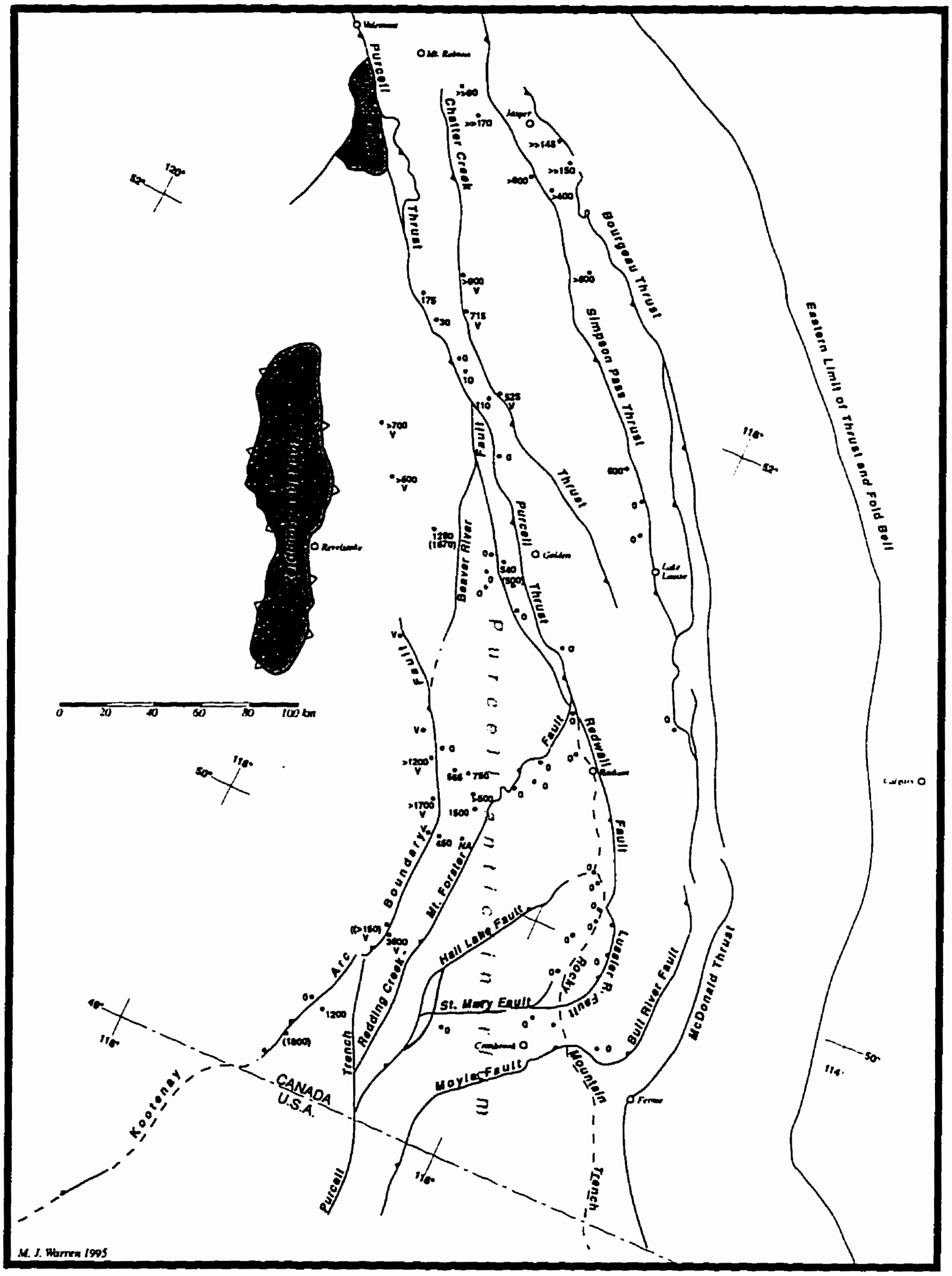


**Figure 3-8:** Thickness data for the lower part of the Hamill and Gog Groups in the southeastern Canadian Cordillera, used to construct the palinspastically restored paleogeographic map for lower Hamill/Gog time (Figure 3-9). Thicknesses in meters. **V** = locality of mafic flows or sills. > = thickness exceeds limit of measured section. >> = thickness greatly exceeds limit of measured section. Values in ( ) may be unrelaible due to strong deformation or uncertainty in correlations. Data compiled from Devlin (1986), Evans (1933), Fritz et al. (1991), Höy (1980), Kubli and Simony (1992), Leclair (1988), Lickorish and Simony (1995), Little (1960), Rice (1937, 1941), Root (1987), Simony and Wind (1970), and Warren (this study). Note that north is oblique to edge of figure.

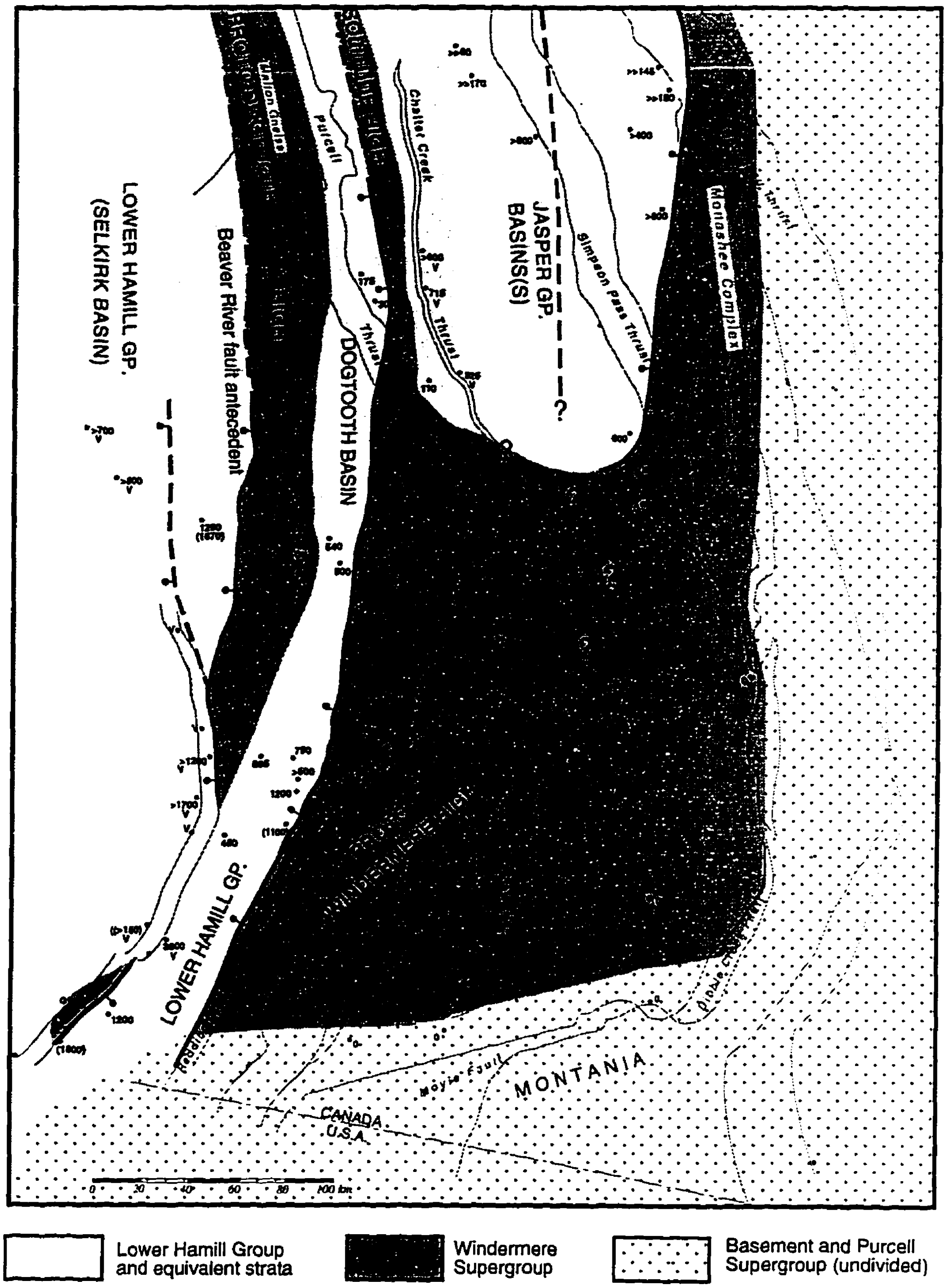



**Figure 3-9:** Paleogeographic interpretation of palinspastically restored thickness data (Figure 3-8) for the lower part of the Hamill Group and equivalent strata (see Figure 3-2 for regional correlations). Palinspastic map was constructed by restoring major thrust faults with respect to a "pin line" at the eastern limit of Cordilleran deformation, using restored cross sections of Price and Mountjoy (1970) and Price and Fermor (1985) for the Rocky Mountains, and restored cross sections and maps of Kubli and Simony (1992, 1995) and Warren (this study, Plate 4) for the Purcell anticlinorium. The Monashee Complex is assumed to be autochthonous. V = location of volcanic sills or flows. Thicknesses in meters. "Solitude High" discussed by Lickorish and Simony (1995). Dashed intrabasinal normal fault to west of "Proto-Dogtooth High" inferred by Devlin (1989).

configurations are constructed from palinspastically restored thickness data from these strata (Fig. 3-8), which include units H1, H2 and H3 of the Hamill Group in the western Purcell anticlinorium and in the Dogtooth Range, unit HA of the Hamill Group in the Kootenay Arc, the Jasper Formation/lower McNaughton Formation in the Rocky Mountains and the Three Sisters Formation and perhaps some of the Quartzite Range Formation in the southern Selkirk Mountains. These are regionally discontinuous strata whose deposition was controlled by a series of roughly north-trending, west-dipping syn-sedimentary normal faults (Lickorish and Simony, 1995; Kubli and Simony, 1992; and this study). These strata record deposition in shallow marine and fluvial basins that were superposed during Neoproterozoic to Early Cambrian regional crustal extension on deeper-water deposits of the older Windermere Supergroup.

The western basin (the "Selkirk basin") shown in Figure 3-9 is represented by the lower part of the Hamill Group in the northern Selkirk Mountains and in the central Kootenay Arc. The fault that bounded the eastern edge of the basin coincides closely with the Beaver River fault to the north, and with the Kootenay Arc boundary fault to the south. During the intitial stage of basin development, abundant sediment was transported and deposited on an alluvial braidplain (Devlin, 1986) that was transitional northward into a shallow marine setting (Devlin and Bond, 1989). Sediment was supplied primarily from the south but also was supplied locally from the bounding fault scarp or scarps to the east (Devlin and Bond, 1989). Subsequent rapid subsidence and deeper marine sedimentation was accompanied by mafic to intermediate volcanism that was more voluminous to the north than to the south (Wheeler, 1963; Devlin and Bond, 1989; Logan and Colpron, 1995; Höy, 1980; and this study). A submerged fault in the northern part of the basin was the locus of volcanism (Devlin, 1986) and may have merged southward with the fault that marked the eastern edge of the basin.

The "Selkirk basin" was separated from a central basin (the "Dogtooth basin") by an eastward-tilted block of Windermere Supergroup strata whose western edge was emergent and bevelled by erosion prior to deposition of the upper unit of the Hamill Group. The central part of this tilted block is the "Dogtooth high" of Kubli and Simony (1992), which records subsequent post-Hamill as well as syn-Hamill tilting and faulting. The southern continuation of this block during lower Hamill time is defined by stratigraphic relationships beneath the upper quartzarenite unit of the Hamill Group along the western flank of the Purcell anticlinorium (this study) and from relationships mapped by Leclair (1988) in the immediate hangingwall of the Kootenay Arc boundary fault (Seeman Creek fault). To the east of this emergent block, strata of western exposures of the Jasper Formation, of the lower units of the Hamill Group, of the Three Sisters Formation and of the Quartzite Range Formation thicken and become coarser eastward and are truncated to the east by pre-Mesozoic normal faults (Rice, 1941; Little, 1960; Kubli and Simony, 1992; Lickorish and Simony, 1995; and this study), recording sedimentation in the fault-bounded half-graben that developed above this tilted block.

Some of the coarsest sediment in the "Dogtooth basin" was derived locally from fault scarps that cut the Windermere and Purcell Supergroups and perhaps the lower Three Sisters Formation, but most of the sediment was derived from a more distant source. Locally, clasts apparently derived from mature siliciclastic and carbonate rocks of the upper part of the Purcell Supergroup were contributed from the fault scarp at the eastern edge of the basin (this study). The sedimentary fill in this basin also thickens toward the south, and conglomerates well above the base of the Three Sisters Formation coarsen markedly to the south and east, suggesting uplift of a source area to the southeast (Little, 1960). Clasts in these conglomerates appear to be locally derived from underlying upper Purcell Supergroup, Windermere Supergroup and basal Three Sisters Formation strata (Little, 1960). However, the pebbles and cobbles of white vein quartz, blue and purple quartz and feldspar that are abundant and widespread throughout the lower Hamill Group indicate a continental basement source that must have been exposed to the south, east or north of the thick succession of Belt-Purcell Supergroup strata. Some of this detritus could have been derived from grits of the immediately underlying Windermere Supergroup, but pebbles and cobbles in the lower Hamill Group are generally larger yet more well rounded than clasts in the underlying Windermere Supergroup. Sediment that was derived from the craton must have been transported northeastward across the Belt-Purcell Supergroup, or it must have entered the basin from the east or north following transport across the underlying Windermere Supergroup. Paleocurrent data imply northerly transport of sediment in the southern part of the basin, and westerly transport of sediment in the northern part (Mountjoy and Aitken, 1963; Ellison, 1967; Devlin, 1986; this study). In either case, a significant transport distance from the primary source area to the basin is implied, perhaps over a low-gradient, alluvial braidplain that is no longer preserved.

The initial phase of basin development involved upward shallowing from the turbiditic environment of the upper pelite unit of the Windermere Supergroup, and increased clastic input. Locally, the drop in relative sea level and the eastward tilting of the underlying Windermere Supergroup may have resulted in a disconformable contact beneath the Hamill Group. A shallow marine environment persisted in this basin, although sediment at the northern and southern limits of exposure may have been deposited in a fluvial setting (Devlin, 1986; Devlin and Bond, 1988; Lickorish and Simony, 1995). The strata of the "Dogtooth basin" do not record an episode of volcanism and of rapid subsidence like that recorded by the strata of the"Selkirk basin" to the west, but instead record a gradual waning of clastic influx due either to a slow rise in relative sea level or to erosional retreat of the source area.

A wide northeastern basin or series of basins (the "Jasper basin") was separated from the "Dogtooth basin" by a narrow horst or eastward-tilted block (the "Solitude high" of Lickorish, 1992). The southern continuation of this block was the antecedent to the Early Paleozoic "Windermere high." Because of gaps in thickness data between overlapping north-striking thrust sheets (Fig. 3-8), it cannot be determined whether the "Jasper basin" was a single continuous basin or whether it comprised a series of north-trending emergent blocks and intervening basins. Its southern boundary is not constrained. The Jasper Formation to the north of the area shown in Figures 3-1 and 3-9 varies abruptly and significantly in thickness and is perhaps locally discontinuous (Slind and Perkins, 1966; Young, 1979). These relationships suggest that a fault-bounded "Jasper basin" or basins probably continued to the north.

The "Jasper basin" comprises intercalated alluvial braidplain and shallow marine deposits (Hein, 1982, 1987; Lickorish and Simony, 1995). These strata record abrupt upward shallowing from deeper marine deposits of the Windermere Supergroup along the southeast margin of the basin (Hein, 1982, 1987). They also record the establishment of these shallow clastic environments on a subaerially exposed Windermere carbonate platforms in the northern part of the basin (Mountjoy, 1962; Aitken, 1969; Teitz and Mountjoy, 1985, 1989). The "Jasper basin" records less subsidence relative to sedimentation rate than the "Dogtooth" or "Selkirk" basins to the west. Sediment in the basin was transported in longitudinal drainage systems that were parallel to the basin margins (Lickorish and Simony, 1995). The sediment composition indicates a cratonic basement source similar to that of the "Dogtooth basin." Cratonic basement may have been exposed to the north and east of the "Jasper basin" as well as to the south and east of the Purcell Supergroup.

The palinspastic restoration of these basins (Fig. 3-9) suggests that there was little tectonic uplift, topographic relief and tectonic subsidence across this portion of the continental margin during regional extension and normal faulting. There is a slight increase in basin subsidence and associated volcanic activity across the margin from east to west, but the strata deposited in these basins rarely exceed 2000 m in thickness, except perhaps immediately south of Jasper, where the entire undivided McNaughton Formation is as much as 3500 m thick (Corneil, 1967; Mountjoy and Price, 1988), and in the central Kootenay Arc, where the lower part of the Hamill Group is as much as 3600 m thick (Höy, 1980). Most of the sediment in the basins must have been derived from a cratonic basement source that was exposed to the east of the Windermere and Purcell Supergroups, rather than from uplifted and eroded local sources. The relative sea level fall that accompanied the onset of normal faulting could indicate a slight, uniform regional tectonic uplift, or it could be eustatic. The palinspastic restoration thus supports the conclusion that there was relatively little attenuation of at least the upper, brittle part of the continental crust on the southern Canadian Cordilleran continental margin during Neoproterozoic to Early Cambrian extension. In contrast, the underlying Neoproterozoic Windermere Supergroup records more significant upper crustal attenuation, local uplift and basin subsidence across the same part of the Cordilleran rifted margin.

## REGIONAL PALEOGEOGRAPHY AND TECTONIC SETTING DURING DEPOSITION OF THE UPPER HAMILL GROUP

An extensive sheet of Lower Cambrian quartzarenite was deposited in a tidally influenced shelf setting across the older fault-bounded basins after regional E-W extension and normal faulting had ceased. A palinspastically restored isopach map (Fig. 3-11) constructed from regional thickness data from this unit (Fig. 3-10) shows its widespread distribution and its gradual westward thickening from the craton. The zero-edge of the quartzarenite unit is roughly N-S north of latitude 50° N, but it swings sharply westward between latitude 50° N and the International Border. The northwestward increase in pelite is consistent with more distal sedimentation to the northwest. The upward increase in pelite in these strata implies gradual flooding of the shelf and/or subsidence that outpaced the supply of sand to the shelf. Above the sheet of quartzarenite, the Lower Cambrian stratigraphic record shows that coarse clastic input to the shelf waned, except locally, adjacent to the craton (southern exposures of the upper Gog Group). The shallow sandy clastic depositional environments were succeeded by the shallow muddy and carbonate depositional environments of the Donald, Mohican, Badshot and Mural formations. Farther to the west, there was an abrupt upward transition from the shallow depostional environments of the Badshot Formation to the deeper fine clastic and carbonate depositional environment of the lower Index Formation. These relationships are consistent with a transition from an actively extending continental margin to a thermally subsiding passive margin following continental separation and the initiation of seafloor spreading (e.g. McKenzie, 1978; Beaumont et al., 1982). The passive margin is inferred to have developed on continental crust that was progressively thinner, and thus showed progressively greater subsidence, toward the west.

However, normal faulting, uplift , tilting and minor mafic volcanism continued during Early Cambrian time on the "Dogtooth high" (Kubli and Simony, 1992) and intermittently throughout

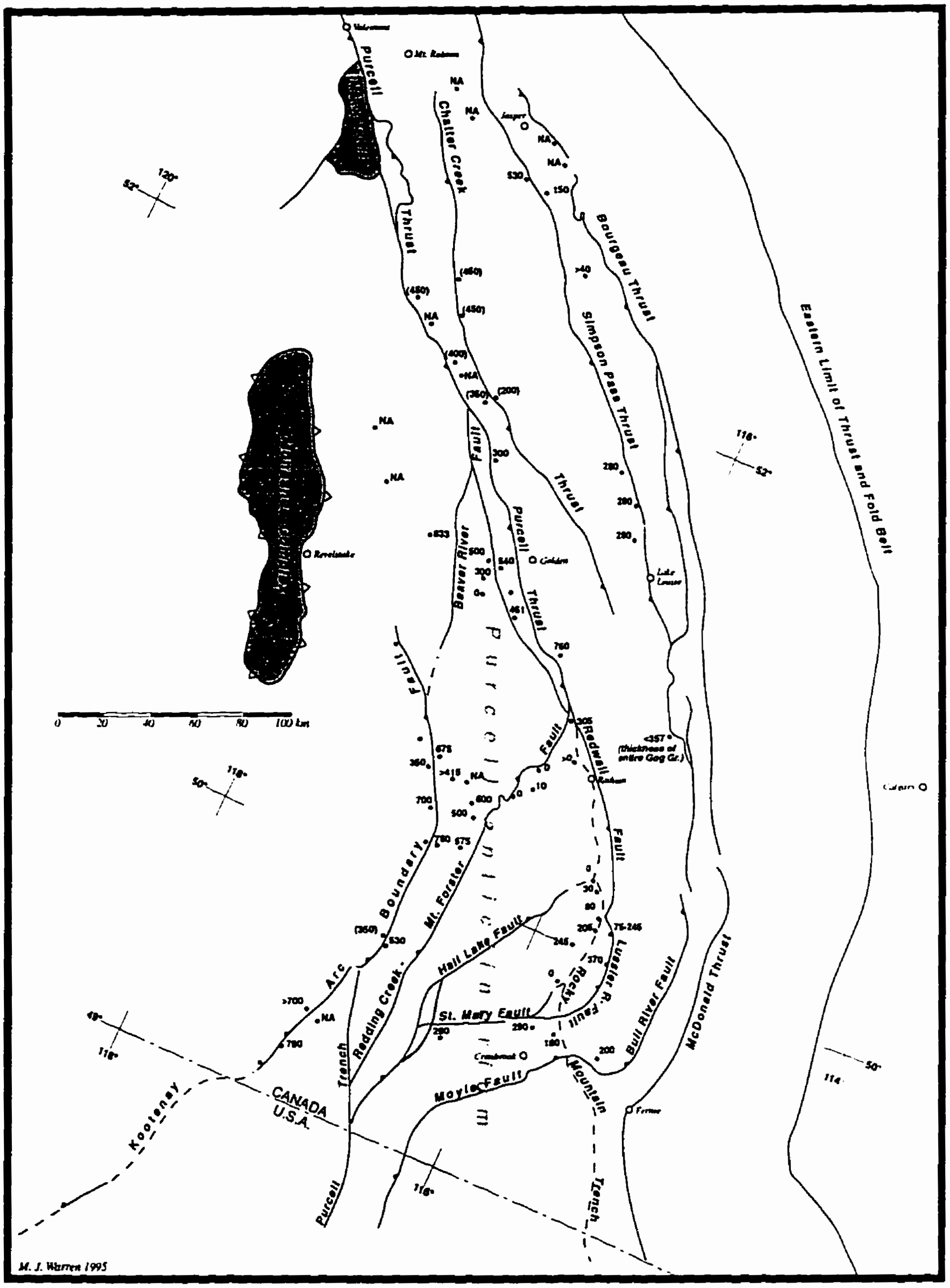



**Figure 3-10:** Thickness data for the upper part of the Hamill Group and equivalent strata (see Figure 3-2 for regional correlations) in the southeastern Canadian Cordillera, used to construct the palinspastically restored paleogeographic map for upper Hamill/Gog time (Figure 3-11). Thicknesses in meters. **NA** = data not available. Values in ( ) may be unrelaible due to intense deformation or uncertainty in correlations. Data compiled from Devlin (1986), Evans (1933), Fritz et al. (1991), Høy (1980), Kubli and Simony (1992), Leclair (1988), Lickorish and Simony (1995), Little (1960), Rice (1937, 1941), Root (1987), Simony and Wind (1970) and Warren (this study). Note that north is oblique to edge of figure.

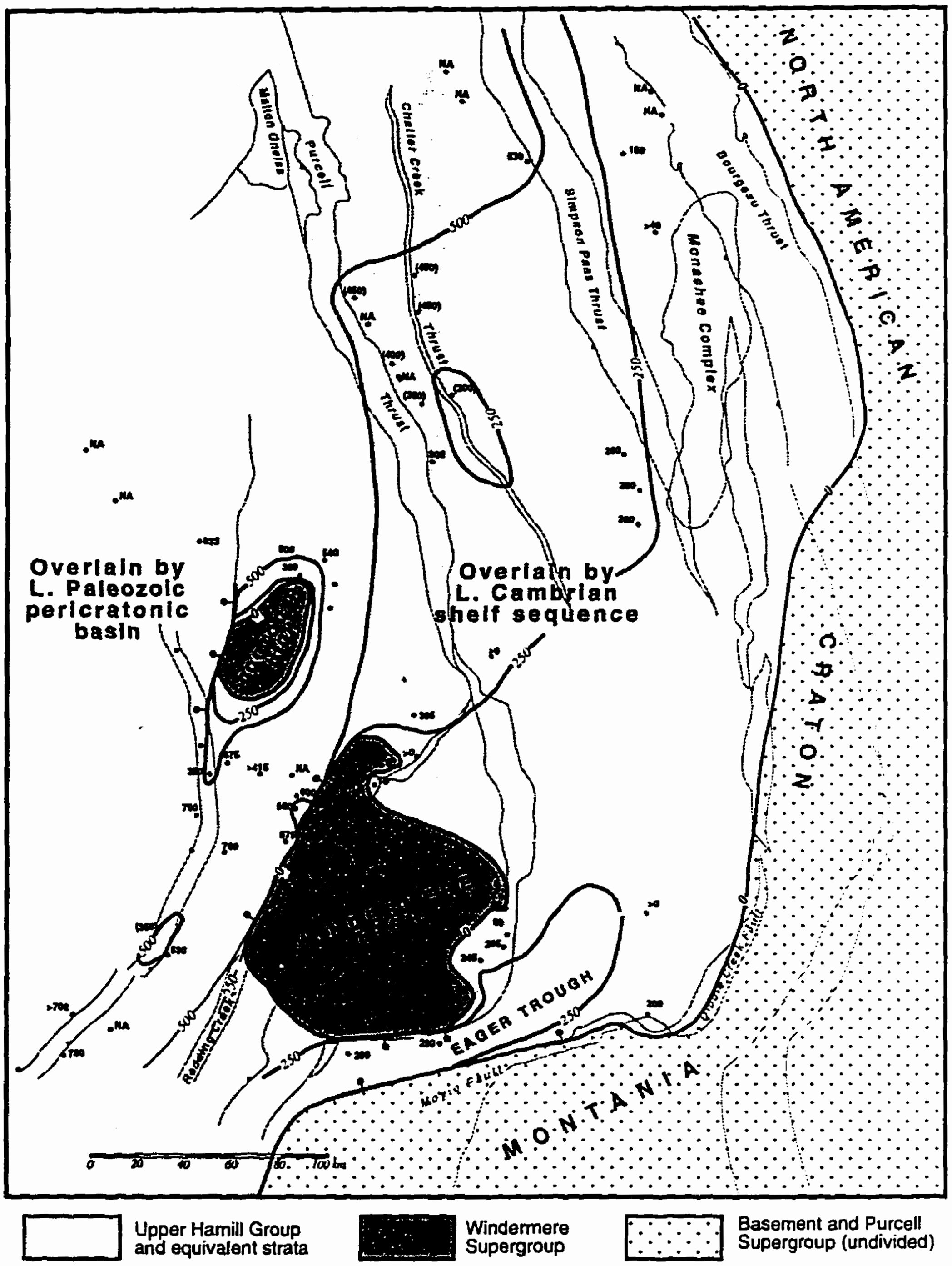



**Figure 3-11:** Isopach map and paleogeographic interpretation of palinspastically restored thickness data (Figure 3-10) for the upper part of the Hamill Group and equivalent strata (see Figure 3-2 for regional correlations). Palinspastic map was constructed by restoring major thrust faults with respect to a "pin line" at the eastern limit of Cordilleran deformation, using restored cross sections of Price and Mountjoy (1970) and Price and Fermor (1985) for the Rocky Mountains, and restored cross sections and maps of Kubli and Simony (1992, 1995) and Warren (this study, Plate 4) for the Purcell anticlinorium. The Monashee Complex is assumed to be autochthonous. Thicknesses in meters. "Dogtooth High" first discussed by Kubli and Simony (1992). "Windermere High" first discussed by Reesor (1973). "Montania" discussed by Deiss (1941).

Early Paleozoic time on the "Windermere high" (Reesor, 1973; Price, 1981; Root, 1987; Pope, 1989). Intermittent uplift also occurred on "Montania" during this time (Deiss, 1940, 1941). Furthermore, the Lower Paleozoic Lardeau Group, above the lower Index Fomration, includes abundant coarse, immature clastic sedimentary rocks and mafic volcanic and volcaniclastic rocks that were deposited in a deep basin to the west of these crustal "highs."

Several relationships suggest that the Lower Cambrian quartzarenite sheet was initially continuous and that the local crustal "highs" did not become active until a tectonically quiescent and uniformly subsiding passive margin had been briefly established. The presence of locally abundant orthoquartzite blocks at the base of the Index Formation to the west of the "Dogtooth high" (Brown, 1991; Logan et al., 1996) suggests that the Lower Cambrian quartzarenite sheet was first deposited across the "Dogtooth high" and subsequently eroded during reactivation of the normal fault on the western side of the uplifted block. The "Eager trough," which lies between the "Windermere high" and "Montania," contains a thick succession of upper Lower Cambrian shale that is distinct from and up to three times as thick as the Lower Cambrian strata that overlie the Hamill Group in the Dogtooth Range and southwestern Purcell anticlinorium. The Eager shale was deposited on a deeply-subsided block following the deposition of the shallow marine Cranbrook Formation. These relationships suggest that the "Windermere high," perhaps "Montania" and the down-dropped block between them were tectonically quiescent during deposition of the Lower Cambrian quartzarenite sheet, but that they became active during the latter part of the Early Cambrian. Magnesite and dolostone intervals near the top of the Cranbrook Formation near Cranbrook (Rice, 1937, 1941; Höy, 1993) could indicate that the upper part of the Cranbrook Formation was deposited in a restricted environment that was partly shut off from the shelf to the north and west. This would imply that the "Windermere high" had begun to emerge during the deposition of the upper part of the Cranbrook Formation. Fine-grained, deep-water siliciclastic and carbonate rocks of the lower Index Formation also indicate that there was a brief hiatus between initial subsidence of the outer shelf and deposition of volcanic and coarse clastic rocks of the rest of the Lardeau Group.

The tectonic setting of these uplifted blocks is enigmatic. Their intermittent Early Paleozoic uplift has been ascribed to regional extension associated with episodic rifting of several continental fragments from the western margin of North America during the Early Paleozoic (Struik, 1988). However, the youngest (Middle Devonian) episode of uplift on the "Windermere high" also has been interpreted as the result of crustal flexure caused in part by loading by thrust sheets from the west during Early Paleozoic orogeny (Root, 1987). The "highs" are located in a reentrant in the Cordilleran miogeocline and their limits are defined both by reactivated NE-trending normal faults of Windermere age (Chapter 2) and by N-trending normal faults of lower Hamill age (Fig. 3-9). I suggest that the crustal "highs" were initially defined in Early Cambrian time, during initially rapid, regional, thermally-driven subsidence of the newly-formed passive margin, and that they may be the result of local differential subsidence between thicker fault-bounded blocks of the underlying basement, defined during syn-Windermere and syn-lower Hamill extension, and more attenuated surrounding continental crust. The "highs" subsequently remained relatively buoyant and were the locus of fault reactivation, uplift and erosion during several periods of poorly understood tectonic activity and eustatic sea level fluctuations along the Cordilleran margin in Late Cambrian to Devonian time.

## LOCATION, TIMING AND PLATE TECTONIC SIGNIFICANCE OF THE RIFT-TO-DRIFT TRANSITION

The stratigraphic record of a rifted continental margin should contain a "break-up" unconformity that records the transition from active extension of the continental crust to continental separation and the initation of seafloor spreading (Falvey, 1974). Several models of passive margin development equate the "break-up" unconformity and the beginning of a continental "drift" phase with the initiation of thermal contraction of the cooling lithosphere (McKenzie, 1978; Beaumont et al., 1982). Thermal contraction, coupled with sediment loading, results in regional tectonic subsidence that decreases exponentially with time. Construction of tectonic subsidence curves and analysis of rates of subsidence of Lower Paleozoic strata of the southern Canadian Cordilleran miogeocline (Bond and Kominz, 1984) implies that thermal subsidence must have been initiated in latest Proterozoic to Early Cambrian time. However, extrapolation of the exponential subsidence curves backward in time does not constrain precisely the timing of breakup.

Embry and Dixon (1990) described three criteria for recognizing the "break-up" unconformity in the stratigraphic record. The most important of these is that the

> "unconformity and its correlative conformity separate strata which were deposited mainly in actively developing half grabens from strata that were deposited over a broad area. Thus the pre-breakup strata are cut by major normal faults that extend into the basement, whereas the post-rift strata are unfaulted or have undergone substantially less faulting (due to minor reactivation of rift faults and differential subsidence over fault scarps)."

The other criteria include volcanic strata that occur mainly beneath the unconformity, and commonly greater subsidence beneath the unconformity than above it.

The regional unconformity in the Hamill and Gog Groups is thus interpreted as a "break-up" unconformity in the Cordilleran stratigraphic record, corresponding to an Early Cambrian rift-to-drift transition. The Neoproterozoic(?) to lower Lower Cambrian (Placentian) strata (Fig. 3-2) that lie below the unconformity were deposited in primarily shallow, N-trending, half-grabens that were separated by uplifted, bevelled and primarily eastward-tilted blocks of Windermere Supergroup strata (Figs. 3-7 and 3-9). Alkalic mafic and minor intermediate volcanism accompanied this episode of regional extension. Upper Lower Cambrian (Waucoban) and perhaps some Placentian quartzarenite and pelite that lie above the unconformity were deposited as a more continuous, regional sheet across the basins and the uplifted blocks in a shallow shelf depositional setting after extension and normal faulting had ceased (Fig. 3-11). Sedimentation may have been continuous in the down-dropped parts of the basins. The initial emergence of the "Dogtooth" and "Windermere" highs may record differential subsidence across pre-existing structures. The local unconformity *beneath* the Hamill and Gog Groups, although lithologically significant, does not match the criteria for a "break-up" unconformity, but rather is interpreted to record the initiation of the pulse of regional extension, faulting and perhaps regional uplift that immediately preceeded seafloor spreading.

Regional fossil data from the Hamill and Gog Groups (Fig. 3-2) show that the strata immediately above the "break-up" unconformity are primarily upper Lower Cambrian (Waucoban) in age, and fall primarily in the *Nevadella* trilobite zone which comprises the lower part of the Waucoban. The overlying Badshot, Donald and Mural Formations, strata that record initial subsidence on the newly-formed continental shelf, contain fauna that lie within the*Nevadella* and *Bonnia-Olenellus* (upper Waucoban) trilobite zones. The parts of the Hamill and Gog Groups that lie beneath the unconformity, as well as the Three Sisters Formation, are unfossiliferous, except in the Dogtooth Range, where strata immediately beneath the upper quartzarenite contain trace fossils consistent with a lower Lower Cambrian (Placentian) age (Kubli and Simony, 1992). However, Kubli and Simony (1992) did not identify an unconformity in this succession, and thus the position of the fossils and the strata that contain them with respect to the inferred unconformity is unconfirmed (see Fig. 3-2). Thus, the regional unconformity is no older than 549 Ma, might be no older that 543 Ma (if the lowest Cambrian trace fossils in the Dogtooth Range are beneath it) and is no younger than 520 Ma. These dates correspond to the absolute ages assigned to the base of the uppermost assemblage of Ediacara fauna (see Chapter 2, pages 69-70), the Precambrian/Cambrian boundary and the Lower/Middle Cambrian boundary, respectively (Grotzinger et al., 1995, and references therein). This age is entirely consistent with the Neoproterozoic to Early Cambrian age for continental breakup and initiation of thermal subsidence proposed by Bond and Kominz (1984) from extrapolation of tectonic subsidence curves.

If full continental separation between Laurentia and Australia (e.g. Hoffman, 1991; Dalziel, 1992) had already occurred by the end of Windermere (Neoproterozoic) time, as paleomagnetic data from both cratons (e.g. Powell et al., 1993, 1994) and sedimentological/stratigraphic data from the northern Canadian Cordillera (Dalrymple and Narbonne, 1996) suggest, then the Lower Cambrian "break-up" unconformity must record separation from Laurentia of an intervening crustal fragment or fragments. However, no such crustal blocks have been identified. Alternatively, intracontinental rifting began in Windermere time and was a protracted and/or diachronous process, similar to the opening of the modern Atlantic Ocean, so that final continental separation did not take place in at least the southern Canadian Cordillera until Early Cambrian time. In either case, the lack of significant subsidence and topographic uplift of the source displayed by the sedimentary fill of the Hamill/Gog rift basins relative to that of the underlying Windermere basin suggests that the underlying upper crust already had been significantly thinned at the onset of the Hamill/Gog rifting, or that the process or geometry of crustal extension was fundamentally different during the two episodes. Future discussions of their plate tectonic settings and evolution should address these fundamental differences in the two "rifting" episodes.

## SUMMARY: LATEST NEOPROTEROZOIC TO EARLY CAMBRIAN TECTONIC EVOLUTION OF THE RIFTED MARGIN

Toward the end of Windermere time, which Ediacaran fauna indicate to be latest Neoproterozoic time (Hofmann et al., 1985; Narbonne et al., 1994) a carbonate bank, slope and basin were established off the western edge of ancestral North America, during a relative highstand (see Chapter 2). The tectonic setting was either a passive margin that had been established during a previous cycle of continental rifting and separation during lower Windermere time (Ross, 1991b; Dalrymple and Narbonne, 1996) or a time of tectonic quiescence during a protracted, multi-stage intracontinental rifting episode that involved significant upper crustal stretching and rift basin subsidence. By the end of Windermere time, a relative sea-level fall, perhaps related to broad, regional tectonic uplift, had resulted in progradation of clastic sediments toward the basin and a shallowing-upward succession in the sedimentary record (Hein, 1987, and this study), and subaerial exposure and karsting of some of the carbonate platform to the northeast (e.g. Mountjoy, 1962; Teitz and Mountjoy, 1985). The relative sea-level fall was in part accompanied by or closely followed by the onset of crustal extension, normal faulting and synchronous deposition of fluvial and shallow marine clastic and perhaps evaporitic sediments in narrow, north-trending, fault-controlled basins (Figs. 3-7 and 3-9). At this time, the lower Hamill and Gog Groups and Three Sisters Formation were deposited unconformably over the previously exposed upper Windermere carbonate platform and conformably over the down-dropped parts of the former slope/basin, and they onlapped onto the uplifted portions of the tilted fault blocks. The basins were filled primarily by sediment that was derived from the craton and was transported by longitudinal drainage systems (Lickorish and Simony, 1995), and also by less abundant sediment that was locally derived from uplifted blocks of Windermere or Purcell Supergroup strata. Local volcanism accompanied this episode of crustal extension.

In early Cambrian time, before the end of of the *Nevadella* trilobite Zone, extension and normal faulting ceased, and a wide, thermally-subsiding continental shelf was established on the newly-formed passive continental margin. A gently westward-thickening sheet of primarily quartz sand and overlying sand and mud was deposited unconformably across the uplifted and bevelled, fault-bounded blocks and perhaps conformably across the down-dropped portions of the basins (Figs. 3-7 and 3-11). However, before the end of Early Cambrian (*Bonnia-Olenellus*) time, regional subsidence and sedimentation on the margin were locally disrupted by reactivation of both NE-trending normal faults that had been active during Windermere time and N-trending normal faults that were active during lower Hamill/Gog time.

## CONCLUSIONS

A regional unconformity in the Hamill and Gog Groups of the southern Canadian Cordillera is the stratigraphic expression of the rift-to-drift transition that occurred in Early Cambrian time between 549 Ma and before about 520 Ma. It marks the change from regional E-W extension of continental crust to seafloor spreading and thermal subsidence of an underlying passive continental margin. The Neoproterozoic(?) to lower Lower Cambrian (Placentian) strata that lie below the unconformity were deposited in primarily shallow, N-trending, fault-bounded basins that were separated by uplifted, bevelled and primarily eastward-tilted blocks of Windermere Supergroup strata. Upper Lower Cambrian (Waucoban) quartzarenite and quartzite that lie above the unconformity were deposited as a more continuous sheet across the basins and the uplifted blocks in a shallow shelf setting after extension and normal faulting had ceased.

The emergence of several paleogeographic "highs" during Lower Paleozoic time is interpreted to record, at least initially, differential subsidence of several fault-bounded blocks of different crustal thickness, and thus of different buoyancy. These features were defined by structures that were inherited both from older northeast-trending syn-Windermere and north-trending syn-lower Hamill/Gog normal faults. These normal faults were reactivated during several enigmatic episodes of Early Paleozoic normal faulting as well as during Mesozoic thrust faulting.

The episode of crustal extension that immediately preceded the rift-to-drift transition involved relatively insignificant stretching of the upper crust, in contrast to significant upper crustal stretching, topographic relief and deep rift basin subsidence that occurred during syn-Windermere extension.

# Chapter 4

# TECTONIC EVOLUTION OF THE TRANSITION FROM THE KOOTENAY ARC TO THE FORELAND THRUST AND FOLD BELT OF THE SOUTHERN CANADIAN CORDILLERA

## ABSTRACT

Neoproterozoic and Early Paleozoic extension faults, related changes in sedimentary thicknesses and lithofacies, and changes in mechanical and thermal properties of underlying lithosphere controlled styles of Mesozoic crustal thickening along the outboard Cordilleran margin.

A zone of structural divergence extending from Alaska to Washington lies at the western edge of the Cordilleran thrust and fold belt, and immediately to the east of the Jurassic "suture" between North America and accreted terranes. In southeastern British Columbia, this zone is marked by sharp structural and stratigraphic contrasts between the Kootenay Arc (an arcuate belt of poly-phase, mainly west-verging ductile deformation) and the Purcell anticlinorium (a broad belt of east-verging, thinned-skinned deformation). The Kootenay Arc is characterized by an early phase of high-amplitude, predominantly west-verging folds, which are coaxially refolded by steep folds with an amphibolite-facies regional foliation. The younger steep folds extend eastward into the Purcell anticlinorium, but the older west-verging structures do not. The boundary between the Kootenay Arc and the Purcell anticlinorium is a regionally significant, locally mylonitic fault zone (the Kootenay Arc boundary fault), congruent with the steep folds and foliation. This fault zone extends for a strike length of at least 200 km. Stratigraphic and structural relationships imply west-side-up thrust displacement; however, stratigraphic throw across the fault (<1 km) and pressure

differences (< 1 kbar) are apparently incompatible with the marked contrast in structural style across this fault.

A restored regional cross section (based on new and published stratigraphic, structural, geobarometric and geochronologic data) clarifies the Mesozoic tectonic evolution of the transition from the metamorphic "hinterland" (Kootenay Arc) to the foreland thrust and fold belt. Compressional collapse of the continental margin began with obduction of Slide Mountain and Quesnel terranes eastward over distal North American strata, prior to 187-178 Ma. By 173 +/-5 Ma, west-verging folding and thrusts were underway within Paleozoic North American supracrustal rocks, as a tectonic "wedge" of Proterozoic North American strata was driven eastward at depth, between these strata and their basement. The wedge subsequently impinged upon a pre-existing ramp within the Neoproterozoic to Early Paleozoic sedimentary basin. Horizontal shortening above the wedge continued with upright refolding of the wedge and overlying strata, and was accompanied by rapid crustal thickening followed by post-metamorphic exhumation of up to 30 km. Both phases of Kootenay Arc structures were juxtaposed against Purcell anticlinorium structures prior to 164 Ma, perhaps as early as 170 Ma, by a steep splay fault from the tip of the wedge (the Kootenay Arc boundary fault). This high strain zone primarily reflects flattening of the vertically thickened Kootenay Arc against the older normal fault ramp, rather than significant horizontal or vertical displacement. Rapid regional exhumation of up to 10 km also occured prior to 164 - 157 Ma. Steeply west-dipping late- to post-metamorphic normal faults with significant displacement in the Kootenay Arc may reflect tectonic denudation that accompanied rapid crustal thickening.

Vertical thickening provided gravitational potential which helped to drive deformation eastward over the older normal fault ramp. The Cordilleran basal detachment and east-verging splay faults subsequently propagated eastward through cooler, stiffer Proterozoic and Paleozoic metasedimentary rocks of the Purcell anticlinorium, during continued compression, retrograde metamorphism and much less rapid exhumation. Out-of-sequence thrusts may have developed in response to eastward motion over a second rift-related crustal ramp. The Kootenay Arc and the

Purcell anticlinorium were subsequently carried passively during Late Cretaceous to Tertiary time as the basal detachment continued to propagate beneath the Paleozoic rocks of the rest of the foreland fold and thrust belt.

Thin, weak, perhaps hot lithosphere west of the ramp responded to crustal thickening by deep, localized subsidence with no foreland basin. Tectonic loading of progressively thicker, stiffer lithosphere to the east is recorded by the narrow Kootenay Group (ca. 155-135 Ma) and much broader Blairmore Group (ca. 115-95 Ma) and Laramide (ca. 90-60 Ma) foreland basins.

## INTRODUCTION

A conspicuous zone of structural divergence occurs near the boundary between North American strata and accreted terranes along the length of the Canadian Cordillera from east-central Alaska to northeastern Washington (Fig. 4-1: Wheeler, 1970; Price, 1986; Eisbacher et al., 1974). This zone of structural divergence separates west-verging faults and high-amplitude, refolded, west-verging folds to the west from upright and east-verging structures to the east. The nature, age(s) and origin(s) of the west-verging structures and their relationship to the thin-skinned, east-verging structures of the Rocky Mountain thrust and fold belt are controversial (e.g. Wheeler, 1970; Price, 1981, 1986; Archibald et al., 1983; Brown et al, 1986).

A segment of this zone of structural divergence, near Duncan Lake, B. C., presents an opportunity both to clarify the structural transition from the metamorphic "hinterland" to the foreland fold and thrust belt of the southern Canadian Cordillera and to evaluate the influence of multi-stage Neoproterozoic and Early Paleozoic crustal attenuation on Mesozoic terrane convergence and crustal thickening. The Kootenay Arc and the Purcell anticlinorium are two distinct structural domains in southeasten British Columbia (Figs. 4-1, 4-2); both expose metasedimentary rocks of North American affinity that were deformed, intruded and regionally metamorphosed during Mesozoic terrane accretion (Monger and Price, 1979). The sharp boundary between these two domains coincides with a segment of the regional zone of structural

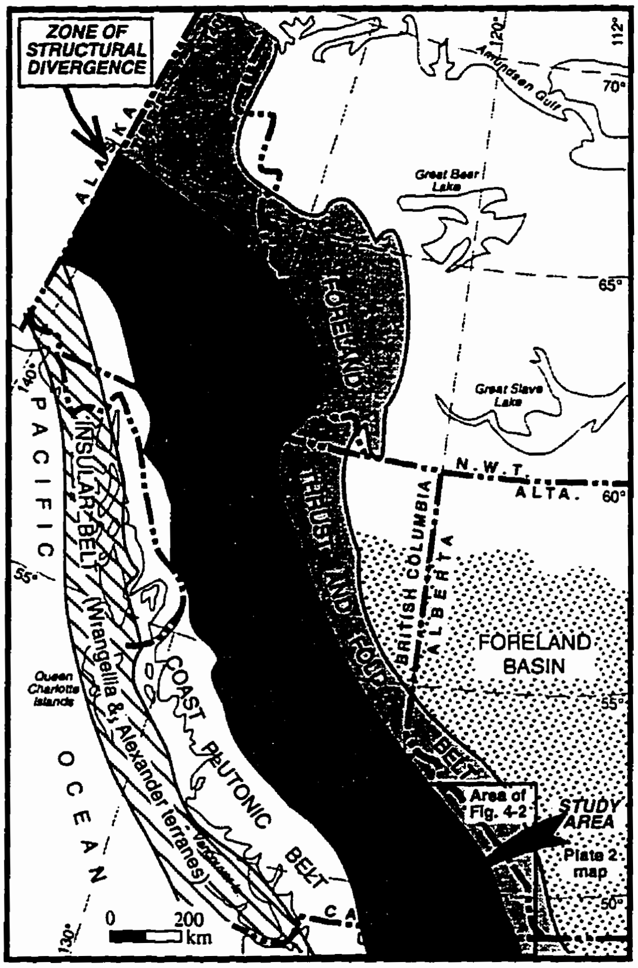



**Figure 4-1:** Morphogeological belts of the Canadian Cordillera (after Wheeler and McFeeley, 1991), showing the location of the zone of structural divergence between east-verging, thin-skinned deformation and west-verging, poly-phase ductile deformation. The Omineca belt comprises rocks of both North America and exotic terranes that were regionally metamorphosed during Mesozoic terrane accretion.

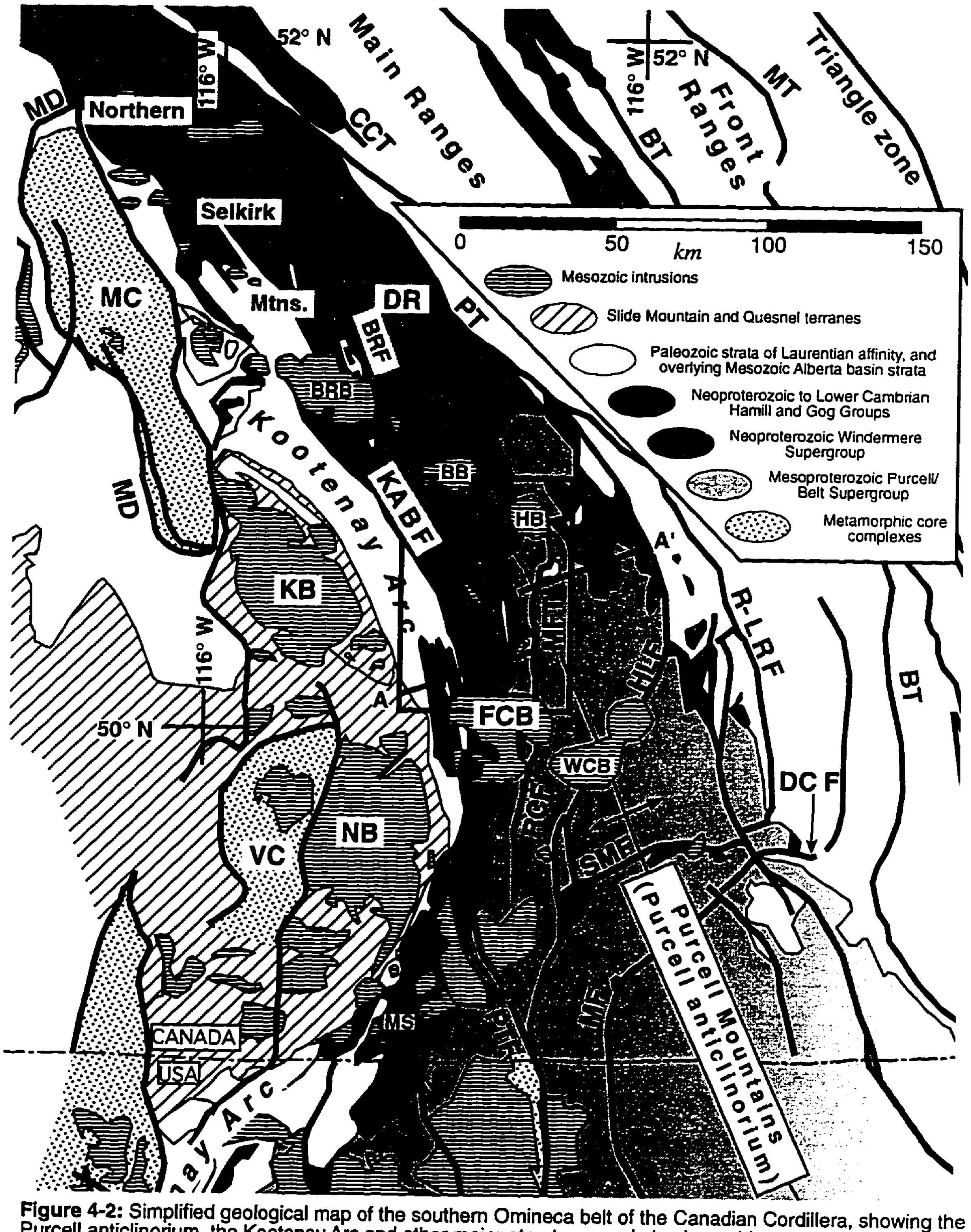



**Figure 4-2:** Simplified geological map of the southern Omineca belt of the Canadian Cordillera, showing the Purcell anticlinorium, the Kootenay Arc and other major structures and physiographic provinces discussed in the text. After Wheeler and McFeeley (1991). Abbreviations: **MT** = McConnell thrust; **BT** = Bourgeau thrust; **CCT** = Chatter Creek thrust; **PT** = Purcell thrust; **R-LRF** = Redwall-Lussier River fault; **DCF** = Dibble Creek fault; **MF** = Moyie fault; **SMF** = St. Mary fault; **HLF** = Hall Lake fault; **RCF** = Redding Creek fault; **MFT** = Mount Forster thrust; **SRMTF** = Southern Rocky Mountain Trench fault; **PTF** = Purcell Trench fault; **BRF** = Beaver River fault; **KABF** = Kootenay Arc boundary fault; **MD** = Monashee decollement; **VC** = Valhalla Complex; **MC** = Monashee Complex; **DR** = Dogtooth Range; **MS** = Mine stock; **BRB** = Battle Range batholith; **NB** = Nelson batholith; **WCB** = White Creek batholith; **FCB** = Fry Creek batholith; **KB** = Kuskanax batholith; **HB** = Horsethief Creek batholith; **BB** = Bugaboo batholith. Line A - A' is regional cross section of Plate 4a.

divergence. The Kootenay Arc is an arcuate belt of ductile, poly-phase deformation and regional metamorphism that contains the "suture" between Slide Mountain/Quesnel terranes and North American strata (e. g. Klepacki, 1985). The Kootenay Arc is bounded to the east by the Purcell anticlinorium, a broad, north-plunging structure that contains less complexly deformed and less metamorphosed strata of ancestral North America. The eastern flank of the Purcell anticlinorium displays a structural style similar to that of the adjacent thrust and fold belt of the Rocky Mountains (Price, 1981; Root, 1987; Kubli and Simony, 1994).

Previous work raised a fundamental question about the relationship between polyphase folding in the Kootenay Arc and the single dominant phase of folding in the Purcell anticlinorium. Older, high-amplitude, predominantly west-verging folds in the Kootenay Arc are coaxially refolded by steep folds and an amphibolite-facies regional foliation (Fyles, 1964; Plate 4a). The younger steep folds extend eastward into lower-grade rocks of the Purcell anticlinorium, but the older west-verging structures do not (Reesor, 1973), thus defining a puzzling structural contrast between the two domains. A critical gap in information remained between the areas mapped by Fyles (1964) and Reesor (1973) that coincided with the change in structural style. Höy (1980) and Leclair (1988) subsequently demonstrated that, farther south in the Kootenay Arc, steeply-dipping faults separate tight, upright folds that deform overturned strata on the lower limbs of recumbent folds in the Kootenay Arc from tight, upright folds that deform equivalent but upright strata in the western Purcell anticlinorium. However, they were unable to resolve the sense of motion on these faults, or to reconcile the contrast in structural style across them.

Another critical question is the timing of the west-verging folding. Most of the upright penetrative deformation and regional metamorphism in the Kootenay Arc and western Purcell anticlinorium is considered of Middle Jurassic age, on the basis of U-Pb zircon and K-Ar age data from a suite of plutons that are syn-kinematic with respect to the upright deformation (Wanless et al., 1968 [Toby stock]; Archibald et al., 1983; Parrish and Wheeler, 1988 [Kuskanax batholith], Armstrong, 1988 [Nelson batholith]: Fig. 4-2). A younger, appparently post-kinematic suite of plutons is mid- to Late-Cretaceous (e. g. Archibald et al., 1984, and references therein). The older suite cuts fault contacts between Quesnel terrane and North America, and the younger suite cuts "out-of-sequence" oblique faults in the Purcell anticlinorium, thus providing younger limits on the timing of terrane accretion and on the cessation of deformation in the Purcell anticlinorium, respectively. However, the timing of the older, west-verging folds in the Kootenay Arc remained poorly constrained. Some workers have argued that the west-verging folds were mid-Paleozoic, because of their apparent restriction to pre-Mississippian strata and because a Mississippian basal conglomerate (Milford Group) that contains foliated clasts rests unconformably on these strata in the Kootenay Arc (Read, 1973, 1976; Read and Wheeler, 1976; Klepacki and Wheeler, 1985). However, the zone of structural divergence extends to the north into the Selkirk fan structure, whose western flank comprises west-verging structures that are of Middle Jurassic age (e.g. Brown et al., 1992; Murphy, 1987; Colpron et al., 1996), and Fyles (1964) suggested that that the coaxially refolded west-verging folds of the Kootenay Arc represented a continuum of ductile deformation rather than two distinct "phases" of folding.

The tectonic setting and cause of the reversal in vergence are also controversial. Recent work on the western flank of the Purcell anticlinorium (Archibald et al., 1983), as well as in similar strata on the western flank of the Selkirk fan structure to the north (Colpron et al., 1996), has shown that the Middle Jurassic deformation and regional metamorphism were associated with a period of rapid crustal thickening followed by rapid denudation and regional exhumation. Two opposing models explain the west-verging structures and the associated crustal thickening. Brown et al. (1986) argued that the west-verging structures developed in response to east-dipping thrusting that extended eastward into the mantle beneath North America. Price (1986) concluded that the west-verging structures developed above a tectonic wedge of allochthonous rock that was driven eastward between the North American basement and its supracrustal cover, along a west-dipping thrust fault that extended westward into the mantle.

This new study in the west-central Purcell Mountains of British Columbia provides new insight into the nature, timing, evolution and tectonic significance of the abrupt contrast in structural style and metamorphism between the Kootenay Arc and the Purcell anticlinorium. In this chapter, I present evidence from the area east of Duncan Lake, that shows that rocks of the Kootenay Arc are separated from rocks of the Purcell anticlinorium by a steeply-dipping, west-side-up, ductile thrust fault. This structure can be linked to the structures previously mapped to the south by Höy (1980) and by Leclair (1988), which implies that it is a regionally significant fault that separates two distinct structural domains along much of the strike length of the Kootenay Arc. This fault is referred to hereafter as the Kootenay Arc boundary fault. I use field observations, coupled with geobarometric and geochronological data, to argue that the west-verging folds in the Kootenay Arc and much of the upright folds and foliation in both the Kootenay Arc and western Purcell anticlinorium record a continuum of Mesozoic deformation that was kinematically linked to the development of this fault zone. Stratigraphic relationships across the fault indicate that the development of the Kootenay Arc boundary fault and other Mesozoic structures in the Purcell anticlinorium were controlled by Proterozoic to Paleozoic extensional faults that marked part of the transition from normal to tectonically attenuated continental crust on the North American rifted margin. I will argue that the structural style that characterizies the eastern Kootenay Arc is primarily a result of poly-phase horizontal shortening and vertical thickening against a pre-existing, west-facing ramp in the Neoproterozoic to Paleozoic sedimentary basin, and that the Kootenay Arc boundary fault is a steep, high-strain zone that records significant flattening, rather than significant horizontal or vertical displacement.

These data and arguments provide the basis for a new tectonic model for the Mesozoic compressional collapse of the outboard margin of ancestral North America. This model differs from and builds upon previously published tectonic models in illustrating that:

1) the restored width of deformed ancestral North American supracrustal strata (at least 550 km), and thus of their underlying continental basement, is much wider than shown in previous reconstructions;

2) the west-verging deformation at the latitude of northern Kootenay Lake developed during eastward displacement of a tectonic wedge of Proterozoic North American strata that was inserted between the Paleozoic strata of the Kootenay Arc and the underlying North American

basement. This conclusion contrasts with relationships at the latitude of southern Kootenay Lake that indicate that the west-verging deformation developed above a wedge of accreted rocks, and is more compatible with previous models (Price, 1986; Colpron et al., in review) for west-verging deformation along the western flank of the Selkirk fan structure to the north;

3) the eastern limit of the tectonic wedge and of the west-verging deformation above it were controlled by a Neoproterozoic to Early Paleozoic normal fault ramp that was marked by an abrupt eastward increase in the slope of the basal detachment and by an increase in lithospheric thickness.

The model also illustrates the link between this transition from west-verging, ductile deformation to east-verging, thin-skinned deformation in the hinterland with foreland basin subsidence and sedimentation that recorded tectonic loading of progressively thicker, stiffer lithosphere to the east.

## GEOLOGY OF THE WEST-CENTRAL PURCELL MOUNTAINS, EAST OF DUNCAN LAKE

New data presented in this chapter are the result of new 1:50,000 scale mapping and study of a segment of the west-central Purcell Mountains, east of Duncan and northern Kootenay Lakes, British Columbia (Fig. 4-2). This study links previous detailed work of Fyles (1964) and Klepacki (1985) in the Kootenay Arc with that of Root (1987) and Pope (1989, 1990) in the footwall of the Mount Forster thrust in the eastern Purcell anticlinorium, and thus completes a detailed transect from Slide Mountain terrane through the Kootenay Arc and Purcell antclinorium to the western Rocky Mountains (Plate 4a). The study area (Figs. 4-2, Plate 2) includes the western flank of the Purcell anticlinorium, in the hangingwall of the Mount Forster thrust, and the eastern part of the Kootenay Arc, with specific focus on the nature and evolution of the transition between these two structural domains. New and previously published structural, metamorphic and geochronological data are discussed below for each of these two domains.

## STRATIGRAPHIC CONTRASTS BETWEEN THE PURCELL ANTICLINORIUM AND KOOTENAY ARC

Important changes in the stratigraphic successions occur between the eastern and western Purcell anticlinorium, and between the western Purcell anticlinorium and the eastern Kootenay Arc, but a few critical stratigraphic units provide firm correlation between these three domains (Fig. 4-3). In the southern and eastern part of the anticlinorium the Mesoproterozoic Purcell Supergroup and unconformably overlying Neoproterozoic Windermere Supergroup (basal Toby Formation and overlying Horsethief Creek Group) are overlain unconformably by an anomalously thin Lower Paleozoic succession. Several overstepping unconformities (sub-Lower Cambrian to sub-Middle Devonian) within this thin succession are associated with the eastward tilting and emergence of high-standing crustal blocks known as the "Windermere High" and "Montania" during Early Paleozoic time (Deiss, 1941; Norris and Price, 1966; Price, 1972; Reesor, 1973; Root, 1987; and Chapter 3). Lower Cambrian strata are only a few tens of meters thick or are absent altogether in the succession that was deposited on the "Windermere high" (Reesor, 1973; Fig. 4-3).

In contrast, the latest Neoproterozoic to Lower Cambrian stratigraphic succession that is exposed in the western Purcell anticlinorium is up to 2500 m thick. The succession overlies the Dutch Creek and Mount Nelson Formations of the Mesoproterozoic Purcell Supergroup, and it comprises the Neoproterozoic Windermere Supergroup, the Neoproterozoic to Lower Cambrian Hamill Group and the Lower Cambrian Mohican and Badshot Formations (Fig. 4-3). The uppermost unit of the Hamill Group can be correlated with the Cranbrook Formation in the eastern Purcell anticlinorium (Fig. 4-3; see Chapter 3).

The stratigraphic succession exposed in the adjacent segment of the Kootenay Arc comprises a different, but partly correlative, succession of Hamill Group strata that is overlain by the Badshot Formation and the Lower Paleozoic Lardeau Group. The Lardeau Group comprises deeper-water and primarily less mature facies than the Lower Paleozoic succession that is

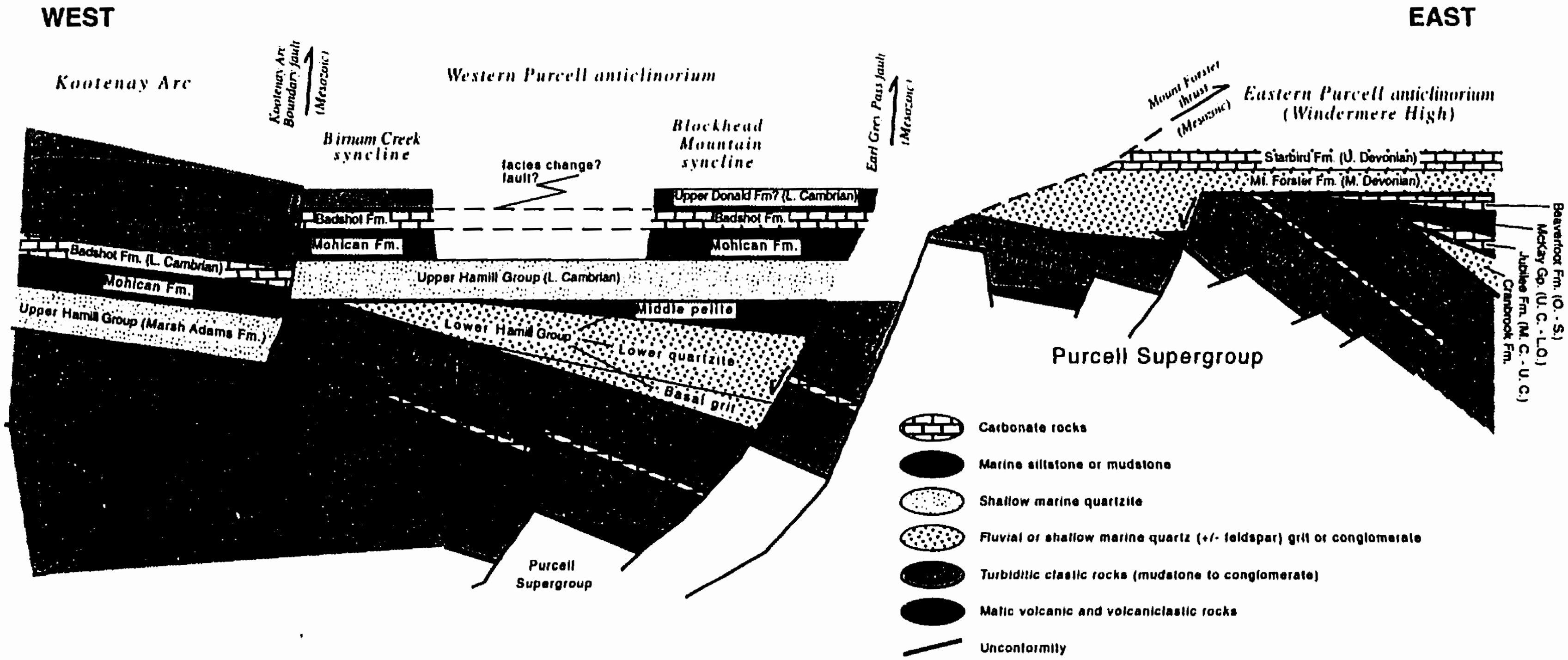



**Figure 4-3:** Diagram showing stratigraphic relationships across the Purcell anticlinorium and eastern Kootenay Arc at the latitude of Invermere - Duncan Lake. Stratigraphic relationships across major Mesozoic thrust faults in the study area indicate that they are reactivated normal faults, which were the result of several distinct episodes of extension, uplift and normal faulting in Neoproterozoic and Early Paleozoic time (see Chapters 2 and 3). Thicknesses not to scale (see Plate 4, Section 6 for pre-Mesozoic margin restored to scale). Relationships compiled from Reesor (1973), Root (1987), and this study.

exposed to the east (Fig. 4-3). The Lardeau Group is unconformably overlain by deep-water metasedimentary rocks of the Mississippian Milford Group. The Milford Group is in turn stratigraphically overlain by Triassic marine volcanic and sedimentary rocks that are associated with Slide Mountain and Quesnel terranes (Read and Wheeler, 1976; Klepacki, 1985).

These stratigraphic contrasts between the Purcell anticlinorium and the Kootenay Arc, and important lateral variations in thickness and lithofacies within each domain, are inherited from syn-depositional faulting related to episodes of Late Proterozoic to Early Paleozoic extension (Chapters 2-4). Deposition of the Neoproterozoic Windermere Supergroup was controlled by north- and northeast-trending normal faults (Root, 1987; Pope, 1989; and Chapter 2). Abrupt northwestward thickening and lithofacies changes that occur in the Horsethief Creek Group at the latitude of Starbird Pass and Horsethief Creek (Plates 2 and 3) are the most conspicuous expression in the study area of this episode of extension.

The lower part of the Hamill Group in the western Purcell anticlinorium and the lower part of the Hamill Group in the eastern Kootenay Arc were deposited during latest Neoproterozoic to earliest Cambrian time in separate half grabens that were bounded to the east by west-dipping normal faults (Chapter 3). The eastern normal fault was reactivated as a thrust fault during Mesozoic compressional deformation and is exposed in the immediate hangingwall of the Mount Forster thrust. The western normal fault coincides with the Kootenay Arc boundary fault, a Mesozoic thrust fault that marks the boundary between the Purcell anticlinorium and the Kootenay Arc. The two basins were separated by a eastward-tilted and bevelled block of Windermere Supergroup strata. The upper unit of the Hamill Group, the Mohican Formation and the Badshot Formation comprise an overlap assemblage that was deposited across the two basins and the intervening tilted blaock after this episode of normal faulting had ceased (Chapter 3).

Localized intermittent normal faulting continued after the deposition of the Badshot Formation (Chapter 3) and resulted in deposition of shallow-water facies to the east (upper Donald Formation and younger strata in the Purcell anticlinorium), in intermittent uplift, eastward tilting and erosion on the Windermere high and in deposition of deep-water, immature clastic and volcanic rocks in a basin to the west (Lardeau Group in the Kootenay Arc).

## STRUCTURAL STYLE OF THE PURCELL ANTICLINORIUM AND KOOTENAY ARC

### *Purcell anticlinorium*

The eastern limb of the Purcell anticlinorium is characterized by early east-verging thrust faults and thrust duplex structures that can be linked with structures in the western Main Ranges of the Rocky Mountains (Root, 1987, Kubli and Simony, 1994). The thrust faults are overprinted by upright folds, cleavage and low-grade regional metamorphism. The upright folds, axial planar foliation and regional metamorphism can be traced westward, where they comprise the single "phase" of more ductile deformation and higher-grade regional metamorphism that characterizes the western limb of the Purcell anticlinorium (Reesor, 1973). The Purcell anticlinorium is cut by several "out-of-sequence" east-verging reverse faults that include oblique northeast-trending, right-hand reverse fault segments. These faults cut the upright folds and their regional axial planar cleavage (e. g. Reesor, 1973; Root, 1987). From south to north these structures are: the Moyie-Dibble Creek, the St. Mary, Hall Lake and the Redding Creek-Mount Forster faults (Fig. 4-2). These faults merge with the Lussier River, Redwall and Purcell thrust faults, which carry the Purcell anticlinorium in their hangingwalls. The oblique right-hand reverse faults juxtapose different Proterozoic and Lower Paleozoic stratigraphic successions in their hangingwalls and footwalls, because they follow older Proterozoic to Early Paleozoic northwest-side-down normal faults (Lis and Price, 1976; Price, 1981; Root, 1987).

This chapter presents new structural and metamorphic data from the hangingwall of the Mount Forster fault. Structural relationships in the footwall of the Mount Forster fault have been described in detail previously by Root (1987) and Pope (1989, 1990).

The structural style of the western Purcell anticlinorium, east of Duncan Lake, is defined primarily by upright, generally north-plunging folds that deform an upward-facing stratigraphic succession (Plates 1,2 and 4a; Photo Plate 4B). The dominant, penetrative regional schistosity or cleavage is axial planar to these folds, and it defines an arcuate trend from northerly strikes in the south to northwesterly strikes in the north. The foliation generally dips to the west. There are also several areas in which minor folds and bedding-cleavage intersection lineations plunge gently to moderately to the south rather than to the north. These plunge reversals result in exposure of the Hamill Group in several doubly-plunging synclines primarily in the southern part of the study area (Plate 2).

Locally developed shear zones occur parallel to the steep dominant foliation. In general, they are more common and more ductile toward the west. A subhorizontal stretching lineation, parallel to minor fold axes and intersection lineations, occurs in these shear zones (Photo Plate 5C). However, where competent strata are tightly folded, (e.g. the Hamill Group in the Blockhead Mountain syncline), down-dip stretching lineations and boudinage due to flexural slip are common on fold limbs (Photo Plate 4C). Younger crenulations or spaced cleavages commonly overprint the dominant regional foliation, but neither their orientation nor their style is regionally consistent, except in parts of the Birnam Creek syncline where the dominant foliation is commonly a composite schistosity. The youngest observed structures in the western Purcell anticlinorium are locally developed but widespread, steep, northeast-striking left-lateral kink bands or spaced cleavage.

A conspicuous northeast-trending zone of change in structural style and in the plunge of folds extends across the western Purcell anticlinorium at the latitude of Starbird Pass and Horsethief Creek (Plates 2 and 3). This zone coincides with the major facies and thickness change in the Horsethief Creek Group, and with a change of strike in the Mount Forster thrust (Plate 2 and Fig. 4-2). Folds in strata at the top of the Horsethief Creek Group (e.g. at Starbird Pass) are gently north-plunging to the north of this zone but are more steeply south-plunging to the south. However, folds in strata at the base of the Horsethief Creek Group (e.g. at Horsethief Creek headwaters) are steeply north-plunging to the south and subhorizontal to the north. The reversals in plunge at different stratigraphic levels reflect the stratigraphic thickening of the Horsethief Creek Group from southeast to northwest, and the subsequent folding and eastward displacement of this basin. To the south, the doubly-plunging synclines expose a thick succession of the Hamill Group and younger strata. The Birnam Creek and Blockhead Mountain synclines are locally overturned to the west, and the Blockhead Mountain syncline is much tighter than other folds along strike to the north. To the north, the amplitude of major folds is less, the axial traces of major folds are more difficult to define and axial surfaces dip consistently to the west and are not as steep (Plate 2; compare sections A-C with E-H). Only the very base of the Hamill Group is exposed in one shallow, open, doubly-plunging syncline to the north of Starbird Pass.

### Blockhead Mountain syncline

The Blockhead Moutain syncline is a moderate to tight structure that exposes the Hamill Group, Badshot Formation and the upper Donald Formation in its core. Its eastern limb comprises a homoclinal sequence of Windermere and Purcell Supergroup that is carried in the hangingwall of the Mount Forster thrust. An adjacent anticline, the Hanging Glacier anticline, lies to the east, above a ramp in the Mount Forster thrust (Plate 2; Section E). The Blockhead Mountain syncline and the adjacent anticline die out abruptly at the latitude of Starbird Pass and Horsethief Creek. At their northern terminations, the anticline plunges abruptly to the north and the syncline abruptly to the south.

The Blockhead Mountain syncline is locally overturned to the west, and its orientation and structural style are different from the more open structures immediately to the west. Strata on the limbs show evidence of significant vertical tectonic attenuation (Photo Plate 4C), and incompetent strata are commonly sheared out. Multiple cleavages are common toward the core of the syncline, and small-scale refolded isoclinal folds occur in competent strata. The Blockhead Mountain syncline contains subvertical stretching or flexural-slip lineations much more commonly than the regional subhorizontal stretching lineation that is found both to the east and to the west toward the Kootenay Arc. The down-dip lineations appear to overprint older subhorizontal lineations, and are less ductile.

An east-verging, west-dipping thrust fault in the Horsethief Creek Group on the eastern limb (the Earl Grey Pass fault; Photo Plate 3F) cuts the cleavage that is axial planar to the Blockhead Mountain syncline. Statigraphic and sedimentological relationships along and across this fault imply that it is a reactivated normal fault that controlled the pattern of deposition of the Hamill Group in the western Purcell anticlinorium (Chapter 3).

Tea Creek anticline

The Tea Creek anticline, and equivalent structures to the south, comprises a complexly folded and faulted zone in the Horsethief Creek Group between the Blockhead Mountain and Mount Cauldron synclines. The folds and their axial planar foliations are vertical to moderately west-dipping, and the anticlines are locally overturned to the east. They are cut by several ductile to brittle fault zones that are coplanar with the axial planar cleavage and are associated with tight to isoclinal mesoscopic folds in the incompetent Horsethief Creek Group strata. None of these faults appears to have significant displacement.

Mount Cauldron syncline

The Mount Cauldron syncline is an open, upright, symmetrical structure that exposes grit and quartzite of the Hamill Group. A subvertical axial planar foliation fans across the structure. The foliation is steeply west-dipping on the eastern limb and steeply east-dipping on the western limb. The syncline shows an overall increase in penetrative strain from east to west, probably in part due to increasing metamorphic grade and ductility toward the west. Discrete ductile shear zones are limited to narrow zones between strata of contrasting competency and to the edges of the synkinematic Glacier Creek stock. These zones are more common on the western limb of the syncline and on the western margin of the stock. The amount of subhorizontal stretching of pebbles in the ductile shear zones appears to increase toward the west and parasitic folds are tighter and more numerous toward the west. The axial planar foliation within the Hamill Group quartzite is a spaced cleavage to the east, but to the west it is a penetrative schistosity, defined by strongly elongated and flattened quartz grains and oriented muscovite.

### Mount Banquo anticline

The axial trace of the Mount Banquo anticline marks an abrupt change in structural style in the western part of the Purcell anticlinorium. Structures on its eastern limb are open, strata are gently to moderately dipping toward the east, and ductile deformation is confined to discrete zones. On its western limb strata are subvertical and tightly to isoclinally folded, ductile deformation is pervasive even within very competent strata of the Hamill Group, and mylonitic fabric is common in ductile deformation zones. The Mount Banquo anticline becomes less distinct to the north toward the Howser Creek drainage, where it appears to be replaced by a zone of complex deformation comprising several tight structures. However, the contrast in structural style from east to west is just as sharp (Photo Plate 4B and Plate 2).

### Birnam Creek synclinorium

The Birnam Creek synclinorium is a composite structure on the western limb of the Purcell anticlinorium that exposes subvertical, isoclinally folded strata of the Hamill Group, Mohican, Badshot and lower Index Formations. It is bounded to the west by the Kootenay Arc boundary fault, which is discussed below. It is transitional in structural style and lithofacies between the Purcell anticlinorium and the Kootenay Arc. It commonly contains two or more tight synclines that expose the Mohican, Badshot and Index formations. The synclines are separated by ductiley folded and sheared strata of the upper quartzite of the Hamill Group. Near the Kootenay Arc boundary fault, the pervasive schistosity is a composite or transposed foliation. Isoclinally refolded folds are common in thin section but rare at the outcrop scale. They appear to be related to inhomogeneous high strain in sequences of interbedded competent and incompetent strata on the limbs of upright folds. Map-scale and outcrop-scale folds are primarily symmetrical, suggesting a significant component of E-W ductile flattening. Subhorizontal stretching lineations are conspicuous.

*Eastern Kootenay Arc*

Two phases of coaxial ductile folding (Photo Plate 4A) control the map pattern of the eastern Kootenay Arc (Fyles, 1964). The older folds are large-amplitude (>10 km scale) west-verging isoclinal folds that have been deformed by one, and locally two, phases of coaxial, steep, tight to isoclinal folds (Plates 1 and 2) Much of the stratigraphic sequence is overturned, and along the eastern boundary of the Kootenay Arc, the sequence is everywhere overturned (Fyles, 1964). The younger foliation becomes stronger and the younger folds tighter toward the east, and the earlier foliation is progressively overprinted and transposed, so that distinct "phases" of deformation are difficult to distinguish toward the eastern limit of the Kootenay Arc. The younger foliation and axial surfaces dip more gently toward the west at lower structural levels, due to refolding by a third phase of gently west-dipping kink folds that do not much affect the map pattern (Fyles, 1964; Klepacki, 1985). This third phase of folding is more strongly developed toward the eastern limit of the Kootenay Arc.

Deformation associated with both major phases of folding was conspicuously more ductile in the eastern Kootenay Arc than in the adjacent Purcell anticlinorium. Marker beds or stratigraphic units are generally more attenuated on fold limbs. Ductile shear zones are common, and thin sections show evidence of pervasive dynamic recrystallization, particularly toward the east. Primary sedimentary structures are very poorly preserved. The subhorizontal stretching lineation that occurs in discrete high strain zones in the Purcell anticlinorium is more strongly and pervasively developed in the Kootenay Arc. This lineation is parallel to the fold axes and the intersection lineations associated with both phases of folding.

The limbs of several of the older west-verging folds are truncated by two steep west-dipping faults that show normal displacement of up to 2 km (Fyles, 1964; Plates 1 and 4a). These faults are roughly parallel to the axial surfaces of the younger upright folds, but they locally cut the steep foliation and are primarily late- to post-metamorphic, brittle structures.

The eastern Kootenay Arc can be divided into two main structural domains: a western domain in which younger structures are primarily at a high angle to older, recumbent west-verging structures, and an eastern domain in which the older structures have been isoclinally folded about the upright structures, and thus they are more coplanar. The two structural domains are separated by the St. Patrick syncline (Fyles, 1964), a complexly deformed and poorly defined synclinal zone, and by a steeply west-dipping normal fault that truncates the eastern limbs of both the Meadow Creek anticline and the Kootenay Lake antiform (Plates 1 and 4a). The relative age relationships and orientations of foliations in the two domains are critical to subsequent discussions of the relative and absolute ages of intrusions, metamorphism and deformation.

## Western Domain

The western domain comprises the Meadow Creek anticline, a recumbent west-verging structure with an amplitude of at least 6 km (Fyles, 1964), and the the Kootenay Lake antiform, another large-amplitude (10 km), subvertical structure that gently refolds the Meadow Creek anticline (Fyles, 1964; Klepacki, 1985; Plates 1 and 4a). The Meadow Creek anticline is a complex, isoclinal structure with a strong axial planar schistosity that closes to the west of Kootenay Lake and appears to open to the east of Kootenay Lake. It is defined primarily by the Hamill Group and the Mohican and Badshot Formations, but it also contains several lenses of Index Formation. These and the complex map pattern near the hinge region imply that the "anticline" in fact comprises several high-amplitude, isoclinal anticlinal and synclinal hinges (Fyles, 1964).

The axial trace of the Kootenay Lake antiform follows Kootenay Lake and the wetland to the north, so that the limbs and axial planar schistosity of the Meadow Creek anticline dip gently to west, on the west side of Kootenay Lake, and moderately to the east, on the east side of Kootenay Lake. The cleavage or schistosity that is axial planar to the younger Kootenay Lake antiform dips moderately to the west on the west side of the Lake and becomes steeper through the hinge region to the eastern limb (Fig. 4-4).

The two foliations are easy to distinguish on the west side of the lake within the Hamill Group, Mohican Formation and Badshot Formation. The gently west-dipping schistosity or compositional layering that is axial planar to the Meadow Creek anticline is the most apparent foliation within these strata. The younger, steeper cleavage or schistosity more strongly overprints the older in the less competent Lardeau Group, where it is locally the dominant fabric. In the hinge region and adjacent eastern limb of the Kootenay Lake antiform, at the south end of Duncan Lake, the two foliations are also easy to distinguish, even within the Lardeau Group. The dominant, bedding-parallel schistosity is flat to moderately east-dipping and folded by a steeply west-dipping cleavage. However, the two foliations are more difficult to distinguish on the eastern limb of the Kootenay Lake antiform at lower structural levels, toward the south, because: 1) the older foliation is more tightly folded by parasitic folds; 2) the antiform is slightly overturned to the east so that the two schistosities are nearly parallel; and 3) the two foliations are defined by the same metamorphic assemblage and thus appear similar.

The two generations of folds are coaxial. Bedding-schistosity intersection lineations and minor folds associated with the Meadow Creek anticline have the same orientation as bedding-cleavage intersection lineations and minor folds associated with the Kootenay Lake antiform.

### Eastern domain

To the east of the Meadow Creek anticline, the steeply west-dipping folds and axial planar schistosity very tightly refold and strongly overprint the earlier isoclinal folds and schistosity (Plate 4a). Bedding and both schistosities are most commonly subparallel and they appear as a composite or transposed compositional layering. Stratigraphic contacts are strongly transposed at a scale of centimeters to hundreds of meters (Fyles, 1964). The hinge regions of early folds are rarely preserved. Therefore, the geometry of the earlier structures is defined primarily from

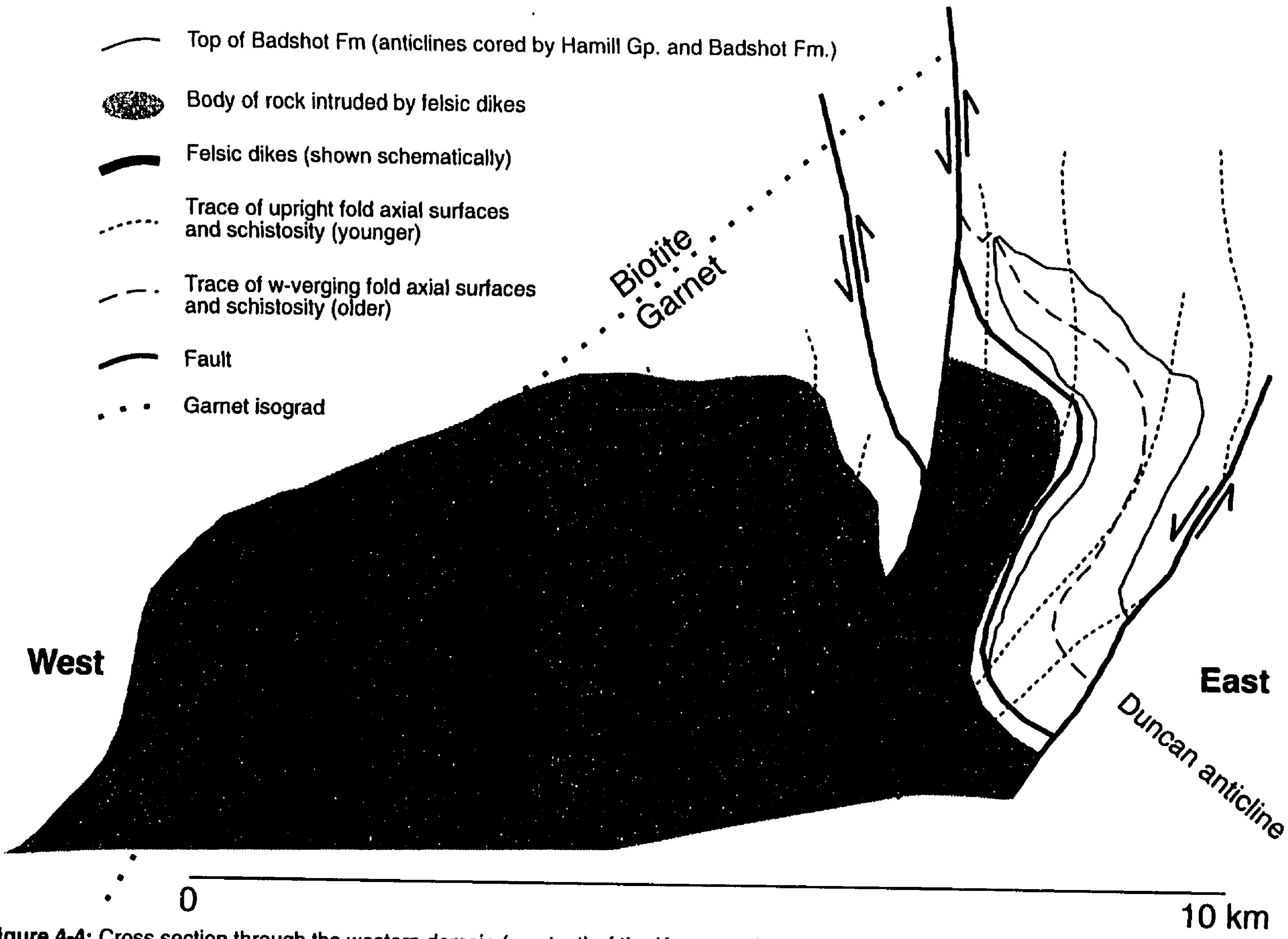



**Figure 4-4:** Cross section through the western domain (see text) of the Kootenay Arc, illustrating structural relationships between syn-kinematic felsic dikes, older west-verging folds, younger upright folds, late normal faults and regional metamorphic isograds. Individual dikes locally cut bedding at the hinges of the older folds, yet they contain a weak foliation that is parallel to the axial surfaces of the older folds, indicating that the dikes were intruded during the episode of older, west-verging folding. Cross section (see Plate 4) modified after Fyles (1964) and Klepacki (1985). Upper limit of area intruded by dikes projected from data in Fyles (1964). Lower limit not exposed. Truncation of garnet isograd by normal fault is inferred from outcrop-scale relationships and from map relationships shown in Fyles (1964).

reversals in the stratigraphic sequence that must pre-date the dominant, steep folds (Fyles, 1964).

The Duncan anticline and the Howser syncline are the major early structures that affect the map pattern (Fyles, 1964; Plates 1 and 4a). The Duncan anticline lies to the west of and below the Howser syncline. The Duncan anticline is an isoclinal structure with an amplitude of about 10 km. It is tightly refolded by the Glacier Creek synform, to the west, and the Comb Mountain antiform, to the east. A normal fault that is congruent with the axial surface of the Glacier Creek synform is interpreted to offset the Duncan anticline by about 2 km (Fyles, 1964). The hinge line of the Howser syncline is tightly refolded by the Glacier Creek synform and the Comb Mountain antiform and is more openly refolded by the Lavina synform to the east. A second early anticlinal hinge is inferred to the east of the Howser syncline, because the overturned sequence of Hamill Group, Mohican Formation and Badshot Formation that comprises the eastern limb of the Howser syncline must at one time have been continuous with the equivalent upright sequence that is exposed a few kilometers to the east in the Birnam Creek synclinorium of the Purcell anticlinorium. However, the hinge and upright limb of the inferred anticline must be everywhere above the erosion surface in the northern Kootenay Arc.

The Lavina synform (Plate 4a) is an open fold and relatively strain-free "window" close to the Kootenay Arc boundary fault. It reveals the relationships between the two fold phases. The axial planar cleavage dips moderately to steeply to the east at high structural levels and steeply to the west at low structural levels. It clearly cross-cuts a gently folded schistosity that is parallel to bedding. Fold interference patterns at outcrop scale reflect clearly the refolding of the early recumbent, isoclinal folds. The hinges of the two phases of folds are parallel or nearly parallel. It is noteworthy that the older bedding-parallel schistosity that is axial planar to the early folds is the more strongly developed foliation, as it is in the Meadow Creek anticline to the west.

Strain associated with the upright folds increases markedly eastward between the Lavina synform and the Kootenay Arc boundary fault to the east. In the corresponding antiform to the east (called Lavina antiform here for simplicity), the older foliation is again more tightly folded and overprinted by a progressively stronger upright schistosity that is locally mylonitic. The eastern limb of the "Lavina antiform" is subvertical, in contrast to the gently-dipping western limb that it shares with the Lavina synform.

## THE KOOTENAY ARC BOUNDARY FAULT

The subvertical, locally mylonitic fault that is the structural boundary between the Purcell anticlinorium and the Kootenay Arc is referred to here as the Kootenay Arc boundary fault (Figs. 2 and 3). The Kootenay Arc boundary fault is well exposed in the study area on several east-west ridges to the east of Duncan Lake. It is a ductile high strain zone (Photo Plate 5A) up to several tens of meters wide that marks the culmination of a ductile strain gradient that decreases eastward across the Birnam Creek syncline and decreases westward toward the Meadow Creek anticline. The fault juxtaposes an upward-facing stratigraphic sequence to the east against an overturned sequence to the west.

Similar structural relationships have been described to the south across the West Bernard fault (Höy, 1974, 1980) and across the Seeman Creek fault (Leclair, 1989) and can be inferred from map relationships in the region immediately to the north and west of Duncan Lake (Read and Wheeler, 1976). These relationships indicate that the Kootenay Arc boundary fault is a regionally significant structure or structural zone with a strike length of at least 120 km (Fig. 4-2). Leclair (1988) concluded that a structural zone equivalent to the Seeman Creek fault extends to the south of the International Border for at least another 50 km. Thus the newly-proposed name "Kootenay Arc boundary fault" is intended as a regional name that includes the West Bernard and Seeman Creek faults, as well as un-named, inferred segments of this structure. Along the entire length of the Kootenay Arc boundary fault, coaxially refolded, large-amplitude folds occur to the west but not to the east of the fault.

The amount and sense of motion along the southern parts of the Kootenay Arc boundary fault have been long-standing enigmas. Kinematic indicators are lacking on both the West Bernard (Höy, 1980) and the Seeman Creek (Leclair, 1988) faults. The West Bernard fault is subparallel to tightly to isoclinally folded bedding, exposes primarily the Hamill Group in its hangingwall and footwall and, because it cuts folded strata, the fault does not consistently place younger subdivisions of the Hamill Group against older, or vice versa. Therefore, the amount and direction of displacement cannot be determined. Höy (1980) considered the possibilities of both normal (west-side-down) and thrust (west-side-up) displacement, but he favored the thrust interpretation, primarily due to the decrease in regional metamorphic grade eastward from the fault. However, both interpretations required that he invoke a complexly deformed and locally downward-facing "root zone" for the recumbent folds to the east of the West Bernard fault, either at depth or above the erosion surface, in order to reconcile the juxtaposition of refolded recumbent folds to the west against a ubiquitously upright sequence to the east. Despite the regional northerly plunge and the significant local relief (more than 2 km), no such "root zone" is observed along the entire western flank of the Purcell anticlinorium.

The subvertical Seeman Creek fault, which is exposed farther to the south, juxtaposes a thick Neoproterozoic to Lower Cambrian succession of Windermere Supergroup and Hamill Group strata against an equivalent but distinctly different succession to the east. Thus, hangingwall and footwall cutoffs cannot be matched across the fault because it apparently coincides with a Neoproterozoic to Early Paleozoic syn-sedimentary normal fault (Leclair, 1988; and Chapter 3). However, the repetition of equivalent, thick successions across the fault implies thrust motion. Leclair (1988) also argued that the syn-metamorphic upright folds with which the Seeman Creek fault is congruent show west-over-east asymmetry and therefore that the fault is a coeval east-verging thrust fault.

The displacement history of the Kootenay Arc boundary fault east of Duncan and northern Kootenay lakes remains puzzling. The enigmatic subhorizontal stretching lineation that occurs pervasively in the Kootenay Arc and locally in the Purcell anticlinorium (Photo Plate 5C) is most strongly developed in and near the Kootenay Arc boundary fault. Down-dip stretching would be expected for simple thrust or normal diplacement. The fault is more ductile in this area than it is to the south, but most of the mylonitic fabric is well-annealed (Photo Plate 5B) and contains no conclusive kinematic indicators. Shear bands and non-cylindrical asymmetric folds locally deform the annealed fabric and most commonly indicate transcurrent dextral motion. Therefore, the earlier displacement history of the fault has been obscured, and the dextral displacement can only be treated as the latest ductile displacement.

New stratigraphic evidence from the Duncan Lake area implies that earlier west-side-up thrust displacement occurred on the Kootenay Arc boundary fault and supports the conclusions of Höy (1980) and Leclair (1988). Between the Fry Creek batholith and Duncan Lake (Plate 1) the greenstone- and pelite-bearing unit of the Hamill Group, which underlies the upper quartzite unit in the Kootenay Arc, is exposed on the west side of the Kootenay Arc boundary fault. It is juxtaposed against the upper quartzite unit of the Hamill Group on the east side of the fault everywhere except on the southeast shore of Duncan Lake, where it is juxtaposed against previously unrecognized Index Formation exposed in the core of the Birnam Creek syncline. Thus, older rocks on the west side are juxtaposed against younger rocks on the east side everywhere along this segment of the fault. The Kootenay Arc boundary fault is congruent with the steep eastern limb of the Lavina antiform (Plates 1 and 4a), the Birnam Creek synclinorium and the schistosity that is axial planar to both structures and which is most strongly developed in the fault zone. These relationships imply that it is a west-side-up thrust fault that developed during the upright folding. However, the amount of dip separation across the fault cannot be determined because the hangingwall cutoffs that correspond to observed footwall cutoffs are an unknown distance above the present erosion surface. The anomalously thick section of the upper Hamill Group that occurs to the east of the Kootenay Arc boundary fault in the southern part of the study area (Plate 2; section H) and the presence of several sub-parallel ductile shear zones in this unit suggest that there may have been thickening or repetition by thrusting on possible splays or related faults in the immediate footwall of the Kootenay Arc boundary fault.

## RELATIONSHIPS BETWEEN REGIONAL METAMORPHISM AND DEFORMATION

Regional metamorphic isograds define a sharp metamorphic culmination that lies to the east of northern Kootenay and Duncan lakes (Fig. 4-5). Regional metamorphism increases from the chlorite zone in the east and north to the staurolite-kyanite zone in the west and south of the study area, and it decreases to the biotite zone on the west side of Kootenay and Duncan lakes.

The staurolite-kyanite zone that comprises the core of the regional metamorphic culmination is very poorly defined, and the isograd is largely inferred. It's dip is unconstrained but is assumed to be steep (Plate 4a). The garnet isograd in the western Purcell anticlinorium, on the eastern flank of the culmination, must be nearly vertical, although its precise location and orientation are poorly constrained. The garnet isograd on the western flank of the culmination must dip steeply to the west. On the slope to the north of Duncan Lake, where it is well constrained (Fyles, 1964), it appears to dip at about 55° to the west, which is less steep than both foliations in this locality. At the south end of Duncan Lake, the isograd is inferred to show a few thousand meters of left-lateral separation along a subvertical post-metamorphic normal fault with 2 km of normal displacement. Therefore, the isograd probably also dips steeply to the west in this locality. The biotite isograd in the western Purcell anticlinorium appears to dip moderately to steeply to the east, based on its relationship with the topography. The biotite isograd is inferred to be offset by steep syn-metamorphic faults as shown in Plate 4a, because the faults locally cut the dominant foliation. However, because the isograd is so nearly parallel to the faults, the precise relationships between the faults and the isograd are uncertain.

In the western Purcell anticlinorium, the peak metamorphic assemblages are associated with the pervasive regional cleavage or schistosity that is axial planar to the map-scale folds. In pelitic rocks this foliation is defined by metamorphic mineral assemblages that include chlorite, muscovite, biotite, quartz, albite and, in the Birnam Creek syncline, retrograded garnet. In calc-silicate rocks in the Birnam Creek syncline, this foliation is defined by metamorphic mineral ssemblages that include dolomite, quartz, muscovite, actinolite and locally hornblende. In rare amphibolites in eastern exposures of the Horsethief Creek Group the dominant foliation is

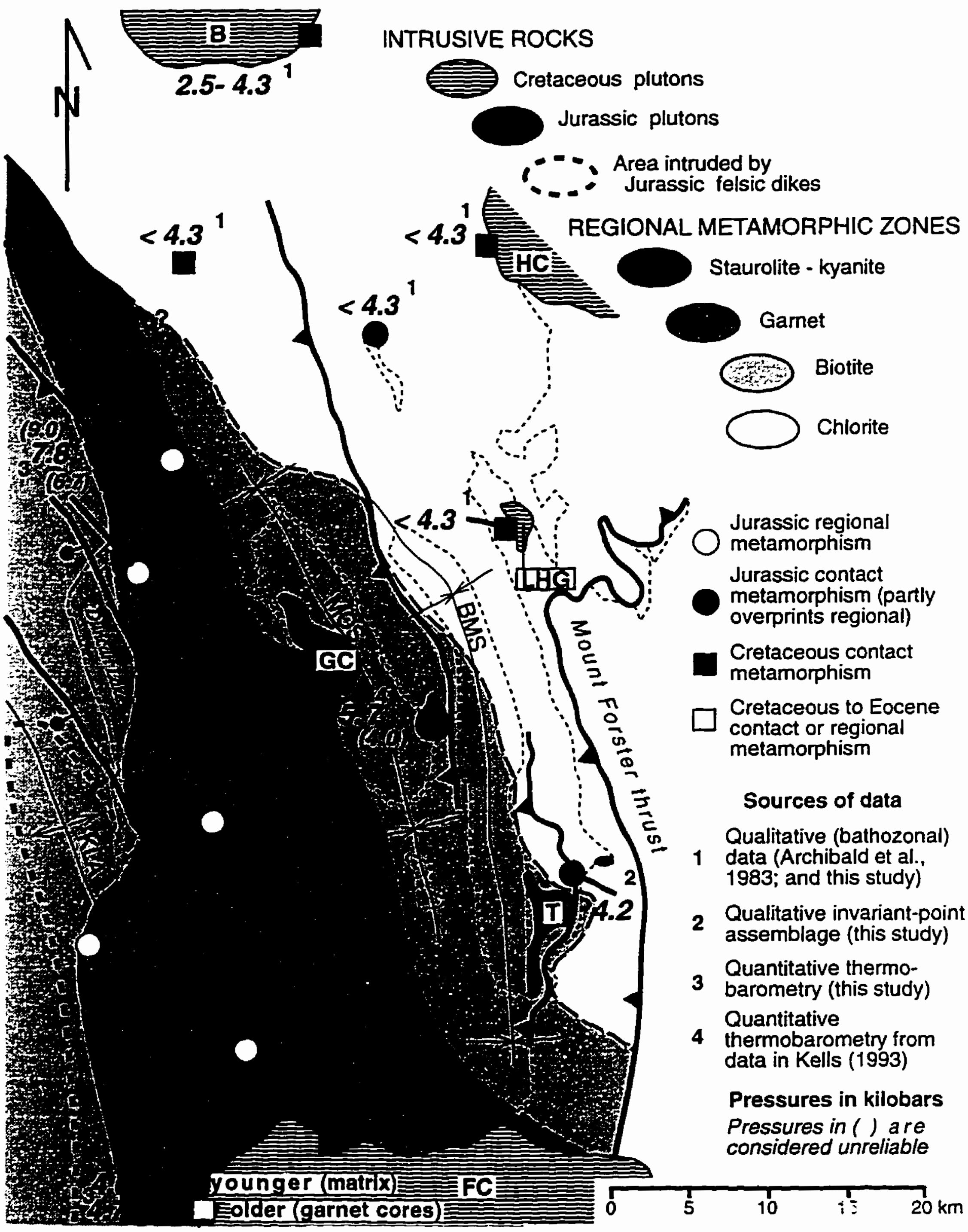



**Figure 4-5:** Simplified map of the study area showing distribution of regional metamorphic isograds (bold dashed lines; compiled from Reesor, 1973, and this study), their relationships to major faults (bold solid lines) and the locations of regional and contact metamorphic pressure data. **B** = Bugaboo batholith; **HC** = Horsethief Creek batholith; **LHG** = Lake of the Hanging Glacier stock; **GC** = Glacier Creek stock; **T** = Toby stock; **FC** = Fry Creek batholith; **BCS** = Birnam Creek syncline; **MCS** = Mt. Cauldron syncline; **BMS** = Blockhead Mtn. syncline; **KLA** = Kootenay Lake antiform. Contacts (thin dotted lines) are the base of the Hamill Group in the Birnam Creek, Mt. Cauldron and Blockhead Mtn. synclines and the base of the Windermere Supergroup in the immediate hangingwall of the Mt. Forster thrust. See text for discussion of data.

defined by hornblende, biotite, albite and fresh garnet. Garnet porphyroblasts in the amphibolites and garnet pseudomorphs in the pelites are syn-kinematic with respect to the dominant regional foliation, but the development of the this foliation in the pelites outlasted garnet growth. Inclusion trails in garnet suggest that garnet growth was also syn- to post-kinematic with respect to a locally-preserved, older coplanar foliation in the Birnam Creek syncline. Thus, peak metamorphic conditions in the western Purcell anticlinorium were achieved after regional folding began and before it ended.

Younger schistosities that post-date the dominant regional foliation and map-scale folds in the western Purcell anticlinorium are defined by muscovite and chlorite. To the west of the Mount Banquo anticline, where the younger foliations become penetrative, garnet pophyroblasts in pelitic rocks are most commonly retrograded to biotite and chlorite, and the retrograde assemblage and foliation strongly overprints and transposes the older steep schistosity. Virtually no fresh garnet is preserved in pelites in the footwall of the Kootenay Arc boundary fault, except near the northern contact of the Fry Creek batholith. Retrograde metamorphic mineral assemblages (chlorite and chloritoid) and weak foliations are also superposed on contact metamorphic assemblages and the older regional foliation in the contact aureoles of the synkinematic plutons (discussed below).

Prograde regional metamorphism in the eastern Kootenay Arc accompanied both the west-verging and upright phases of deformation. Peak P-T conditions were achieved during the younger, upright phase of deformation. Garnet growth in the metamorphic culmination (Photo Plate 7A) was associated with both the earlier west-verging folds and the younger, steep folds and their axial planar foliations. On the west side of Kootenay Lake, where the schistosity axial planar to the recumbent Meadow Creek anticline is at a higher angle to younger foliations, garnet growth was synchronous with but outlasted the development of the schistosity that is associated with the west-verging structures.

Metamorphism outlasted most of the deformation at lower structural levels in the Kootenay Arc. Retrograde metamorphism locally accompanied the latter stages of penetrative deformation, but it is much less severe than in the adjacent western Purcell anticlinorium. In general, the higher-grade assemblages to the south more commonly overprint the foliations and show only slight post-kinematic alteration, whereas the prograde assemblages to the north are more strongly deformed and retrograded (Fyles, 1964; and this study). Staurolite and kyanite in the core of the metamorphic culmination were syn- to post-kinematic with respect to the younger, upright deformation (Fyles, 1964; Reesor, 1973).

## RELATIONSHIPS BETWEEN INTRUSIONS, CONTACT METAMORPHISM AND DEFORMATION

Two suites of granitoid plutons have intruded the region: 1) a predominantly granodioritic suite that is partly deformed and is semi-concordant with map-scale folds, and 2) a granitic or quartz monzonitic suite that cross cuts major folds and thrust faults and is unfoliated (Reesor, 1973). The contact aureoles of both suites of plutons overprint the regional metamorphism in the northeastern part of the region, but contact metamorphism becomes increasingly difficult to distinguish from regional prograde metamorphism in the southwestern part of the region. A suite of deformed granitic dykes and sills in the Kootenay Arc (Photo Plate 6B) has been interpreted by Fyles (1964) as mainly syn-kinematic with respect to the second-phase upright folds, but as having ambiguous relationships with the older, high-amplitude, west-verging folds in the Kootenay Arc. New observations provide further insight into relationships between granitoid intrusions, regional metamorphism and deformation, as discussed below.

### *Syn-Kinematic Plutons*

#### Granodioritic stocks: Purcell anticlinorium

Several granodioritic plutons that cross-cut major structures in the western Purcell anticlinorium (Plate 1) are elongated and locally foliated parallel to the axial surfaces of the regional folds. These include the Toby stock, which intrudes the southern part of the Blockhead Mountain syncline, the Glacier Creek stock, which intrudes the Mount Cauldron syncline, and several smaller bodies. Toby stock ranges in composition from pyroxene syenodiorite to hornblende-biotite granodiorite, and the other plutons range from quartz diorite to quartz monzonite (Reesor, 1973), although hornblende-biotite granodiorite appears to be most common. The rims of both Glacier Creek and Toby stocks are more mafic than the cores. They contain more abundant hornblende and clinopyroxene and lesser amounts of orthopyroxene. Several small, previously unmapped plutons that are less than 1 km in diameter are located to the northeast of Toby stock and are similar in composition to its rim. The "tail" at the south end of the Toby stock and the margins of the Glacier Creek stock are strongly sheared.

Contact metamorphic aureoles are well developed over horizontal distances of up to 1200 m from the plutons in pelitic rocks of the Horsethief Creek Group (Reesor, 1973; Kells, 1993), but are less easily recognized in pelitic or semi-pelitic strata in the Hamill Group. Pelitic contact metamorphic assemblages most commonly include staurolite, biotite, muscovite, garnet, plagioclase, quartz, and rarely, kyanite and/or andalusite (Photo Plate 7B and 7C). Sillimanite (fibrolite) is common within 500 m of the contacts (Kells, 1993, and this study). Chloritoid porphyroblasts occur in the outer parts of the Glacier Creek and Toby stock aureoles (Photo Plate 7D). Calc-silicate assemblages containing tremolite, diopside, epidote, dolomite and quartz occur in siliceous marble of the lower Horsethief Creek Group in the contact aureoles of Toby stock and smaller outlying bodies.

Structural relationships indicate that the plutons were intruded during the penetrative deformation, after regional folding and metamorphism had begun but before they had ended. The Glacier Creek stock has steep contacts where it intrudes the core of the Mount Cauldron syncline, and it cuts sub-horizontal stratigraphic contacts at high angles, but the similarly steep-sided, "tadpole-shaped" Toby stock intrudes the steeply-dipping limb of the Blockhead Mountain syncline subparallel to the stratigraphic contacts. These relationships indicate that the plutons were intruded after the folds were at least partly formed. Dikes associated with both Glacier Creek and Toby stocks cut smaller folds that are parasitic to the Mount Cauldron and Blockhead Mountain synclines, but the dikes are themselves folded about the axial surfaces, indicating that the dikes are syn-kinematic (Photo Plate 6A). A thrust fault on the east limb of the Blockhead Mountain syncline cuts the cleavage that is axial planar to the syncline and truncates the Toby stock in its hangingwall. The contact metamorphic mineral assemblages partially overgrow the axial planar cleavage that is cut by the fault. These relationships imply that Toby stock was intruded during regional folding and prior to local thrust faulting.

The relationships between contact metamorphic mineral assemblages and regional metamorphic foliations also imply that the plutons were intruded during the regional deformation and metamorphism. Contact metamorphic porphyroblasts most commonly overgrow the prograde regional metamorphic mineral assemblage that defines the dominant regional foliation. However, this foliation also is deflected around the porphyroblasts, or in some cases they appear to have been rotated parallel to it (Photo Plate 7C).

Younger crenulation cleavages and regional retrograde metamorphic foliations are superposed on the contact metamorphic assemblages, particularly on the outer margins of the contact aureoles (Kells, 1993, and this study). However, chloritoid porphyroblasts, which are confined to the outer contact aureoles, also locally have overgrown this foliation, suggesting that contact metamorphism locally may have post-dated regional retrograde metamorphism. This relationship also reinforces the conclusion that the prograde metamorphism and some of the retrograde regional metamorphism occurred concurrently or overlapped in age. Regional retrograde metamorphic foliations most strongly overprint the contact metamorphism on the western edge of the Glacier Creek stock, which is closest to the Kootenay Arc. Contact metamorphism most clearly post-dates regional metamorphism and deformation at the eastern side of Toby Creek stock, and at a small plug north of Starbird Pass, which are farthest to the east.

East-northeast-trending, sub-vertical left-lateral kink bands, the youngest "phase" of deformation observed in the study area, clearly cut the Glacier Creek stock. They are brittle features in the pluton as well as in the country rock, indicating that they were superposed on the pluton after it had cooled below the temperature at which it was ductile. The kink bands occur regionally throughout the northern Purcell anticlinorium (Kubli, 1990) and the adjacent western Main Ranges of the Rocky Mountains (M. McDonough, pers. comm., 1994).

Felsic dikes: Kootenay Arc

Abundant deformed biotite- and hornblende-bearing aplite and feldspar porphyry dikes intrude the Hamill Group, Badshot Formation and Lardeau Group in the recumbent Meadow Creek anticline (Fyles, 1964 and 1967). They are sheet-like bodies that are generally one to several metres thick but locally up to 100 m thick (Photo Plate 6B). Locally, their volume exceeds that of the country rock. Most of the dikes are subparallel to the dominant schistosity, which is axial planar to the Meadow Creek anticline and which is folded by the upright Kootenay Lake antiform. They are weakly lineated and foliated and are boudinaged. A few dikes cross-cut this schistosity at low angles or cross-cut other dikes. Their upper limit lies primarily within the Lardeau Group on the upright limb of the Meadow Creek anticline, and it is subparallel to the limbs and axial surface of the anticline (Fig. 4-4). The volume of rock that they intrude appears to plunge to the north, parallel to the axis of the Kootenay Lake antiform.

Fyles (1964, 1967) inferred that the dikes were emplaced primarily during the latter stages of upright folding in the Kootenay Arc, but his reports are ambiguous about cross-cutting relationships with the earlier, refolded folds and their axial planar schistosity. His argument was based in part on the fact that the dikes are not strongly deformed by the upright structures. A U-Pb zircon upper intercept age of 173 +/- 5 Ma from one of these dikes on the west side of Kootenay Lake, immediately to the south of the study area, has been used an an emplacement age for the dikes (Smith et al., 1992). Thus it is critical to determine the relative age relationships between the dikes and both major phases of folding in this part of the Kootenay Arc. Smith et al. (1992) followed Fyles (1964, 1967) in concluding that dikes such as the one they dated were intruded during the upright folding, but their discussion of structural relationships is also ambiguous because, toward the south, the west-verging folds and their axial planar schistosity are more strongly overprinted by one or more phases of tight, upright folding, and the dikes themselves are more strongly deformed.

Most of the individual dikes, as well as the sheet-like volume of country rock that they intrude, clearly have been deformed by the Kootenay Lake antiform, because they are weakly deformed by the foliation that is axial planar to the older, recumbent Meadow Creek anticline, and both the dikes and the early foliation dip in opposite directions on either side of the hinge of the antiform (Fig. 4-4). The critical cross-cutting relationships between the dikes and the older, west-verging Meadow Creek anticline are exposed on the western limb of the Kootenay Lake antiform and near its hinge, where the two foliations are at a high angle to each other and the younger folds and foliation are weak. The dike dated by Smith et al. (1992) can be traced northward for several km toward this area, where it is parallel to the dominant, older schistosity. The weak foliation in the dikes is also parallel to the dominant schistosity. The upright cleavage is much more weakly developed in the country rock and does not appear to cut the dikes at all. Some of the dikes cut across early isoclinal hinges defined by bedding. Because the dikes locally cut the older folds but are also deformed by a foliation that is parallel to their axial planar schistosity, they must have intruded during the development of that early axial planar foliation. They were emplaced as sill-like bodies parallel to the gently-dipping axial planar foliation. They were not strongly deformed by either phase of deformation in this area because they are so much more competent than the mainly schistose country rock.

However, a few smaller dikes also cross-cut the early foliation. On the west side of Kootenay Lake, some dikes cut earlier dikes that are folded by the upright folds (Fyles, 1967). The younger dikes are not folded, but they are lineated. A few dikes on the east side of Kootenay Lake clearly do cut the steep, composite foliation on the east limb of the Kootenay Lake antiform. Therefore, dike emplacement must have continued during the younger, upright folding. Based on map relationships shown by Fyles (1964), one of the late- to post-metamorphic normal faults in the Kootenay Arc appears to truncate the southeastern edge of the sheet-like volume of rock that is intruded by the dykes (Plate 1 and Fig. 4-4). Dikes exposed in outcrops on the east side of

Kootenay Lake are truncated by small-scale, north-trending, west-side-down normal faults that are parallel to this larger fault.

Some of the dikes show obvious chilled margins (Fyles, 1964, 1967), indicating that they were intruded into rocks that were relatively cool. Since the peak metamorphic mineral assemblages generally have overgrown the dominant schistosity, the dikes must have been intruded before peak regional metamorphic conditions were achieved. The superposition of the steeply-dipping garnet isograd across the gently-dipping limbs of the recumbent Meadow Creek anticline and the tabular body of rock intruded by the dikes reinforces this conclusion. It is noteworthy that the structural culmination in the volume of rock intruded by the dykes does not coincide with the regional metamorphic culmination to the east, indicating that the metamorphic culmination was not controlled by the intrusion of a large volume of dikes.

Late- or Post-Kinematic Plutons

The Fry Creek, Horsethief Creek and Bugaboo batholiths and Lake of the Hanging Glacier stock (Plate 1) comprise a suite of quartz monzonitic to granitic plutons that cut map-scale folds in all parts of the study area (Reesor, 1973). The plutons in this suite are more aluminous than those in the syn-kinematic suite. They generally lack hornblende and locally contain both biotite and muscovite.

The western part of the Fry Creek batholith truncates both the older and the younger map-scale folds in the eastern part of the Kootenay Arc. The eastern part of the Fry Creek batholith and the other plutons truncate but locally deform the regional foliation and upright folds in the western Purcell anticlinorium. Both bedding and the dominant regional foliation are deflected from their northerly/northwesterly trends near the margins of the Fry Creek and Bugaboo batholiths, and they are warped upward into an orientation that strikes parallel to and dips steeply away from the sub-vertical batholith margins (Reesor, 1973; Plate 2). A north-trending cliff exposure of the Lake of the Hanging Glacier stock shows that the Toby Formation is warped upward adjacent to and above the pluton (Plate 2).

The plutons also post-date regional thrust-faulting in the study area. The Horsethief Creek batholith cuts the Mount Forster fault (Walker, 1926; Reesor, 1973) which in turn cuts the regional cleavage. The Fry Creek batholith is inferred to cut the Kootenay Arc boundary fault. Similar cross-cutting relationships between the Fry Creek batholith, the structures of the Kootenay Arc and the Kootenay Arc boundary fault (West Bernard fault) were reported by Höy (1980) from the southern margin of the batholith.

The batholiths and stocks of this suite are rarely internally foliated, and quartz grains in thin sections are strain free (Reesor, 1973). However, the northwestern part of the Fry Creek batholith is foliated by a north-trending cataclastic and ductile fabric, defined by muscovite and flattened or elongate strained quartz (Reesor, 1973; and this study). Fyles (1964) suggested that this section of the batholith may have been affected by the latest episode of upright folding in the Kootenay Arc, but the structures in the pluton have distinctly different orientations from those in the intruded metasedimentary rocks. The foliation in the batholith has a similar strike to the regional schistosity, but it dips only gently to moderately to the west. More importantly, the ductile mineral lineation associated with this foliation is a down-dip lineation, in contrast to the pervasive subhorizontal lineation in the metasedimentary rocks of the Kootenay Arc. The northwestern margin of the pluton is along strike from the trace of one of the late- to post-metamorphic, west-dipping normal faults on Fyles' (1964) map, and it is possible that these faults post-date the pluton. Geobarometric and geochronological data that indicate that the western part of the pluton has been exhumed in the footwall of a normal fault and tilted relative to the eastern part are discussed below.

The contact metamorphic aureoles associated with these plutons clearly overprint the regional metamorphism and deformation, except at the northwestern margin of the Fry Creek batholith where it is difficult to distinguish contact metamorphism from regional metamorphism, and deformation associated with pluton emplacement from regional deformation. Within a few meters to tens of meters from the pluton in this locality, sillimanite in part defines a schistosity that is parallel to the pluton margin and axial planar to cm- to m-scale isoclinal folds. These folds appear to merge away from the pluton with folds that are parasitic to the north-trending, upright folds and amphibolite-facies foliation in the Kootenay Arc. To the east, along the northern margin of the Fry Creek batholith, the north-trending regional isograds deflect to the east and appear to merge with the outer limit of the contact aureole (Fig. 4-5). Yet Reesor (1973) noted that in this area the contact and regional assemblages remain distinct compositionally and texturally. The regional metamorphic porphyroblasts include staurolite and kyanite that contain rotated inclusions, and the contact metamorphic porphyroblasts include staurolite, andalusite and cordierite whose inclusions are parallel to the schistosity (Fyles, 1964; Reesor, 1973). These relationships show that the contact metamorphism occurred following the regional metamorphism and deformation, and that it occurred at lower pressure (2.5 - 4.3 kbar vs. greater than 5.6 kbar; Archibald et al., 1984; Carmichael, *unpublished*;). The "deflection" of the isograds, coupled with the upward deflection of bedding and schistosity in the same area and around other plutons, may then suggest that these post-kinematic plutons were emplaced diapirically into relatively stiff, brittle country rock, thus warping the previous regional metamorphic isograds as well as the regional foliation close to the batholith margins (Fig. 4-6b).

In contrast, the syn-kinematic plutons were intruded at greater depths (4.3 - 5.6 kbar; Archibald et al., 1984; Carmichael, *unpublished*), and they were intruded without regional deflection of the country rock, which instead shows ductile, isoclinal folding and shearing within a few meters to tens of meters from the pluton margins (Fig. 4-6a). The anomalous structural relationships along the northwestern margin of the post-kinematic Fry Creek batholith are more similar to relationships along the margins of the syn-kinematic plutons and contribute to subsequent arguments that the western portion of the Fry Creek batholith has been exhumed from deeper crustal levels than the eastern portion.

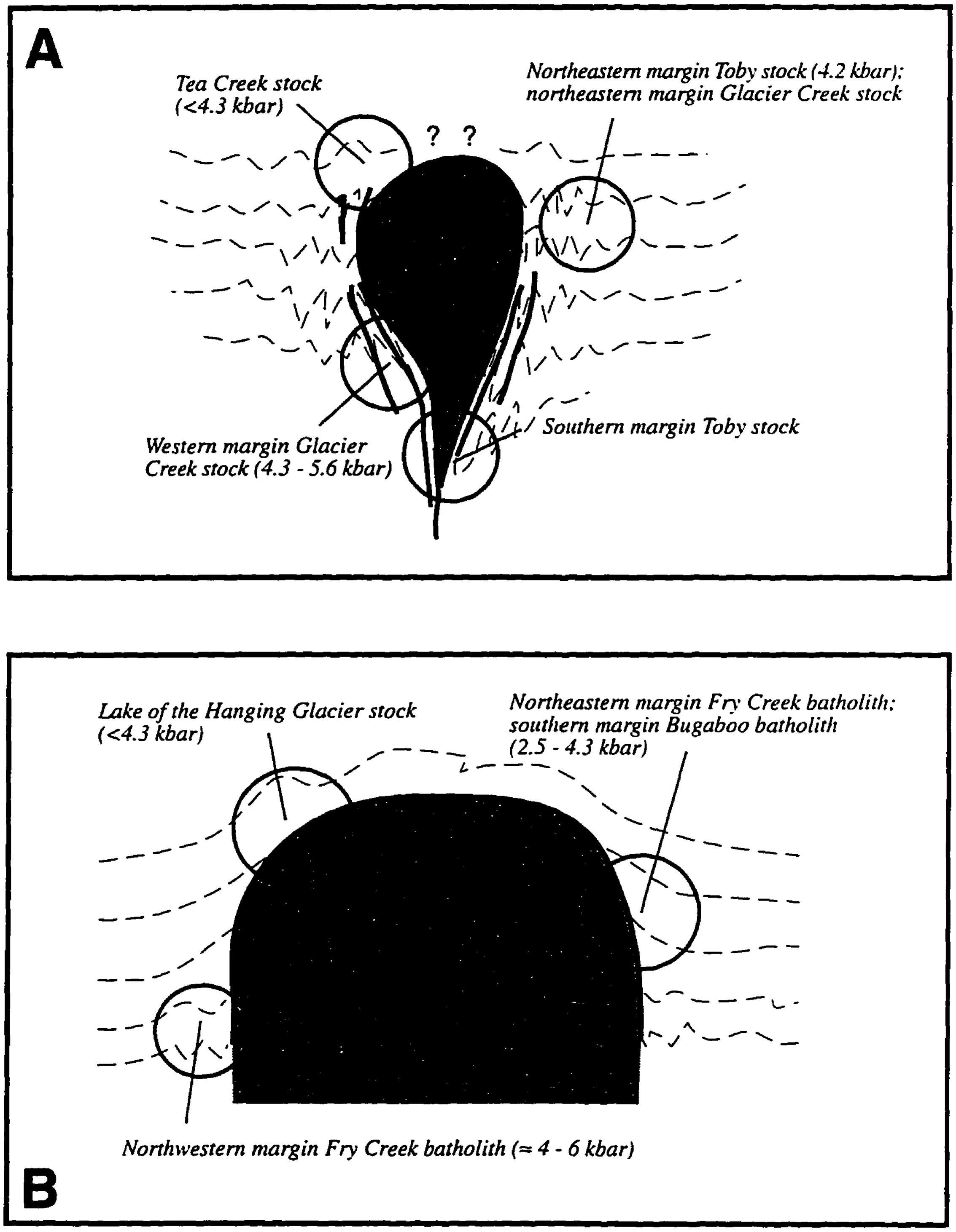



**Figure 4-6**: Schematic diagram showing structural relationships at the margins of: A) syn-kinematic plutons; and B) apparently post-kinematic plutons. Anisotropy in country rock represents bedding. Relationships with map-scale folds not shown (see Plates 2 and 3). Syn-kinematic plutons were emplaced generally at deeper crustal levels (higher pressure) into more ductile country rock. Post-kinematic plutons were emplaced generally to shallower crustal levels (lower pressure) into stiffer country rock. See text for discussion of pressure data and Figure 4-5 for locations and sources of pressure data.

## NEW METAMORPHIC PRESSURE DATA

Metamorphic pressure data from the study area (Figs. 4-5, 4-7 to 4-10) show that regional metamorphism, syn-kinematic contact metamorphism and post-kinematic contact metamorphism occurred at progressively lower pressures, respectively. These data include qualitative pressure data (Archibald et al., 1983, 1984; and this study) from appropriate metamorphic mineral assemblages (bathozones of Carmichael, 1978) and quantitative thermobarometric data (discussed below) obtained from microprobe analyses (Kells, 1993, and this study). Pressure limits for bathozones as originally defined by Carmichael (1978) have been modified according to a revised petrogenetic grid for pelitic rocks (D. M. Carmichael, unpublished; see Fig. 4-7).

## QUANTITATIVE THERMOBAROMETRY: ANALYTICAL METHODS AND DATA REDUCTION

Several specimens of metamorphic rock were selected for quantitative thermobarometric study of regional and contact metamorphism. Primarily garnet-grade pelitic specimens were selected, for application of the garnet-biotite geothermometer and the garnet-plagioclase-muscovite-biotite-quartz geobarometers (Berman, 1991, and references therein). Analytical data are presented in Appendix 2.

Mineral analyses were performed at Queen's University using an ARL-SEMQ electron microprobe with an energy-dispersive spectrometer (EDS). Operating conditions were maintained at an accelerating voltage of 15 kV and a beam current of 75 nA. Glass NBS 470 (National Bureau of Standards) was used as primary standard for analyses. Mineral compositions listed in Appendix 2 represent an average of three analyses, each having a count time of 100 seconds, from different points within a single grain or from neighboring grains of the same mineral species.

Data reduction was completed according to the procedure of Bence and Albee (1969). Structural formulae and ideal activities were computed with MINPROBE, an unpublished APL software package developed by D. M. Carmichael at Queen's University. P-T phase diagrams

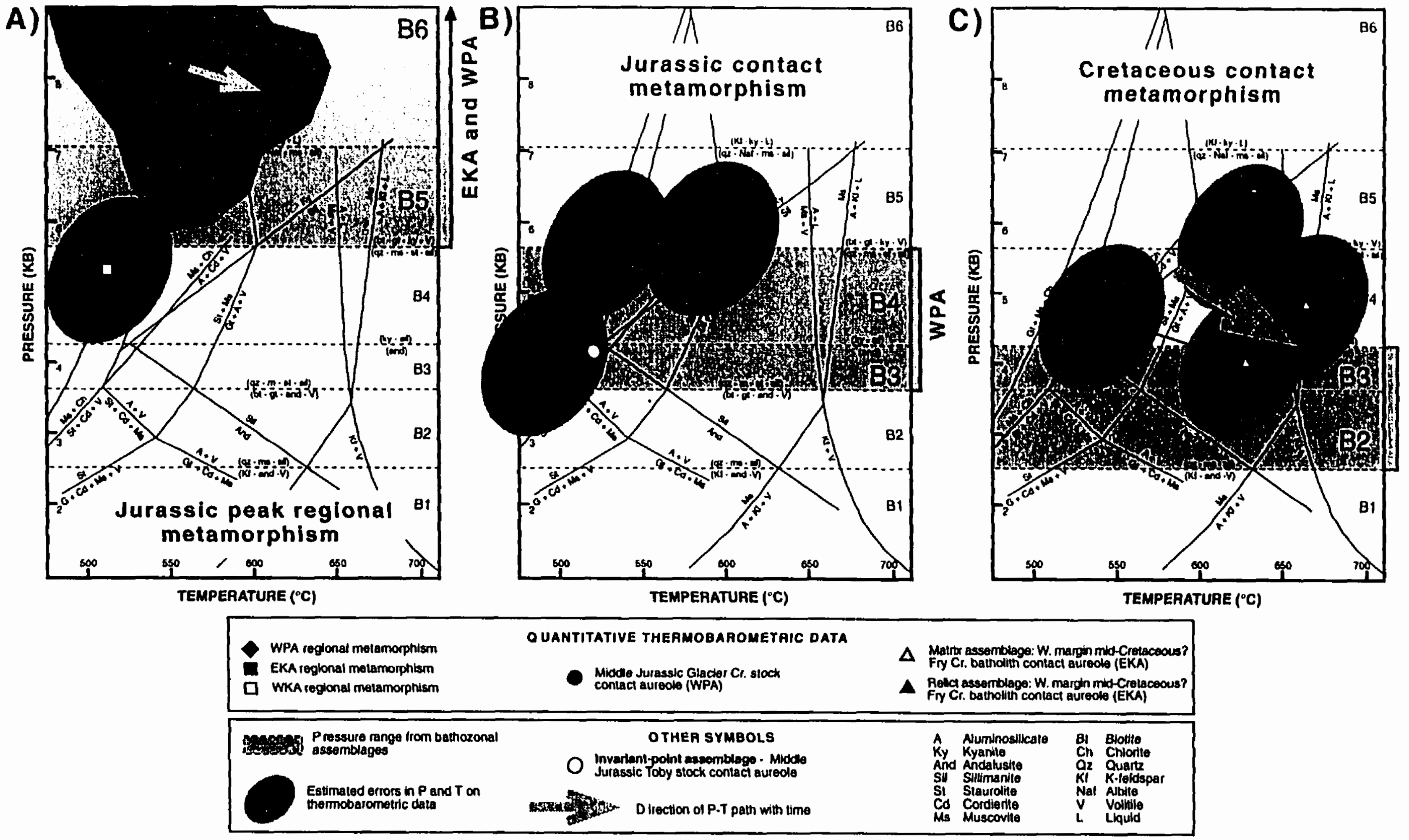



**Figure 4-7:** P-T petrogenetic grids for part of the model pelitic system SiO2-Al2O3-FeO-MgO-Na2O-K2O-H2O (D.M. Carmichael, unpublished), showing qualitatively and quantitatively determined P-T conditions for rocks in the eastern Kootenay Arc and western Purcell anticlinorium during: A) Middle Jurassic regional metamorphism; B) Middle to Late Jurassic contact metamorphism; C) mid-Cretaceous contact metamorphism. Assemblages in parentheses define bathozones (modified after Carmichael, 1978), shown as horizontal dotted lines. Ranges of pressure conditions determined from bathozonal assemblages (Reesor, 1973; Archibald et al., 1983, 1984; Kells, 1993 and this study) are shown by brackets at right. P-T values for Glacier Creek stock contact aureole calculated from data in Kells (1993). See text for discussion of uncertainties in quantitative data. WPA = western Purcell anticlinorium; EKA = eastern Kootenay Arc; WKA = western Kootenay Arc (hinge region of Meadow Creek anticline).

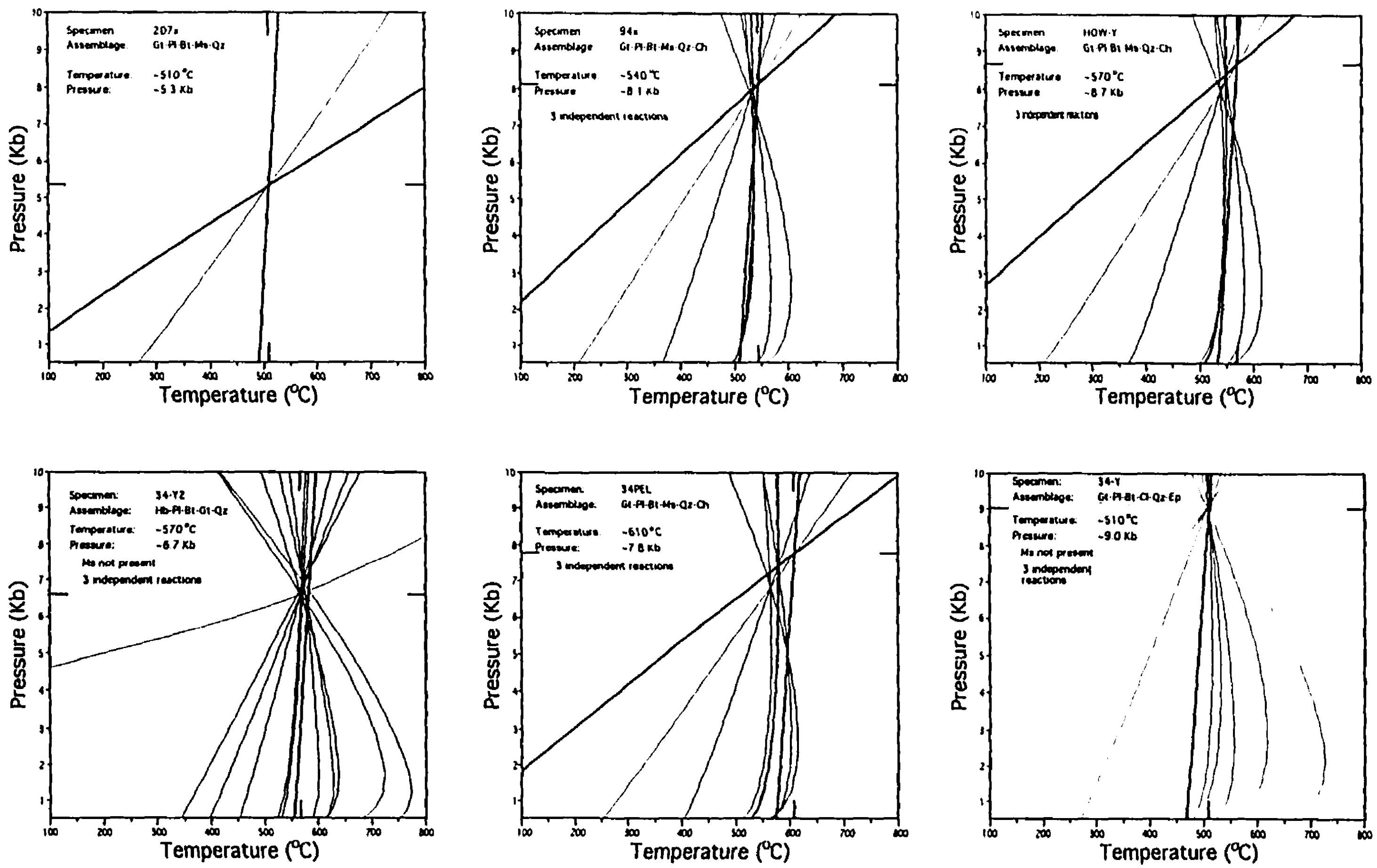



**Figure 4-8:** P-T diagrams for regionally metamorphosed pelitic and semi-pelitic rocks of the Kootenay Arc and westernmost Purcell anticlinorium. (A) Pelitic schist, hinge zone of Meadow Creek anticline near Meadow Creek. (B) Pelitic schist, 4 km east of Kootenay Arc boundary fault (Lavina Ridge). (C) Pelitic schist, 2 km east of Kootenay Arc boundary fault (lower Howser Creek). (D - F) Specimens from intercalated schistose quartzite and pelite, 500 m west of Kootenay Arc boundary fault (near Duncan Lake); (E) is pelitic schist. Diagrams generated with TWQ software (Berman, 1991) using data given in Appendix 2 and solid-solution models of Fuhrman and Lindsley (1988) for feldspars, of Berman (1990) for garnets and of Mader and Berman (1992) for hornblende. P-T values given in diagrams computed from the intersection of the garnet-biotite thermometer and the garnet-plagioclase barometers (thermometer and least steep barometer highlighted). Note that mineral assemblages in (D) and (F) lack muscovite and thus do not permit application of the garnet-plagioclase barometers; these P-T values are graphic estimates of reaction intersections. See text for interpretation and discussion of uncertainties in P-T values. Calculations assume activity of H2O = 1.

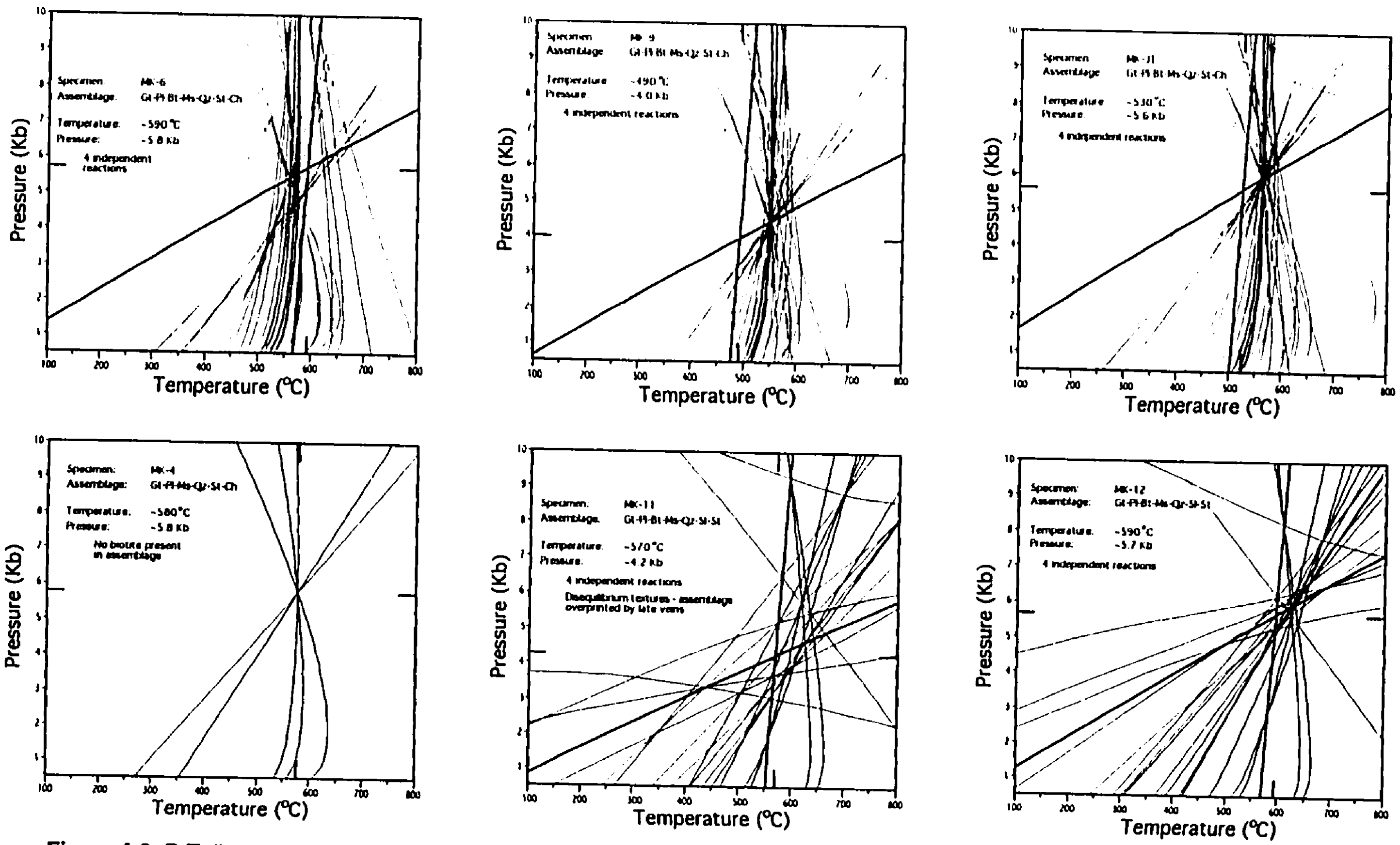



**Figure 4-9:** P-T diagrams for pelitic rocks from the contact aureole of the syn- to late-kinematic Glacier Creek stock (southwestern margin of pluton). Specimens MK-4 through MK-12 were collected by D. M. Carmichael on a transect of the contact aureole (1 km south of Glacier Creek) from the outer staurolite zone (MK-4) to the inner sillimanite zone (MK-12). Specimen J-1 collected by M. J. Warren from the inner staurolite zone (adjacent to Glacier Creek). Specimens analysed by M. P. Kells. Diagrams generated with TWQ software (Berman, 1991) using data from Kells (1993), given in Appendix 2, and solid-solution models of Fuhrman and Lindsley (1988) for feldspars and of Berman (1990) for garnets. P-T values given in diagrams computed from the intersection of the garnet-biotite thermometer and the garnet-plagioclase barometers (thermometer and least steep barometer highlighted). Note that mineral assemblage in Specimen MK-4 lacks biotite and thus does not permit application of the garnet-biotite thermometer or garnet-plagioclase barometers; P-T value is intersection of reactions involving gt, pl, st, qz, ch and water. See text for interpretation and discussion of uncertainties in P-T values. Calculations assume that activity of $H_2O = 1$.

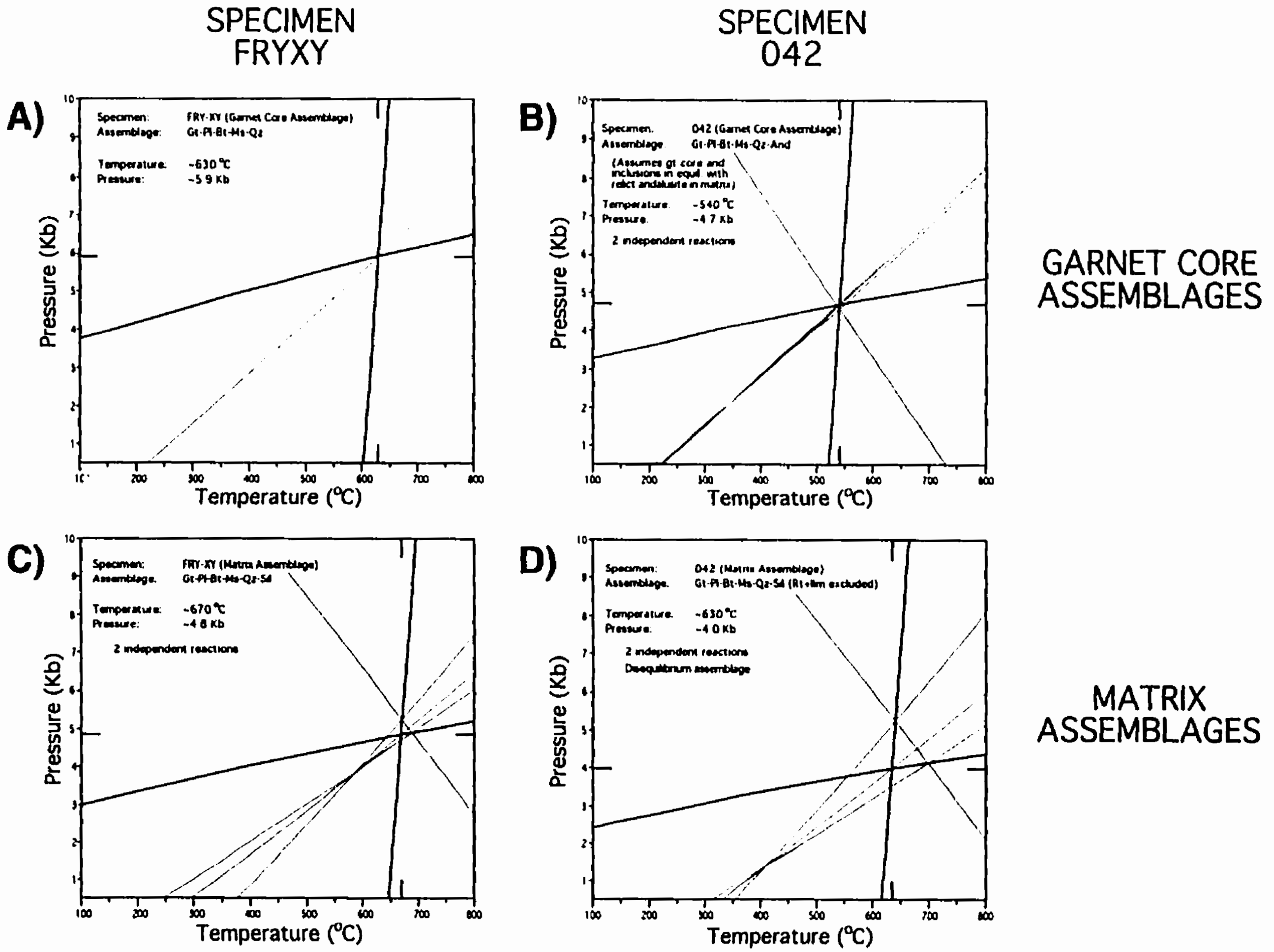



**Figure 4-10:** P-T diagrams for pelitic rocks from the contact aureole of the late- to post-kinematic Fry Creek batholith (northwestern margin, lower Fry Creek). Specimen FRY-XY (A and C) collected within a few m of the batholith; Specimen 042 (B and D) collected within a few tens of m at same locality. Diagrams generated with TWQ software (Berman, 1991) using data given in Appendix 2, and solid-solution models of Fuhrman and Lindsley (1988) for feldspars and of Berman (1990) for garnets. P-T values given in diagrams computed from the intersection of the garnet-biotite thermometer and the garnet-plagioclase barometers (thermometer and least steep barometer highlighted). See text for interpretation and discussion of uncertainties in P-T values. Calculations assume activity of H2O = 1.

(Figs. 4-8 to 4-10, and Appendix 2) were generated with the TWQ software of Berman (1991) using ideal activities computed by MINPROBE and the thermodynamic database of Berman (1988, 1990). Tightly clustered reaction intersections on the phase diagrams are consistent with a postulated close approach to equilibrium; scattered intersections indicate disequilibrium. Errors in pressure and temperature are estimated at about +/- 1.0 - 1.5 kbar and +/- 30 - 50 °C for pelitic specimens containing the assemblage gar-plag-bt-ms-qtz; these uncertainties increase with additional phases or with different metamorphic mineral assemblages (Berman, 1991; Gordon et al., 1994). Pressures and temperatures obtained from pelitic specimens from the contact aureole of the Glacier Creek stock were re-calculated from data obtained by Kells (1993) at Queen's University, using the same analytical methods.

*Regional metamorphic assemblages*

Data from six regional metamorphic specimens indicate that peak metamorphic conditions in the westernmost Purcell anticlinorium and the easternmost Kootenay Arc occurred at pressures of 5.3 to 9 kbar (Figs. 4-5, 4-7 and 4-8). Previously published qualitative metamorphic data (Reesor, 1973; Archibald et al., 1984) had indicated only that peak regional metamorphism occurred at pressures of greater than 5.6 kbar (bathozone 5 or higher of Carmichael, 1978). In all six specimens the growth of metamorphic porphyroblasts overlapped with the development of the locally dominant foliation.

Thermobarometric results obtained from a garnet-plagioclase-quartz-muscovite-biotite-chlorite schist from the Horsethief Creek Group (specimen HOW-Y) in the footwall (east side) of the Kootenay Arc boundary fault indicate that regional metamorphism occurred at 8.7 kbar (Fig. 4-8C). Both petrographic textures and comparison of garnet and plagioclase core and rim compositions suggest that this rock equilibrated on a retrograde metamorphic path. Thermobarometric results obtained from an equivalent metamorphic mineral assemblage in a pelitic schist from the lower Hamill Group that is exposed a few kilometers to the west (Specimen 34PEL), in the immediate hangingwall (west side) of the Kootenay Arc boundary fault, indicate that regional metamorphism occurred at 7.8 kbar (Fig. 4-8E). The results shown in Fig. 4-8D and F were obtained from specimens at the same locality (34 Y-2 and 34-Y), but Figure 4-8D and F show results computed from mineral assemblages that lack muscovite and thus are not appropriate for application of the garnet-plagioclase-muscovite-biotite-quartz geobarometer. Therefore, the pressure of 7.8 kbar obtained from pelitic specimen 34PEL (Fig. 4-8E) is considered the most reliable of the three and most appropriate for comparison with other results. Specimen 94x (Fig. 4-8B) is from the same unit of pelitic schist in the lower Hamill Group, exposed along strike to the south, also in the immediate hangingwall of the Kootenay Arc boundary fault. The metamorphic mineral assemblage garnet-plagioclase-quartz-muscovite-biotite-chlorite has yielded a pressure of 8.1 kbar. Thus the three pressures obtained from pelitic schists in the immediate hangingwall and footwall of the Kootenay Arc boundary fault are the same within error. The implications of this result are discussed below.

A pressure of 5.3 kbar was obtained from the hinge region of the Meadow Creek anticline (Fig. 4-8A). This pelitic schist has been juxtaposed against higher-pressure regional metamorphic rocks to the east by at least two significant, late- to post-metamorphic west-dipping normal faults (Plate 4a). However, the combined displacement on the two normal faults shown on Plate 4a and on Plate 2 (about 5 km) accounts for only about half of the total depth difference of about 10 km implied by the differences in pressure data.

*Contact metamorphic assemblages: syn-kinematic Glacier Creek stock*

Previous qualitative pressure data from pelitic metamorphic mineral assemblages in the contact aureoles of the Glacier Creek and Toby stocks indicate that these plutons were intruded into bathozones 3 or 4, at pressures of 3.7 - 5.6 kb (Archibald et al., 1983; Kells, 1993; Figs. 4-5 and 4-7B). The Cauldron stock was intruded at the same range of pressure conditions, but the small, unnamed granodiorite stock in the Tea Creek drainage was intruded into bathozones 2 or 3, at pressures of 2.5 - 4.3 kbar. A pelitic schist from the contact aureole of the Toby stock (Specimen 506.2) contains the assemblage quartz-plagioclase-biotite-muscovite-staurolite-kyanite-andalusite-chlorite, which plots on an invariant point on a petrogenetic grid for pelitic rocks D. M. Carmichael , unpublished). This assemblage indicates that this stock intruded at 4.2 kbar (Figs. 4-5 and 4-7B).

Quantitative geothermobarometry on six pelitic specimens that represent a transect through the Glacier Creek stock contact aureole (Kells, 1993) indicate pressures of 4.0 - 5.8 kbar (Fig. 4-9) and are in close agreement with the upper limit of pressure obtained from qualitative data (Fig. 4-7B). In all six specimens, contact metamorphic porphyroblasts have overgrown or were synchronous with the regional foliation, but they are commonly rotated or overprinted by a younger retrograde foliation or crenulation. Specimens MK-4, MK-6, MK-9 and MK-J1 (Fig. 4-9A-D) are from the outside to the inside of the staurolite+chlorite zone of the aureole. Specimens MK-11 and MK-12 (Fig. 4-9E-F) are from the staurolite+sillimanitie zone of the contact aureole that is adjacent to the pluton. The specimens are discussed further in Kells (1993). A pressure of 5.7 kbar and a temperature of 590°C obtained from a specimen from the sillimanite zone 300m from the pluton are considered most reliable (Specimen MK-12; Fig. 4-9F), because the assemblage in this specimen is least overprinted by retrograde metamorphism and by younger veining, and because the intersection of the garnet-biotite thermometer and garnet-plagioclase barometers is in close agreement with other reaction intersections, indicating a record of conditions that approached equilibrium. Pressures and temperatures obtained from specimens MK-4, MK-6 and MK-J1 agree closely with those obtained from specimen MK-12. However, the mineral assemblage in specimen MK-4 lacks biotite and thus is considered less reliable because the garnet-biotite thermometer and the garnet-plagioclase-muscovite-biotite-quartz barometer cannot be applied. Both textural evidence (strong alteration of minerals, late veinlets) and the analytical results (scattered reaction intersections) from specimen MK-11 clearly indicate disequilibrium conditions, and these results are considered unreliable. The contact metamorphic mineral assemblage in specimen MK-9 is synchronous with a strong schistosity that includes retrograde chlorite, and it is possible that the significantly lower pressure and temperature (4.0 kbar, 490°C) obtained from this rock are due to partial re-equilibration of the contact metamorphic assemblage along a retrograde path at some time after the emplacement of the pluton, but during continuing regional metamorphism and deformation.

*Post-kinematic Fry Creek batholith (northwest margin)*

Contact metamorphic mineral assemblages from the contact aureoles of the post-kinematic plutons in the western Purcell anticlinorium indicate that the plutons intruded into bathozones 2 or 3, at pressures of 2.5 - 4.3 kbar (Figs. 4-5 and 4-7C). However, the northwestern margin of the Fry Creek batholith appears to have intruded country rock that was far more ductile than the rock intruded by the same batholith in the Purcell anticlinorium, and contact and regional metamorphism are more difficult to distiguish along the northwestern margin of the pluton, suggesting perhaps a greater depth of emplacement on this side.

Two pelitic specimens were selected for application of quantitative thermobarometry from a sillimanite zone that is confined to rocks that lie adjacent to the northwest margin of the Fry Creek batholith. Specimen 042 was collected from a pelitic schist in the Hamill Group, about ten meters from the northwestern margin of the Fry Creek batholith. The specimen contains a single large garnet that contains inclusions of quartz, muscovite, biotite and plagioclase. The dominant foliation deflects around this garnet and is defined by quartz, muscovite, biotite, plagioclase, ilmenite and pseudomorphs of fine white mica that were perhaps relict andalusite. Retrograde chlorite has overgrown the prograde assemblage in the foliation and has replaced some of the garnet. Quantitative thermobarometry was applied to the assemblage in the core of the garnet, using analyses from the core of the garnet and from the inclusions, and it yielded a pressure of 4.7 kbars and a temperature of 540°C. (The P-T phase diagram in Fig. 4-10B was generated by assuming that possible relict andalusite in the matrix had been in equilibrium with the garnet core and its inclusions; calculations that exclude andalusite yielded the same P-T values) Results of thermobarometry on the matrix assemblage, using analyses from the rim of the garnet and from matrix minerals, indicate disequilibrium at perhaps lower pressure and higher temperature (Fig. 4-10D).

Specimen FRY-XY was collected a few meters from the Fry Creek batholith, from the same pelitic schist and near the same locality as specimen 042. The foliation is defined by sillimanite, quartz, plagioclase, ilmenite, muscovite and biotite. The foliation deflects around garnet porphyroblasts that contain rotated inclusion trails of plagioclase and quartz. Retrograde chlorite is not as abundant as in specimen 042. Thermobarometry was applied to the assemblage quartz-muscovite-biotite-plagioclase-garnet, using analyses from the core of the garnet, from plagioclase inclusions and from matrix biotite and muscovite. The results yielded a pressure of 5.9 kbars and a temperature of 630°C (Fig. 4-10C). Because the geobarometer used is dependent on Ca-Mg exchange between garnet and plagioclase (Ghent and Stout, 1981), the difference in pressures are considered meaningful, even though analyses of muscovite and biotite inclusions from the garnet core were not obtained. Quantitative thermobarometry on the matrix assemblage quartz-muscovite-sillimanite-biotite-plagioclase-garnet yielded a pressure estimate of 4.8 kbars and a temperature of 675°C (Fig. 4-10C). Thus this rock was being unroofed and heated during the growth of the garnets.

These pressures are significantly higher than the qualitative data (Bathozones 2 to 3; 2.5 - 4.3 kb) obtained from the contact aureole to the east (Fig. 4-5), and they also indicate that the rock was being exhumed at the time that the pluton was emplaced. The differences in pressures from east to west are interpreted to reflect differential post-emplacement tilting and regional uplift (west side up). Post-regional metamorphic tilting and normal faulting might also explain the distribution of regional metamorphic isograds (Fig. 4-5), which are closely spaced on the western flank of the metamorphic culmination in the Kootenay Arc and more widely spaced on the eastern flank of the metamorphic culmination in the Purcell anticlinorium.

# REGIONAL CONSTRAINTS ON TIMING AND CONDITIONS OF MESOZOIC DEFORMATION

## EMPLACEMENT OF QUESNEL AND/OR SLIDE MOUNTAIN TERRANES

The eastward emplacement of Quesnel and Slide Mountain terranes over the distal part of the former North American margin occurred between 187 and 168 Ma. The Hall Formation of the Rossland Group represents the latest pre-orogenic or earliest syn-orogenic (flysch) sediments that were deposited on Quesnel terrane. These strata contain Toarcian fauna (Tipper, 1984), which dates them at 190-187 Ma (time scale of Gradstein et al., 1994). U-Pb zircon ages of 183 Ma to 178 Ma have been obtained from plutons that are synkinematic with respect to east-verging deformation in eastern Quesnel terrane, west of northern Kootenay Lake (Fig. 4-2; Klepacki, 1985) and farther to the south near the International Border (Andrew et al., 1990; T. Höy, pers. comm., 1996). The east-verging thrust faults that place Slide Mountain and Quesnel terranes over the North American rocks to the west of Duncan Lake are intruded by the Kuskanax batholith (Fig. 4-2; Read, 1973; Klepacki, 1985), which has yielded a U-Pb zircon age of 173 +/- 5 Ma (Parrish and Wheeler, 1983), thus placing a younger limit on the time of emplacement of these terranes.

Quesnel and/or Slide Mountain terranes must have overlapped the former North American passive margin at least as far east as the rocks that host the Toby stock, because estimates of metamorphic pressures obtained from the contact aureoles of this and other syn-kinematic plutons (Glover, 1978; Archibald et al., 1983; Kells, 1993, and this study) in the western Purcell anticlinorium indicate that there was a minimum of about 15 km of material above the Horsethief Creek Group at the time of their emplacement. This amount of burial far exceeds what might reasonably be expected for the thickness of deep-water Phanerozoic rocks that were deposited on this distal segment of the continental margin. Therefore, the burial must have been tectonic. The tectonic burial is interpreted as the result of eastward thrusting of Slide Mountain and/or Quesnel terranes, which has been well documented farther north along strike in the Cariboo Mountains (Struik, 1988). This interpretation is also compatible with interpretation of seismic data (Cook et al., 1992) suggesting that Quesnel terrane is a tectonic slab that is only 12-15 km thick.

## THE KOOTENAY ARC

The relationships between the felsic dikes and the Meadow Creek anticline indicate that the west-verging deformation in the Kootenay Arc was under way by 173 +/- 5 Ma. This conclusion is supported by similar relationships between west-verging deformation and the Kuskanax batholith (also 173 +/- 5 Ma; Parrish and Wheeler, 1983), which was emplaced prior to or during the early stages of the west-verging folding (Read, 1973). The younger limit of the west-verging deformation is poorly constrained by relationships in the Kootenay Arc, but Colpron et al. (1996) showed that along strike in the northern Selkirk Mountains the west-verging deformation ended by about 168 Ma.

The upright folding in the Kootenay Arc occurred between an older limit of 173 +/- 5 Ma and 164 Ma, the age of a post-tectonic part (Mt. Carlyle stock) of the Nelson batholith (Fig. 4-2; Read and Wheeler, 1975; and Archibald et al., 1983). The overlap in age between prograde regional metamorphism, felsic dike emplacement and both phases of deformation, discussed previously, strongly suggest that there was little if any time gap between development of the west-verging folds and of the younger, coaxial upright folds and their axial planar schistosities. Thus, most of the penetrative deformation in the Kootenay Arc is bracketed between 173 +/- 5 Ma and 164 Ma, and it is reasonable to conclude that it represents one continuous episode of deformation.

This conclusion conflicts with the hypothesis that the large-amplitude, west-verging folds in the Kootenay Arc were produced by a Lower Paleozoic orogenic event (Read, 1973). A conglomerate at the base of the Mississippian Milford Group contains, at several localities, previously foliated clasts which are interpreted to have been eroded from the unconformably underlying Lower Paleozoic Lardeau Group (Read and Wheeler, 1976; D. Klepacki, pers. comm., 1997). These relationships, coupled with the apparent restriction of the west-verging folds to the pre-Mississippian strata, are the evidence cited to support the hypothesis that the west-verging structures formed in Paleozoic time.

However, it should be noted that the Milford Group everywhere rests on the uppermost part of the Lardeau Group (Read and Wheeler, 1976). It rests either on the Broadview Formation or, in a few localities, on metavolcanic rocks which are similar to the underlying Jowett Formation, but which may be part of the Broadview Formation. If the Lardeau Group had been deformed by isoclinal folds with 10 km of amplitude prior to the deposition of the Milford Group, then surely the sub-Milford unconformity would have overlapped substantial relief, and would now rest on different stratigraphic levels in the Lardeau Group and perhaps older units. It should also be noted that by Late Devonian to Mississippian time, island arc terranes had developed off the western margin of Laurentia (e.g. Mortensen et al., 1981; Schweickert et al., 1984; Poole et al., 1992), and therefore, foliated clasts of deep-water strata in the Mississippian Milford Group might have been derived from a metamorphic terrane that was exposed to the west. Thus the sub-Milford unconformity may well record a mid-Paleozoic orogenic event related to the well-documented Antler orogeny to the south, as has been proposed by Smith and Gehrels (1989) and Smith et al. (1993), but the west-verging folds in the Kootenay Arc are more readily explained as a consequence of Mesozoic, rather than Paleozoic, deformation.

## THE PURCELL ANTICLINORIUM

Plutons that intrude the western Purcell anticlinorium provide critical constraints on the timing and conditions of deformation of the country rock strata. The synkinematic Mine stock (Fig. 4-2), which intrudes strata equivalent to the Horsethief Creek and Hamill Groups on the western flank of the Purcell anticlinorium south and west of Kootenay Lake (Glover, 1978), has been dated by a U-Pb zircon upper intercept of 171 Ma and concordant muscovite and biotite K-Ar cooling dates of 166 Ma (Archibald et al., 1983). These dates bracket the time of emplacement and indicate that penetrative deformation and regional metamorphism was underway in the western flank of the Purcell anticlinorium by about 170 Ma. The Horsethief Creek batholith cuts the Mount

Forster thrust and has yielded a K-Ar biotite date of 95 Ma (Archibald et al., 1984) and a $^{40}Ar/^{39}Ar$ muscovite date of 93 Ma (D. A. Archibald, unpublished data; Fig. 4-11). The White Creek batholith cuts the Hall Lake fault and has yielded a K-Ar hornblende date of 89 Ma (Wanless et al., 1968). These data place a younger limit on both the out-of-sequence thrusts and on the regional folds and cleavage that are cut by these faults.

New 40Ar/39Ar step-heating data from the syn- to late-kinematic Toby and Glacier Creek stocks (Warren, unpublished data; Appendix 1) provide additional constraints on the cooling history of the western Purcell anticlinorium (Fig. 4-11) and, indirectly, better constraints on the timing of the termination of prograde regional metamorphism and penetrative deformation in the Purcell anticlinorium. A plateau date of 157 +/-1 Ma from muscovite in the Glacier Creek stock contact aureole and a biotite date of 160 +/- 2 Ma from the Toby stock place a younger limit on the initiation of deformation and regional metamorphism and imply that these rocks had cooled below about 350°C and 275-300°C, respectively, by these dates (McDougall and Harrison, 1988, and references therein). The regional metamorphic assemblages in pelitic country rocks indicate that penetrative deformation occurred in the chlorite, biotite and garnet zones of metamorphism, or at temperatures of at least 350°C at the range of pressure at which the plutons intruded (e.g. Yardley, 1989; p. 50, 94). Thus it is reasonable to propose that the regional metamorphism and the formation of the axial planar cleavage that is defined by the regional metamorphic assemblages had ceased by 159 Ma in the rocks which are intruded by Toby stock, and by 157 Ma in rocks which are intruded by Glacier Creek stock. The fact that the biotite date from Toby stock is older than the muscovite date from Glacier Creek stock could be explained by the fact that Toby stock was emplaced at lower pressure (4.2 kbar) into lower-grade rocks (chlorite and biotite zones) than Glacier Creek stock (emplaced at as much as 5.8 kbar, in biotite and garnet zones), and so was uplifted and cooled through a lower temperature earlier.

Additional $^{40}Ar/^{39}Ar$ biotite age spectra from Glacier Creek stock (located on Fig. 4-11) have yielded integrated cooling ages of 146 and 119 Ma, but they are significantly disturbed spectra

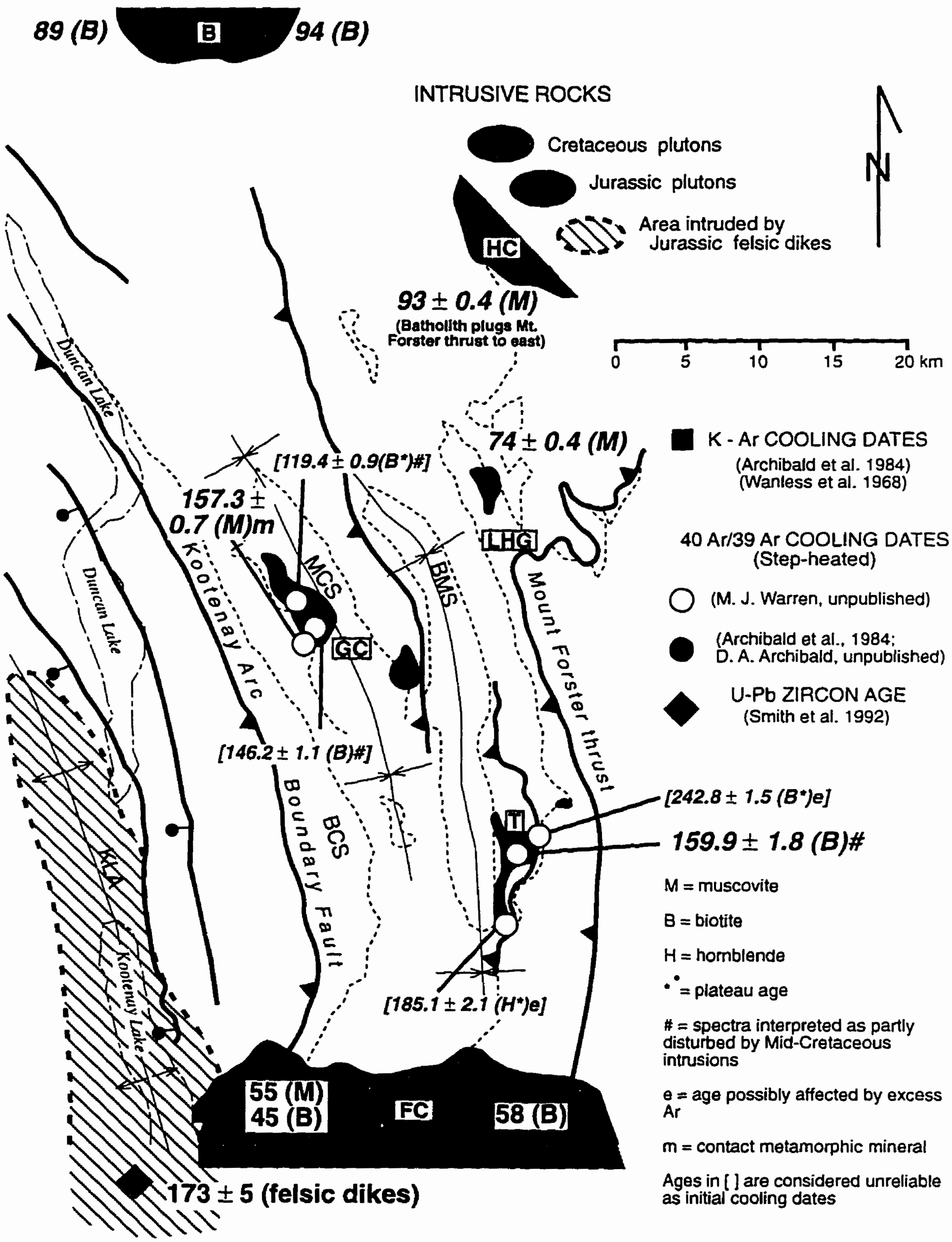



**Figure 4-11:** Simplified map of the study area showing locations for new and previously published geochronological data. New data from this study are presented and discussed in Appendix 1. **B** = Bugaboo batholith; **HC** = Horsethief Creek batholith; **LHG** = Lake of the Hanging Glacier stock; **GC** = Glacier Creek stock; **T** = Toby stock; **FC** = Fry Creek batholith; **BCS** = Birnam Creek syncline; **MCS** = Mt. Cauldron syncline; **BMS** = Blockhead Mtn. syncline; **KLA** = Kootenay Lake antiform. Contacts (thin dotted lines) are the base of the Hamill Group in the Birnam Creek, Mt. Cauldron and Blockhead Mtn. synclines and the base of the Windermere Supergroup in the immediate hangingwall of the Mt. Forster thrust.

that indicate different degrees of thermal overprinting by mid-Cretaceous magmatism (M. Warren, unpublished data; Appendix 1).

## THE MAIN RANGES OF THE ROCKY MOUNTAINS, FRONT RANGES, AND FOOTHILLS

The time of initation of deformation in the Main Ranges of the Rocky Mountains is poorly constrained. Thrusting and folding in the Main Ranges of the Rocky Mountains has been correlated with tectonic subsidence and sediment accumulation in the foreland basin (e.g. Price and Mountjoy, 1970; Price, 1994). The oldest westerly-derived syn-orogenic sediments in the foreland basin are the fine-grained clastic rocks of the Kimmeridgian Passage beds (Upper Jurassic; 154 - 151 Ma) of the Fernie Group (Poulton, 1984). The oldest syn-orogenic sediments that contain recognizable fragments of Paleozoic miogeoclinal strata are the uppermost Jurassic to Lower Cretaceous Kootenay Group (Price and Mountjoy, 1970). The western limit of these strata is contained in thrust sheets of the Front Ranges of the Rocky Mountains. The deformation on the western flank of the Porcupine Creek anticlinorium in the Western Main Ranges of the Rocky Mountains can be linked kinematically with the deformation in the Dogtooth Range of the Purcell anticlinorium, which pre-dates the Purcell thrust (Kubli and Simony, 1994). Faults that branch from the Purcell thrust include the Mount Forster, Hall Lake and St. Mary faults, all of which are cut by mid-Cretaceous plutons (e.g. Archibald et al., 1984). Thus at least some of the deformation in the Western Main Ranges is pre-mid-Cretaceous and is likely as old as Latest Jurassic.

Strata preserved in the footwalls of the Bourgeau and Lewis thrusts indicate that these major structures of the Eastern Main Ranges and Front Ranges were active during Late Cretaceous time (Price, 1994, and references therein). The youngest sediments deformed by the triangle zone in the foothills at the front of the foreland thrust and fold belt are the Paleocene (63 - 58 Ma) Paskapoo and Porcupine Hills formations (Dawson et al., 1994).

## TIMING OF MOTION ON THE BASAL DÉCOLLEMENT OF THE OROGEN

The Proterozoic and Paleozoic North American supracrustal strata that comprise the Kootenay Arc, the Purcell anticlinorium and the Main Ranges of the Rocky Mountains are all allochthonous with respect to their North American crystalline basement (Bally et al., 1966; Price and Mountjoy, 1970; Price, 1981). West-dipping thrust faults in the thrust and fold belt flatten at depth and "root" into a basal décollement at or near the contact between the supracrustal rocks and the underlying crystalline basement. In the Omineca Belt, which is the metamorphic "hinterland" of the orogen, Eocene extension has resulted in exhumation of the Monashee décollement, a mid-crustal, east-verging shear zone that juxtaposes the distal part of the North American continental margin over the North American basement and its thin, autochthonous sedimentary cover sequence that comprises the Monashee Complex (Brown, 1981; Brown et al., 1992; Fig. 4-2). Isotopic dating of deformed and undeformed pegmatites along the Monashee décollement shows that east-verging thrusting ceased by 59 - 58 Ma (Carr, 1992; Parrish, 1995). This matches the age of the youngest sediments preserved in the foreland basin and supports the interpretation that the Monashee décollement is the deeper and more westerly extension of the detachment beneath the foreland thrust and fold belt (Parrish et al., 1988; Carr, 1991; Cook et al., 1992; Parrish, 1995).

Rocks in the hangingwall of the Monashee décollement record a protracted and apparently continuous history of regional metamorphism and deformation from 173 Ma to 58 Ma (Parrish, 1995, and references therein). Younger deformation and younger peak metamorphic conditions are recorded with increasing paleo-depth in the crust (e.g. Carr, 1991; Scammell, 1993; Leclair et al., 1993; Parrish, 1995). Thus, in a vertical cross-section through the hangingwall of the Monashee décollement, rocks at higher crustal levels cooled and "froze" earlier as they were moved eastward and upward on the basal décollement. Similarly, in a horizontal cross-section through the hangingwall of the Monashee décollement, rocks in the west "froze" earlier and were then carried passively as the deformation front above the basal décollement propagated eastward from beneath the Kootenay Arc, the Purcell anticlinorium and then beneath the Rocky

Mountains. The thermal history of the exposed footwall of the Monashee décollement is interpreted to record burial beneath a foreland basin from prior to ca. 135 Ma to around 100 Ma, followed by tectonic loading beneath the advancing thrust belt that culminated in maximum tectonic burial in Latest Cretaceous to Paleocene time (Parrish, 1995). The documentation of continuous deformation in the allochthonous strata and of the timing of the advance of the thrust belt over this outlier of apparently autochthonous basement in the footwall is critical to the discussions of shortening rates and orogenic dynamics that are presented below.

## DISCUSSION

### DISPLACEMENT HISTORY AND TECTONIC INHERITANCE OF THE KOOTENAY ARC BOUNDARY FAULT

Geobarometric data from regional metamorphic assemblages on either side of the Kootenay Arc boundary fault (Figs. 4-5 and 4-8) indicate that there was no significant post-peak metamorphic vertical motion on this fault. At least some of the motion on this fault occurred at amphibolite facies conditions. Hornblende and biotite in metabasite define a strong subhorizontal lineation within a few hundred metres of the fault, and schistosity and mylonite associated with the fault locally contain syn-kinematic garnet porphyroblasts. Late, dextral shear bands are commonly defined by chlorite and muscovite, indicating that any dextral motion occurred during cooling through the greenschist facies P-T conditions.

The relationships described above support two alternatives: 1) that there was significant vertical motion on the fault prior to peak metamorphism, or 2) that there was relatively little thrust motion on the fault at all. The apparent continuity across the fault of upright folds that are cogenetic with the fault suggests that there is little syn-metamorphic displacement on the Kootenay Arc boundary fault. This observation, coupled with the stratigraphic links between the Kootenay Arc and the Purcell anticlinorium, implies that it is unlikely that the Kootenay Arc is far-travelled with respect to the western Purcell anticlinorium. The regional cross-section (Plate 4a) therefore illustrates the second alternative and shows only about a kilometer of dip-slip displacement on the Kootenay Arc boundary fault.

However, the abrupt contrast between the large west-verging structures in the Kootenay Arc and the ubiquitously upright sequence to the east remains puzzling. If the Kootenay Arc boundary fault did not juxtapose two distinct structural domains across a great distance, then the fault must coincide with a profound pre-existing structural contrast. Since structural relationships discussed above indicate that dikes dated at 173 +/- 5 Ma (Smith et al., 1992) were syn-kinematic with respect to the west-verging folds, then this profound structural contrast must have developed after Mesozoic deformation had begun but before the development of the upright folds in the Kootenay Arc and Purcell anticlinorium.

The Kootenay Arc boundary fault coincides with a previous normal fault that was active during at least two periods of extension in latest Neoproterozoic to Early Paleozoic time (pre-upper Hamill Group and post-Badshot Formation; see Fig. 4-3). This normal fault had juxtaposed a thick succession of primarily deep-water, incompetent strata, that were deposited probably over highly attenuated continental crust to the west, against a succession to the east that includes more competent, shallow-water strata and is underlain by competent Purcell Supergroup and probably less attenuated continental crust as "basement". I therefore suggest that this rift-related ramp acted as a buttress against which west-verging folds formed above a basal décollement during initial Mesozoic compression and collapse of the outboard North American margin. The axis of the Selkirik fan structure, the continuation of the regional zone of structural divergence to the north, also coincides with a previous normal fault (antecedent to the Beaver River fault; Fig. 4-2) that was active intermittently during Neoproterzoic, Early Cambrian and Early Paleozoic (post-Badshot Formation) time and similarly separates primarily deep-water facies to the west from shallow-water facies to the east (Kubli and Simony, 1992. 1994; Colpron et al., *in review*). These relationships have been interpreted similarly to the north (Colpron et al., *in review*), and they indicate that the correspondence between the eastern limit of west-verging deformation and a pre-existing, tectonically significant normal fault persists along strike. This interpretation, coupled with constraints on timing and conditions of deformation, is part of the basis for the tectonic evolution model presented below.

## NEW TECTONIC MODEL FOR EVOLUTION OF THE STRUCTURAL CONTRAST BETWEEN THE HINTERLAND AND FORELAND THRUST AND FOLD BELT OF THE SOUTHERN CANADIAN CORDILLERA

Stratigraphic and structural relationships, metamorphic data and regional geochronological data provide the basis for illustrating the tectonic evolution of the segment of the former North American margin that is exposed in the Purcell anticlinorium and the eastern Kootenay Arc. The tectonic evolution of this segment of the margin is illustrated with a palinspastically restored regional cross section (Plate 4 and Appendix 4) and with a plot of time vs. depth for rocks of the eastern Kootenay Arc and western Purcell anticlinorium (Fig. 4-12). Mesozoic deformation of these strata clearly involved two distinct phases with contrasting tectonic styles:

Phase 1) an early phase of rapid horizontal shortening, west-verging folding and faulting (Plate 4, sections 5 and 4), rapid tectonic burial and vertical thickening with little emergent topography, and rapid exhumation (Fig. 4-12)

Phase 2) a later phase of less rapid horizontal shortening, east-verging, thin-skinned thrusting and folding, passive eastward transport of the metamorphic hinterland, slow, isostatic uplift and erosion, and emergent topography that provided sediment for the foreland basin (Plate 4; Sections 3, 2, 1; Fig. 4-12).

The evolution of the zone of structural divergence can be linked with deformation in the thrust and fold belt to the east and with the development of the foreland basin (Fig. 4-14), providing the basis for a model for the Mesozoic tectonic evolution of the southern Canadian Cordilleran rifted margin (Fig. 4-13).

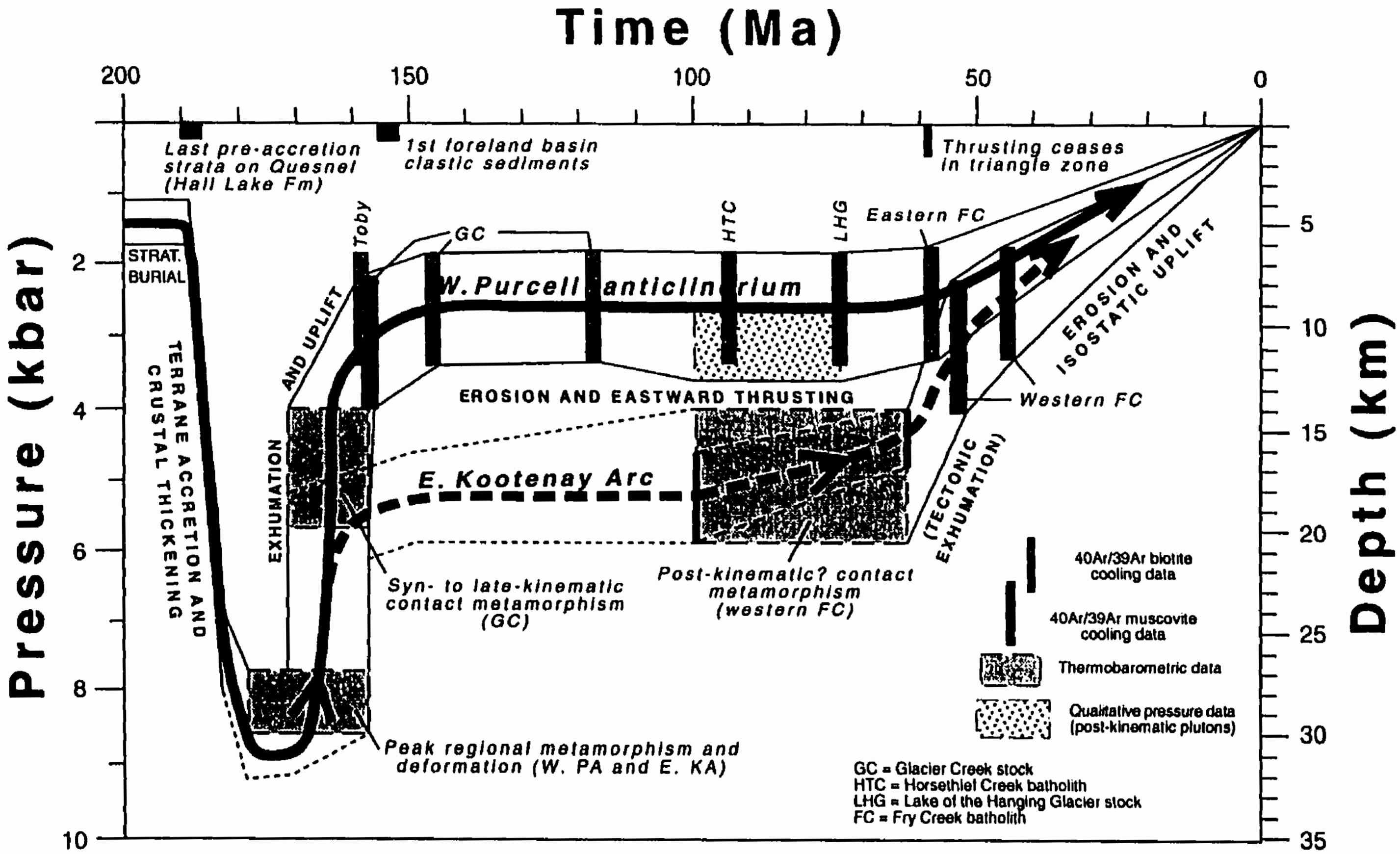



**Figure 4-12:** Depth vs. time plot for rocks of the eastern Kootenay Arc and western Purcell anticlinorium. Pressure converted to depth using equivalency of 3.5 km = 1 kbar. 40Ar/39Ar cooling ages converted to depths by using closure temperatures of 350°C for muscovite and 300°C for biotite (McDougall and Harrison, 1988) and upper and lower limits for geothermal gradients of 45°C/km and 25°C/km. Uncertainties in closure temperatures were not considered. Pressure ranges from thermobarometric data do not include uncertainties (unquantified) in P-T values. Thermobarometric data for Glacier Creek stock calculated from data in Kells (1993). Bathozonal data for mid-Cretaceous plutonic suite in western Purcell anticlinorium from Archibald et al. (1984) and this study. 40Ar/39Ar data from Horsethief Creek, Lake of the Hanging Glacier and Fry Creek plutons from D. A. Archibald (unpublished) and Archibald et al. (1984). *See text for discussion of age constraints for regional and contact metamorphism.*

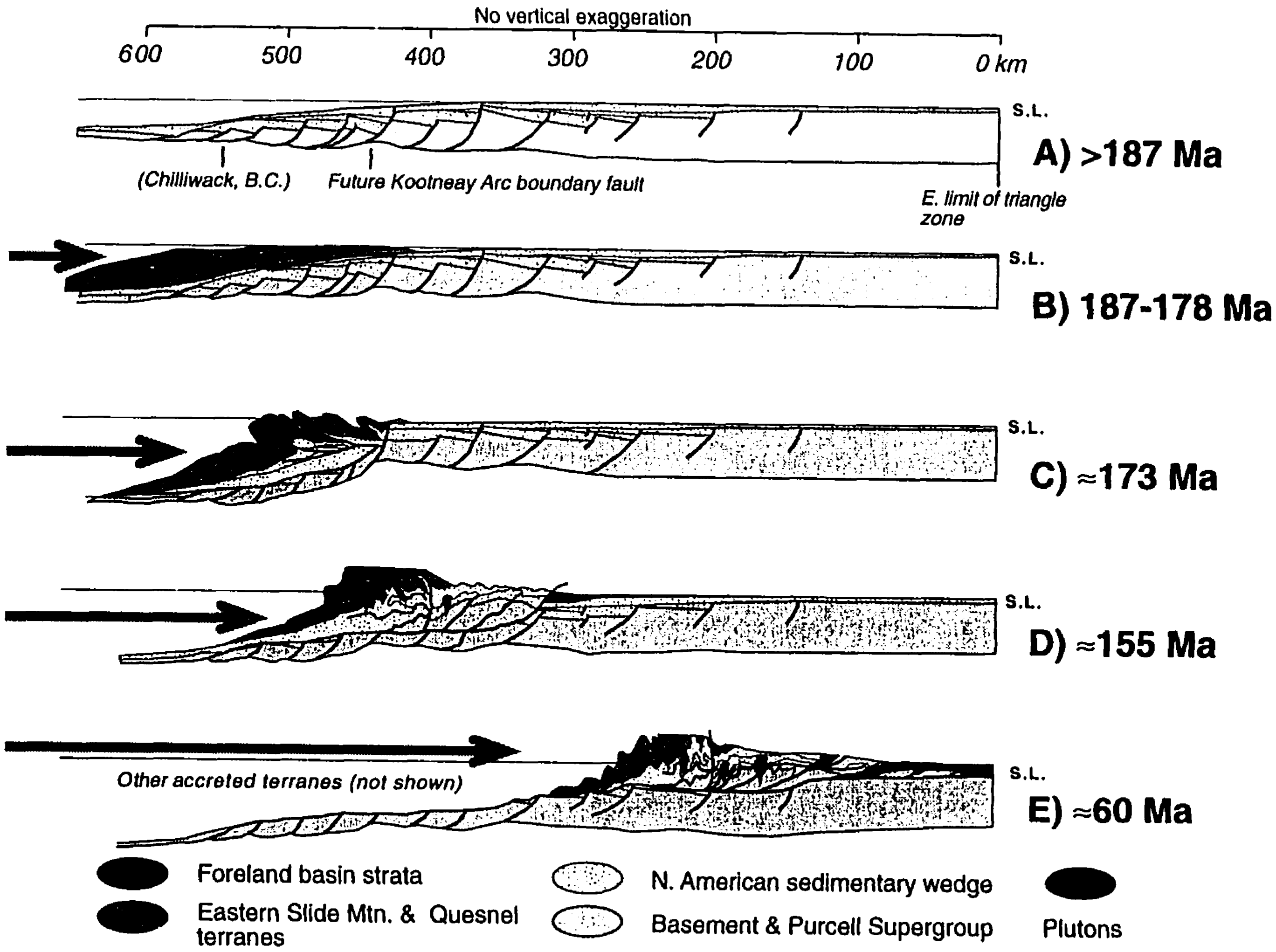
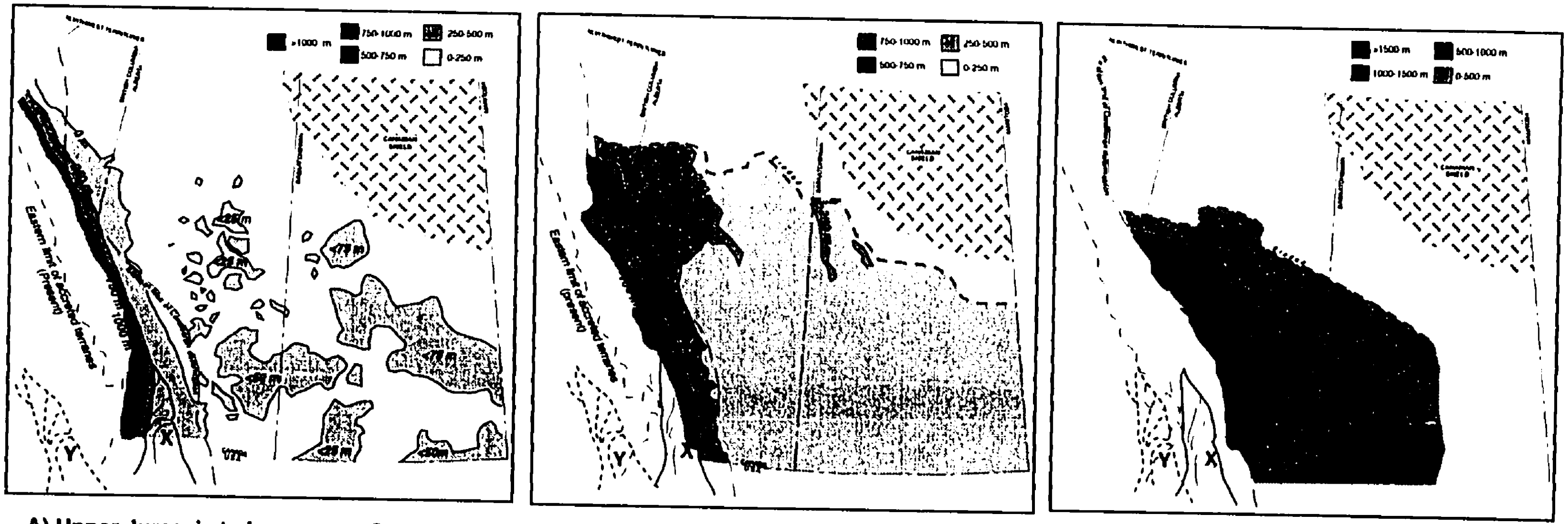



**Figure 4-12:** Tectonic model illustrating the pre-Mesozoic configuration of the North American continental marginand its subsequent Mesozoic tectonic evolution, at the latitude of northern Kootenay Lake. Sections are approximately to scale and are based on the palinspastically restored maps of the rifted margin in Chapters 2 and 3 and on the restored sections through the Purcell anticlinorium nd Kootenay Arc on Plate 4. The foreland thrust-and-fold-belt segment of section E is modified from Price (1994).




A) Upper Jurassic to Lowermost Cretaceous

B) Lower Cretaceous

C) Uppermost Cretaceous

**Figure 4-14:** Isopach maps of *foreland* basin strata of the southern Canadian Cordillera: **A)** Upper Jurassic to Lowermost Cretaceous foreland basin strata (Kootenay Group and equivalent strata). Modified after Poulton et al. (1994); **B)** *Lower Cretaceous* (Aptian and Albian) foreland basin strata (Blairmore and Mannville Groups). Modified after Hayes et al. (1994); **C)** Uppermost Cretaceous (Campanian and Maastrichtian) foreland basin strata, representative of locus of *foreland basin subsidence* in Late Cretaceous and Early Tertiary time. Modified from Price (1994). Strata west of the Cordilleran *deformation front* are palinspastically restored to original positions. The solid outline marked by "X" and the dotted outline marked by "Y" represent, respectively, the Purcell anticlinorium in its present location and approximate restored locations for the beginning of each time interval; the belt of greatest Mesozoic crustal thickening (i.e. tectonic loading) was between the restored positions and the western edge of the foreland basin.

*Phase 1: Development of the metamorphic hinterland*

The west-verging deformation in the Kootenay Arc is inferred to have developed above an eastward-propagating "tectonic wedge" (Plate 4), similar to the "triangle zone" that marks the toe of the foreland thrust and fold belt (e.g. Price, 1986). Regional relationships imply that the wedge developed within distal North American supracrustal strata following initial eastward obduction of Slide Mountain and Quesnel terranes over the outboard margin of North America. The cores of high-amplitude anticlines in the Kootenay Arc contain strata no older than the Hamill Group, but relationships both to the north and to the south along strike imply that the Hamill Group depositionally overlay the Windermere Supergroup (Devlin, 1989; Leclair, 1988; Colpron et al., *in review*) to the west of the antecedent to the Kootenay Arc boundary fault. Therefore the upper detachment of the tectonic wedge is inferred to have developed beneath the competent Hamill Group strata and to have stripped them from the underlying Windermere Supergroup. The lower detachment of the tectonic wedge is inferred to have developed at or near the contact between the supracrustal rocks and their crystalline basement, as the initial segment of the Cordilleran basal décollement, and to have ramped up through the Windermere Supergroup to the base of the Hamill Group.

Material was progressively coupled to the upper part of wedge as it propagated eastward, so that recumbent west-verging folds and splay backthrusts developed above the upper detachment. Charlesworth and Gagnon (1985) and McKay (1991) have demonstrated similar accretion of material to the upper surface of the allochthonous wedge in the frontal triangle zone of the Canadian Rockies. Felsic dikes intruded the deforming strata above the upper detachment as sill-like bodies sub-parallel with it, perhaps contributing to the ductility of deformation and weakness of the detachment. As the tip of the wedge followed the detachment surface at the base of the Hamill Group, it impinged upon the major west-facing ramp in the Neoproterozoic to Paleozoic sedimentary basin. Its eastward progress was impeded, resulting in internal deformation of the wedge and refolding of the west-verging structures about steep axial surfaces. This resulted in rapid vertical thickening of the hinterland of the orogen and rapid tectonic burial of rocks in the Kootenay Arc (Fig. 4-12). Although the foreland basin may have begun to subside tectonically as early as about 180 Ma (Poulton et al., 1984), the time of initial eastward terrane obduction (Fig. 4-13B), the foreland basin received no westerly-derived clastic sediment before about 155 Ma (Fig. 4-12; Fig. 4-13D), suggesting that there was no significant emergent topography before this time (Plate 4 and Fig. 4-13C). This also implies that there was little elastic communication between the loaded, subsided lithosphere to the west of the rift-related ramp and the thicker, stiffer lithosphere to the east.

This phase of the orogeny involved both large amounts and high rates of horizontal shortening. The present width of the upper unit of the Hamill Group in the Kootenay Arc is about 15 km. The restored width of the upper unit of the Hamill Group in the Kootenay Arc (Plate 4; Section 6) is shown as about 90-100 km. Uncertainties in the restored length are estimated at +25 km/-10 km by considering uncertainties in the geometry of the Meadow Creek anticline and in the unexposed portion of the anticline in the immediate hangingwall of the Kootenay Arc boundary fault. These values correspond to a shortening (change in length/restored length) of about 85%. The timing of deformation in the Kootenay Arc is best constrained between 173+/-5 Ma and 164 Ma, as discussed above. Using a lower limit of 85 km for restored length and an upper limit of 14 Ma for duration of deformation yields a lower limit for shortening rate of 6 mm/yr. Using an upper limit of 120 km for restored length and a lower limit of 4 Ma for duration of deformation yields a probably unrealistic result of 30 mm/yr for horizontal shortening rate. Using the well-constrained and intermediate value of 10 Ma for the duration of west-verging folding on the western flank of the Selkirk fan structure (Colpron et al., 1996) and a restored length of 95 km yields a probably realistic shortening rate of about 10 mm/yr. These rates are compatible with paleomagnetic data of Engebretson et al. (1985) that indicate relative convergence rates of about 14 mm/yr between the North American and Farallon plates, sub-orthogonal to the southern Canadian Cordilleran margin, during Middle Jurassic time.

The present interpretation differs from or clarifies previous models for the formation of the west-verging structures in the metamorphic hinterland of the southern Canadian Cordillera. Brown et al. (1986, 1993) proposed that the zone of structural divergence developed above a plate boundary at which oceanic lithosphere was being subducted eastward beneath North America. However, the magmatic arc associated with Early and Middle Jurassic subduction (Quesnel terrane) occurs to the west of the zone of structural divergence. This relationship, coupled with petrologic and geochemical data from the Quesnel arc, indicates that the active plate boundary during Middle Jurassic time was an east-dipping subduction zone that lay to the west of the Quesnel arc.

The interpretation presented in this study also helps to refine the previous interpretation of Price (1986), which involved tectonic wedging of accreted and "suspect" terranes between Paleozoic strata of the southern Kootenay Arc and western Selkirk fan structure, respectively, and their crystalline basement. The interpretation that I present for the intermediate latitude of northern Kootenay Lake is more compatible with the interpretations of Price (1986) and Colpron et al. (in review) for west-verging deformation on the west flank of the Selkirk fan structure, in which the tectonic wedge comprises supracrustal strata that were deposited on the distal part of the North American margin. A tectonic wedge of accreted rocks at the latitude of northern Kootenay Lake would require that the shallow-water strata of the upper Hamill Group and Badshot Formation had been deposited directly on continental or oceanic basement, rather than on Windermere Supergroup strata.

However, the interpretation that I present differs from that of Colpron et al. (in review) about the relative timing and mechanics of initiating the west-verging structures. Their interpretation shows that the west-verging structures did not begin to form until the basal décollement had impinged upon the rifted margin ramp. The direction of propagation of west-verging structures above the upper detachment surface of the wedge is not constrained, either in this segment of the Kootenay Arc or along the western flank of the Selkirk fan structure to the north (M. Colpron, pers. comm., 1997). Workers in triangle zones in the Canadian Rockies and other mountain belts have shown that backthrusts form above the upper detachment both from foreland to hinterland (Butler, 1985) and from hinterland to foreland (Banks and Warburton, 1986). In either case, the very large amount of shortening in the Kootenay Arc implies that west-verging deformation above the wedge was initiated far to the west of the hinge zone in the rifted margin. Similarly, cross sections across the Selkirk fan structure and the Monashee Complex, restored prior to Eocene extension (Johnson, 1994), indicate that west-verging faults with significant displacement extend far to the west of the western flank of the Selkirk fan structure as discussed by Colpron et al. (*in review*).

I suggest that west-verging deformation above a very weak basal décollement was favored over east-verging deformation because eastward terrane obduction caused tectonic loading and flexure of underlying thin North American lithosphere, resulting in west-dipping anisotropies and lack of foreland-dipping emergent topography (Plate 4, Section 5). The hinge zone in the rifted margin, therefore, indirectly influenced the initiation of the west-verging structures by controlling the rheology of the rocks to the west. The hinge zone also directly controlled the eastern limit of west-verging deformation by acting as a mechanical "backstop" or "buttress" that impeded eastward progress of the west-verging deformation and the underlying tectonic wedge.

*Phase 2) Gravitationally-driven thrust and fold belt*

The conceptual models of Jamieson and Beaumont (1988, 1989) describe the effects of a pre-existing ramp in a rifted continental margin on subsequent compressional collapse of the supracrustal strata along that margin. Their models can be used to help illustrate the transition from development of a metamorphic hinterland to initiation of thin-skinned, foreland-propagating thrusting and folding. The abrupt ramp in the previously rifted margin resulted in vertical thickening that eventually produced enough gravitational potential to advance the basal décollement upward and eastward into cooler, stiffer strata to the east. The behavior of material above the basal décollement can be described according to critical Coulomb wedge models (e.g. Davis et al., 1983; Dahlen et al., 1984). These models describe how a foreland-sloping orogenic wedge at "critical taper" can slide gravitationally along its basal décollement without internal deformation. Changes in internal cohesion of material in the orogenic wedge or an increase in the slope of the basal décollement would cause the wedge to deform internally, in order to maintain its critical taper.

Vertical crustal thickening and initial movement of the orogenic wedge up the ramp resulted in rapid uplift, exhumation and cooling of deeply-buried rocks (Fig. 4-12). The high exhumation rates suggest that much of the exhumation was tectonic as well as erosional (Colpron et al., 1996), perhaps due to syn-orogenic extension of an overly-steepened orogenic wedge. No significant normal faults of Middle Jurassic age have been observed directly in the southern Canadian Cordillera. However, the age of the west-dipping normal faults in the eastern Kootenay Arc (Plate 1) is poorly constrained, and they could be as old as late-Middle Jurassic.

Following uplift and cooling, the metamorphic hinterland was transported passively without much further internal deformation as the basal décollement propagated eastward beneath the rocks of the western Purcell anticliorium (Plate 4, sections 3, 2; Fig. 4-13D-E). Exhumation rates were much slower (Fig. 4-12), and are interpreted to record both gradual eastward shallowing of the basal décollement with time and isostatic uplift due to erosion. However, the western edge of the Early Cambrian to Devonian Windermere high (Fig. 4-3 and Plate 4; section 6) marks a second west-dipping ramp in this segment of the rifted margin. The stratigraphic succession that defines the Windermere high must have been deposited above a significantly thicker crustal block than that to the west. This ramp might have resulted in another local increase in the gradient of the basal décollement, again requiring internal thickening of the orogenic wedge to maintain its critical taper. Because the strata in the Purcell anticlinorium were stiffer, cooler and more brittle than those in the hinterland, and because the orogenic wedge now had a foreland-dipping upper surface, the mechanism of crustal thickening and internal deformation of the wedge was different. Out-of-sequence, east-verging thrusting (e.g. Mount Forster and Purcell thrusts) may have been the mechanism of vertical thickening, rather than steep or west-verging poly-phase folding.

The basal décollement continued to propagate eastward during late Cretaceous to Paleocene time beneath the strata of the Main Ranges, Front Ranges and Foothills of the Canadian Rockies (Fig. 4-13E). Compresssive deformation ceased at the beginning of Eocene time (59-58 Ma) as a result of a change in relative plate motions between North America and Farallon and Kula plates (e.g. Price, 1994). Subsequent regional extension in Eocene time exhumed deeper crustal levels of the metamorphic hinterland (e.g. Parrish et al., 1988), including perhaps the rocks intruded by the western part of the Fry Creek batholith (Fig. 4-12).

The amounts and rates of horizontal shortening during this second phase of the orogeny were nearly an order of magnitude less than during the development of the metamorphic hinterland. The strata of the Purcell anticlinorium as shown in Plate 4 restore from a width of about 60 km to about 100 km, a shortening of only 40%. This deformation occurred between about 170 Ma and a younger limit of about 95 Ma, yielding a lower limit on shortening rate of about 0.5 mm/yr. The supracrustal strata that were deposited between the triangle zone and the Chatter Creek fault in the Western Main Ranges (Fig. 4-2) were shortened by 50% (Price and Mountjoy, 1970). Restoration of the entire southern Canadian Cordilleran margin (Fig. 4-13A, based on palinspastic maps in Chapters 2 and 3) also indicates that the rocks to the east of the Kootenay Arc boundary fault were shortened from about 400 km to 200 km, or by about 50%. Deformation of these strata was initiated at about 170 Ma and ended about 100 Ma later, implying an overall shortening rate of a little more than 2 mm/yr. A comparable shortening rate of 1.7 mm/yr has been calculated for the Early Cretaceous to Paleocene history of the thrust and fold belt in Idaho, Wyoming and northern Utah (Allmendinger, 1992). The record of nearly continuous deformation in Cretaceous and Paleocene time from deeper crustal levels near the Monashee décollement (Parrish, 1995, and references therein) implies that there were no significant gaps in motion on the basal décollement and thus no major periods of significantly more rapid shortening.

The advance of the orogenic wedge onto progressively stiffer, thicker lithosphere to the east also is recorded by changes in the morphology of the foreland basin with time. The deposition of the first westerly-derived clastic sediments in the foreland basin (Upper Jurassic upper Fernie/Kootenay groups) coincided closely with rapid uplift and cooling of rocks in the metamorphic hinterland and the initiation of east-verging thrusting and folding in the Purcell anticlinorium (Fig. 4-12). However, the Kootenay Group was deposited in a relatively narrow, deep foreland basin (Fig. 4-14A). In contrast, the Lower to Uppermost Cretaceous foreland basin sediments were deposited in a broader basin (Fig. 4-14B and C), indicating that the lithosphere toward the east was stiffer and more elastic during tectonic loading.

## SUMMARY AND CONCLUSIONS

Abrupt and significant east-west changes in the thickness and mechanical properties of the Neoproterozoic to Early Proterozoic continental margin of Laurentia profoundly influenced mechanisms of crustal thickening and styles of deformation during Mesozoic compressional tectonics in the southern Canadian Cordillera. The Kootenay Arc boundary fault is a regionally significant thrust fault that coincides with a regional zone of structural divergence. It coincides with a pre-exisitng normal fault that marks perhaps the most significant "hinge zone" in the former rifted margin, separating significantly attenuated continental crust to the west from less attenuated crust to the east. Deformation of the supracrustal rocks that were deposited on the former North American continental margin can be divided into two distinct stages that reflect the nature of the lithosphere that underlay them on either side of this "hinge zone":

1) development of a metamorphic hinterland above significantly attenuated, weak, plastic lithosphere. The deformation was characterizied by:

- west-verging, poly-phase ductile folding and faulting above an eastward-propagating tectonic wedge of North American strata
- rapid and large horizontal shortening
- little emergent topography

- rapid vertical thickening, rapid exhumation and probably syn-orogenic extension (Colpron et al., 1996) in response to the abrupt increase in basal décollement gradient at the hinge zone.
- initial behavior not according to Coulomb wedge theory of Davis et al. (1983) and Dahlen et al. (1984)

2) development of a foreland thrust and fold belt above less attenuated, stronger, more elastic lithosphere. The deformation was characterized by:

- thin-skinned, east-verging thrusting and folding above a basal décollement
- less rapid and less overall horizontal shortening
- moderate vertical thickening with foreland-sloping topography
- behavior according to Coulomb wedge theory

The initial development of the foreland basin recorded the propagation of the basal décollement across the hinge zone. Subsequent changes in the morphology of the foreland basin recorded the propagation of the basal décollement across progressively stiffer, thicker lithosphere to the east.

# Chapter 5

# CONCLUSIONS

## CHAPTER 2

The Windermere Supergroup of the west-central Purcell Mountains is exposed in two contrasting stratigraphic domains: a southern/eastern domain characterized by a thinner succession of homogeneous and laterally continuous units, and a northern/western domain characterized by a thicker succession of more heterogeneous and laterally variable units. Horsethief Creek Group strata in both domains comprise distinct lower and upper clastic sequences, separated by a regional carbonate-bearing marker unit and possibly significant unconformity at the base of the upper clastic sequence.

Two episodes of normal faulting controlled the sedimentation of the Windermere Supergroup in the Purcell Mountains and are recorded by each of the clastic sequences in the Horsethief Creek Group. The first is part of a major intracontinental rifting event that was initiated immediately prior to the deposition of the Toby Formation and is recorded throughout the Canadian Cordillera. The development of the Windermere basin was controlled by two sets of extensional faults: a N-trending, primarily west-side-down set, and a northeast-trending, northwest-side-down set. The northeast-trending set is recorded by conspicuous thickness and lithofacies changes in the Windermere Supergroup. It accommodated significant upper crustal stretching along deep, probably listric normal faults, from southeast to northwest. Regional uplift of the basin margins to the east and south along both sets of faults is recorded by the provenance of sediment in the Windermere Supergroup of the Purcell anticlinorium.

The second episode of normal faulting was recorded by faults that cut the underlying lower clastic sequence and by relative sea level fall and rapid distribution of a large volume of coarse, immature sediment to the basin. This episode of normal faulting appears to be restricted to the southern Canadian Cordillera, and its tectonic setting is questionable. I have argued that a eustatic sea level fall alone cannot account for such a large thickness of grit over such a large area of the southern Canadian Cordillera, and that the basin records renewed regional tectonic uplift of the basin margins, again along major structures both to the south and to the east.

This conclusion apparently conflicts witth previous arguments for continental separation between the northwestern margin of Laurentia and the eastern margin of Australia by 750-720 Ma. I propose that seafloor spreading might have occurred diachronously from north to south, and/or involved separation of more than one continental block from north to south, as it did from south to north in the northern Atlantic Ocean.

## CHAPTER 3

The Hamill Group in the west-central Purcell Mountains is divisible into several lithostratigraphic units. The lowermost fluvial and marine units are discontinous, commonly immature and locally contain mafic volcanic rocks. The uppermost shallow marine quartzite unit is continuous and rests unconformably on the lower units or directly on the Horsethief Creek Group. Lateral stratigraphic variations indicate that the distinct stratigraphic successions of lower Hamill Group in the Kootenay Arc and in the Purcell anticlinorium were deposited in separate half-grabens that were bounded to the east by normal faults.

The unconformity beneath the upper Hamill Group in the west-central Purcell Mountains is a regional unconformity in the Hamill and Gog Groups of the southern Canadian Cordillera. It is the stratigraphic expression of the rift-to-drift transition that occurred in Early Cambrian time between 549 Ma and less than about 520 Ma. It marks the change from regional E-W extension of continental crust to seafloor spreading and thermal subsidence of an underlying passive continental margin. The Neoproterozoic(?) to lower Lower Cambrian (Placentian) strata that lie below the unconformity were deposited in primarily shallow, N-trending, fault-bounded basins that were separated by uplifted, bevelled and primarily eastward-tilted blocks of Windermere Supergroup strata. Upper Lower Cambrian (Waucoban) quartz arenite and quartzite that lie above the unconformity were deposited as a more continuous sheet across the basins and the uplifted blocks in a shallow shelf setting after extension and normal faulting had ceased.

The emergence of several paleogeographic "highs" during Lower Paleozoic time is interpreted to record, at least initially, differential subsidence of several fault-bounded blocks of different crustal thickness, and thus of different buoyancy. These features were defined by structures that were inherited both from older northeast-trending syn-Windermere and north-trending syn-lower Hamill/Gog normal faults. These normal faults were reactivated during several enigmatic episodes of Early Paleozoic normal faulting as well as during Mesozoic thrust faulting. The episode of crustal extension that immediately preceded the rift-to-drift transition involved relatively insignificant stretching of the upper crust, in contrast to significant upper crustal stretching, topographic relief and deep rift basin subsidence that occurred during syn-Windermere extension.

## CHAPTER 4

Abrupt and significant east-west changes in the thickness and mechanical properties of the Neoproterozoic to Early Proterozoic continental margin of Laurentia profoundly influenced mechanisms of crustal thickening and styles of deformation during Mesozoic compressional tectonics in the southern Canadian Cordillera.

The Kootenay Arc boundary fault is a regionally significant thrust fault that separates rocks of the Purcell anticlinorium from rocks of the Kootenay Arc and coincides with a regional zone of structural divergence. It also coincides with a pre-exisitng Neoproterozoic to Early Paleozoic normal fault that marks perhaps the most significant "hinge zone" in the former rifted margin, separating significantly attenuated continental crust to the west from less attenuated crust to the east.

Deformation of the supracrustal rocks that were deposited on the former North American continental margin can be divided into two distinct stages that reflect the nature of the lithosphere that underlay them on either side of this "hinge zone":

Phase 1) short-lived development of a metamorphic hinterland above significantly attenuated, weak, plastic lithosphere. The deformation was characterized by:

- west-verging, poly-phase ductile folding and faulting above an eastward-propagating tectonic wedge of North American strata, following initial eastward obduction of accreted terranes
- rapid and large horizontal shortening of about 10 mm/yr and about 85 percent
- little emergent topography
- rapid vertical thickening and deep tectonic burial, rapid exhumation of as much as 10 km, and probably syn-orogenic extension (Colpron et al., 1996), in response to the abrupt increase in basal décollement gradient at the hinge zone.
- initial behavior not according to Coulomb wedge theory of Davis et al. (1983) and Dahlen et al. (1984)

West-verging folding was underway in the Kootenay Arc by 173 +/- 5 Ma, and rocks of the Kootenay Arc and westernmost Purcell anticlinorium were uplifted and cooled by 159-157 Ma. Thus the high-amplitude folds in the Kootenay Arc are not mid-Paleozoic, as previously argued (Read and Wheeler, 1976; Klepacki and Wheeler, 1985).

Phase 2) development of a foreland thrust and fold belt above less attenuated, stronger, more elastic lithosphere. The deformation was characterized by:

- thin-skinned, east-verging thrusting and folding above an eastward-propagating basal décollement

- less rapid and less overall horizontal shortening of about 1-2 mm/yr and about 40-50%
- moderate vertical thickening with foreland-sloping topography
- behavior according to Coulomb wedge theory

The propagation of the basal décollement across the hinge zone is recorded by the initial development of the foreland basin, and by uplift and cooling of the metamorphic hinterland. Subsequent changes in the morphology of the foreland basin recorded the propagation of the basal décollement over progressively stiffer, thicker lithosphere to the east.

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# Appendix 1

# $^{40}Ar/^{39}Ar$ GEOCHRONOLOGY

## PURPOSE

Previous geochronology on the syn-kinematic plutons in the study area was limited to several K-Ar dates on hornblende and biotite from Toby and Glacier Creek stocks (Reesor, 1973; Wanless et al., 1968). The dates obtained ranged from 232 Ma (Middle Triassic) to 127 Ma (Early Cretaceous) and were clearly inconsistent with Midddle Jurassic crystallization and/or cooling dates obtained from similar syn-kinematic granodiorites to the west, south and north (e.g Archibald et al., 1983, and references therein; Parrish and Wheeler, 1983). Three mineral separates of hornblende and biotite from Glacier Creek and Toby stocks were obtained from the Geological Survey of Canada, from specimens previously dated by the K-Ar method, with the intent of improving upon the previous K-Ar dates and of better elucidating the thermal history of these rocks with $^{40}Ar/^{39}Ar$ step-heating. These specimens were supplemented by hornblende and biotite mineral separates from three new samples of these stocks, and by metamorphic muscovite from the contact aureole of the Glacier Creek stock. Samples collected from the Tea Creek and Mt. Cauldron stocks were altered following emplacement and were not considered appropriate to date. Several additional samples from Glacier Creek and Toby stocks were also considered inappropriate due to obvious fine intergrowth of biotite and chlorite, or of biotite and hornblende.

An additional biotite separate was obtained from a sample of a biotite-bearing quartz monzonite or granite that was collected from a small stock exposed to the north of the study area in the Duncan River valley. The stock is deformed by a brittle to cataclastic west-side-down fault fabric, similar to the fabric present in the northwestern portion of the Fry Creek batholith.

## ANALYTICAL METHODS

The mineral separates and neutron flux monitors were irradiated in three batches with fast neutrons for 29 hours in site 5C of the water-moderated, enriched-uranium reactor at McMaster University, Hamilton, Ontario. Samples were contained in aluminum-foil pouches, stacked into aluminum irradiation canisters 11.5 cm long and 2.0 cm in diameter. Flux monitors (LP-6: biotite; 128.5 Ma) were evenly spaced (about 1 cm) between the samples of unknown age. The J-values for samples of unknown age were determined by interpolation between those of the monitors.

The samples were dated at Queen's University, Kingston, Ontario. For converntional step-heating, the samples were heated in a quartz furnace tube using a Lindberg resistance furnace, and they were analyzed using an ultra-high vacuum, stainless steel, argon extraction system that is on-line to a substantially modified A. E. I. MS-10 mass spectrometer run in the static mode.

For step-heating and spot-dating using a laser, the samples were mounted in an aluminum sample-holder, beneath the sapphire view-port of a small stainless-steel chamber connected to an ultra-high vacuum purification system. Step-heating was accomplished by de-focusing the beam of an 8 Watt Lexel 3500 laser to cover the entire sample and by heating for approximately two minutes at increasing power settings for each step. A focused beam was used to produce melt pits ca. 50 microns in diameter. The evolved gas, after purification using as SAES C50 getter, was admitted to an on-line, MAP 216 mass spectrometer, with a Bäur Signer source and an electron multiplier (set to a gain of 100 over the Faraday), and analysed in the static mode. Blanks, measured routinely, were subtracted from the subsequent sample gas-fractions. Extraction-blank volumes during this series of analyses were: 4.0 - 17.0 x $10^{-13}$, 0.2 - 0.9 x $10^{-13}$, 0.4 - 0.6 x $10^{-13}$, and 0.2 - 1.0 x $10^{-13}$ $cm^3$ STP for masses 40, 39, 37 and 36, respectively.

Measured mass-spectrometric ratios were extrapolated to zero-time and corrected to a $^{40}Ar/^{39}Ar$ atmospheric ratio 295.5. They were corrected for neutron-induced $^{40}Ar$ from potassium, for $^{39}Ar$ and $^{36}Ar$ from calcium and for $^{36}Ar$ from chlorine. Dates and errors were calculated using formulae of Dalrymple et al. (1981) and the constants recommended by Steiger and Jäger (1977). The calculated errors represent the 2-sigma analytical precision assuming that the error in the J-value is zero. This assumption is suitable for comparing steps from a single analysis. A conservative estimate of the error in the J-value is 0.5%, which could be included for comparison of samples irradiated in different pouches.

## RESULTS

Age spectra and complete $^{40}Ar/^{39}Ar$ age data are presented for each analysis following this discussion. The data are presented in the order in which they are discussed below. Samples were step-heated in the furnace unless it is noted that they were laser step-heated or spot-dated.

### TOBY STOCK

Specimen GSC 66-49 (Wanless et al., 1968; Reesor, 1973) is from a weakly foliated mafic-rich hornblende-biotite granodiorite from near the center of the widest part of the pluton. Step-heating of hornblende from this rock yielded an integrated age of 185 +/- 2 Ma. The specimen was dated in only six steps because of the small amount of mineral separate remaining, but the spectrum shows a slight "saddle" shape, with an age of 182 +/- 3 Ma for the low part of the saddle. However, an initial $^{40}Ar/^{36}Ar$ ratio of 1633 indicates excess Ar, and thus this age is considered unreliable and only a maximum age for the pluton. This age is nonetheless consistent with U-Pb zircon ages of 181 +/- 7 Ma from Cooper Creek stock (Klepacki, 1985) and of 183 to 178 Ma in the Rossland area (Andrew et al., 1990; T. Höy, pers. comm., 1995) from other syn-kinematic plutons that stitch North America and accreted terranes. It should be noted that the $^{40}Ar/^{39}Ar$ date differs markedly from and casts doubt on the seemingly reasonable previous K-Ar age of 162 +/- 8 Ma obtained from the same mineral separate of this specimen (Wanless et al., 1968).

Specimen GSC 62-14 (Leech et al., 1968; Reesor, 1973) is from the strongly foliated and lineated, fault-bounded "tail" of the Toby stock. The specimen is an epidote-bearing biotite granodiorite augen gneiss. Step-heating on biotite yields a disturbed spectrum with an integrated age of 160 +/- 2 Ma. The lower-temperature steps are interpreted to reflect Ar loss as a result of reheating during regional intrusion of the ca. 100 Ma suite of post-kinematic plutons; of these the voluminous Fry Creek batholith is exposed less than 10 km to the south of Toby stock. Post-crystallization ductile deformation superposed on the "tail" of the pluton might also have resulted in Ar loss. The lower-temperature steps step up to a "plateau" age of 163 +/- 2 Ma, interpreted as a minimum cooling age through the biotite closure temperature. The higher-temperature steps or "saddle" in the spectrum might reflect two-phased release of Ar from biotite (York and Lopez-Martinez, 1986), with some Ar loss from each phase. However, Ca/K values for the higher-temperature steps suggest contamination by a Ca-bearing phase for these steps. Textural relationships and mineralogy in the contact aureole indicate that the Toby stock was emplaced into rocks undergoing biotite-grade regional metamorphism, at 4.2 kbar and temperatures of at least 350°C, well above the closure temperature of Ar in biotite (McDougall and Harrison, 1988), so it is reasonable to conclude that the pluton probably underwent initial slow cooling through the biotite closure temperature as well as subsequent reheating.

Specimen TOBY-1 was collected from the mafic northeastern rim of the Toby stock. Specimen TOBY-1 is a pyroxene-epidote-bearing hornblende-biotite granodiorite to hornblendite. Coarse (1-2 mm) hornblende and biotite were hand-picked from crushed rock. Hornblende is finely intergrown with biotite, pyroxene and plagioclase, and analysis of hornblende yielded completely meaningless results. Analysis of biotite yielded both a plateau age and a correlation age of 243 +/- 1 Ma. However, a previous K-Ar biotite date of 232 Ma from Toby stock, (Wanless et al., 1968; Reesor, 1973) as well as anomolously old K-Ar biotite dates from the circa 170 Ma-old Adamant pluton (Shaw, 1980) have been interpreted to reflect excess Ar (Wanless et al., 1968).

Specimen TOBY-2 is compositionally similar to TOBY-1, except that there is significantly more chlorite intergrown with biotite along the cleavage planes. Both samples were collected from the same locality. Step heating of biotite yielded a disturbed concave-upward spectrum with an integrated age of 285 +/- 2 Ma. This age is also inferred to reflect excess Ar, and the age from this specimen supports the interpretation that the 243 Ma age obtained from TOBY-1 biotite is a geologically meaningless one, despite the fact that it is a plateau age.

## GLACIER CREEK STOCK

Specimen GSC 61-18 (Wanless et al., 1968; Reesor, 1973) is a weakly to non-foliated biotite granodiorite. Step-heating of this very fine-grained biotite separate produced a disturbed spectrum with a "plateau" segment at 120 +/- 1Ma. This age is interpreted to reflect almost complete resetting during mid-Cretaceous intrusion and reheating.

Specimen GC-1 is a weakly- to non-foliated hornblende-biotite granodiorite from near the center of the pluton. Step-heating of biotite produced a disturbed spectrum with an integrated age is 146 +/- 1 Ma, that climbs progressively from a low-temperature age of 119 Ma to a high-temperature age of 151 Ma. The spectrum is also interpreted to reflect Ar loss during mid-Cretaceous intrusion, but the biotite in this specimen is not as severely reset, probably because it is a larger size fraction. Thus the oldest step of 151 Ma is interpreted as a minimum age for cooling of Glacier Creek stock through the biotite closure temperature.

Analysis of hornblende from specimen GC-1 yielded a highly disturbed spectrum with an integrated age of 200 +/- 10 Ma. Thin sections show that hornblende in this specimen is finely intergrown with clinopyroxene, biotite and plagioclase. Thirteen total fusion laser spot analyses from two single grains (D. A. Archibald, unpublished data) from the same sample yielded a range of ages similar to those from these individual steps, and they confirm that the hornblende is not homogeneous. The Glacier Creek stock was emplaced at lower amphibolite facies conditions, and thin sections of the stock from nearby localities also show that igneous hornblende is commonly overgrown with perhaps metamorphic hornblende of a different composition, as well as less abundant actinolite. Thus hornblende from Glacier Creek stock can be expected to yield poor results.

Specimen GC-3-7 is a pelitic schist from the sillimanite zone of the contact aureole of the Glacier Creek stock. Coarse (> 1 mm) muscovite porphyroblasts clearly overgrow an earlier regional foliation. Muscovite was step-heated with the laser, and it produced a distrubed spectrum with an integrated age of 157 +/-1 Ma. This spectrum does not appear to be severely overprinted following cooling. The age is interpreted as a cooling age associated with uplift of the western Purcell anticlinorium during and immediately following the emplacement of the syn-kinematic plutons (see Chapter 4). It is compatible with the 160 +/- 2 Ma biotite cooling age from Toby stock. Toby stock might be expected to have cooled through the biotite closure temperature before Glacier Creek stock cooled through the higher muscovite closure temperature because Toby stock was emplaced at lower pressures into rocks that were undergoing lower-grade regional metamorphism. Alternatively, the plutons might have different emplacement ages.

**UNNAMED STOCK IN DUNCAN RIVER VALLEY**

Specimen DR-2 was collected from a biotite quartz monzonite or granite that intrudes the Horsethief Creek Group and cuts the regional cleavage. It is similar in composition to the nearby mid-Cretaceous Battle Range and Bugaboo batholiths. The pluton is deformed by a west-side-down brittle to cataclastic fault fabric that occurs in the Duncan River and Beaver River valleys and is similar to the fabric that deforms the northwest margin of the Fry Creek batholith. The biotite separate from this rock was contaminated with chlorite, but it gives an age of 69 +/- 1 Ma, an age compatible with the suite of post-kinematic plutons that includes the Lake of the Hanging Glacier stock (74 Ma; D. A. Archibald, unpublished). This age implies that the normal faulting is younger, most likely Eocene, as Archibald et al. (1984) and I (Chapter 4) suggest for the post-emplacement normal faulting on the west side of the Fry Creek batholith. Therefore, Eocene normal faults related to the Purcell trench fault system to the south continue to the north as several west-dipping normal faults along northern Kootenay and Duncan Lakes (Fyles, 1964, 1967; and this study) and into the Beaver River fault zone in the Duncan and Beaver River valleys.

## MW-26: GSC-66-49 Hornblende 60/80

Run date: 1993/11/05  Can/Pos: 124/31  J Value: 0.007116
Recalc date: 1995/08/28  Mass: 124.0 mg  ± 0.000046

Volume 39Ar: 98.17 x 1E-9 cm3 NTP  Approx. 1.57% K
Integrated Age: 185.06 ± 2.05 Ma  7.32% Ca

Initial 40/36: 1633.03 ±3121.22 (MSWD = 5.29, isochron between -0.41 and 3.83)
Correlation Age: 156.27 ± 62.46 Ma ( 24.4% of 39Ar, steps marked by >)

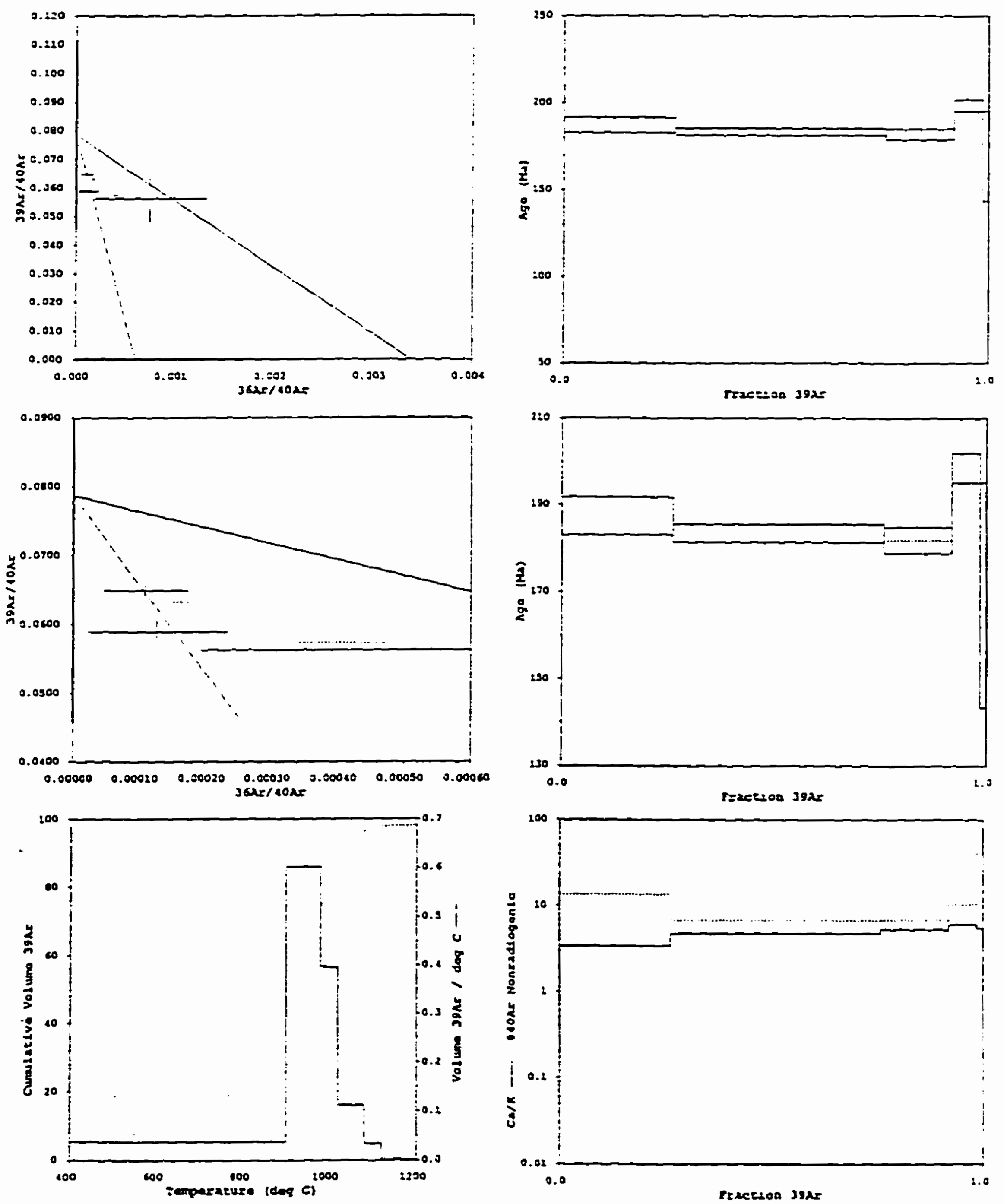
Plateau Age: 181.90 ± 3.21 Ma ( 16.1% of 39Ar, steps marked by <)

| Temp | 36Ar/40Ar | 39Ar/40Ar | r | Ca/K | %40Ar* | %39Ar | 40Ar*/39K | Age |
|---|---|---|---|---|---|---|---|---|
| 900 | 0.00040244±0.00006726 | 0.057278±0.000645 | -0.076 | 3.45 | 86.27 | 26.42 | 15.392± 0.376 | 187.40± 4.35 |
| 980 | 0.00016441±0.00002200 | 0.063231±0.000663 | -0.125 | 4.37 | 93.09 | 49.24 | 15.047 0.179 | 183.51 2.06 |
| <1020> | 0.00010988±0.00006470 | 0.064900±0.000844 | -0.501 | 5.39 | 93.03 | 16.14 | 14.906 0.242 | 181.90 3.01 |
| 1080> | 0.00012755±0.00010595 | 0.058397±0.001707 | -0.331 | 6.10 | 99.59 | 6.95 | 16.339 0.299 | 198.43 3.43 |
| 1120> | 0.00075158±0.00056049 | 0.056293±0.009753 | -0.663 | 5.60 | 96.91 | 1.38 | 13.319 2.212 | 169.22 25.36 |
| 1200 | -0.00346153±0.03494951 | 0.165219±0.854030 | -0.957 | 3.41 | 18.60 | 0.17 | 12.244 18.539 | 150.71 216.33 |

| Temp | 40Ar | 39Ar | 38Ar | 37Ar | 36Ar | Blank 40Ar | Atmos 40/36 |
|---|---|---|---|---|---|---|---|
| 900 | 457.644±3.262 | 25.936±0.190 | 33.340±0.237 | 49.835±0.350 | 0.210±0.030 | 4.652±2.326 | 310.20±1.00 |
| 980 | 766.629 5.410 | 48.144 0.332 | 91.330 0.567 | 127.991 0.883 | 0.175 0.014 | 5.160 2.580 | 310.28 1.00 |
| <1020> | 249.721 1.105 | 15.349 0.069 | 29.879 0.131 | 46.655 0.209 | 0.057 0.013 | 5.547 2.773 | 310.36 1.00 |
| 1080> | 120.571 0.636 | 6.729 0.034 | 12.868 0.077 | 22.438 0.153 | 0.042 0.006 | 6.383 3.191 | 310.44 1.00 |
| 1120> | 31.225 0.918 | 1.354 0.025 | 2.485 0.055 | 4.141 0.117 | 0.043 0.010 | 7.178 3.589 | 310.52 1.00 |
| 1200 | 10.648 1.702 | 0.164 0.020 | 0.226 0.039 | 0.305 0.089 | 0.029 0.008 | 9.653 4.827 | 310.60 1.00 |

All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

## MW-26: GSC-66-49 Hornblende 60/80




All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

## MW-15: GSC-62-14 Biotite 60/80

| | | | | | |
|---|---|---|---|---|---|
| Run date: | 1993/10/27 | Can/Pos: | 124/29 | J Value: | 0.007123 |
| Recalc date: | 1995/08/28 | Mass: | 151.0 mg | | ± 0.000044 |

| | | | |
|---|---|---|---|
| Volume 39Ar: | 613.25 x 1E-9 cm3 NTP | Approx. | 8.04% K |
| Integrated Age: | 159.92 ± 1.83 Ma | | 0.12% Ca |

Initial 40/36: 292.55 ± 40.66 (MSWD = 3.39, isochron between 0.37 and 2.26)
Correlation Age: 161.99 ± 2.11 Ma ( 69.9% of 39Ar, steps marked by >)

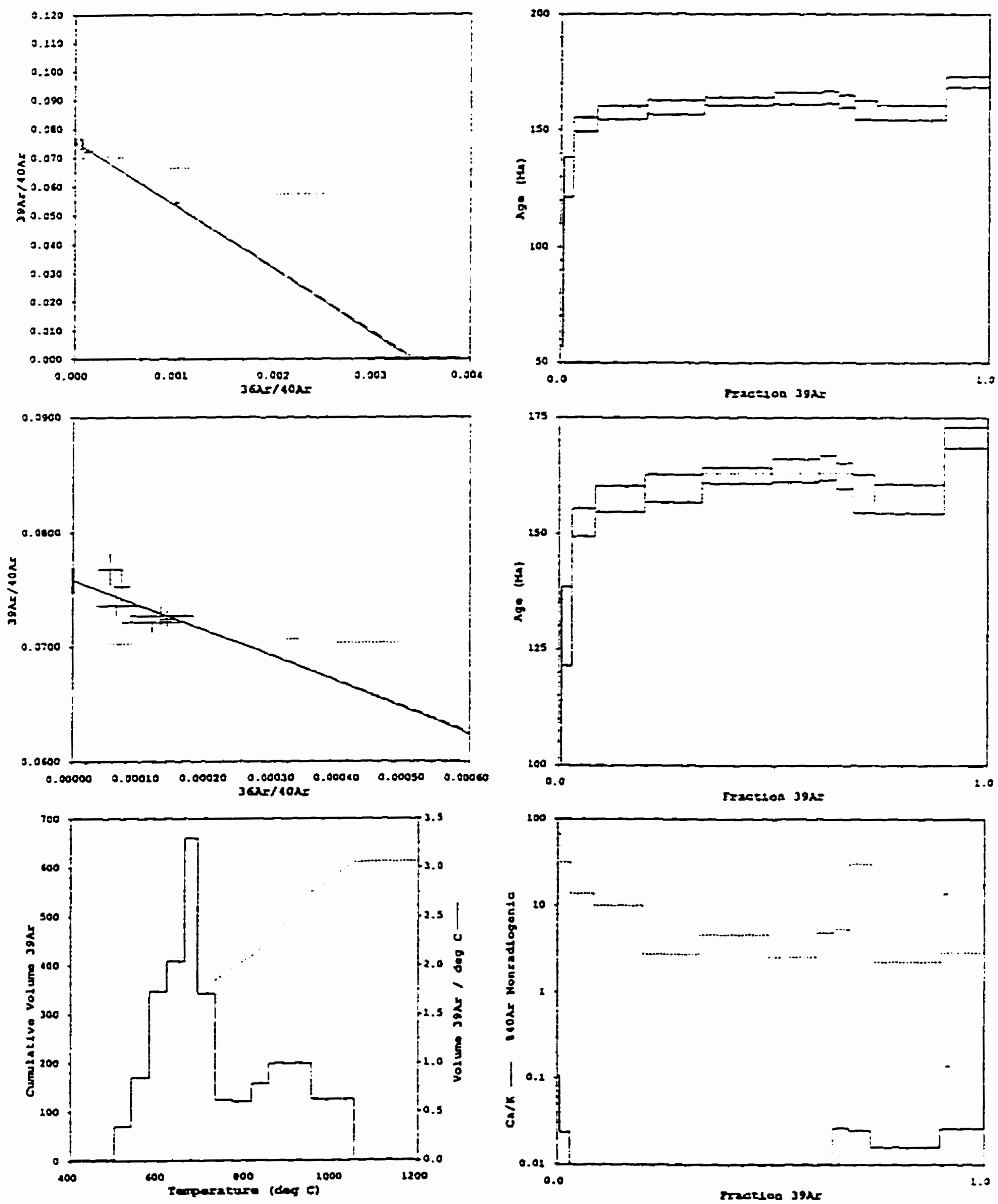
Plateau Age: 163.03 ± 1.55 Ma ( 35.3% of 39Ar, steps marked by <)
170.76 ± 2.47 Ma ( 10.2% of 39Ar, steps marked by [)

| Temp | 36Ar/40Ar | 39Ar/40Ar | r | Ca/K | %40Ar* | %39Ar | 40Ar*/39K | Age |
|---|---|---|---|---|---|---|---|---|
| 500 | 0.00221739±0.00026316 | 0.057441±0.003580 | 0.329 | 0.11 | 32.54 | 0.66 | 6.002± 1.519 | 75.53± 18.72 |
| 540 | 0.00102463±0.00015088 | 0.066522±0.001491 | 0.012 | 0.02 | 69.26 | 2.28 | 10.481 0.713 | 129.39 8.53 |
| 580 | 0.00043895±0.00005101 | 0.070335±0.000808 | -0.036 | 0.01 | 86.13 | 5.54 | 12.375 0.253 | 150.38 2.99 |
| 620 | 0.00032293±0.00001321 | 0.070648±0.001224 | 0.153 | 0.01 | 89.90 | 11.30 | 12.804 0.237 | 157.46 2.73 |
| 660> | 0.00007404±0.00001300 | 0.075267±0.001452 | 0.013 | 0.01 | 97.23 | 13.31 | 12.998 0.257 | 159.71 3.02 |
| < 690> | 0.00014135±0.00001353 | 0.072407±0.000741 | -0.000 | 0.00 | 95.32 | 16.16 | 13.232 0.147 | 162.43 1.72 |
| < 730> | 0.00006566±0.00003158 | 0.073575±0.000977 | -0.026 | 0.01 | 97.41 | 11.17 | 13.328 0.216 | 163.62 2.51 |
| < 770> | 0.00011384±0.00004623 | 0.072133±0.000894 | -0.206 | 0.01 | 95.11 | 4.05 | 13.378 0.225 | 164.19 2.64 |
| < 810> | 0.00013271±0.00004746 | 0.072657±0.000949 | -0.215 | 0.03 | 94.62 | 3.90 | 13.224 0.230 | 162.40 2.70 |
| 850> | 0.00101024±0.00002747 | 0.054359±0.001100 | 0.352 | 0.02 | 69.50 | 5.12 | 12.304 0.344 | 156.65 4.25 |
| 950> | 0.00005635±0.00001974 | 0.076771±0.001519 | -0.001 | 0.02 | 97.74 | 16.19 | 12.809 0.265 | 157.52 3.12 |
| [1050 | 0.00006910±0.00002137 | 0.070296±0.000912 | -0.113 | 0.03 | 97.11 | 10.16 | 13.837 0.193 | 170.76 2.25 |
| 1200 | 0.00329440±0.00027674 | 0.001293±0.000103 | 0.346 | 1.35 | 2.90 | 0.16 | 22.752 44.737 | 271.30 716.31 |

| Temp | 40Ar | 39Ar | 38Ar | 37Ar | 36Ar | Blank 40Ar | Atmos 40/36 |
|---|---|---|---|---|---|---|---|
| 500 | 74.725± 3.772 | 4.054±0.053 | 1.308±0.042 | 0.236±0.117 | 0.170±0.016 | 4.037±0.016 | 311.30±1.00 |
| 540 | 213.023 3.342 | 13.975 0.171 | 5.038 0.094 | 0.179 0.091 | 0.227 0.031 | 4.049 0.024 | 312.00 1.00 |
| 580 | 489.263 4.057 | 33.988 0.224 | 12.642 0.093 | 0.140 0.030 | 0.226 0.024 | 4.068 0.033 | 312.80 1.00 |
| 620 | 988.394 11.985 | 69.399 0.839 | 26.071 0.498 | 0.245 0.074 | 0.331 0.011 | 4.097 0.043 | 313.50 1.00 |
| 660> | 1090.667 15.508 | 81.606 1.043 | 30.833 0.405 | 0.266 0.143 | 0.094 0.012 | 4.116 0.059 | 314.30 1.00 |
| < 690> | 1375.410 9.968 | 99.093 0.702 | 37.507 0.296 | 0.202 0.026 | 0.208 0.017 | 4.144 0.072 | 315.00 1.00 |
| < 730> | 937.287 8.741 | 68.509 0.624 | 26.146 0.247 | 0.262 0.093 | 0.075 0.029 | 4.192 0.096 | 315.80 1.00 |
| < 770> | 348.972 2.642 | 24.814 0.186 | 9.404 0.079 | 0.113 0.076 | 0.055 0.014 | 4.256 0.128 | 316.50 1.00 |
| < 810> | 333.796 2.639 | 23.887 0.189 | 9.076 0.072 | 0.344 0.157 | 0.058 0.014 | 4.341 0.171 | 317.30 1.00 |
| 850> | 583.241 8.731 | 31.413 0.405 | 12.092 0.160 | 0.424 0.139 | 0.599 0.012 | 4.455 0.227 | 318.00 1.00 |
| 950> | 1301.321 13.608 | 89.306 1.333 | 37.909 0.520 | 0.861 0.126 | 0.090 0.024 | 4.334 0.467 | 318.80 1.00 |
| [1050 | 894.297 7.903 | 62.316 0.549 | 23.808 0.229 | 0.893 0.068 | 0.082 0.016 | 5.020 0.960 | 319.60 0.00 |
| 1200 | 781.918 58.261 | 0.999 0.025 | 0.852 0.032 | 0.737 0.101 | 2.569 0.095 | 5.683 4.827 | 320.61 0.00 |

All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

## MW-15: GSC-62-14 Biotite 60/80




All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

## MW-27: TOBY-1 Hornblende 60/120

Run date: 1993/11/05 Can/Pos: 124/27 J Value: 0.007133
Recalc date: 1995/08/28 Mass: 80.0 mg ± 0.000042

Volume 39Ar: 19.80 x 1E-9 cm3 NTP Approx. 0.49% K
Integrated Age: 568.74 ± 6.67 Ma 8.01% Ca

Initial 40/36: 2411.20 ±2583.47 (MSWD = 22.47, isochron between 0.18 and 2.63)
Correlation Age: 282.79 ± 160.02 Ma (100.0% of 39Ar, steps marked by >)

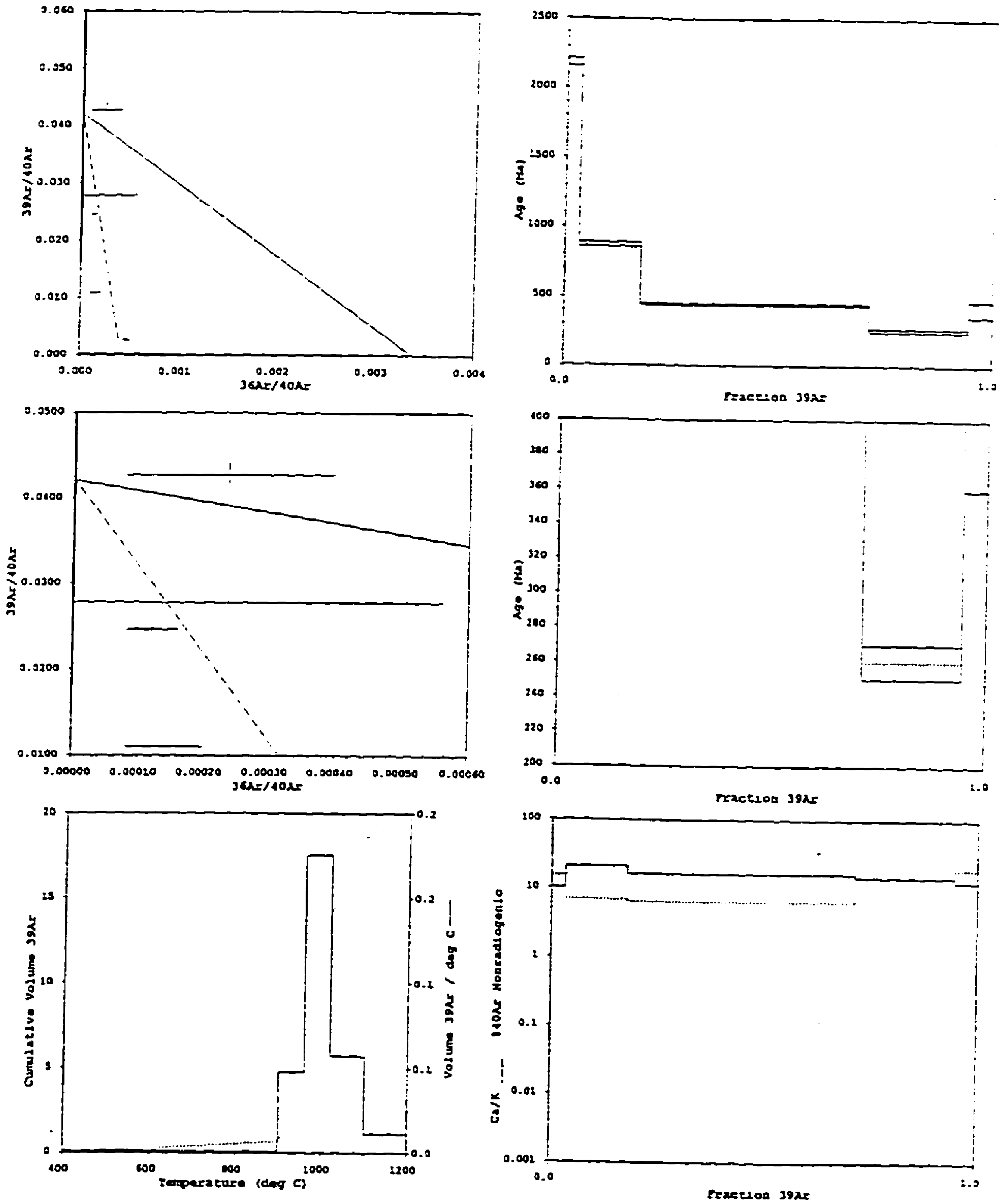
Plateau Age: 260.44 ± 9.90 Ma ( 23.1% of 39Ar, steps marked by <)

| Temp | 36Ar/40Ar | 39Ar/40Ar | r | Ca/K | %40Ar* | %39Ar | 40Ar*/39K | Age |
|---|---|---|---|---|---|---|---|---|
| 900 | 0.00046636±0.00003721 | 0.002609±0.000052 | -0.162 | 10.42 | 94.57 | 3.43 | 330.439±7.273 | 2194.88±27.88 |
| 960 | 0.00014006±0.00005929 | 0.011007±0.000222 | -0.233 | 21.09 | 93.08 | 14.52 | 87.195 2.293 | 871.72 18.83 |
| 1020 | 0.00012002±0.00003902 | 0.024622±0.000367 | -0.219 | 16.60 | 93.48 | 53.24 | 38.173 0.666 | 444.54 6.70 |
| <1100 | 0.00023479±0.00015931 | 0.042744±0.001395 | -0.604 | 14.80 | 94.97 | 23.13 | 22.772 1.873 | 260.44 9.90 |
| 1200 | -0.00004358±0.00060462 | 0.027783±0.003577 | -0.635 | 12.44 | 80.49 | 5.67 | 36.451 5.634 | 416.95 57.54 |

| Temp | 40Ar | 39Ar | 38Ar | 37Ar | 36Ar | Blank 40Ar | Atmos 40/36 |
|---|---|---|---|---|---|---|---|
| 900 | 264.247±4.035 | 0.680±0.006 | 1.046±0.025 | 3.870±0.174 | 0.138±0.007 | 4.452±2.326 | 310.68±1.00 |
| 960 | 264.271 3.347 | 2.974 0.035 | 3.545 0.057 | 33.106 0.418 | 0.062 0.013 | 5.204 2.522 | 310.77 1.00 |
| 1020 | 431.449 4.072 | 10.542 0.099 | 12.046 0.117 | 95.595 0.301 | 0.094 0.014 | 5.547 2.773 | 310.85 1.00 |
| <1100 | 113.466 0.732 | 4.590 0.020 | 5.070 0.044 | 37.041 0.230 | 0.057 0.013 | 4.752 3.276 | 310.93 1.00 |
| 1200 | 49.927 1.672 | 1.123 0.024 | 1.463 0.051 | 7.637 0.184 | 0.033 0.015 | 9.853 4.327 | 311.00 1.00 |

All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

MW-27: TOBY-1 Hornblende 60/120




All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

## MW-10: TOBY-1 Biotite 1.5 mm

| | | | | | |
|---|---|---|---|---|---|
| Run date: | 1993/10/14 | Can/Pos: | 124/25 | J Value: | 0.007138 |
| Recalc date: | 1995/08/28 | Mass: | 180.0 mg | | ± 0.000042 |

Volume 39Ar: 796.83 x 1E-9 cm3 NTP — Approx. 8.74% K
Integrated Age: 242.94 ± 1.49 Ma — 0.08% Ca

Initial 40/36: 299.57 ± 8.27 (MSWD = 1.79, isochron between 0.55 and 1.89)
Correlation Age: 242.83 ± 0.78 Ma (100.0% of 39Ar, steps marked by >)

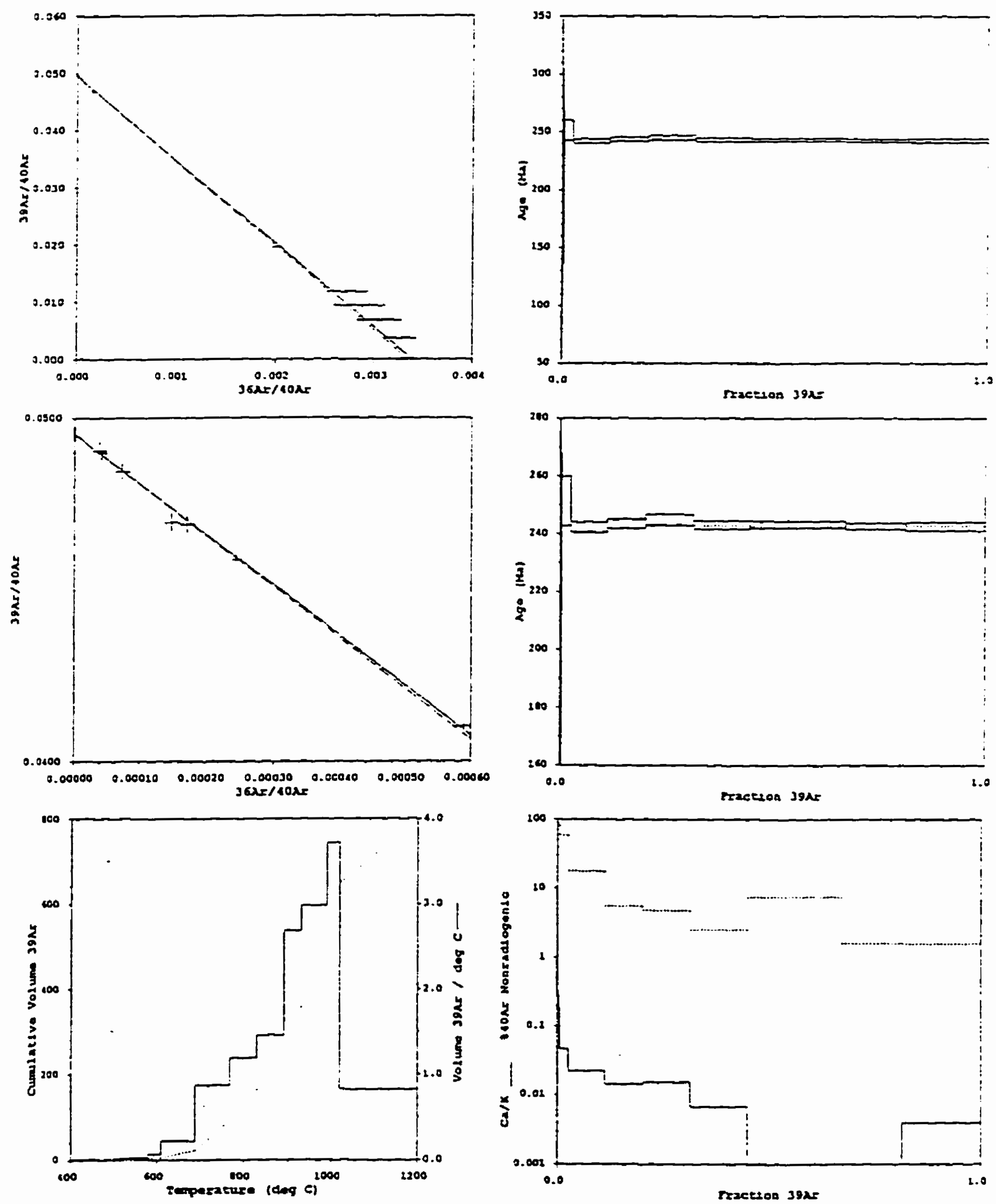
Plateau Age: 242.86 ± 1.49 Ma ( 68.5% of 39Ar, steps marked by <)

| Temp | 36Ar/40Ar | 39Ar/40Ar | r | Ca/K | %40Ar* | %39Ar | 40Ar*/39K | Age |
|---|---|---|---|---|---|---|---|---|
| 500 | 0.00326834±0.00015540 | 0.003717±0.000167 | 0.696 | 0.33 | 3.35 | 0.10 | 9.201±12.674 | 114.76±153.15 |
| 540 | 0.00304672±0.00023342 | 0.006841±0.000476 | 0.553 | 0.32 | 9.61 | 0.09 | 14.572 11.686 | 178.52 124.64 |
| 580 | 0.00294420±0.00026731 | 0.009394±0.000320 | 0.147 | 0.27 | 15.39 | 0.13 | 16.981 8.818 | 206.42 97.83 |
| 610 | 0.00271985±0.00020990 | 0.011821±0.000709 | 0.411 | 0.18 | 19.12 | 0.24 | 16.604 8.734 | 202.07 96.11 |
| 690 | 0.00199630±0.00004636 | 0.019576±0.000187 | 0.178 | 0.05 | 40.78 | 2.19 | 20.939 0.770 | 251.30 9.63 |
| 770 | 0.00058981±0.00001932 | 0.041014±0.000113 | -0.039 | 0.02 | 82.27 | 8.77 | 20.132 0.151 | 242.24 1.71 |
| 830 | 0.00017216±0.00001363 | 0.046689±0.000273 | -0.047 | 0.01 | 94.51 | 8.96 | 20.242 0.144 | 243.47 1.62 |
| 890 | 0.00014852±0.00001324 | 0.046951±0.000331 | -0.014 | 0.02 | 95.24 | 10.99 | 20.364 0.169 | 244.85 1.87 |
| < 930 | 0.00007343±0.00001195 | 0.048432±0.000255 | -0.046 | 0.01 | 97.48 | 13.43 | 20.220 0.127 | 243.30 1.43 |
| < 990 | 0.00024607±0.00000836 | 0.045870±0.000170 | 0.045 | 0.00 | 92.48 | 22.47 | 20.214 0.097 | 243.19 1.09 |
| <1020 | 0.00004153±0.00000904 | 0.048978±0.000210 | -0.113 | 0.00 | 98.40 | 14.02 | 20.167 0.099 | 242.63 1.10 |
| <1200 | 0.00003820±0.00001181 | 0.043040±0.000287 | -0.114 | 0.00 | 98.42 | 18.55 | 20.161 0.131 | 242.57 1.47 |

| Temp | 40Ar | 39Ar | 38Ar | 37Ar | 36Ar | Blank 40Ar | Atmos 40/36 |
|---|---|---|---|---|---|---|---|
| 500 | 221.268± 8.378 | 0.808±0.017 | 0.297±0.034 | 0.147±0.105 | 0.724±0.020 | 4.237±2.018 | 307.60±1.00 |
| 540 | 114.402 6.028 | 0.755 0.030 | 0.207 0.009 | 0.133 0.055 | 0.350 0.018 | 4.249 2.024 | 307.50 1.00 |
| 580 | 117.868 2.619 | 1.069 0.019 | 0.219 0.060 | 0.159 0.189 | 0.337 0.029 | 4.265 2.033 | 307.40 1.00 |
| 610 | 162.748 6.989 | 1.875 0.072 | 0.367 0.016 | 0.185 0.054 | 0.445 0.027 | 4.081 2.040 | 307.20 1.00 |
| 690 | 894.414 5.852 | 17.419 0.113 | 2.684 0.063 | 0.446 0.136 | 1.791 0.033 | 4.144 2.072 | 307.10 1.00 |
| 770 | 1710.431 3.357 | 69.896 0.103 | 9.037 0.013 | 0.955 0.096 | 1.020 0.032 | 4.256 2.128 | 307.00 1.00 |
| 830 | 1529.548 6.317 | 71.371 0.272 | 8.630 0.058 | 0.569 0.124 | 0.277 0.029 | 4.394 2.137 | 306.90 1.00 |
| 890 | 1871.391 9.205 | 87.529 0.428 | 10.582 0.133 | 0.744 0.256 | 0.293 0.023 | 4.607 2.303 | 306.80 1.00 |
| < 930 | 2226.461 8.152 | 107.449 0.389 | 12.839 0.052 | 0.403 0.081 | 0.179 0.025 | 4.839 2.405 | 306.70 1.00 |
| < 990 | 3914.111 13.231 | 179.064 0.242 | 21.251 0.036 | 0.043 0.118 | 0.978 0.031 | 5.246 2.623 | 306.50 1.00 |
| <1020 | 2290.277 6.861 | 111.744 0.312 | 13.000 0.041 | 0.051 0.023 | 0.124 0.018 | 5.547 2.773 | 306.40 1.00 |
| <1200 | 3029.739 12.657 | 147.848 0.554 | 17.098 0.082 | 0.321 0.105 | 0.148 0.032 | 9.653 4.827 | 306.30 0.00 |

All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

MW-10: TOBY-1 Biotite 1.5 mm




All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

## MW-32: TOBY-2 Biotite 1.0 mm

| | | | | | |
|---|---|---|---|---|---|
| Run date: | 1993/12/21 | Can/Pos: | 124/26 | J Value: | 0.007135 |
| Recalc date: | 1995/08/28 | Mass: | 175.0 mg | | ± 0.000042 |

Volume 39Ar: 617.94 x 1E-9 cm3 NTP — Approx. 6.98% K
Integrated Age: 285.15 ± 1.86 Ma — 0.10% Ca

Initial 40/36: 296.46 ± 7.86 (MSWD = 11.72, isochron between 0.29 and 2.41)
Correlation Age: 280.93 ± 3.41 Ma ( 44.2% of 39Ar, steps marked by >)

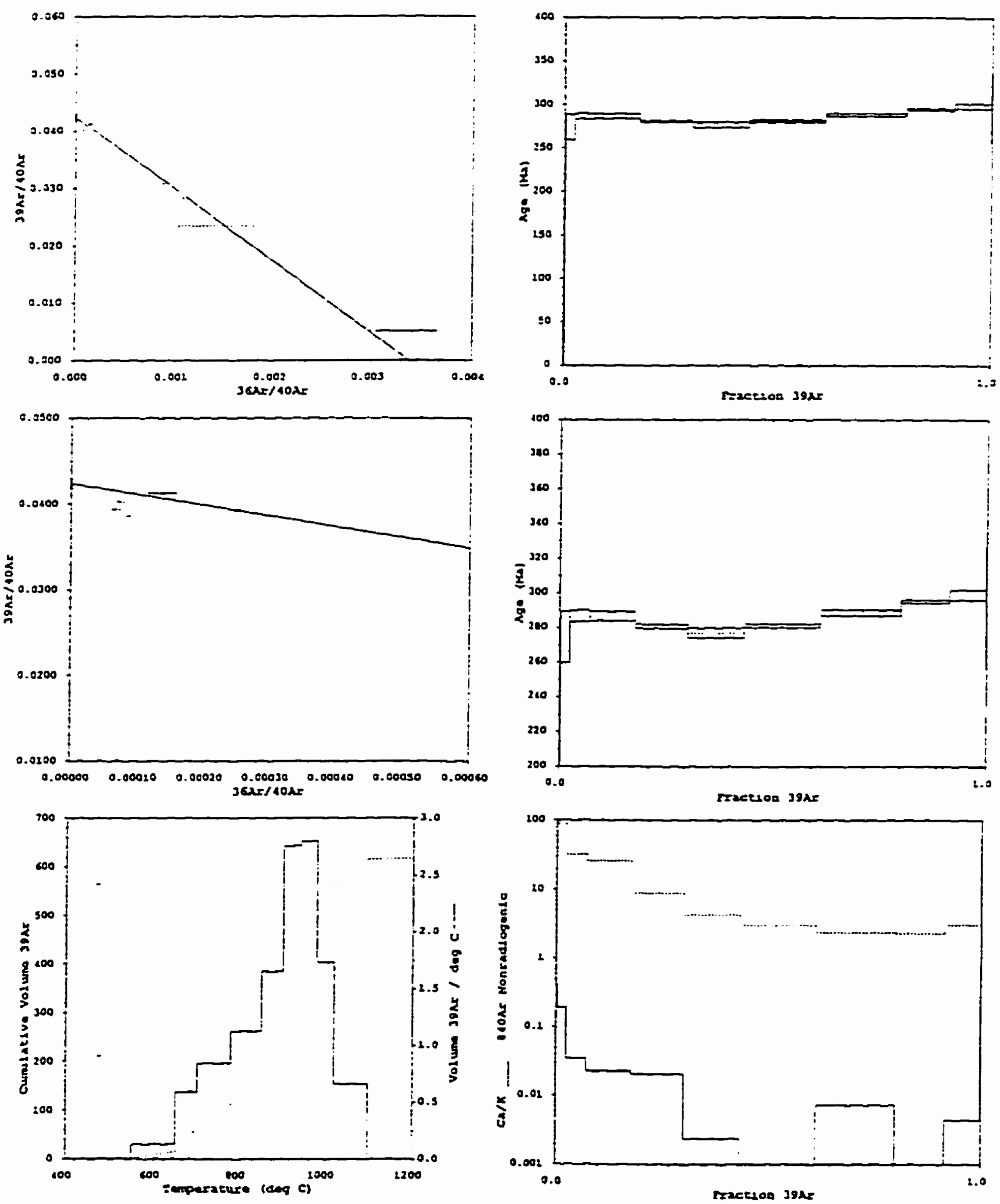
Plateau Age: 277.09 ± 3.27 Ma ( 13.4% of 39Ar, steps marked by <)

| Temp | 36Ar/40Ar | 39Ar/40Ar | r | Ca/K | %40Ar* | %39Ar | 40Ar*/39K | Age |
|---|---|---|---|---|---|---|---|---|
| 550> | 0.00334364±0.00030515 | 0.005156±0.000435 | 0.586 | 0.41 | 1.17 | 0.31 | 2.319±17.615 | 29.60±223.05 |
| 650> | 0.00299184±0.00002126 | 0.005162±0.000013 | 0.165 | 0.19 | 11.56 | 2.10 | 23.025 1.339 | 274.45 14.31 |
| 700> | 0.00107330±0.00001822 | 0.029270±0.000216 | 0.152 | 0.04 | 67.35 | 4.77 | 24.154 0.259 | 286.85 3.17 |
| 780> | 0.00056526±0.00001453 | 0.030833±0.000209 | 0.207 | 0.02 | 74.22 | 10.90 | 24.140 0.240 | 286.70 2.63 |
| 850> | 0.00029245±0.00000925 | 0.039684±0.000126 | -0.328 | 0.02 | 91.05 | 12.72 | 23.614 0.110 | 280.94 1.31 |
| < 900> | 0.00013479±0.00002250 | 0.041257±0.000361 | 0.013 | 0.00 | 95.68 | 13.37 | 23.257 0.262 | 277.09 2.99 |
| 940 | 0.00009244±0.00000656 | 0.041130±0.000129 | -0.111 | 0.00 | 96.98 | 17.53 | 23.649 0.085 | 281.30 0.94 |
| 980 | 0.00007040±0.00000910 | 0.040247±0.000226 | -0.031 | 0.01 | 97.62 | 18.12 | 24.330 0.151 | 288.75 1.66 |
| 1020 | 0.00006364±0.00000957 | 0.039366±0.000059 | -0.374 | -0.01 | 97.70 | 11.21 | 24.928 0.075 | 295.29 0.82 |
| 1100 | 0.00008492±0.00000897 | 0.038534±0.000416 | -0.159 | 0.00 | 96.90 | 8.55 | 25.267 0.272 | 299.03 2.97 |
| 1200 | 0.00139198±0.00040474 | 0.023546±0.003445 | -0.697 | 0.78 | 45.56 | 0.13 | 25.001 3.644 | 296.13 39.80 |

| Temp | 40Ar | 39Ar | 38Ar | 37Ar | 36Ar | Blank 40Ar | Atmos 40/36 |
|---|---|---|---|---|---|---|---|
| 550> | 373.908±24.843 | 1.907±0.097 | 0.740±0.057 | 0.429±0.286 | 1.250±0.076 | 4.052±2.026 | 329.98±1.00 |
| 650> | 2517.724 4.390 | 12.973 0.021 | 5.169 0.075 | 1.380 0.420 | 7.508 0.052 | 4.108 2.054 | 329.85 1.00 |
| 700> | 1046.555 5.529 | 29.445 0.150 | 6.138 0.032 | 0.566 0.372 | 1.132 0.017 | 4.154 2.077 | 323.71 1.00 |
| 780> | 2189.741 11.213 | 67.325 0.291 | 13.668 0.144 | 0.846 0.210 | 1.304 0.030 | 4.275 2.137 | 329.56 1.00 |
| 850> | 2038.447 6.290 | 78.596 0.101 | 15.146 0.129 | 0.891 0.355 | 0.609 0.017 | 4.455 2.227 | 329.45 1.00 |
| < 900> | 2009.005 14.439 | 82.616 0.399 | 15.821 0.115 | 0.107 0.407 | 0.286 0.044 | 4.652 2.326 | 329.31 1.00 |
| 940 | 2686.521 5.313 | 110.167 0.229 | 20.453 0.050 | 0.028 0.166 | 0.264 0.016 | 4.870 2.435 | 329.18 1.00 |
| 980 | 2791.162 11.796 | 111.999 0.399 | 19.298 0.076 | 0.450 0.115 | 0.213 0.024 | 5.160 2.580 | 329.04 1.00 |
| 1020 | 1767.721 2.509 | 69.292 0.054 | 11.104 0.038 | -0.431 -0.561 | 0.131 0.014 | 5.547 2.773 | 328.91 1.00 |
| 1100 | 1377.567 10.261 | 52.833 0.392 | 7.704 0.059 | 0.129 0.577 | 0.139 0.005 | 6.752 3.376 | 328.78 1.00 |
| 1200 | 42.993 0.640 | 0.785 0.005 | 0.114 0.035 | 0.334 0.150 | 0.079 0.009 | 9.653 4.327 | 328.64 1.00 |

All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

MW-32: TOBY-2 Biotite 1.0 mm




All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

# MW-05: GSC-61-18 F.G. Biotite 80/120

| | | | | | |
|---|---|---|---|---|---|
| Run date: | 1993/10/13 | Can/Pos: | 124/28 | J Value: | 0.007130 |
| Recalc date: | 1995/08/28 | Mass: | 200.0 mg | | ± 0.000044 |

Volume 39Ar: 873.22 x 1E-9 cm3 NTP — Approx. 8.63% K
Integrated Age: 119.44 ± 0.87 Ma — 0.28% Ca

Initial 40/36: 256.55 ± 23.78 (MSWD = 1.54, isochron between 0.47 and 2.07)
Correlation Age: 120.57 ± 0.67 Ma ( 83.6% of 39Ar, steps marked by >)

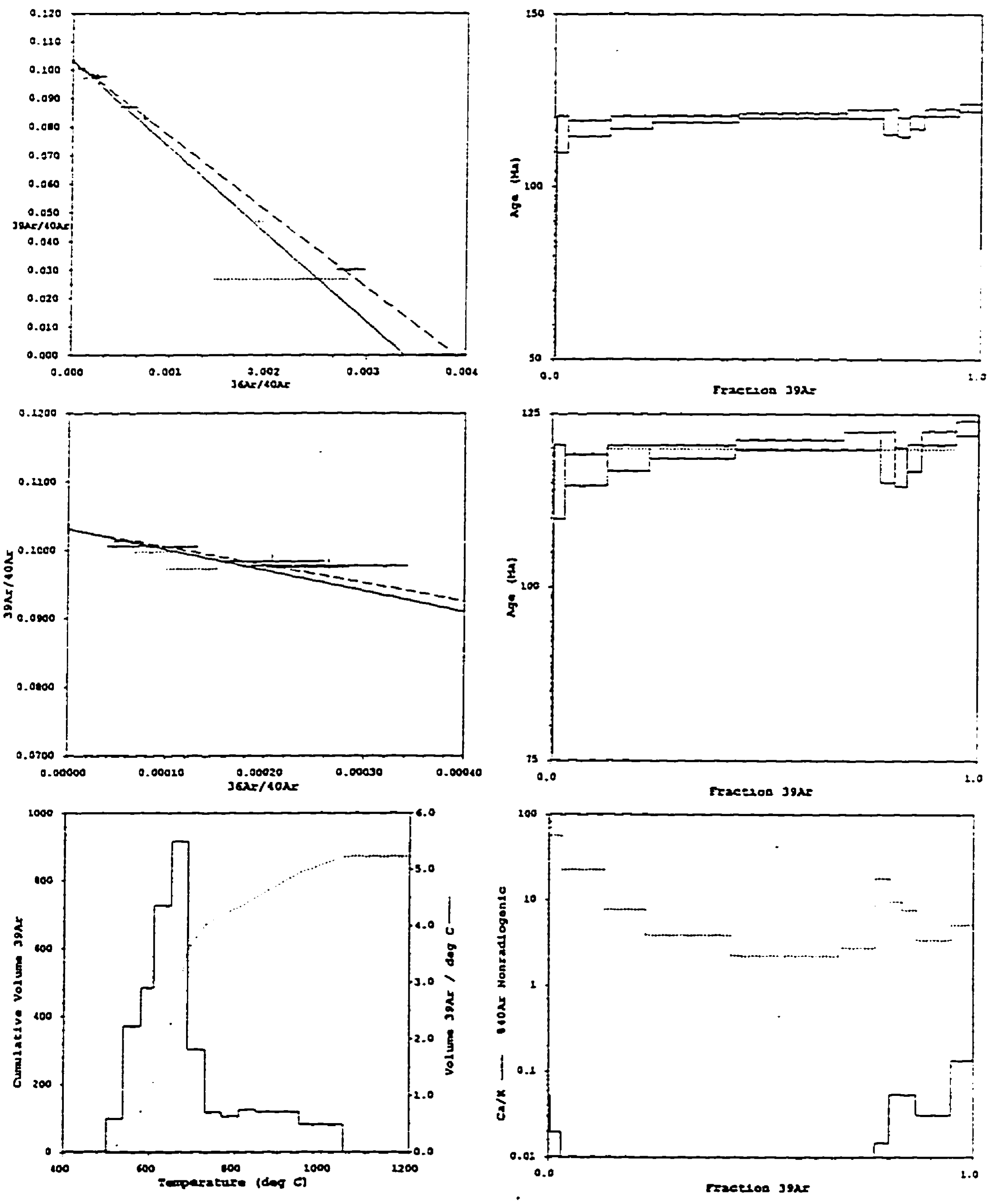
Plateau Age: 119.96 ± 0.86 Ma ( 81.0% of 39Ar, steps marked by <)

| Temp | 36Ar/40Ar | 39Ar/40Ar | r | Ca/K | %40Ar* | %39Ar | 40Ar*/39K | Age |
|---|---|---|---|---|---|---|---|---|
| 500> | 0.00282978±0.00014461 | 0.030176±0.000507 | 0.044 | 0.05 | 15.96 | 0.55 | 5.428±1.426 | 68.51±17.66 |
| 540 | 0.00191052±0.00005199 | 0.047086±0.000991 | 0.377 | 0.02 | 43.13 | 2.70 | 9.248 0.440 | 115.20 5.30 |
| 580> | 0.00074256±0.00003401 | 0.083161±0.001160 | 0.134 | 0.01 | 77.58 | 10.22 | 9.356 0.190 | 116.97 2.29 |
| < 610> | 0.00024019±0.00003929 | 0.097527±0.000967 | -0.012 | 0.01 | 92.22 | 9.95 | 9.526 0.111 | 118.55 1.32 |
| < 650> | 0.00011570±0.00001573 | 0.100556±0.000666 | -0.020 | 0.00 | 96.07 | 19.97 | 9.605 0.078 | 119.50 0.94 |
| < 690> | 0.00006004±0.00001406 | 0.101330±0.000464 | -0.038 | 0.01 | 97.75 | 25.17 | 9.694 0.060 | 120.57 0.72 |
| < 730> | 0.00006590±0.00002712 | 0.100644±0.000791 | -0.123 | 0.01 | 97.20 | 8.29 | 9.743 0.104 | 121.16 1.25 |
| < 770> | 0.00057324±0.00009054 | 0.086954±0.000748 | -0.145 | 0.01 | 81.77 | 3.20 | 9.552 0.227 | 118.87 2.70 |
| < 810> | 0.00026409±0.00007645 | 0.097764±0.001010 | -0.267 | 0.06 | 90.40 | 2.88 | 9.430 0.231 | 117.40 2.73 |
| < 850> | 0.00020752±0.00005139 | 0.098387±0.001079 | -0.230 | 0.05 | 92.22 | 3.40 | 9.541 0.161 | 118.73 1.94 |
| < 950 | 0.00008436±0.00002104 | 0.099771±0.000710 | -0.255 | 0.03 | 96.55 | 8.10 | 9.773 0.081 | 121.53 0.99 |
| 1050 | 0.00012574±0.00002620 | 0.097297±0.000866 | -0.493 | 0.14 | 94.82 | 5.48 | 9.896 0.085 | 123.00 1.02 |
| 1200 | 0.00210812±0.00069900 | 0.026925±0.005453 | -0.192 | 15.83 | 26.36 | 0.08 | 14.004 7.682 | 171.70 89.50 |

| Temp | 40Ar | 39Ar | 38Ar | 37Ar | 36Ar | Blank 40Ar | Atmos 40/36 |
|---|---|---|---|---|---|---|---|
| 500> | 164.000± 1.330 | 4.823±0.035 | 0.560±0.050 | 0.137±0.119 | 0.466±0.023 | 4.037±2.018 | 306.90±1.00 |
| 540 | 506.024 7.536 | 23.604 0.327 | 2.521 0.058 | 0.253 0.146 | 0.972 0.022 | 4.049 2.024 | 309.70 1.00 |
| 580> | 1079.474 10.616 | 59.220 0.859 | 8.951 0.102 | 0.446 0.113 | 0.811 0.035 | 4.065 2.033 | 308.50 1.00 |
| < 610> | 897.905 6.200 | 86.929 0.590 | 8.596 0.072 | 0.318 0.185 | 0.228 0.034 | 4.081 2.040 | 308.30 1.00 |
| < 650> | 1743.070 3.096 | 174.361 0.790 | 17.249 0.089 | 0.413 0.147 | 0.215 0.026 | 4.108 2.054 | 308.10 1.00 |
| < 690> | 2179.597 6.311 | 219.801 0.689 | 21.506 0.073 | 0.629 0.136 | 0.144 0.030 | 4.144 2.072 | 307.90 1.00 |
| < 730> | 725.198 3.854 | 72.356 0.357 | 7.000 0.048 | 0.372 0.034 | 0.062 0.019 | 4.192 2.096 | 307.70 1.00 |
| < 770> | 326.553 1.365 | 27.955 0.097 | 2.726 0.042 | 0.226 0.057 | 0.199 0.028 | 4.256 2.129 | 307.50 1.00 |
| < 810> | 262.417 1.164 | 25.161 0.098 | 2.440 0.055 | 0.757 0.129 | 0.083 0.019 | 4.341 2.171 | 307.30 1.00 |
| < 850> | 307.172 1.831 | 29.700 0.160 | 2.876 0.050 | 0.879 0.023 | 0.078 0.014 | 4.455 2.227 | 307.10 1.00 |
| < 950 | 715.889 3.256 | 70.731 0.295 | 6.923 0.034 | 1.190 0.107 | 0.077 0.012 | 4.934 2.467 | 306.90 1.00 |
| 1050 | 499.183 2.358 | 47.862 0.214 | 4.627 0.024 | 3.538 0.113 | 0.083 0.009 | 5.920 2.960 | 306.70 1.00 |
| 1200 | 36.265 2.229 | 0.720 0.023 | 0.118 0.010 | 6.229 0.208 | 0.090 0.017 | 9.653 4.827 | 306.50 1.00 |

All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

MW-05: GSC-61-18 F.G. Biotite 80/120




All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

## MW-35: GC-1 Biotite 40/60

| | | | | | |
|---|---|---|---|---|---|
| Run date: | 1994/06/11 | Can/Pos: | 132/41 | J Value: | 0.007238 |
| Recalc date: | 1995/08/28 | Mass: | 156.0 mg | | ± 0.000020 |

Volume 39Ar: 527.09 x 1E-9 cm3 NTP — Approx. 6.58% K
Integrated Age: 146.18 ± 1.14 Ma — 0.72% Ca

Initial 40/36: 326.08 ± 56.39 (MSWD = 1.86, isochron between 0.29 and 2.41)
Correlation Age: 145.26 ± 1.63 Ma ( 63.9% of 39Ar, steps marked by >)

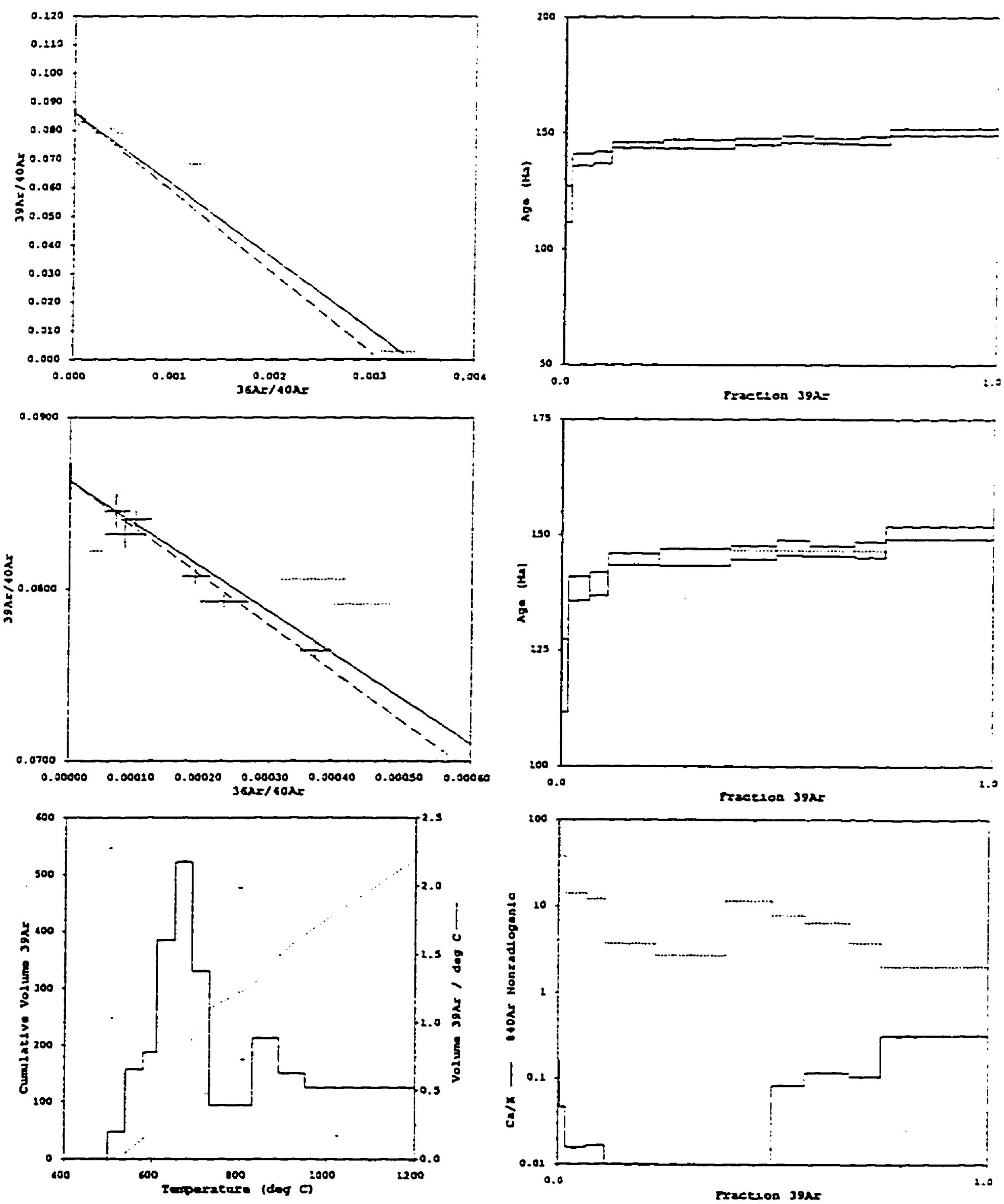
Plateau Age: 146.67 ± 0.83 Ma ( 35.1% of 39Ar, steps marked by <)

| Temp | 36Ar/40Ar | 39Ar/40Ar | r | Ca/K | %40Ar* | %39Ar | 40Ar*/39K | Age |
|---|---|---|---|---|---|---|---|---|
| 500 | 0.00323083±0.00018717 | 0.002952±0.000165 | 0.495 | 0.16 | 4.50 | 0.38 | 15.341±19.202 | 189.85±225.63 |
| 540 | 0.00119709±0.00013883 | 0.068323±0.002029 | -0.066 | 0.05 | 62.36 | 1.53 | 9.459 1.646 | 119.47 7.30 |
| 580 | 0.00043483±0.00004466 | 0.079141±0.001089 | -0.116 | 0.02 | 85.90 | 4.99 | 11.012 0.212 | 139.35 2.57 |
| 610 | 0.00036100±0.00004995 | 0.080570±0.001014 | -0.203 | 0.02 | 87.89 | 4.46 | 11.088 0.227 | 139.27 2.52 |
| 650> | 0.00009961±0.00002276 | 0.084100±0.000525 | -0.155 | 0.01 | 96.30 | 12.18 | 11.541 0.100 | 144.73 1.20 |
| 690> | 0.00007038±0.00001953 | 0.084576±0.001051 | -0.026 | 0.01 | 97.29 | 16.53 | 11.578 0.183 | 145.18 1.90 |
| < 730> | 0.00036737±0.00002504 | 0.076446±0.000552 | -0.087 | 0.00 | 88.43 | 10.45 | 11.661 0.123 | 146.18 1.49 |
| < 830> | 0.00023136±0.00003724 | 0.079290±0.000416 | -0.312 | 0.08 | 92.11 | 7.44 | 11.751 0.134 | 147.25 1.61 |
| < 890> | 0.00018939±0.00002100 | 0.080759±0.000492 | -0.255 | 0.12 | 93.52 | 10.11 | 11.693 0.183 | 146.57 1.29 |
| < 950> | 0.00008348±0.00003130 | 0.083215±0.000868 | -0.232 | 0.11 | 96.22 | 7.15 | 11.721 0.142 | 146.90 1.71 |
| 1200 | 0.00003648±0.00001335 | 0.082230±0.000817 | -0.228 | 0.32 | 97.39 | 24.79 | 12.031 0.119 | 150.62 1.43 |

| Temp | 40Ar | 39Ar | 38Ar | 37Ar | 36Ar | Blank 40Ar | Atmos 40/36 |
|---|---|---|---|---|---|---|---|
| 500 | 680.386±27.101 | 1.999±0.079 | 0.315±0.115 | 0.176±0.196 | 2.201±0.092 | 4.037±2.018 | 312.00±1.00 |
| 540 | 122.585 2.077 | 8.093 0.135 | 1.563 0.035 | 0.205 0.086 | 0.155 0.016 | 4.249 2.224 | 312.08 1.00 |
| 580 | 336.866 3.141 | 26.279 0.207 | 4.941 0.051 | 0.229 0.208 | 0.158 0.013 | 4.265 2.233 | 312.16 1.00 |
| 610 | 296.341 2.195 | 23.493 0.171 | 4.344 0.044 | 0.213 0.022 | 0.119 0.013 | 4.282 2.240 | 312.25 1.00 |
| 650> | 769.319 3.153 | 64.199 0.245 | 11.716 0.056 | 0.306 0.090 | 0.090 0.016 | 4.108 2.354 | 312.33 1.00 |
| 690> | 1036.833 10.359 | 87.130 0.609 | 15.838 0.122 | 0.295 0.153 | 0.087 0.019 | 4.144 2.072 | 312.42 1.00 |
| < 730> | 726.302 3.725 | 55.082 0.226 | 9.973 0.093 | 0.143 0.136 | 0.279 0.017 | 4.192 2.096 | 312.50 1.00 |
| < 830> | 499.821 1.189 | 39.194 0.054 | 7.226 0.011 | 1.770 0.173 | 0.130 0.017 | 4.334 2.137 | 312.58 1.00 |
| < 890> | 665.833 2.412 | 53.279 0.180 | 10.133 0.050 | 3.435 0.188 | 0.141 0.012 | 4.607 2.303 | 312.67 1.00 |
| < 950> | 458.780 3.279 | 37.678 0.194 | 7.530 0.056 | 2.197 0.140 | 0.055 0.012 | 4.934 2.467 | 312.75 1.00 |
| 1200 | 1602.354 11.759 | 130.674 0.768 | 29.073 0.165 | 22.753 0.229 | 0.096 0.014 | 9.653 4.827 | 312.84 1.00 |

All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

MW-35: GC-1 Biotite 40/60




All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

## MW-40: GC-1 Hornblende 40/60

| | | | | | |
|---|---|---|---|---|---|
| Run date: | 1994/06/12 | Can/Pos: | 132/43 | J Value: | 0.007234 |
| Recalc date: | 1995/08/28 | Mass: | 433.0 mg | | ± 0.000020 |

Volume 39Ar: 229.71 x 1E-9 cm3 NTP  Approx. 1.03% K
Integrated Age: 200.28 ± 9.87 Ma  5.96% Ca

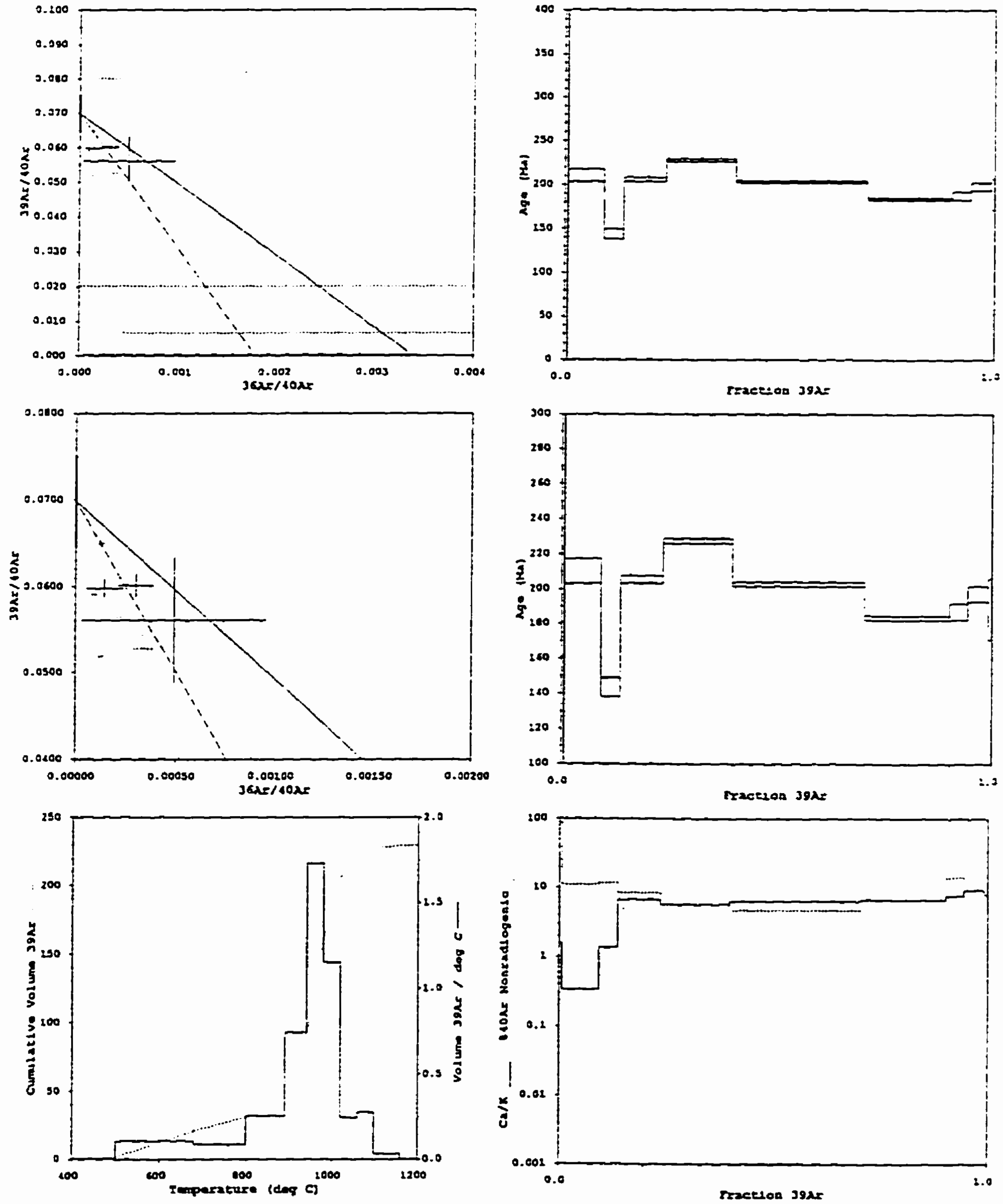
Initial 40/36: 564.03 ± 527.08 (MSWD = 5.15, isochron between 0.00 and 3.00)
Correlation Age: 177.75 ± 12.72 Ma ( 29.9% of 39Ar, steps marked by >)

| Temp | 36Ar/40Ar | 39Ar/40Ar | r | Ca/K | %40Ar* | %39Ar | 40Ar*/39K | | Age | |
|---|---|---|---|---|---|---|---|---|---|---|
| 500 | 0.00301777±0.00263434 | 0.006651±0.005655 | 0.532 | 1.55 | 10.60 | 0.67 | 16.276±124.362 | | 200.81±1459.07 | |
| 680 | 0.00033455±0.00005131 | 0.052751±0.001596 | 0.047 | 0.34 | 88.91 | 5.43 | 17.083 | 0.603 | 210.21 | 7.01 |
| 800 | 0.00027600±0.00013316 | 0.080135±0.001435 | -0.365 | 1.35 | 88.23 | 4.49 | 11.453 | 0.458 | 143.60 | 5.52 |
| 890 | 0.00019507±0.00003751 | 0.056457±0.000367 | -0.412 | 6.68 | 91.50 | 10.04 | 16.692 | 0.182 | 205.66 | 2.12 |
| 940 | 0.00012326±0.00001464 | 0.051986±0.000336 | -0.380 | 5.66 | 94.37 | 16.16 | 18.571 | 0.119 | 227.41 | 1.37 |
| 980 | 0.00008901±0.00001584 | 0.059113±0.000325 | -0.170 | 6.27 | 95.26 | 30.15 | 16.472 | 0.111 | 203.10 | 1.29 |
| 1020> | 0.00012603±0.00001909 | 0.064396±0.000451 | -0.264 | 6.72 | 93.57 | 20.03 | 14.315 | 0.139 | 183.67 | 1.28 |
| 1060> | 0.00030229±0.00008338 | 0.060161±0.001416 | -0.433 | 7.64 | 85.53 | 4.27 | 15.137 | 0.418 | 187.47 | 4.80 |
| 1100> | 0.00014496±0.00009435 | 0.059798±0.001143 | -0.588 | 9.23 | 90.03 | 4.86 | 16.207 | 0.379 | 197.66 | 4.42 |
| 1160> | 0.00049106±0.00046334 | 0.056077±0.007329 | -0.797 | 7.88 | 66.47 | 0.78 | 15.245 | 1.510 | 188.73 | 17.75 |
| 1200 | 0.00329366±0.00744158 | 0.020246±0.038246 | 0.572 | 4.45 | 0.75 | 0.35 | 1.232 | 103.964 | 16.01 | 1422.43 |

| Temp | 40Ar | | 39Ar | | 38Ar | | 37Ar | | 36Ar | | Blank 40Ar | | Atmos 40/36 | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 500 | 235.479±145.454 | | 1.540±0.391 | | 1.232±0.708 | | 1.308±0.784 | | 0.712±0.423 | | 4.037±2.018 | | 312.73±1.00 | |
| 680 | 374.422 | 8.272 | 19.506 | 0.331 | 5.360 | 0.133 | 3.595 | 0.262 | 0.139 | 0.018 | 4.134 | 2.067 | 312.80 | 1.00 |
| 800 | 133.173 | 0.625 | 10.315 | 0.039 | 2.984 | 0.089 | 7.623 | 0.068 | 0.052 | 0.016 | 4.317 | 2.159 | 312.87 | 1.00 |
| 890 | 412.717 | 1.069 | 23.058 | 0.043 | 18.240 | 0.075 | 84.167 | 0.231 | 0.116 | 0.013 | 4.607 | 2.303 | 312.94 | 1.00 |
| 940 | 720.005 | 2.826 | 37.125 | 0.141 | 48.005 | 0.188 | 114.735 | 0.460 | 0.134 | 0.007 | 4.870 | 2.435 | 313.01 | 1.00 |
| 980 | 1176.257 | 4.390 | 69.265 | 0.232 | 110.850 | 0.395 | 237.370 | 1.004 | 0.152 | 0.016 | 5.160 | 2.580 | 313.08 | 1.00 |
| 1020> | 713.131 | 2.865 | 46.007 | 0.194 | 79.627 | 0.338 | 168.913 | 0.717 | 0.151 | 0.010 | 5.547 | 2.773 | 313.15 | 1.00 |
| 1060> | 169.067 | 1.695 | 9.816 | 0.097 | 17.562 | 0.177 | 40.982 | 0.407 | 0.080 | 0.011 | 6.063 | 3.032 | 313.22 | 1.00 |
| 1100> | 193.320 | 0.824 | 11.174 | 0.046 | 20.735 | 0.090 | 56.342 | 0.332 | 0.064 | 0.014 | 6.752 | 3.376 | 313.28 | 1.00 |
| 1160> | 40.320 | 0.672 | 1.801 | 0.017 | 3.240 | 0.041 | 7.757 | 0.214 | 0.046 | 0.009 | 8.239 | 4.119 | 313.35 | 1.00 |
| 1200 | 14.887 | 9.200 | 0.106 | 0.054 | 0.145 | 0.084 | 0.258 | 0.181 | 0.050 | 0.028 | 9.653 | 4.827 | 313.43 | 1.00 |

All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

MW-40: GC-1 Hornblende 40/60




All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

## D-215 : GC-1 Hornblende 18/40

Run date: 1995/11/11
Recalc date: 1996/07/19

Can/Pos: 132/42
Mass: 20.0 mg

J Value: 0.007236
± 0.000020

Volume 39Ar: 10.38 x 1E-9 cm3 NTP
Integrated Age: 171.79 ± 0.83 Ma

Approx. 1.01% K
1.92% Ca

Initial 40/36: Linear regression has positive slope.
Correlation Age:

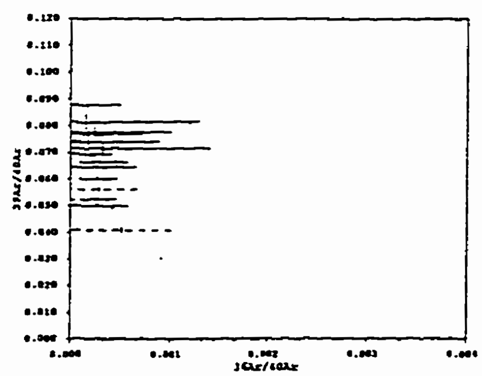
Plateau Age: 170.99 ± 2.44 Ma ( 7.6% of 39Ar, steps marked by <)
145.83 ± 0.99 Ma ( 56.5% of 39Ar, steps marked by [)

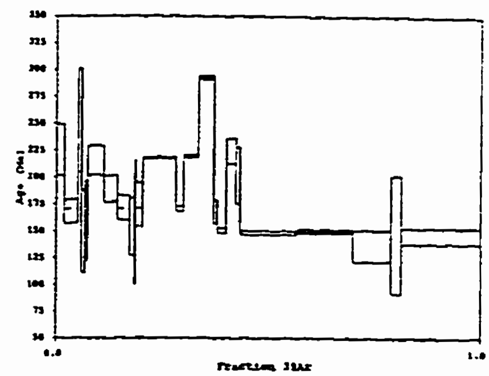
| Temp | 36Ar/40Ar | 39Ar/40Ar | r | Ca/K | %40Ar* | %39Ar | 40Ar*/39K | Age |
|---|---|---|---|---|---|---|---|---|
| 2500 | 0.00026842±0.00030400 | 0.049906±0.000794 | 0.929 | 0.00 | 91.94 | 1.95 | 18.448±2.076 | 226.06±23.91 |
| <5000 | 0.00021124±0.00019444 | 0.069285±0.000557 | 0.774 | 0.00 | 93.57 | 3.09 | 13.532 0.916 | 168.53 10.89 |
| 750 | 0.00051144±0.00050095 | 0.040815±0.001275 | 0.959 | 0.00 | 84.79 | 0.94 | 20.798 4.256 | 252.92 48.29 |
| 1000 | 0.00024200±0.00077350 | 0.077581±0.001903 | 0.898 | 0.56 | 92.46 | 0.77 | 11.968 3.214 | 149.84 38.62 |
| 1300 | 0.00017968±0.00070747 | 0.073878±0.001696 | 0.857 | 0.00 | 94.49 | 0.71 | 12.817 3.087 | 160.01 36.88 |
| 1700 | 0.00027407±0.00018173 | 0.052385±0.000511 | 0.873 | 0.00 | 91.76 | 3.72 | 17.544 1.178 | 215.61 13.64 |
| 2100 | 0.00027403±0.00018903 | 0.060061±0.000537 | 0.832 | 0.00 | 91.74 | 3.15 | 15.301 1.047 | 189.45 12.30 |
| <2500 | 0.00028418±0.00019639 | 0.066390±0.000565 | 0.815 | 0.07 | 91.41 | 2.96 | 13.798 0.973 | 171.68 11.54 |
| 2900 | 0.00018096±0.00053843 | 0.076900±0.001213 | 0.771 | 1.07 | 94.10 | 1.08 | 12.309 2.223 | 153.92 26.65 |
| 3400 | 0.00032549±0.00108462 | 0.071542±0.002408 | 0.816 | 2.24 | 89.53 | 0.49 | 12.633 4.837 | 157.81 57.85 |
| <6000 | 0.00032106±0.00034034 | 0.064459±0.000879 | 0.878 | 0.98 | 90.08 | 1.55 | 14.042 1.732 | 174.58 20.52 |
| 1000 | 0.00020830±0.00000993 | 0.052846±0.000093 | 0.005 | 1.72 | 93.33 | 8.04 | 17.758 0.066 | 218.09 0.76 |
| 750 | 0.00020603±0.00004470 | 0.068639±0.000162 | 0.006 | 4.82 | 92.36 | 2.02 | 13.682 0.195 | 170.31 2.32 |
| 750 | 0.00041721±0.00001588 | 0.048965±0.000097 | 0.031 | 5.55 | 86.42 | 3.49 | 17.905 0.104 | 219.79 1.20 |
| 1000 | 0.00090826±0.00001206 | 0.030067±0.000060 | 0.069 | 5.77 | 72.38 | 3.76 | 24.333 0.134 | 292.59 1.49 |
| 750 | 0.00035388±0.00020756 | 0.066408±0.000315 | 0.002 | 6.09 | 87.68 | 0.88 | 13.484 0.926 | 167.95 11.01 |
| 750 | 0.00026811±0.00005080 | 0.076540±0.000189 | 0.003 | 2.95 | 90.94 | 2.03 | 12.030 0.199 | 150.58 2.39 |
| 800 | 0.00015180±0.00015797 | 0.052069±0.000488 | 0.760 | 6.79 | 93.91 | 2.25 | 18.344 1.033 | 224.86 11.91 |
| 500 | 0.00028631±0.00037791 | 0.056054±0.000994 | 0.880 | 4.62 | 90.32 | 1.00 | 16.331 2.253 | 201.50 26.30 |
| [1500 | 0.00042648±0.00002983 | 0.073740±0.000402 | 0.157 | 1.60 | 86.73 | 13.27 | 11.852 0.145 | 148.44 1.75 |
| [2000 | 0.00026966±0.00002525 | 0.076753±0.000417 | 0.129 | 1.84 | 91.25 | 13.23 | 11.991 0.124 | 150.11 1.49 |
| [2000 | 0.00016939±0.00032629 | 0.087610±0.000980 | 0.825 | 1.10 | 94.36 | 8.83 | 10.843 1.203 | 136.27 14.56 |
| [1500 | 0.00015329±0.00114620 | 0.081451±0.003040 | 0.935 | 1.58 | 94.72 | 2.44 | 11.721 4.571 | 146.87 55.00 |
| [1500 | 0.00020344±0.00015056 | 0.081084±0.000591 | 0.628 | 1.61 | 93.23 | 18.95 | 11.591 0.605 | 145.31 7.29 |

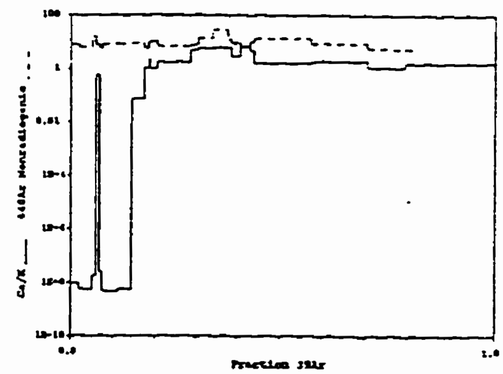
| Temp | 40Ar | 39Ar | 38Ar | 37Ar | 36Ar | Blank 40Ar | Atmos 40/36 |
|---|---|---|---|---|---|---|---|
| 2500 | 4.036±0.000 | 0.202±0.001 | 0.294±0.003 | 0.000±-0.233 | 0.001±0.000 | 0.007±0.000 | 287.67±1.00 |
| <5000 | 4.622 0.000 | 0.321 0.001 | 0.228 0.002 | 0.000 -0.226 | 0.001 0.000 | 0.006 0.000 | 287.67 1.00 |
| 750 | 2.387 0.000 | 0.098 0.001 | 0.035 0.000 | 0.000 -0.186 | 0.002 0.000 | 0.006 0.000 | 287.67 1.00 |
| 1000 | 1.032 0.000 | 0.080 0.000 | 0.020 0.000 | 0.000 0.224 | 0.001 0.000 | 0.006 0.000 | 287.67 1.00 |
| 1300 | 0.994 0.000 | 0.074 0.000 | 0.026 0.000 | 0.000 -0.181 | 0.001 0.000 | 0.006 0.000 | 287.67 1.00 |
| 1700 | 7.348 0.000 | 0.386 0.002 | 0.404 0.003 | 0.000 -0.268 | 0.002 0.000 | 0.006 0.000 | 287.67 1.00 |
| 2100 | 5.420 0.000 | 0.327 0.001 | 0.433 0.004 | 0.000 -0.231 | 0.002 0.000 | 0.006 0.000 | 287.67 1.00 |
| <2500 | 4.617 0.000 | 0.308 0.001 | 0.400 0.003 | 0.000 0.225 | 0.002 0.000 | 0.006 0.000 | 287.67 1.00 |
| 2900 | 1.459 0.000 | 0.113 0.001 | 0.067 0.001 | 0.001 0.194 | 0.001 0.000 | 0.007 0.000 | 287.67 1.00 |
| 3400 | 0.714 0.000 | 0.051 0.000 | 0.043 0.000 | 0.001 0.181 | 0.001 0.000 | 0.006 0.000 | 287.67 1.00 |
| <6000 | 2.478 0.000 | 0.160 0.001 | 0.177 0.002 | 0.001 0.200 | 0.001 0.000 | 0.006 0.000 | 287.67 1.00 |
| 1000 | 15.707 0.000 | 0.835 0.001 | 0.519 0.002 | 0.043 0.005 | 0.004 0.000 | 0.015 0.000 | 297.86 1.00 |
| 750 | 3.031 0.000 | 0.209 0.000 | 0.294 0.001 | 0.031 0.003 | 0.001 0.000 | 0.013 0.000 | 297.86 1.00 |
| 750 | 7.349 0.000 | 0.363 0.001 | 0.419 0.001 | 0.061 0.006 | 0.004 0.000 | 0.014 0.000 | 297.86 1.00 |
| 1000 | 12.864 0.000 | 0.390 0.001 | 0.563 0.002 | 0.068 0.007 | 0.013 0.000 | 0.015 0.000 | 297.86 1.00 |
| 750 | 1.368 0.000 | 0.092 0.000 | 0.152 0.001 | 0.017 0.003 | 0.001 0.000 | 0.016 0.000 | 297.86 1.00 |
| 750 | 2.744 0.000 | 0.211 0.001 | 0.185 0.001 | 0.019 0.003 | 0.001 0.000 | 0.015 0.000 | 297.86 1.00 |
| 800 | 4.450 0.000 | 0.233 0.001 | 0.346 0.004 | 0.003 0.137 | 0.001 0.000 | 0.008 0.000 | 297.86 1.00 |
| 500 | 1.836 0.000 | 0.103 0.001 | 0.158 0.002 | 0.001 0.145 | 0.001 0.000 | 0.007 0.000 | 297.86 1.00 |
| [1500 | 18.619 0.000 | 1.377 0.007 | 0.571 0.006 | 0.004 0.146 | 0.009 0.000 | 0.007 0.000 | 297.86 1.00 |
| [2000 | 17.837 0.000 | 1.373 0.007 | 0.406 0.004 | 0.004 0.123 | 0.005 0.000 | 0.007 0.000 | 297.86 1.00 |
| [2000 | 10.439 0.000 | 0.916 0.006 | 0.145 0.002 | 0.001 1.148 | 0.002 0.001 | 0.004 0.000 | 297.86 1.00 |
| [1500 | 3.110 0.000 | 0.254 0.003 | 0.047 0.001 | 0.000 1.117 | 0.001 0.001 | 0.004 0.000 | 297.86 1.00 |
| [1500 | 24.201 0.001 | 1.967 0.011 | 0.322 0.004 | 0.002 1.135 | 0.006 0.001 | 0.004 0.000 | 297.86 1.00 |

All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

D-215 : GC-1 Hornblende 18/40







All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

## D-270 : GC-3-7 Muscovite H.P.

Run date: 1995/11/11    Can/Pos: 145/38    J Value: 0.007061
Recalc 18-Jun-97    Mass: 5.0 mg    ± 0.000028

Volume 39Ar: 22.59 x 1E-9 cm3 NTP    Approx. 9.02% K
Integrated Age: 157.31 ± 0.68 Ma    0.03% Ca

Initial 40/36: 368.67 ± 38.37 (MSWD = 17.90, isochron between 0.65 and 1.71)
Correlation Age: 155.47 ± 1.37 Ma (100.0% of 39Ar, steps marked by >)

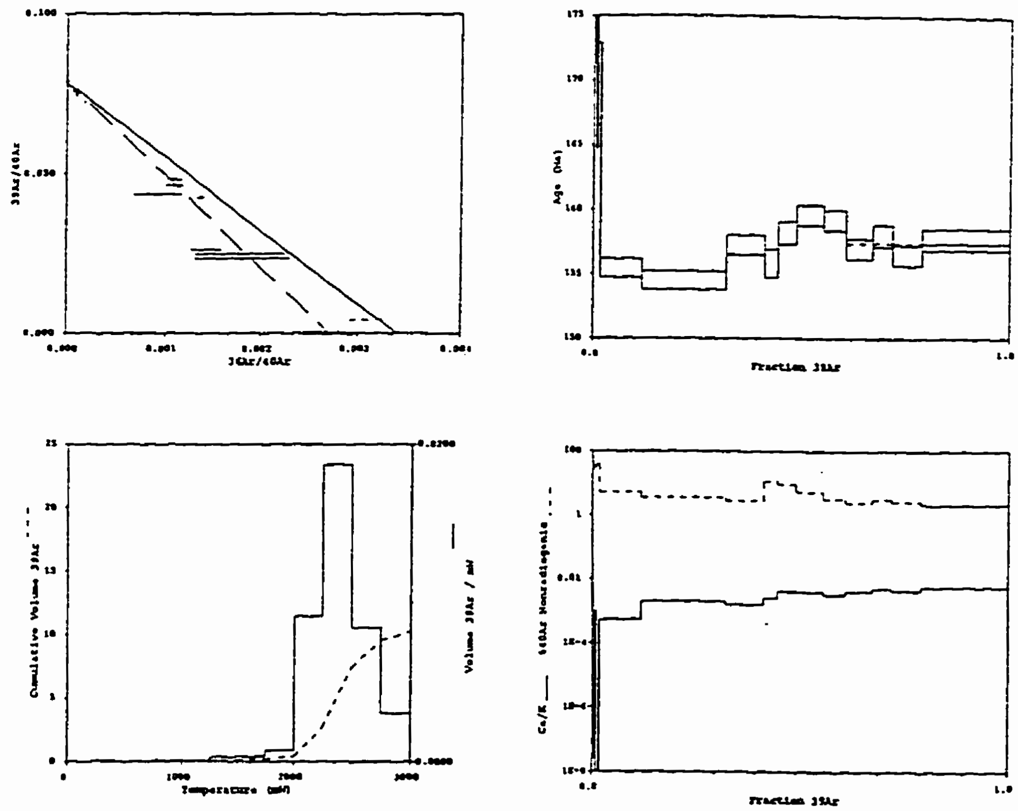
Plateau Age: 157.43 ± 0.79 Ma ( 38.4% of 39Ar, steps marked by <)
154.84 ± 0.80 Ma ( 30.6% of 39Ar, steps marked by [)

| Temp | 36Ar/40Ar | 39Ar/40Ar | r | Ca/K | %40Ar* | %39Ar | 40Ar*/39K | Age |
|---|---|---|---|---|---|---|---|---|
| 250 | 0.00304565±0.00019272 | 0.004246±0.000421 | 0.502 | 0.00 | 10.00 | 0.02 | 23.555±10.294 | 277.57±112.43 |
| 500> | 0.00177683±0.00047320 | 0.025077±0.000839 | 0.012 | 0.04 | 47.46 | 0.04 | 18.940 5.592 | 226.44 62.83 |
| 750> | 0.00180333±0.00049455 | 0.023583±0.000654 | 0.016 | 0.02 | 46.68 | 0.05 | 19.807 6.196 | 236.16 69.25 |
| 1000> | 0.00142362±0.00016417 | 0.026373±0.000329 | 0.028 | 0.01 | 57.89 | 0.11 | 21.967 1.841 | 260.13 20.30 |
| 1250> | 0.00091840±0.00024792 | 0.043700±0.000532 | 0.007 | 0.00 | 72.77 | 0.11 | 16.673 1.684 | 200.79 19.20 |
| 1500> | 0.00108220±0.00009088 | 0.046310±0.000271 | 0.006 | 0.00 | 67.93 | 0.32 | 14.688 0.585 | 178.03 6.76 |
| 1750> | 0.00109179±0.00007964 | 0.048265±0.000270 | 0.002 | 0.00 | 67.64 | 0.37 | 14.035 0.495 | 170.47 5.73 |
| 2000> | 0.00136392±0.00003811 | 0.042739±0.000197 | 0.003 | 0.00 | 59.62 | 0.80 | 13.968 0.274 | 169.69 3.18 |
| [2250> | 0.00017327±0.00000664 | 0.074425±0.000320 | 0.000 | 0.00 | 94.68 | 10.12 | 12.748 0.062 | 155.50 0.73 |
| [2500> | 0.00011149±0.00000529 | 0.076362±0.000328 | 0.000 | 0.00 | 96.49 | 20.72 | 12.664 0.060 | 154.51 0.70 |
| 2750> | 0.00008820±0.00000891 | 0.075488±0.000325 | 0.000 | 0.00 | 97.18 | 9.29 | 12.902 0.067 | 157.29 0.78 |
| 3000> | 0.00038938±0.00001769 | 0.069248±0.000299 | 0.000 | 0.00 | 88.32 | 3.35 | 12.779 0.095 | 155.86 1.11 |
| 3250> | 0.00029653±0.00001073 | 0.070284±0.000303 | 0.000 | 0.00 | 91.05 | 4.67 | 12.981 0.074 | 158.22 0.86 |
| 3500> | 0.00017213±0.00000867 | 0.072462±0.000312 | 0.000 | 0.00 | 94.72 | 6.64 | 13.098 0.068 | 159.59 0.79 |
| 4000> | 0.00009867±0.00000837 | 0.074304±0.000320 | 0.000 | 0.00 | 96.88 | 5.38 | 13.066 0.067 | 159.21 0.78 |
| <4500> | 0.00007731±0.00000990 | 0.075884±0.000327 | 0.000 | 0.00 | 97.50 | 6.36 | 12.877 0.069 | 157.00 0.80 |
| <5000> | 0.00009771±0.00001124 | 0.074913±0.000323 | 0.000 | 0.01 | 96.90 | 4.87 | 12.963 0.072 | 158.01 0.85 |
| <5500> | 0.00008296±0.00000928 | 0.076016±0.000327 | 0.000 | 0.00 | 97.34 | 6.94 | 12.833 0.067 | 156.48 0.78 |
| <7000> | 0.00006380±0.00001118 | 0.075818±0.000326 | 0.000 | 0.01 | 97.90 | 20.48 | 12.941 0.072 | 157.75 0.84 |

| Temp | 40Ar | 39Ar | 38Ar | 37Ar | 36Ar | Blank 40Ar | Atmos 40/36 |
|---|---|---|---|---|---|---|---|
| 250 | 1.145±0.000 | 0.005±0.000 | 0.001±0.000 | 0.000±-0.001 | 0.004±0.000 | 0.006±0.000 | 287.67±1.00 |
| 500> | 0.367 0.000 | 0.009 0.000 | 0.001 0.001 | 0.001 0.000 | 0.001 0.000 | 0.006 0.000 | 287.67±1.00 |
| 750> | 0.444 0.000 | 0.011 0.000 | 0.001 0.001 | 0.000 0.000 | 0.001 0.000 | 0.006 0.000 | 287.67±1.00 |
| 1000> | 0.905 0.000 | 0.024 0.000 | 0.001 0.000 | 0.000 0.000 | 0.002 0.000 | 0.005 0.000 | 287.67±1.00 |
| 1250> | 0.589 0.000 | 0.026 0.000 | 0.001 0.001 | 0.000 -0.000 | 0.001 0.000 | 0.006 0.000 | 287.67±1.00 |
| 1500> | 1.554 0.000 | 0.072 0.000 | 0.001 0.001 | 0.000 -0.000 | 0.002 0.000 | 0.006 0.000 | 287.67±1.00 |
| 1750> | 1.718 0.000 | 0.083 0.000 | 0.002 0.001 | 0.000 0.000 | 0.002 0.000 | 0.006 0.000 | 287.67±1.00 |
| 2000> | 4.219 0.000 | 0.181 0.001 | 0.004 0.000 | 0.000 -0.000 | 0.006 0.000 | 0.006 0.000 | 287.67±1.00 |
| [2250> | 30.577 0.000 | 2.285 0.010 | 0.030 0.001 | 0.001 0.001 | 0.006 0.000 | 0.006 0.000 | 287.67±1.00 |
| [2500> | 61.038 0.000 | 4.680 0.020 | 0.059 0.001 | 0.002 0.001 | 0.007 0.000 | 0.006 0.000 | 287.67±1.00 |
| 2750> | 27.692 0.000 | 2.099 0.009 | 0.027 0.001 | 0.001 0.000 | 0.003 0.000 | 0.006 0.000 | 287.67±1.00 |
| 3000> | 10.888 0.000 | 0.757 0.003 | 0.011 0.000 | 0.001 0.000 | 0.005 0.000 | 0.006 0.000 | 287.67±1.00 |
| 3250> | 14.957 0.000 | 1.056 0.005 | 0.015 0.000 | 0.001 0.000 | 0.005 0.000 | 0.006 0.000 | 287.67±1.00 |
| 3500> | 20.598 0.000 | 1.499 0.006 | 0.019 0.000 | 0.001 0.000 | 0.004 0.000 | 0.006 0.000 | 287.67±1.00 |
| 4000> | 16.302 0.000 | 1.216 0.005 | 0.016 0.000 | 0.001 0.000 | 0.002 0.000 | 0.006 0.000 | 287.67±1.00 |
| <4500> | 18.843 0.000 | 1.436 0.006 | 0.019 0.000 | 0.001 0.000 | 0.002 0.000 | 0.006 0.000 | 287.67±1.00 |
| <5000> | 14.634 0.000 | 1.101 0.005 | 0.015 0.000 | 0.001 0.000 | 0.002 0.000 | 0.006 0.000 | 287.67±1.00 |
| <5500> | 20.531 0.000 | 1.567 0.007 | 0.020 0.000 | 0.001 0.000 | 0.002 0.000 | 0.006 0.000 | 287.67±1.00 |
| <7000> | 60.769 0.000 | 4.626 0.020 | 0.060 0.001 | 0.004 0.001 | 0.004 0.001 | 0.006 0.000 | 287.67±1.00 |

All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

D-270 : GC-3-7 Muscovite H.P.



All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

## MW-21: DR-2 Biotite 60/80

Run date: 1993/10/28 Can/Pos: 124/50 J Value: 0.006985
Recalc date: 1995/08/28 Mass: 276.0 mg ± 0.000044

Volume 39Ar: 1012.50 x 1E-9 cm3 NTP Approx. 7.41% K
Integrated Age: 68.75 ± 0.86 Ma 0.97% Ca

Initial 40/36: 222.85 ± 186.77 (MSWD = 1.66, isochron between 0.18 and 2.63)
Correlation Age: 69.95 ± 1.28 Ma ( 99.8% of 39Ar, steps marked by >)

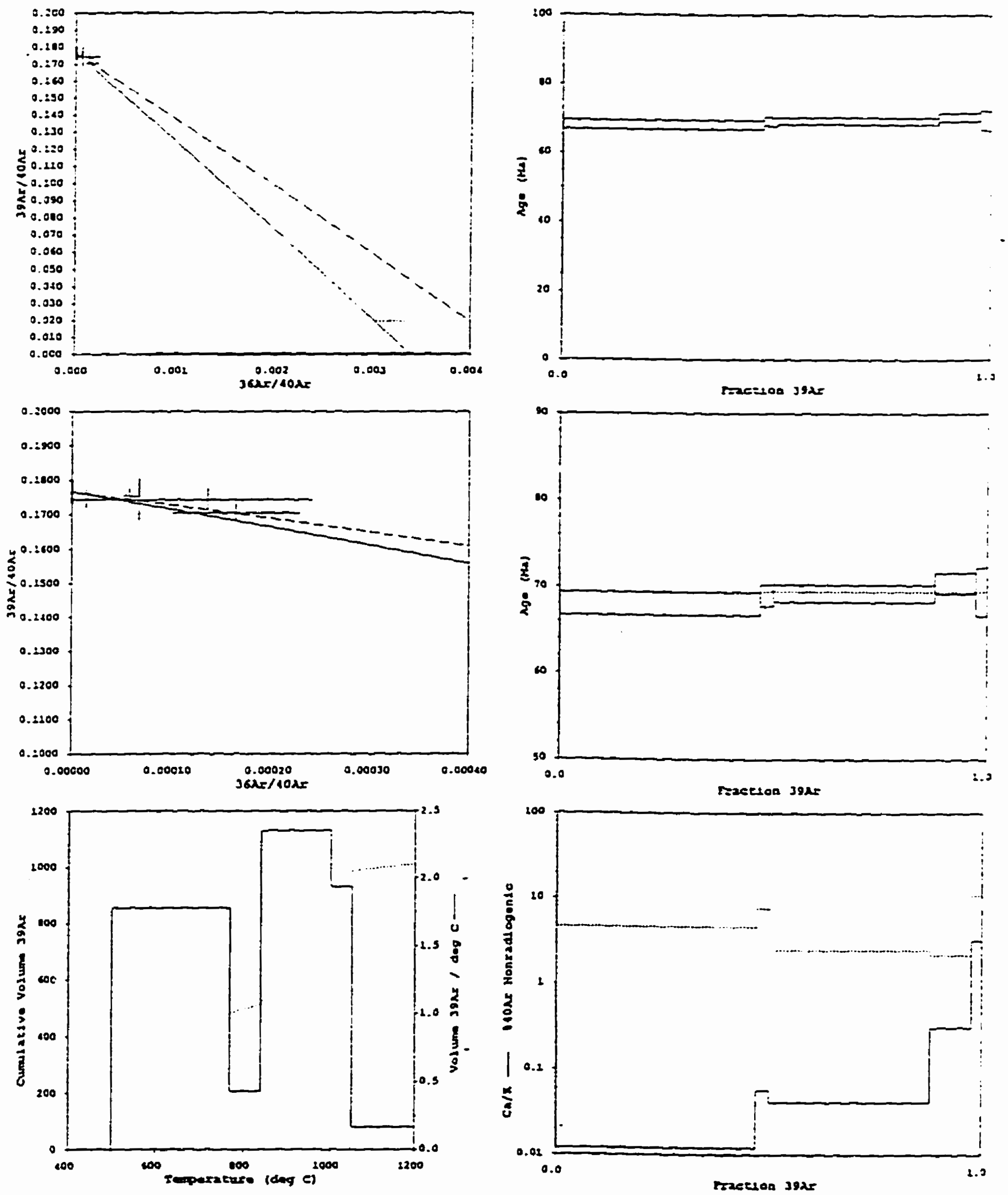
Plateau Age: 69.50 ± 0.89 Ma ( 52.2% of 39Ar, steps marked by <)

| Temp | 36Ar/40Ar | 39Ar/40Ar | r | Ca/K | %40Ar* | %39Ar | 40Ar*/39K | Age |
|---|---|---|---|---|---|---|---|---|
| 500 | 0.00316666±0.00016175 | 0.019671±0.000827 | 0.466 | 0.36 | 6.16 | 0.21 | 3.266±1.502 | 40.70±30.82 |
| 770> | 0.00013464±0.00000555 | 0.174458±0.003369 | 0.219 | 0.01 | 95.39 | 47.56 | 5.534 0.109 | 68.06 1.32 |
| < 840> | 0.00016275±0.00006425 | 0.170471±0.002745 | -0.479 | 0.06 | 95.35 | 2.98 | 5.594 0.135 | 69.03 1.27 |
| <1000> | 0.00005800±0.00000737 | 0.175262±0.002586 | 0.015 | 0.04 | 97.52 | 37.23 | 5.605 0.084 | 69.29 1.12 |
| <1050> | 0.00001445±0.00003253 | 0.174560±0.002820 | -0.173 | 0.31 | 97.32 | 9.58 | 5.704 0.100 | 70.49 1.22 |
| <1200> | 0.00006742±0.00017152 | 0.174353±0.006356 | -0.612 | 3.19 | 99.21 | 2.45 | 5.621 0.232 | 69.48 2.81 |

| Temp | 40Ar | 39Ar | 38Ar | 37Ar | 36Ar | Blank 40Ar | Atmos 40/36 |
|---|---|---|---|---|---|---|---|
| 500 | 111.255± 3.429 | 2.109±0.041 | 0.116±0.018 | 0.411±0.038 | 0.353±0.024 | 4.037±2.018 | 323.75±1.00 |
| 770> | 2778.387 37.799 | 481.559 6.514 | 9.744 0.136 | 3.197 0.078 | 0.387 0.013 | 4.256 2.128 | 323.12 1.00 |
| < 840> | 182.047 1.298 | 30.133 0.211 | 0.610 0.035 | 0.957 0.063 | 0.044 0.009 | 4.423 2.212 | 323.00 1.00 |
| <1000> | 2165.613 23.246 | 376.941 3.730 | 7.549 0.079 | 8.622 0.243 | 0.145 0.013 | 5.339 2.670 | 322.62 1.00 |
| <1050> | 564.157 6.042 | 96.971 1.031 | 1.988 0.030 | 16.476 0.173 | 0.032 0.016 | 5.920 2.960 | 322.50 1.00 |
| <1200> | 152.360 1.402 | 24.796 0.215 | 0.528 0.007 | 43.174 0.408 | 0.053 0.018 | 9.653 4.827 | 322.37 1.00 |

All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

MW-21: DR-2 Biotite 60/80




All volumes are x 1E-9 cm3 NTP. All errors are 2 x standard error.

# Appendix 2

# METAMORPHIC THERMOBAROMETRY

## ANALYTICAL METHODS

Mineral analyses were performed at Queen's University using an ARL-SEMQ electron microprobe with an energy-dispersive spectrometer (EDS). Operating conditions were maintained at an accelerating voltage of 15 kV and a beam current of 75 nA. Glass NBS 470 (National Bureau of Standards) was used as primary standard for analyses. Mineral compositions represent an average of three analyses, each having a count time of 100 seconds, from different points within a single grain or from neighboring grains of the same mineral species.

Data reduction was completed according to the procedure of Bence and Albee (1969). Structural formulae and ideal activities of were computed with MINPROBE, an unpublished APL software package developed by D. M. Carmichael at Queen's University. P-T phase diagrams were generated with the TWQ software of Berman (1991) using ideal activities computed by MINPROBE and the thermodynamic database of Berman (1988, 1990). Tightly clustered reaction intersections on the phase diagrams indicate a close approach to equilibrium; scattered intersections indicate disequilibrium. Errors in pressure and temperature are estimated at about +/- 1.0 - 1.5 kbar and +/- 30 - 50 °C for pelitic specimens containing the assemblage gar-plag-bt-ms-qtz; these uncertainties increase with additional phases or with different metamorphic mineral assemblages (Berman, 1991; Gordon et al., 1994). Pressures and temperatures obtained from pelitic specimens from the contact aureole of the Glacier Creek stock were re-calculated from data obtained by Kells (1993) at Queen's University, using the same analytical methods.

The validity and quality of the results are dependent on the proper choice of the equilibrium mineral assemblage for the calculations, and on errors in the thermodynamic properties of the mineral phases. Textural evidence in thin section for the equilibrium assemblage was in some specimens ambiguous (the ambiguities most commonly involve the relationship between chlorite and the rest of the assemblage). The thermodynamic properties of some mineral phases, such as staurolite and epidote (Berman, 1991), are not well constrained, and thus may contribute to greater error in the P-T calculation. For most specimens, several P-T diagrams were generated for a single specimen using different choices for the equilibrium assemblage.The diagrams were compared to see which yielded the best results. In nearly all cases, the assemblage that appeared to be the equilibrium assemblage, based on textural evidence, also yielded the best P-T results. All P-T results are presented here, but only the results considered to be the most reliable from each rock are presented in Chapter 4.

## MINERAL ANALYSES: SPECIMEN MK-J1

### AVERAGED MICROPROBE ANALYSES (wt %) [Kells, 1993]

| | Garnet | Biotite | Plagioclase | Muscovite | Chlorite | Staurolite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|
| SiO2 | 37.80 | 36.33 | 59.73 | 46.96 | 24.72 | 27.87 | 0.48 | |
| Al2O3 | 21.88 | 19.42 | 35.87 | 35.53 | 23.98 | 54.16 | 0.29 | |
| TiO2 | 0.03 | 1.25 | 0.00 | 0.25 | 0.00 | 0.45 | 49.47 | |
| Fe2O3 | 0.00 | 0.99 | 0.17 | 0.06 | 0.00 | 0.76 | 2.51 | |
| FeO | 32.59 | 16.92 | 0.00 | 0.99 | 22.65 | 12.96 | 44.60 | |
| MgO | 2.42 | 11.54 | 0.00 | 0.80 | 15.64 | 1.54 | 0.04 | |
| MnO | 2.84 | 0.03 | 0.07 | 0.05 | 0.08 | 0.10 | 0.26 | |
| CaO | 4.17 | 0.05 | 7.26 | 0.00 | 0.00 | 0.04 | 0.00 | |
| Na2O | 0.03 | 0.29 | 7.23 | 1.10 | 0.07 | 0.12 | 0.00 | |
| K2O | 0.00 | 7.95 | 0.03 | 8.82 | 0.05 | 0.02 | 0.04 | |
| Cr2O3 | 0.03 | 0.06 | 0.00 | 0.10 | 0.06 | 0.04 | 0.12 | |
| H2O | 0.00 | 4.00 | 0.00 | 4.52 | 11.50 | 1.02 | 0.00 | |
| TOTAL | 101.79 | 98.83 | 100.36 | 99.18 | 98.75 | 99.07 | 97.81 | |

### STRUCTURAL FORMULAE (Calculated by MINPROBE)

| | Garnet | Biotite | Plagioclase | Muscovite | Chlorite | Staurolite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|
| Si | 2.965 | 2.722 | 2.651 | 3.111 | 2.576 | 3.938 | 0.012 | |
| Al(IV) | 0.000 | 1.278 | 1.349 | 0.889 | 1.424 | 0.062 | 0.000 | |
| Al(VI) | 2.023 | 0.436 | 0.000 | 1.885 | 1.522 | 8.957 | 0.009 | |
| Ti | 0.002 | 0.070 | 0.006 | 0.012 | 0.000 | 0.048 | 0.958 | |
| Fe3+ | 0.000 | 0.056 | 0.000 | 0.003 | 0.000 | 0.081 | 0.049 | |
| Fe2+ | 2.138 | 1.060 | 0.000 | 0.055 | 1.974 | 1.531 | 0.961 | |
| Mg | 0.283 | 1.289 | 0.003 | 0.079 | 2.430 | 0.324 | 0.002 | |
| Mn | 0.189 | 0.002 | 0.345 | 0.003 | 0.007 | 0.012 | 0.006 | |
| Ca | 0.350 | 0.004 | 0.622 | 0.000 | 0.000 | 0.006 | 0.000 | |
| Na | 0.005 | 0.042 | 0.002 | 0.141 | 0.014 | 0.033 | 0.000 | |
| K | 0.000 | 0.760 | 0.000 | 0.745 | 0.007 | 0.004 | 0.001 | |
| Cr | 0.002 | 0.004 | 8.000 | 0.005 | 0.005 | 0.004 | 0.002 | |
| O | 12.000 | 10.000 | 8.000 | 10.000 | 10.000 | 23.039 | 3.000 | |
| OH | 0.000 | 2.000 | 0.000 | 2.000 | 8.000 | 0.961 | 0.000 | |

### IDEAL ACTIVITIES (values for garnet end members are cationic ratios)

Input for TWQ software, calculated by MINPROBE

| | Garnet | | Biotite | | Plagioclase | Muscovite | Chlorite | Staurolite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|---|---|
| alm | 0.722 | Xphl | 0.060 | Xan | 0.352 | 0.638 | 0.041 | 0.343 | 0.963 | 1.000 |
| prp | 0.096 | Xann | 0.034 | Xab | 0.646 | | | | | |
| sps | 0.064 | | | Xor | 0.002 | | | | | |
| grs | 0.117 | | | | | | | | | |

MINERAL ANALYSES: SPECIMEN MK-12

AVERAGED MICROPROBE ANALYSES (wt %) [Kells, 1993]

| | Garnet | Biotite | Plagioclase | Muscovite | Staurolite | Ilmenite | Quartz | Sillimanite |
|---|---|---|---|---|---|---|---|---|
| SiO2 | 37.98 | 36.45 | 61.39 | 47.00 | 28.30 | 0.96 | | |
| Al2O3 | 22.16 | 19.71 | 25.45 | 36.63 | 54.98 | 1.09 | | |
| TiO2 | 0.03 | 1.61 | 0.00 | 0.49 | 0.57 | 50.47 | | |
| Fe2O3 | 0.04 | 1.04 | 0.18 | 1.33 | 0.80 | 4.12 | | |
| FeO | 35.08 | 7.82 | 0.00 | 0.00 | 3.74 | 45.14 | | |
| MgO | 3.14 | 10.67 | 0.00 | 0.60 | 1.56 | 0.33 | | |
| MnO | 2.97 | 0.04 | 0.00 | 0.00 | 0.06 | 0.16 | | |
| CaO | 2.39 | 0.00 | 6.58 | 0.01 | 0.00 | 0.07 | | |
| Na2O | 0.03 | 0.23 | 7.79 | 1.36 | 0.00 | 0.12 | | |
| K2O | 0.03 | 8.99 | 0.05 | 8.44 | 0.00 | 0.00 | | |
| Cr2O3 | 0.10 | 0.03 | 0.06 | 0.02 | 0.00 | 0.00 | | |
| H2O | 0.00 | 4.04 | 0.00 | 4.59 | 1.01 | 0.00 | | |
| TOTAL | 103.94 | 100.63 | 101.50 | 100.48 | 101.02 | 102.45 | | |

STRUCTURAL FORMULAE (Calculated by MINPROBE)

| | Garnet | Biotite | Plagioclase | Muscovite | Staurolite | Ilmenite |
|---|---|---|---|---|---|---|
| Si | 2.902 | 2.706 | 2.688 | 3.066 | 3.931 | 0.023 |
| Al(IV) | 0.000 | 1.294 | 1.312 | 0.934 | 0.069 | 0.000 |
| Al(VI) | 1.996 | 0.430 | 0.002 | 1.882 | 8.931 | 0.031 |
| Ti | 0.002 | 0.090 | 0.000 | 0.024 | 0.060 | 0.926 |
| Fe3+ | 0.002 | 0.058 | 0.006 | 0.065 | 0.084 | 0.076 |
| Fe2+ | 2.241 | 1.106 | 0.000 | 0.000 | 1.596 | 0.921 |
| Mg | 0.358 | 1.181 | 0.000 | 0.058 | 0.323 | 0.012 |
| Mn | 0.192 | 0.003 | 0.000 | 0.000 | 0.007 | 0.003 |
| Ca | 0.196 | 0.000 | 0.309 | 0.001 | 0.000 | 0.002 |
| Na | 0.004 | 0.033 | 0.661 | 0.172 | 0.000 | 0.006 |
| K | 0.003 | 0.851 | 0.003 | 0.702 | 0.000 | 0.000 |
| Cr | 0.006 | 0.002 | 0.002 | 0.001 | 0.000 | 0.000 |
| O | 12.000 | 10.000 | 8.000 | 10.000 | 23.065 | 3.000 |
| OH | 0.000 | 2.000 | 0.000 | 2.000 | 0.935 | 0.000 |

*IDEAL ACTIVITIES* (values for garnet end members are cationic ratios)
Input for TWQ software, calculated by MINPROBE

| | Garnet | | Biotite | | Plagioclase | Muscovite | Staurolite | Ilmenite | Quartz | Sillimanite |
|---|---|---|---|---|---|---|---|---|---|---|
| alm | 0.750 | Xphl | 0.052 | Xan | 0.312 | 0.604 | 0.401 | 0.971 | 1.000 | 1.000 |
| prp | 0.120 | Xann | 0.043 | Xab | 0.679 | | | | | |
| sps | 0.064 | | | Xor | 0.009 | | | | | |
| grs | 0.066 | | | | | | | | | |

## MINERAL ANALYSES: SPECIMEN MK-9

### AVERAGED MICROPROBE ANALYSES (wt %) [Kells, 1993]

| | Garnet | Biotite | Plagioclase | Muscovite | Chlorite | Staurolite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|
| SiO2 | 38.03 | 37.60 | 63.43 | 46.83 | 25.70 | 28.26 | 0.73 | |
| Al2O3 | 22.14 | 19.32 | 24.34 | 36.73 | 23.31 | 55.92 | 0.41 | |
| TiO2 | 0.00 | 1.45 | 0.00 | 0.54 | 0.00 | 0.59 | 50.80 | |
| Fe2O3 | 0.00 | 0.97 | 0.16 | 1.00 | 0.00 | 0.76 | 3.70 | |
| FeO | 32.77 | 16.53 | 0.00 | 0.00 | 22.59 | 13.07 | 44.28 | |
| MgO | 2.56 | 11.81 | 0.00 | 0.56 | 17.40 | 1.71 | 0.27 | |
| MnO | 6.29 | 0.12 | 0.13 | 0.04 | 0.10 | 0.36 | 1.69 | |
| CaO | 1.89 | 0.00 | 5.11 | 0.00 | 0.00 | 0.00 | 0.06 | |
| Na2O | 0.00 | 0.14 | 8.81 | 2.06 | 0.00 | 0.00 | 0.00 | |
| K2O | 0.00 | 8.67 | 0.00 | 8.13 | 0.05 | 0.00 | 0.00 | |
| Cr2O3 | 0.00 | 0.06 | 0.00 | 0.00 | 0.10 | 0.07 | 0.00 | |
| H2O | 0.00 | 4.08 | 0.00 | 4.59 | 11.78 | 1.03 | 0.00 | |
| TOTAL | 103.68 | 100.75 | 101.98 | 100.48 | 101.03 | 101.78 | 101.94 | |

### STRUCTURAL FORMULAE (Calculated by MINPROBE)

| | Garnet | Biotite | Plagioclase | Muscovite | Chlorite | Staurolite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|
| Si | 2.936 | 2.759 | 2.755 | 3.056 | 2.613 | 3.888 | 0.018 | |
| Al(IV) | 0.000 | 1.241 | 1.245 | 0.944 | 1.387 | 0.112 | 0.000 | |
| Al(VI) | 2.015 | 0.430 | 0.001 | 1.880 | 1.407 | 8.956 | 0.012 | |
| Ti | 0.000 | 0.080 | 0.000 | 0.026 | 0.000 | 0.061 | 0.942 | |
| Fe3+ | 0.000 | 0.053 | 0.005 | 0.049 | 0.000 | 0.079 | 0.069 | |
| Fe2+ | 2.116 | 1.014 | 0.000 | 0.000 | 1.921 | 1.504 | 0.913 | |
| Mg | 0.295 | 1.292 | 0.000 | 0.054 | 2.637 | 0.351 | 0.010 | |
| Mn | 0.411 | 0.007 | 0.005 | 0.002 | 0.009 | 0.042 | 0.035 | |
| Ca | 0.156 | 0.000 | 0.238 | 0.000 | 0.000 | 0.000 | 0.002 | |
| Na | 0.000 | 0.020 | 0.742 | 0.261 | 0.000 | 0.000 | 0.000 | |
| K | 0.000 | 0.812 | 0.000 | 0.677 | 0.006 | 0.000 | 0.002 | |
| Cr | 0.000 | 0.003 | 0.000 | 0.000 | 0.008 | 0.008 | 0.000 | |
| O | 12.000 | 10.000 | 8.000 | 10.000 | 10.000 | 23.052 | 3.000 | |
| OH | 0.000 | 2.000 | 0.000 | 2.000 | 8.000 | 0.948 | 0.000 | |

### IDEAL ACTIVITIES (values for garnet end members are cationic ratios)
Input for TWQ software, calculated by MINPROBE

| | Garnet | | Biotite | | Plagioclase | Muscovite | Chlorite | Staurolite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|---|---|
| alm | 0.711 | Xphl | 0.065 | Xan | 0.242 | 0.592 | 0.057 | 0.320 | 0.989 | 1.000 |
| prp | 0.099 | Xann | 0.031 | Xab | 0.757 | | | | | |
| sps | 0.138 | | | Xor | 0.001 | | | | | |
| grs | 0.052 | | | | | | | | | |

# MINERAL ANALYSES: SPECIMEN MK-6

AVERAGED MICROPROBE ANALYSES (wt %) [Kells, 1993]

| | Garnet | Biotite | Plagioclase | Muscovite | Chlorite | Staurolite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|
| SiO2 | 37.76 | 35.53 | 59.92 | 47.52 | 25.13 | 28.22 | 0.56 | |
| Al2O3 | 22.20 | 19.60 | 25.97 | 35.79 | 23.61 | 55.22 | 0.40 | |
| TiO2 | 0.04 | 1.30 | 0.05 | 0.24 | 0.07 | 0.62 | 49.56 | |
| Fe2O3 | 0.00 | 1.03 | 0.07 | 0.05 | 0.00 | 0.70 | 1.79 | |
| FeO | 31.39 | 17.61 | 0.00 | 0.86 | 21.30 | 11.88 | 43.44 | |
| MgO | 2.96 | 11.48 | 0.00 | 0.64 | 16.88 | 1.34 | 0.13 | |
| MnO | 4.80 | 0.00 | 0.03 | 0.04 | 0.14 | 0.29 | 1.39 | |
| CaO | 2.80 | 0.03 | 7.27 | 0.05 | 0.01 | 0.03 | 0.00 | |
| Na2O | 0.00 | 0.13 | 7.25 | 1.05 | 0.05 | 0.07 | 0.00 | |
| K2O | 0.00 | 8.06 | 0.01 | 9.01 | 0.03 | 0.02 | 0.05 | |
| Cr2O3 | 0.04 | 0.00 | 0.04 | 0.07 | 0.04 | 0.05 | 0.00 | |
| H2O | 0.00 | 3.98 | 0.00 | 4.56 | 11.58 | 0.78 | 0.00 | |
| TOTAL | 101.99 | 98.75 | 100.61 | 99.88 | 98.84 | 99.22 | 97.32 | |

STRUCTURAL FORMULAE (Calculated by MINPROBE)

| | Garnet | Biotite | Plagioclase | Muscovite | Chlorite | Staurolite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|
| Si | 2.939 | 2.678 | 2.651 | 3.124 | 2.600 | 3.967 | 0.014 | |
| Al(IV) | 0.000 | 1.322 | 1.349 | 0.876 | 1.400 | 0.033 | 0.000 | |
| Al(VI) | 2.037 | 0.418 | 0.006 | 1.896 | 1.479 | 9.116 | 0.012 | |
| Ti | 0.002 | 0.074 | 0.002 | 0.012 | 0.005 | 0.066 | 0.963 | |
| Fe3+ | 0.000 | 0.058 | 0.002 | 0.003 | 0.000 | 0.074 | 0.035 | |
| Fe2+ | 2.043 | 1.110 | 0.000 | 0.048 | 1.843 | 1.397 | 0.939 | |
| Mg | 0.343 | 1.289 | 0.000 | 0.063 | 2.603 | 0.281 | 0.005 | |
| Mn | 0.316 | 0.000 | 0.001 | 0.002 | 0.012 | 0.035 | 0.030 | |
| Ca | 0.234 | 0.002 | 0.345 | 0.004 | 0.001 | 0.005 | 0.000 | |
| Na | 0.000 | 0.019 | 0.622 | 0.134 | 0.010 | 0.019 | 0.000 | |
| K | 0.000 | 0.775 | 0.001 | 0.756 | 0.004 | 0.004 | 0.002 | |
| Cr | 0.002 | 0.000 | 0.001 | 0.004 | 0.003 | 0.006 | 0.000 | |
| O | 12.000 | 10.000 | 8.000 | 10.000 | 10.000 | 23.270 | 3.000 | |
| OH | 0.000 | 2.000 | 0.000 | 2.000 | 8.000 | 0.730 | 0.000 | |

IDEAL ACTIVITIES (values for garnet end members are cationic ratios)
Input for TWQ software, calculated by MINPROBE

| | Garnet | | Biotite | | Plagioclase | Muscovite | Chlorite | Staurolite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|---|---|
| alm | 0.696 | Xphl | 0.062 | Xan | 0.351 | 0.663 | 0.057 | 0.238 | 1.000 | 1.000 |
| prp | 0.117 | Xann | 0.039 | Xab | 0.649 | | | | | |
| sps | 0.108 | | | Xor | 0.001 | | | | | |
| grs | 0.078 | | | | | | | | | |

## MINERAL ANALYSES: SPECIMEN MK-4

AVERAGED MICROPROBE ANALYSES (wt %) [Kells, 1993]

| | Garnet | Plagioclase | Muscovite | Chlorite | Staurolite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|
| SiO2 | 37.83 | 59.91 | 46.74 | 25.27 | 27.96 | 1.68 | |
| Al2O3 | 21.69 | 26.06 | 35.81 | 23.07 | 53.85 | 0.33 | |
| TiO2 | 0.09 | 0.00 | 0.45 | 0.04 | 0.58 | 49.02 | |
| Fe2O3 | 0.00 | 0.29 | 0.06 | 0.00 | 0.00 | 0.00 | |
| FeO | 30.67 | 0.00 | 1.07 | 23.26 | 9.67 | 43.21 | |
| MgO | 2.61 | 0.00 | 0.70 | 15.32 | 1.18 | 0.08 | |
| MnO | 4.57 | 0.00 | 0.00 | 0.11 | 0.26 | 2.22 | |
| CaO | 3.95 | 7.44 | 0.00 | 0.02 | 0.05 | 0.00 | |
| Na2O | 0.03 | 7.47 | 1.48 | 0.03 | 1.89 | 0.00 | |
| K2O | 0.00 | 0.00 | 8.59 | 0.03 | 0.00 | 0.00 | |
| Cr2O3 | 0.02 | 0.00 | 0.03 | 0.00 | 0.11 | 0.00 | |
| H2O | 0.00 | 0.00 | 4.53 | 11.46 | 2.14 | 0.00 | |
| TOTAL | 101.46 | 101.17 | 99.47 | 98.61 | 97.69 | 96.54 | |

STRUCTURAL FORMULAE (Calculated by MINPROBE)

| | Garnet | Plagioclase | Muscovite | Chlorite | Staurolite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|
| Si | 2.981 | 2.642 | 3.089 | 2.643 | 3.967 | 0.044 | |
| Al(IV) | 0.000 | 1.354 | 0.911 | 1.357 | 0.033 | 0.000 | |
| Al(VI) | 2.014 | 0.000 | 1.879 | 1.488 | 8.970 | 0.010 | |
| Ti | 0.005 | 0.000 | 0.022 | 0.003 | 0.062 | 0.955 | |
| Fe3+ | 0.000 | 0.010 | 0.003 | 0.000 | 0.057 | 0.000 | |
| Fe2+ | 2.021 | 0.000 | 0.059 | 2.035 | 1.090 | 0.936 | |
| Mg | 0.307 | 0.000 | 0.069 | 2.389 | 0.250 | 0.003 | |
| Mn | 0.305 | 0.000 | 0.000 | 0.010 | 0.031 | 0.049 | |
| Ca | 0.333 | 0.351 | 0.000 | 0.002 | 0.008 | 0.000 | |
| Na | 0.005 | 0.639 | 0.190 | 0.006 | 0.520 | 0.000 | |
| K | 0.000 | 0.000 | 0.724 | 0.004 | 0.000 | 0.000 | |
| Cr | 0.001 | 0.000 | 0.002 | 0.000 | 0.012 | 0.000 | |
| O | 12.000 | 8.000 | 10.000 | 10.000 | 22.610 | 3.000 | |
| OH | 0.000 | 0.000 | 2.000 | 8.000 | 1.390 | 0.000 | |

IDEAL ACTIVITIES (values for garnet end members are cationic ratios)
Input for TWQ software, calculated by MINPROBE

| | Garnet | | Plagioclase | Muscovite | Chlorite | Staurolite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|
| alm | 0.681 | Xan | 0.358 | 0.618 | 0.037 | 0.088 | 1.000 | 1.000 |
| prp | 0.103 | Xab | 0.642 | | | | | |
| sps | 0.103 | Xor | 0.000 | | | | | |
| grs | 0.112 | | | | | | | |

## MINERAL ANALYSES: SPECIMEN MK-11

AVERAGED MICROPROBE ANALYSES (wt %) (Kells, 1993)

| | Garnet | Biotite | Plagioclase | Muscovite | Staurolite | Ilmenite | Quartz | Sillimanite |
|---|---|---|---|---|---|---|---|---|
| SiO2 | 37.36 | 35.98 | 59.63 | 46.10 | 28.14 | 0.56 | | |
| Al2O3 | 22.47 | 20.06 | 26.40 | 35.11 | 54.49 | 0.52 | | |
| TiO2 | 0.00 | 1.63 | 0.04 | 1.01 | 0.70 | 49.95 | | |
| Fe2O3 | 0.00 | 0.94 | 0.21 | 0.07 | 0.75 | 0.51 | | |
| FeO | 31.56 | 16.13 | 0.00 | 1.19 | 12.84 | 43.68 | | |
| MgO | 3.25 | 11.38 | 0.00 | 0.87 | 1.59 | 0.22 | | |
| MnO | 4.99 | 0.17 | 0.00 | 0.08 | 0.31 | 1.42 | | |
| CaO | 2.04 | 0.02 | 8.03 | 0.04 | 0.00 | 0.06 | | |
| Na2O | 0.00 | 0.29 | 7.02 | 0.90 | 0.09 | 0.00 | | |
| K2O | 0.00 | 8.89 | 0.01 | 9.47 | 0.00 | 0.00 | | |
| Cr2O3 | 0.15 | 0.05 | 0.00 | 0.03 | 0.07 | 0.08 | | |
| H2O | 0.00 | 4.02 | 0.00 | 4.50 | 0.97 | 0.00 | | |
| TOTAL | 101.82 | 99.56 | 101.34 | 99.37 | 99.95 | 97.00 | | |

STRUCTURAL FORMULAE (Calculated by MINPROBE)

| | Garnet | Biotite | Plagioclase | Muscovite | Staurolite | Ilmenite |
|---|---|---|---|---|---|---|
| Si | 2.890 | 2.682 | 2.626 | 3.069 | 3.943 | 0.014 |
| Al(IV) | 0.000 | 1.318 | 1.370 | 0.931 | 0.057 | 0.000 |
| Al(VI) | 2.049 | 0.444 | 0.000 | 1.824 | 8.941 | 0.016 |
| Ti | 0.000 | 0.091 | 0.001 | 0.051 | 0.074 | 0.972 |
| Fe3+ | 0.000 | 0.053 | 0.007 | 0.003 | 0.079 | 0.010 |
| Fe2+ | 2.042 | 1.006 | 0.000 | 0.066 | 1.505 | 0.945 |
| Mg | 0.375 | 1.264 | 0.000 | 0.086 | 0.332 | 0.008 |
| Mn | 0.327 | 0.011 | 0.000 | 0.005 | 0.037 | 0.031 |
| Ca | 0.169 | 0.002 | 0.379 | 0.003 | 0.000 | 0.002 |
| Na | 0.000 | 0.042 | 0.599 | 0.116 | 0.024 | 0.000 |
| K | 0.000 | 0.845 | 0.001 | 0.804 | 0.000 | 0.000 |
| Cr | 0.009 | 0.003 | 0.000 | 0.002 | 0.008 | 0.002 |
| O | 12.000 | 10.000 | 8.000 | 10.000 | 23.094 | 3.000 |
| OH | 0.000 | 2.000 | 0.000 | 2.000 | 0.906 | 0.000 |

IDEAL ACTIVITIES (values for garnet end members are cationic ratios)
Input for TWQ software, calculated by MINPROBE

| | Garnet | | Biotite | | Plagioclase | Muscovite | Staurolite | Ilmenite | Quartz | Sillimanite |
|---|---|---|---|---|---|---|---|---|---|---|
| alm | 0.701 | Xphl | 0.063 | Xan | 0.378 | 0.646 | 0.321 | 0.965 | 1.000 | 1.000 |
| prp | 0.129 | Xann | 0.032 | Xab | 0.621 | | | | | |
| sps | 0.112 | | | Xor | 0.001 | | | | | |
| grs | 0.053 | | | | | | | | | |

# MINERAL ANALYSES: SPECIMEN 34-Y2

AVERAGED MICROPROBE ANALYSES (wt %)

| | Garnet | Biotite | Plagioclase | Chlorite | Epidote | Ilmenite | Hornblende | Quartz |
|---|---|---|---|---|---|---|---|---|
| SiO2 | 38.11 | 36.51 | 62.39 | 25.61 | 38.07 | 0.59 | 43.13 | |
| Al2O3 | 21.26 | 17.79 | 24.23 | 22.51 | 25.48 | 0.27 | 15.36 | |
| TiO2 | 0.08 | 1.75 | 0.05 | 0.02 | 0.03 | 50.36 | 0.33 | |
| Fe2O3 | 0.65 | 1.52 | 0.10 | 0.00 | 11.87 | 3.74 | 2.81 | |
| FeO | 25.57 | 17.68 | 0.00 | 21.04 | 0.01 | 43.48 | 13.82 | |
| MgO | 2.18 | 11.23 | 0.00 | 17.60 | 0.00 | 0.11 | 9.24 | |
| MnO | 9.35 | 0.08 | 0.00 | 0.20 | 0.64 | 2.14 | 0.53 | |
| CaO | 5.61 | 0.02 | 5.41 | 0.04 | 23.09 | 0.04 | 10.75 | |
| Na2O | 0.00 | 0.05 | 8.39 | 0.03 | 0.00 | 0.00 | 1.80 | |
| K2O | 0.00 | 8.82 | 0.10 | 0.00 | 0.00 | 0.03 | 0.25 | |
| Cr2O3 | 0.00 | 0.07 | 0.00 | 0.06 | 0.07 | 0.13 | 0.04 | |
| H2O | 0.00 | 3.98 | 0.00 | 11.57 | 1.92 | 0.00 | 2.04 | |
| TOTAL | 102.82 | 99.50 | 100.67 | 98.68 | 101.17 | 100.89 | 100.10 | |

STRUCTURAL FORMULAE, calculated by MINPROBE

| | Garnet | Biotite | Plagioclase | Chlorite | Epidote | Ilmenite | Hornblende | Quartz |
|---|---|---|---|---|---|---|---|---|
| Si | 2.969 | 2.747 | 2.745 | 2.652 | 2.974 | 0.015 | 6.345 | |
| Al(IV) | 0.000 | 1.253 | 1.255 | 1.348 | 0.026 | 0.000 | 1.655 | |
| Al(VI) | 1.952 | 0.325 | 0.002 | 1.400 | 2.321 | 0.008 | 1.009 | |
| Ti | 0.005 | 0.099 | 0.002 | 0.002 | 0.002 | 0.945 | 0.037 | |
| Fe3+ | 0.038 | 0.086 | 0.003 | 0.000 | 0.698 | 0.070 | 0.311 | |
| Fe2+ | 1.666 | 1.112 | 0.000 | 1.822 | 0.001 | 0.908 | 1.701 | |
| Mg | 0.253 | 1.259 | 0.000 | 2.717 | 0.000 | 0.004 | 2.026 | |
| Mn | 0.617 | 0.005 | 0.000 | 0.018 | 0.042 | 0.045 | 0.066 | |
| Ca | 0.468 | 0.002 | 0.255 | 0.004 | 1.933 | 0.001 | 1.695 | |
| Na | 0.000 | 0.007 | 0.716 | 0.006 | 0.000 | 0.000 | 0.513 | |
| K | 0.000 | 0.847 | 0.006 | 0.000 | 0.000 | 0.001 | 0.047 | |
| Cr | 0.000 | 0.004 | 0.000 | 0.005 | 0.003 | 0.003 | 0.005 | |
| O | 12.000 | 10.000 | 8.000 | 10.000 | 12.000 | 3.000 | 22.000 | |
| OH | 0.000 | 2.000 | 0.000 | 8.000 | 1.000 | 0.000 | 2.000 | |

Hornblende site occupancy
(input for TWQ software)

| | |
|---|---|
| T1: Si/4 | 0.586 |
| T1: Al/4 | 0.552 |
| M2: Al/2 | 0.505 |
| M2: Ti/2 | 0.019 |
| M2: Fe3+/2 | 0.156 |
| M2: Fe2+/2 | 0.146 |
| M2: Mg/2 | 0.174 |
| M1,M3: Fe2+/3 | 0.456 |
| M1,M3:Mg/3 | 0.544 |
| M1,M3: Mn/3 | 0.000 |
| M4: Fe2+/2 | 0.020 |
| M4: Mg/2 | 0.024 |
| M4: Mn/2 | 0.033 |
| M4: Ca/2 | 0.848 |
| M4: Na/2 | 0.076 |
| A: Na | 0.362 |
| A: K | 0.047 |
| A: vacancy | 0.591 |

IDEAL ACTIVITIES (values for garnet end members are cationic ratios):
Input for TWQ software, calculated by MINPROBE

| | Garnet | | Biotite | | Plagioclase | Chlorite | Epidote | Ilmenite | Hornblende | Quartz |
|---|---|---|---|---|---|---|---|---|---|---|
| alm | 0.554 | Xphl | 0.063 | Xan | 0.258 | 0.066 | 0.300 | | | 1.000 |
| prp | 0.084 | Xann | 0.043 | Xab | 0.737 | | | | | |
| sps | 0.205 | | | Xor | 0.006 | | | | | |
| grs | 0.156 | | | | | | | | | |

## MINERAL ANALYSES; SPECIMEN 34-Y

### AVERAGED MICROPROBE ANALYSES (wt %)

| | Garnet | Biotite | Plagioclase | Chlorite | Epidote | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|
| SiO2 | 38.11 | 36.74 | 63.77 | 25.33 | 38.14 | 0.54 | |
| Al2O3 | 21.52 | 17.48 | 23.44 | 22.30 | 25.06 | 0.31 | |
| TiO2 | 0.02 | 1.56 | 0.04 | 0.07 | 0.06 | 49.51 | |
| Fe2O3 | 0.04 | 2.04 | 0.07 | 0.00 | 11.69 | 5.46 | |
| FeO | 29.60 | 15.83 | 0.00 | 22.76 | 0.55 | 42.16 | |
| MgO | 2.41 | 11.45 | 0.00 | 16.22 | 0.00 | 0.20 | |
| MnO | 6.82 | 0.03 | 0.00 | 0.13 | 0.57 | 1.78 | |
| CaO | 3.76 | 0.03 | 4.65 | 0.03 | 22.73 | 0.00 | |
| Na2O | 0.00 | 0.09 | 8.71 | 0.00 | 0.04 | 0.17 | |
| K2O | 0.04 | 9.63 | 0.04 | 0.05 | 0.00 | 0.02 | |
| Cr2O3 | 0.00 | 0.00 | 0.04 | 0.08 | 0.00 | 0.12 | |
| H2O | 0.00 | 3.97 | 0.00 | 11.44 | 1.91 | 0.00 | |
| TOTAL | 102.31 | 98.95 | 100.76 | 98.41 | 100.75 | 100.27 | |

### STRUCTURAL FORMULAE (Calculated by MINPROBE)

| | Garnet | Biotite | Plagioclase | Chlorite | Epidote | Ilmenite |
|---|---|---|---|---|---|---|
| Si | 2.998 | 2.774 | 2.794 | 2.653 | 2.995 | 0.014 |
| Al(IV) | 0.000 | 1.226 | 1.206 | 1.347 | 0.005 | 0.000 |
| Al(VI) | 1.995 | 0.329 | 0.004 | 1.406 | 2.313 | 0.009 |
| Ti | 0.001 | 0.089 | 0.001 | 0.006 | 0.004 | 0.934 |
| Fe3+ | 0.002 | 0.116 | 0.002 | 0.000 | 0.691 | 0.103 |
| Fe2+ | 1.947 | 1.000 | 0.000 | 1.994 | 0.036 | 0.884 |
| Mg | 0.283 | 1.288 | 0.000 | 2.532 | 0.000 | 0.007 |
| Mn | 0.454 | 0.008 | 0.000 | 0.012 | 0.038 | 0.038 |
| Ca | 0.317 | 0.002 | 0.218 | 0.003 | 1.912 | 0.000 |
| Na | 0.000 | 0.013 | 0.740 | 0.000 | 0.006 | 0.008 |
| K | 0.004 | 0.927 | 0.002 | 0.007 | 0.000 | 0.001 |
| Cr | 0.000 | 0.000 | 0.001 | 0.007 | 0.000 | 0.002 |
| O | 12.000 | 10.000 | 8.000 | 10.000 | 12.000 | 3.000 |
| OH | 0.000 | 2.000 | 0.000 | 8.000 | 1.000 | 0.000 |

### IDEAL ACTIVITIES (values for garnet end members are cationic ratios):

Input for TWQ software, calculated by MINPROBE

| | Garnet | | Biotite | | Plagioclase | Chlorite | Epidote | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|---|
| alm | 0.649 | Xphl | 0.073 | Xan | 0.215 | 0.048 | 0.287 | | 1.000 |
| prp | 0.094 | Xann | 0.034 | Xab | 0.782 | | | | |
| sps | 0.151 | | | Xor | 0.002 | | | | |
| grs | 0.106 | | | | | | | | |

## MINERAL ANALYSES: SPECIMEN 34PEL

### AVERAGED MICROPROBE ANALYSES (wt %)

| | Grt rims | Grt cores | Biotite | Plagioclase | Muscovite | Chlorite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|
| SiO2 | 37.94 | 38.05 | 36.59 | 64.17 | 46.39 | 25.15 | 0.67 | |
| Al2O3 | 21.53 | 21.55 | 17.86 | 23.15 | 34.13 | 22.41 | 0.39 | |
| TiO2 | 0.02 | 0.09 | 1.51 | 0.00 | 0.41 | 0.00 | 49.10 | |
| Fe2O3 | 0.12 | 0.20 | 1.92 | 0.10 | 0.00 | 0.00 | 4.64 | |
| FeO | 30.02 | 26.74 | 17.65 | 0.00 | 2.78 | 24.97 | 42.86 | |
| MgO | 2.76 | 2.13 | 10.72 | 0.00 | 0.82 | 14.95 | 0.07 | |
| MnO | 6.70 | 9.54 | 0.07 | 0.05 | 0.05 | 0.24 | 1.70 | |
| CaO | 3.20 | 4.64 | 0.04 | 4.09 | 0.07 | 0.02 | 0.00 | |
| Na2O | 0.00 | 0.02 | 0.14 | 8.76 | 1.08 | 0.10 | 0.00 | |
| K2O | 0.02 | 0.05 | 9.64 | 0.09 | 8.77 | 0.12 | 0.08 | |
| Cr2O3 | 0.05 | 0.00 | 0.07 | 0.03 | 0.04 | 0.03 | 0.00 | |
| H2O | 0.00 | 0.00 | 3.99 | 0.00 | 4.47 | 11.43 | 0.00 | |
| TOTAL | 102.36 | 103.01 | 100.19 | 100.44 | 99.01 | 99.42 | 99.51 | |

### STRUCTURAL FORMULAE (Calculated by MINPROBE)

| | Grt rims | Grt cores | Biotite | Plagioclase | Muscovite | Chlorite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|
| Si | 2.971 | 2.963 | 2.749 | 2.815 | 3.110 | 2.637 | 0.017 | |
| Al(IV) | 0.000 | 0.000 | 1.251 | 1.185 | 0.890 | 1.363 | 0.000 | |
| Al(VI) | 1.987 | 1.978 | 0.331 | 0.011 | 1.806 | 1.407 | 0.012 | |
| Ti | 0.001 | 0.005 | 0.085 | 0.000 | 0.021 | 0.000 | 0.934 | |
| Fe3+ | 0.007 | 0.012 | 0.108 | 0.003 | 0.000 | 0.000 | 0.088 | |
| Fe2+ | 1.966 | 1.742 | 1.109 | 0.000 | 0.156 | 2.190 | 0.907 | |
| Mg | 0.322 | 0.247 | 1.201 | 0.000 | 0.082 | 2.337 | 0.003 | |
| Mn | 0.444 | 0.629 | 0.004 | 0.002 | 0.003 | 0.021 | 0.036 | |
| Ca | 0.269 | 0.387 | 0.003 | 0.192 | 0.005 | 0.002 | 0.000 | |
| Na | 0.000 | 0.003 | 0.020 | 0.745 | 0.140 | 0.020 | 0.000 | |
| K | 0.002 | 0.005 | 0.924 | 0.005 | 0.750 | 0.016 | 0.003 | |
| Cr | 0.003 | 0.000 | 0.004 | 0.001 | 0.002 | 0.002 | 0.000 | |
| O | 12.000 | 12.000 | 10.000 | 8.000 | 10.000 | 10.000 | 3.000 | |
| OH | 0.000 | 0.000 | 2.000 | 0.000 | 2.000 | 8 | 0.000 | |

### IDEAL ACTIVITIES (values for garnet end members are cationic ratios):
Input for TWQ software, calculated by MINPROBE

| | Grt rims | Grt cores | | Biotite | | Plagioclase | Muscovite | Chlorite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|---|---|
| alm | 0.655 | 0.579 | Xphl | 0.059 | Xan | 0.196 | 0.57 | 0.031 | | 1.000 |
| prp | 0.107 | 0.082 | Xann | 0.047 | Xab | 0.798 | | | | |
| sps | 0.148 | 0.209 | | | Xor | 0.005 | | | | |
| grs | 0.089 | 0.129 | | | | | | | | |

## MINERAL ANALYSES: SPECIMEN 94x

### AVERAGED MICROPROBE ANALYSES (wt %)

| | Garnet 1 | Garnet 2 | Biotite | Plagioclase | Muscovite | Chlorite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|
| SiO2 | 37.12 | 37.10 | 33.63 | 65.59 | 47.00 | 24.51 | 0.87 | |
| Al2O3 | 21.34 | 21.57 | 28.43 | 21.73 | 36.34 | 22.39 | 0.38 | |
| TiO2 | 0.08 | 0.00 | 0.22 | 0.00 | 0.50 | 0.08 | 50.74 | |
| Fe2O3 | 2.15 | 0.20 | 0.50 | 0.48 | 0.00 | 0.00 | 1.26 | |
| FeO | 24.68 | 32.61 | 17.44 | 0.00 | 1.45 | 30.25 | 44.21 | |
| MgO | 0.75 | 1.36 | 6.06 | 0.00 | 0.55 | 9.75 | 0.00 | |
| MnO | 11.86 | 7.58 | 0.07 | 0.09 | 0.03 | 0.19 | 2.24 | |
| CaO | 5.02 | 2.47 | 0.07 | 2.57 | 0.00 | 0.05 | 0.00 | |
| Na2O | 0.00 | 0.00 | 0.35 | 9.76 | 0.80 | 0.05 | 0.00 | |
| K2O | 0.00 | 0.00 | 2.78 | 0.05 | 6.48 | 0.25 | 0.06 | |
| Cr2O3 | 0.00 | 0.00 | 0.00 | 0.07 | 0.03 | 0.00 | 0.00 | |
| H2O | 0.00 | 0.00 | 3.93 | 0.00 | 0.00 | 11.04 | 0.00 | |
| TOTAL | 102.99 | 102.89 | 93.48 | 100.34 | 97.7 | 98.56 | 99.77 | |

### STRUCTURAL FORMULAE (Calculated by MINPROBE)

| | Garnet 1 | Garnet 2 | Biotite | Plagioclase | Muscovite | Chlorite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|
| Si | 2.900 | 2.901 | 2.564 | 2.875 | 3.114 | 2.660 | 0.022 | |
| Al(IV) | 0.000 | 0.000 | 1.436 | 1.122 | 0.886 | 1.340 | 0.000 | |
| Al(VI) | 1.965 | 1.988 | 1.119 | 0.000 | 1.953 | 1.525 | 0.011 | |
| Ti | 0.005 | 0.000 | 0.013 | 0.000 | 0.025 | 0.007 | 0.961 | |
| Fe3+ | 0.026 | 0.012 | 0.029 | 0.016 | 0.000 | 0.000 | 0.024 | |
| Fe2+ | 1.712 | 2.132 | 1.112 | 0.000 | 0.080 | 2.746 | 0.932 | |
| Mg | 0.087 | 0.159 | 0.689 | 0.000 | 0.054 | 1.577 | 0.000 | |
| Mn | 0.785 | 0.502 | 0.005 | 0.003 | 0.002 | 0.017 | 0.048 | |
| Ca | 0.420 | 0.207 | 0.006 | 0.121 | 0.000 | 0.006 | 0.000 | |
| Na | 0.000 | 0.000 | 0.052 | 0.829 | 0.103 | 0.011 | 0.000 | |
| K | 0.000 | 0.000 | 0.270 | 0.003 | 0.548 | 0.035 | 0.002 | |
| Cr | 0.000 | 0.000 | 0.000 | 0.002 | 0.002 | 0.000 | 0.000 | |
| O | 12.000 | 12.000 | 10.000 | 8.000 | 10.000 | 10.000 | 3.000 | |
| OH | 0.000 | 0.000 | 2.000 | 0.000 | 2.000 | 8.000 | 0.000 | |

### IDEAL ACTIVITIES (values for garnet end members are cationic ratios)
Input for TWQ software, calculated by MINPROBE

| | Garnet 1 | Garnet 2 | | Biotite | | Plagioclase | Muscovite | Chlorite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|---|---|
| alm | 0.570 | 0.711 | Xphl | 0.003 | Xan | 0.129 | 0.463 | 0.005 | | 1.000 |
| prp | 0.029 | 0.053 | Xann | 0.014 | Xab | 0.868 | | | | |
| sps | 0.261 | 0.167 | | | Xor | 0.003 | | | | |
| grs | 0.140 | 0.069 | | | | | | | | |

MINERAL ANALYSES: SPECIMEN HOW-Y

AVERAGED MICROPROBE ANALYSES (wt %)

| | Garnet | Biotite | Plagioclase | Muscovite | Chlorite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|
| SiO2 | 37.78 | 36.45 | 64.79 | 48.38 | 24.35 | 0.67 | |
| Al2O3 | 21.35 | 17.91 | 23.62 | 35.30 | 22.64 | 0.39 | |
| TiO2 | 0.10 | 1.64 | 0.00 | 0.53 | 0.09 | 49.10 | |
| Fe2O3 | 0.36 | 1.47 | 0.19 | 0.08 | 0.79 | 4.64 | |
| FeO | 31.41 | 20.68 | 0.00 | 1.15 | 28.86 | 42.86 | |
| MgO | 1.47 | 8.91 | 0.00 | 0.89 | 12.63 | 0.07 | |
| MnO | 1.75 | 0.13 | 0.00 | 0.00 | 0.23 | 1.70 | |
| CaO | 7.99 | 0.13 | 4.60 | 0.03 | 0.00 | 0.00 | |
| Na2O | 0.00 | 0.17 | 6.89 | 0.75 | 0.00 | 0.00 | |
| K2O | 0.04 | 9.20 | 0.14 | 7.37 | 0.10 | 0.08 | |
| Cr2O3 | 0.04 | 0.00 | 0.00 | 0.10 | 0.00 | 0.00 | |
| H2O | 0.00 | 3.96 | 0.00 | 4.57 | 11.38 | 0.00 | |
| TOTAL | 102.29 | 100.65 | 100.23 | 99.15 | 101.07 | 99.51 | |

STRUCTURAL FORMULAE (Calculated by MINPROBE)

| | Garnet | Biotite | Plagioclase | Muscovite | Chlorite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|
| Si | 2.950 | 2.755 | 2.829 | 3.169 | 2.564 | 0.017 | |
| Al(IV) | 0.000 | 1.245 | 1.171 | 0.831 | 1.436 | 0.000 | |
| Al(VI) | 1.965 | 0.350 | 0.044 | 1.895 | 1.373 | 0.012 | |
| Ti | 0.006 | 0.093 | 0.000 | 0.026 | 0.007 | 0.934 | |
| Fe3+ | 0.021 | 0.083 | 0.006 | 0.004 | 0.062 | 0.088 | |
| Fe2+ | 2.051 | 1.307 | 0.000 | 0.063 | 2.541 | 0.907 | |
| Mg | 0.171 | 1.004 | 0.000 | 0.087 | 1.982 | 0.003 | |
| Mn | 0.116 | 0.008 | 0.000 | 0.000 | 0.021 | 0.036 | |
| Ca | 0.668 | 0.011 | 0.215 | 0.002 | 0.000 | 0.000 | |
| Na | 0.000 | 0.025 | 0.583 | 0.095 | 0.000 | 0.000 | |
| K | 0.004 | 0.887 | 0.008 | 0.616 | 0.013 | 0.003 | |
| Cr | 0.002 | 0.000 | 0.000 | 0.005 | 0.000 | 0.000 | |
| O | 12.000 | 10.000 | 8.000 | 10.000 | 10.000 | 3.000 | |
| OH | 0.000 | 2.000 | 0.000 | 2.000 | 8.000 | 0.000 | |

IDEAL ACTIVITIES (values for garnet end members are cationic ratios)
Input for TWQ software, calculated by MINPROBE

| | Garnet | | Biotite | | Plagioclase | Muscovite | Chlorite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|---|
| alm | 0.684 | Xphl | 0.033 | Xan | 0.218 | 0.512 | 0.013 | | 1.000 |
| prp | 0.057 | Xann | 0.073 | Xab | 0.771 | | | | |
| sps | 0.038 | | | Xor | 0.010 | | | | |
| grs | 0.221 | | | | | | | | |

MINERAL ANALYSES: SPECIMEN 207x

| | Garnet | Biotite | Plagioclase | Muscovite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|
| **AVERAGED MICROPROBE ANALYSES (wt %)** | | | | | | |
| SiO2 | 37.75 | 35.74 | 61.36 | 47.55 | 0.48 | |
| Al2O3 | 21.16 | 16.93 | 24.11 | 32.67 | 0.33 | |
| TiO2 | 0.22 | 1.90 | 0.00 | 0.51 | 50.46 | |
| Fe2O3 | 0.47 | 1.66 | 0.23 | 0.07 | 3.55 | |
| FeO | 20.65 | 19.58 | 0.00 | 3.19 | 41.62 | |
| MgO | 0.98 | 9.43 | 0.00 | 1.25 | 0.11 | |
| MnO | 13.18 | 0.25 | 0.00 | 0.00 | 3.80 | |
| CaO | 8.16 | 0.02 | 6.09 | 0.06 | 0.22 | |
| Na2O | 0.00 | 0.12 | 7.93 | 0.29 | 0.00 | |
| K2O | 0.00 | 9.46 | 0.02 | 8.07 | 0.00 | |
| Cr2O3 | 0.03 | 0.03 | 0.07 | 0.00 | 0.00 | |
| H2O | 0.00 | 3.89 | 0.00 | 4.46 | 0.00 | |
| TOTAL | 102.60 | 99.02 | 99.81 | 98.13 | 100.58 | |
| **STRUCTURAL FORMULAE (Calculated by MINPROBE)** | | | | | | |
| Si | 2.944 | 2.749 | 2.728 | 3.195 | 0.012 | |
| Al(IV) | 0.000 | 1.251 | 1.264 | 0.805 | 0.000 | |
| Al(VI) | 1.945 | 0.284 | 0.000 | 1.781 | 0.010 | |
| Ti | 0.013 | 0.110 | 0.000 | 0.026 | 0.950 | |
| Fe3+ | 0.028 | 0.096 | 0.008 | 0.004 | 0.067 | |
| Fe2+ | 1.347 | 1.260 | 0.000 | 0.179 | 0.871 | |
| Mg | 0.114 | 1.081 | 0.000 | 0.125 | 0.004 | |
| Mn | 0.871 | 0.016 | 0.000 | 0.000 | 0.081 | |
| Ca | 0.682 | 0.002 | 0.290 | 0.004 | 0.006 | |
| Na | 0.000 | 0.018 | 0.684 | 0.038 | 0.000 | |
| K | 0.000 | 0.928 | 0.001 | 0.692 | 0.000 | |
| Cr | 0.002 | 0.002 | 0.002 | 0.000 | 0.000 | |
| O | 12.000 | 10.000 | 8.000 | 10.000 | 3.000 | |
| OH | 0.000 | 2.000 | 0.000 | 8.000 | 0.000 | |

IDEAL ACTIVITIES (values for garnet end members are cationic ratios):
Input for TWQ software, calculated by MINPROBE

| | Garnet | | Biotite | | Plagioclase | Muscovite | Ilmenite | Quartz |
|---|---|---|---|---|---|---|---|---|
| alm | 0.447 | Xphl | 0.043 | Xan | 0.281 | 0.485 | | 1.000 |
| prp | 0.038 | Xann | 0.069 | Xab | 0.718 | | | |
| sps | 0.289 | | | Xor | 0.001 | | | |
| grs | 0.226 | | | | | | | |

## MINERAL ANALYSES: SPECIMEN 042

AVERAGED MICROPROBE ANALYSES (wt %)

| | Grt rims | Grt core | Matrix Bt | Core Bt | Matrix Pl | Core Pl | Matrix Ms | Core Ms | Ilmenite | Rutile | Quartz | Sillimanite | Andalusite? |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| SiO2 | 37.66 | 37.87 | 35.53 | 35.66 | 48.06 | 45.24 | 46.25 | 46.25 | 0.49 | | | | |
| Al2O3 | 21.65 | 21.59 | 19.40 | 19.73 | 33.98 | 35.88 | 35.76 | 36.52 | 0.38 | | | | |
| TiO2 | 0.00 | 0.00 | 2.45 | 2.29 | 0.06 | 0.05 | 0.74 | 0.73 | 51.04 | | | | |
| Fe2O3 | 0.00 | 0.06 | 0.90 | 0.82 | 0.00 | 0.00 | 0.00 | 0.00 | 2.02 | | | | |
| FeO | 32.41 | 34.00 | 20.04 | 18.16 | 0.18 | 0.34 | 1.15 | 1.57 | 44.85 | | | | |
| MgO | 2.38 | 1.95 | 8.35 | 8.97 | 0.00 | 0.00 | 0.64 | 0.44 | 0.36 | | | | |
| MnO | 5.72 | 3.03 | 0.20 | 0.00 | 0.00 | 0.05 | 0.00 | 0.07 | 0.90 | | | | |
| CaO | 2.26 | 4.13 | 0.07 | 0.00 | 17.18 | 19.47 | 0.07 | 0.06 | 0.06 | | | | |
| Na2O | 0.00 | 0.00 | 0.13 | 0.26 | 2.07 | 0.75 | 0.56 | 0.67 | 0.00 | | | | |
| K2O | 0.00 | 0.00 | 9.43 | 8.99 | 0.00 | 0.00 | 9.45 | 9.49 | 0.00 | | | | |
| Cr2O3 | 0.09 | 0.08 | 0.05 | 0.00 | 0.00 | 0.00 | 0.05 | 0.00 | 0.08 | | | | |
| H2O | 0.00 | 0.00 | 3.97 | 3.95 | 0.00 | 0.00 | 4.51 | 4.55 | 0.00 | | | | |
| TOTAL | 102.17 | 102.72 | 100.52 | 98.83 | 101.53 | 101.78 | 99.18 | 100.35 | 100.18 | | | | |

STRUCTURAL FORMULAE (calculated by MINPROBE)

| | Grt rims | Grt core | Matrix Bt | Core Bt | Matrix Pl | Core Pl | Matrix Ms | Core Ms | Ilmenite |
|---|---|---|---|---|---|---|---|---|---|
| Si | 2.961 | 2.964 | 2.682 | 2.703 | 2.174 | 2.058 | 3.075 | 3.047 | 0.012 |
| Al(IV) | 0.000 | 0.000 | 1.318 | 1.297 | 1.812 | 1.923 | 0.925 | 0.953 | 0.000 |
| Al(VI) | 2.007 | 1.991 | 0.408 | 0.466 | 0.000 | 0.000 | 1.877 | 1.883 | 0.011 |
| Ti | 0.000 | 0.000 | 0.139 | 0.131 | 0.002 | 0.002 | 0.037 | 0.036 | 0.962 |
| Fe3+ | 0.000 | 0.004 | 0.051 | 0.047 | 0.000 | 0.000 | 0.000 | 0.000 | 0.038 |
| Fe2+ | 2.131 | 2.225 | 1.265 | 1.151 | 0.007 | 0.013 | 0.064 | 0.086 | 0.940 |
| Mg | 0.279 | 0.227 | 0.940 | 1.013 | 0.000 | 0.000 | 0.063 | 0.043 | 0.013 |
| Mn | 0.381 | 0.201 | 0.013 | 0.000 | 0.000 | 0.002 | 0.000 | 0.004 | 0.019 |
| Ca | 0.190 | 0.346 | 0.006 | 0.000 | 0.833 | 0.949 | 0.005 | 0.004 | 0.002 |
| Na | 0.000 | 0.000 | 0.019 | 0.038 | 0.182 | 0.066 | 0.072 | 0.086 | 0.000 |
| K | 0.000 | 0.000 | 0.908 | 0.869 | 0.000 | 0.000 | 0.801 | 0.798 | 0.000 |
| Cr | 0.006 | 0.005 | 0.003 | 0.000 | 0.000 | 0.000 | 0.003 | 0.000 | 0.002 |
| O | 12.000 | 12.000 | 10.000 | 10.000 | 8.000 | 8.000 | 10.000 | 10.000 | 3.000 |
| OH | 0.000 | 0.000 | 2.000 | 2.000 | 0.000 | 0.000 | 2.000 | 2.000 | 0.000 |

IDEAL ACTIVITIES (values for garnet end members are cationic ratios): Input for TWQ software, calculated by MINPROBE

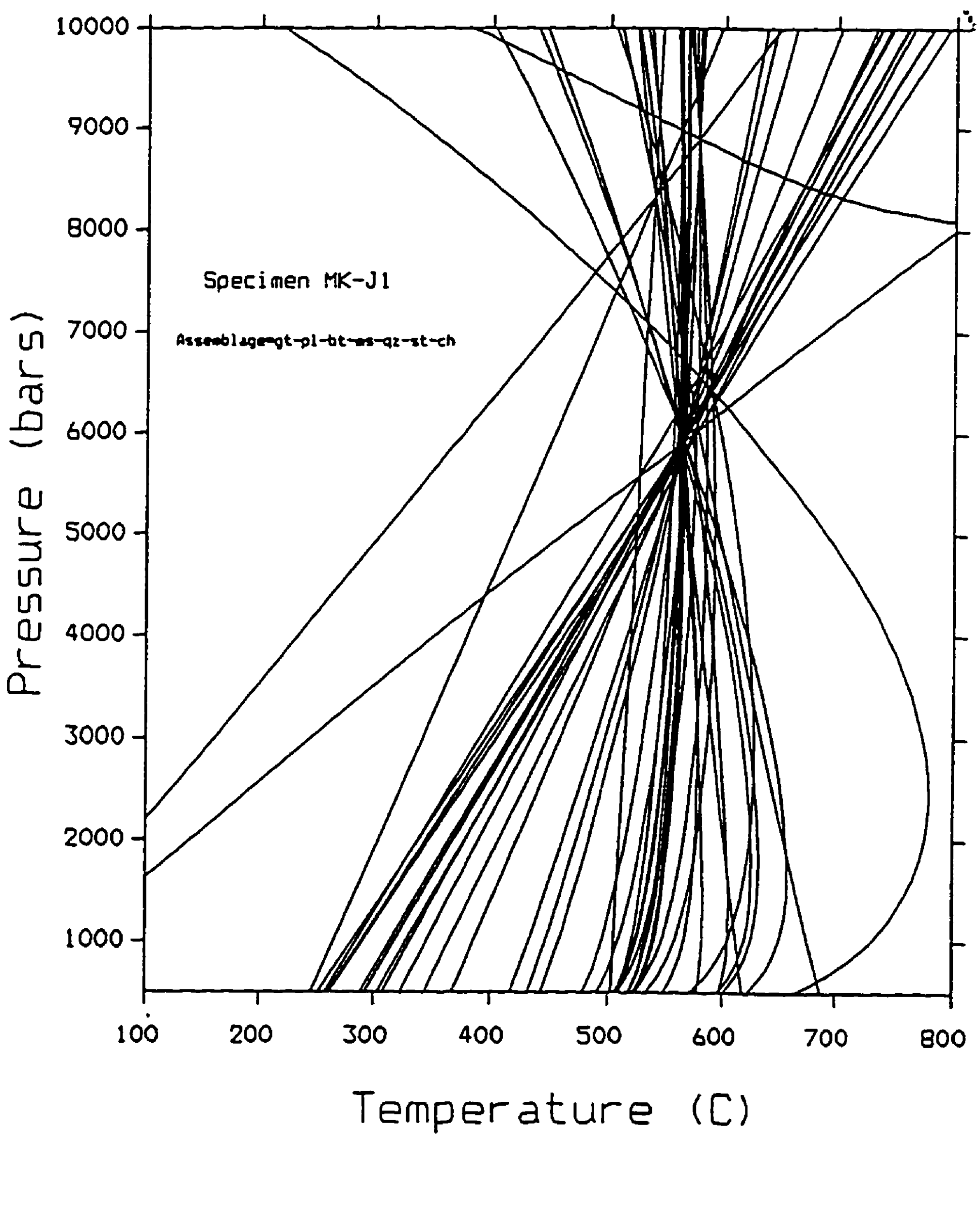
| | Grt rims | Grt core | | Matrix Bt | Core Bt | | Matrix Pl | Core Pl | Matrix Ms | Core Ms | Ilmenite | Rutile | Quartz | Sillimanite | Andalusite? |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| alm | 0.715 | 0.742 | Xphl | 0.028 | 0.034 | Xan | 0.822 | 0.936 | 0.677 | 0.670 | 1.000 | 1.000 | 1.000 | 1.000 | 1.000 |
| prp | 0.094 | 0.076 | Xann | 0.068 | 0.049 | Xab | 0.177 | 0.064 | | | | | | | |
| sps | 0.128 | 0.067 | | | | Xor | 0.000 | 0.000 | | | | | | | |
| grs | 0.064 | 0.115 | | | | | | | | | | | | | |

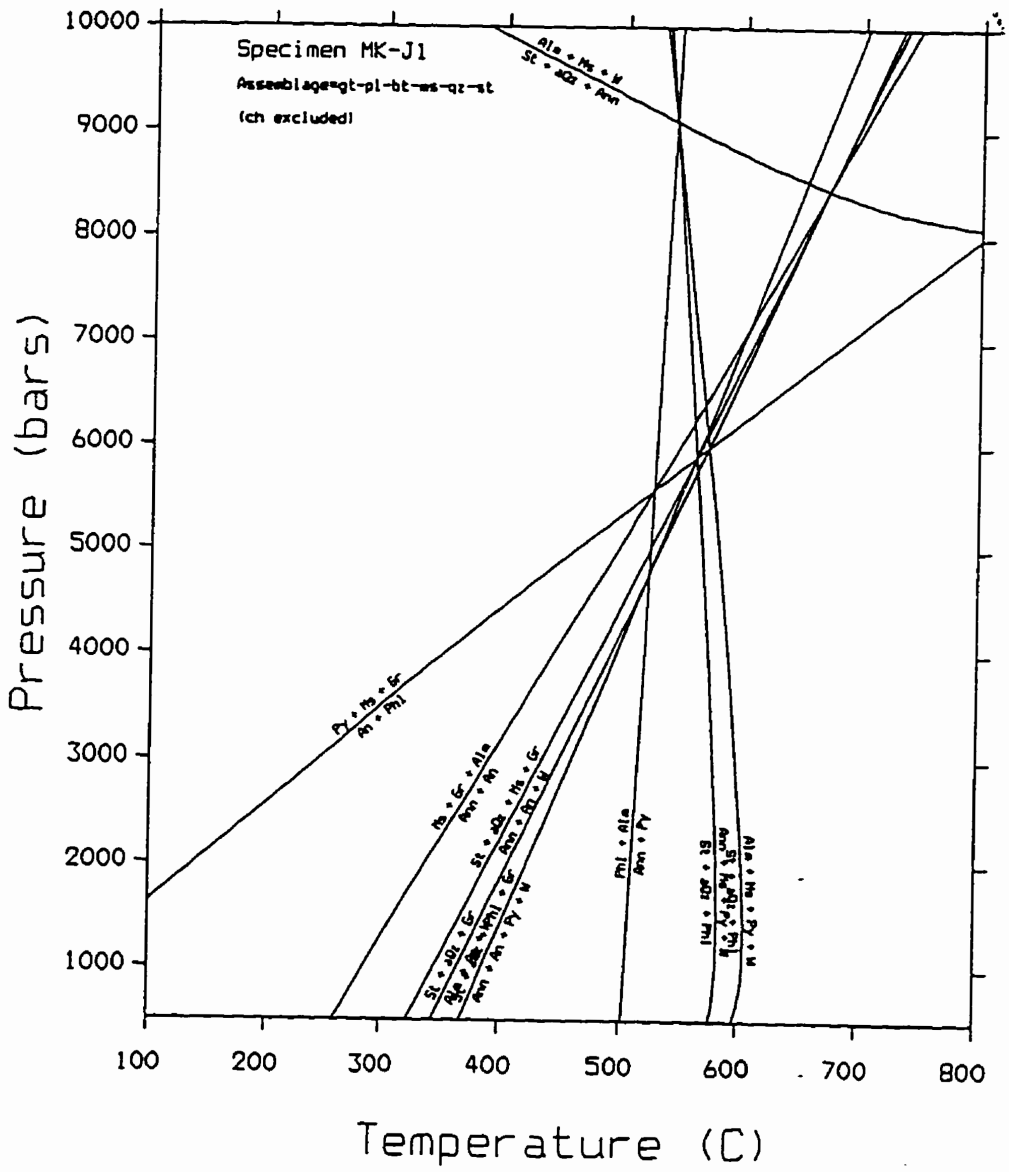
MINERAL ANALYSES: SPECIMEN FRY-XY

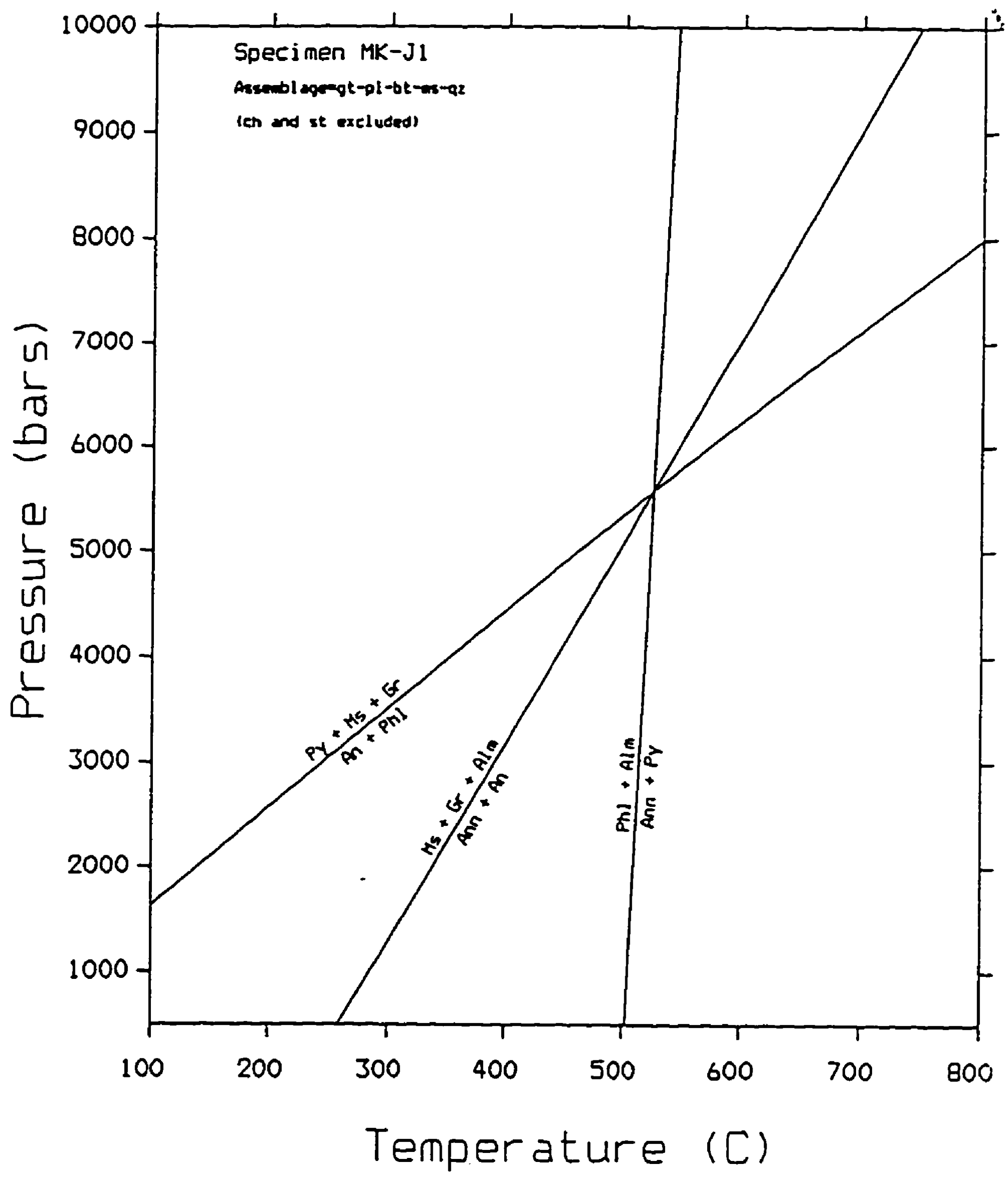
| | Grt rims | Grt core | Biotite | Matrix Pl | Core Pl | Muscovite | Ilmenite | Quartz | Sillimanite |
|---|---|---|---|---|---|---|---|---|---|
| AVERAGED MICROPROBE ANALYSES (wt %) | | | | | | | | | |
| SiO2 | 37.79 | 37.79 | 35.49 | 46.14 | 45.07 | 45.69 | 0.51 | | |
| Al2O3 | 21.66 | 21.88 | 19.22 | 34.85 | 35.93 | 34.94 | 0.40 | | |
| TiO2 | 0.02 | 0.44 | 2.73 | 0.06 | 0.10 | 0.93 | 51.64 | | |
| Fe2O3 | 0.00 | 0.00 | 0.00 | 0.00 | 0.44 | 0.07 | 0.00 | | |
| FeO | 32.70 | 33.99 | 21.47 | 0.08 | 0.00 | 1.14 | 44.25 | | |
| MgO | 2.28 | 1.92 | 7.95 | 0.00 | 0.00 | 0.61 | 0.15 | | |
| MnO | 5.35 | 2.61 | 0.03 | 0.05 | 0.11 | 0.00 | 1.75 | | |
| CaO | 2.85 | 4.71 | 0.00 | 18.30 | 19.19 | 0.00 | 0.15 | | |
| Na2O | 0.00 | 0.00 | 0.10 | 1.49 | 0.85 | 0.66 | 0.00 | | |
| K2O | 0.02 | 0.00 | 9.35 | 0.00 | 0.00 | 10.30 | 0.00 | | |
| Cr2O3 | 0.08 | 0.00 | 0.18 | 0.00 | 0.00 | 0.00 | 0.06 | | |
| H2O | 0.00 | 0.00 | 3.95 | 0.00 | 0.00 | 4.46 | 0.00 | | |
| TOTAL | 102.75 | 103.34 | 100.47 | 100.97 | 101.69 | 98.79 | 98.91 | | |
| STRUCTURAL FORMULAE (Calculated by MINPROBE) | | | | | | | | | |
| Si | 2.954 | 2.911 | 2.689 | 2.108 | 2.051 | 3.070 | 0.013 | | |
| Al(IV) | 0.000 | 0.000 | 1.311 | 1.876 | 1.927 | 0.930 | 0.000 | | |
| Al(VI) | 1.996 | 1.987 | 0.406 | 0.000 | 0.000 | 1.837 | 0.012 | | |
| Ti | 0.001 | 0.025 | 0.156 | 0.002 | 0.003 | 0.047 | 0.984 | | |
| Fe3+ | 0.000 | 0.000 | 0.000 | 0.000 | 0.015 | 0.003 | 0.000 | | |
| Fe2+ | 2.138 | 2.190 | 1.361 | 0.003 | 0.000 | 0.064 | 0.938 | | |
| Mg | 0.266 | 0.220 | 0.898 | 0.000 | 0.000 | 0.061 | 0.006 | | |
| Mn | 0.354 | 0.170 | 0.002 | 0.002 | 0.004 | 0.000 | 0.038 | | |
| Ca | 0.239 | 0.389 | 0.000 | 0.896 | 0.936 | 0.000 | 0.004 | | |
| Na | 0.000 | 0.000 | 0.015 | 0.132 | 0.075 | 0.086 | 0.000 | | |
| K | 0.002 | 0.000 | 0.904 | 0.000 | 0.000 | 0.883 | 0.000 | | |
| Cr | 0.005 | 0.000 | 0.011 | 0.000 | 0.000 | 0.000 | 0.001 | | |
| O | 12.000 | 12.000 | 10.000 | 8.000 | 8.000 | 10.000 | 3.000 | | |
| OH | 0.000 | 0.000 | 2.000 | 0.000 | 0.000 | 2.000 | 0.000 | | |

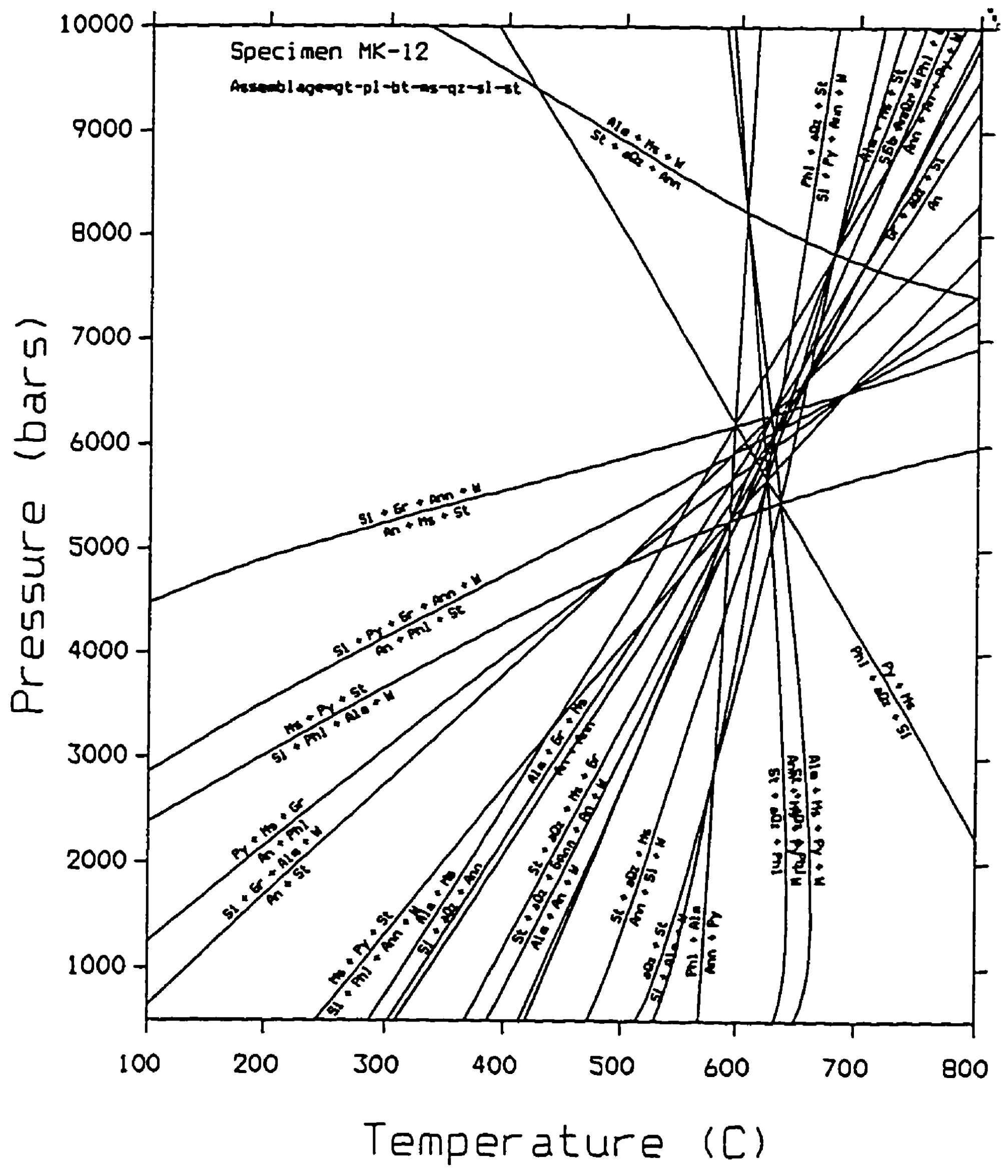
IDEAL ACTIVITIES (values for garnet end members are cationic ratios)
Input for TWQ software, calculated by MINPROBE

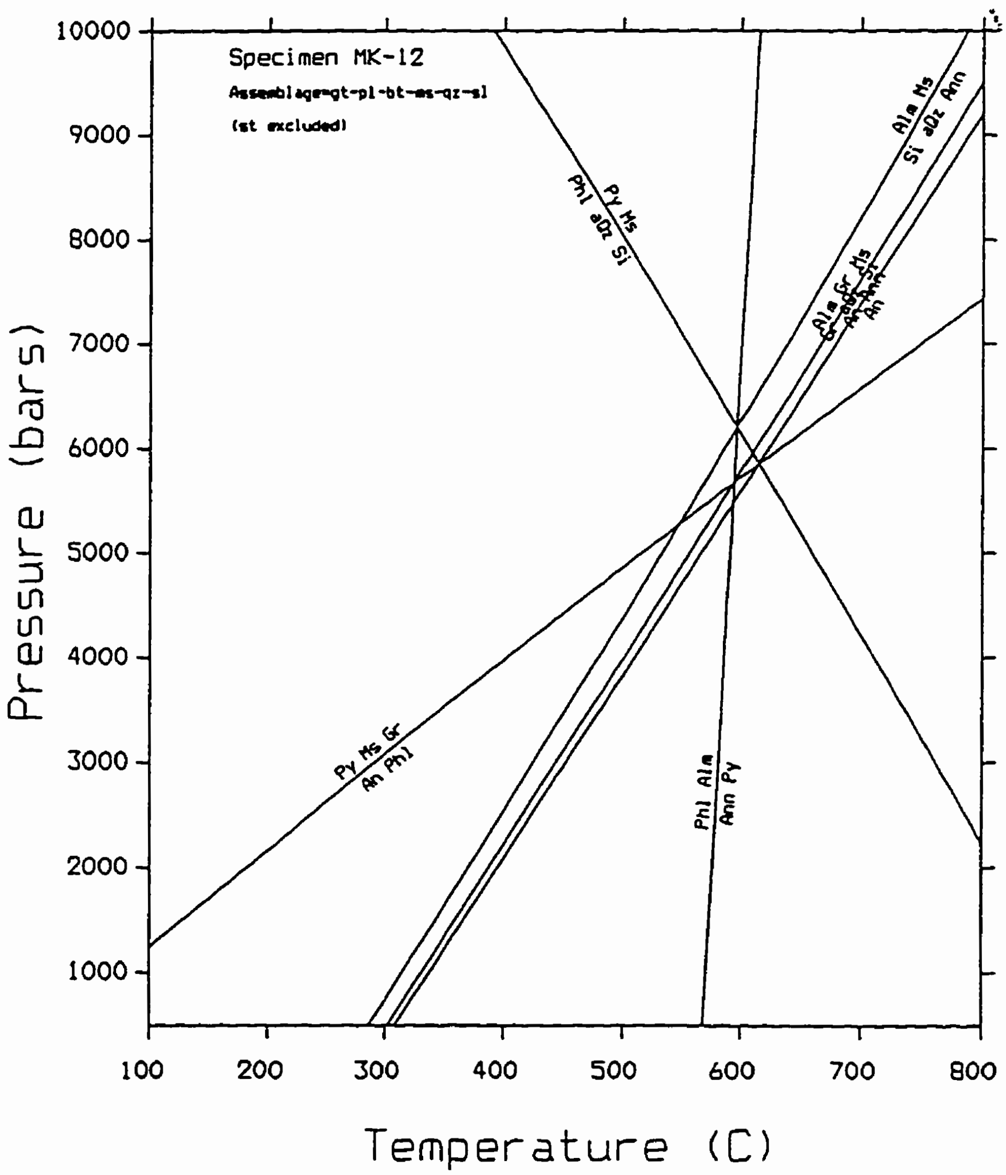
| | Grt rims | Grt core | | Biotite | | Matrix Pl | Core Pl | Muscovite | Ilmenite | Quartz | Sillimanite |
|---|---|---|---|---|---|---|---|---|---|---|---|
| alm | 0.713 | 0.737 | Xphl | 0.024 | Xan | 0.886 | 0.939 | 0.736 | | 1.000 | 1.000 |
| prp | 0.089 | 0.074 | Xann | 0.084 | Xab | 0.114 | 0.061 | | | | |
| sps | 0.118 | 0.057 | | | Xor | 0.000 | 0.000 | | | | |
| grs | 0.008 | 0.131 | | | | | | | | | |

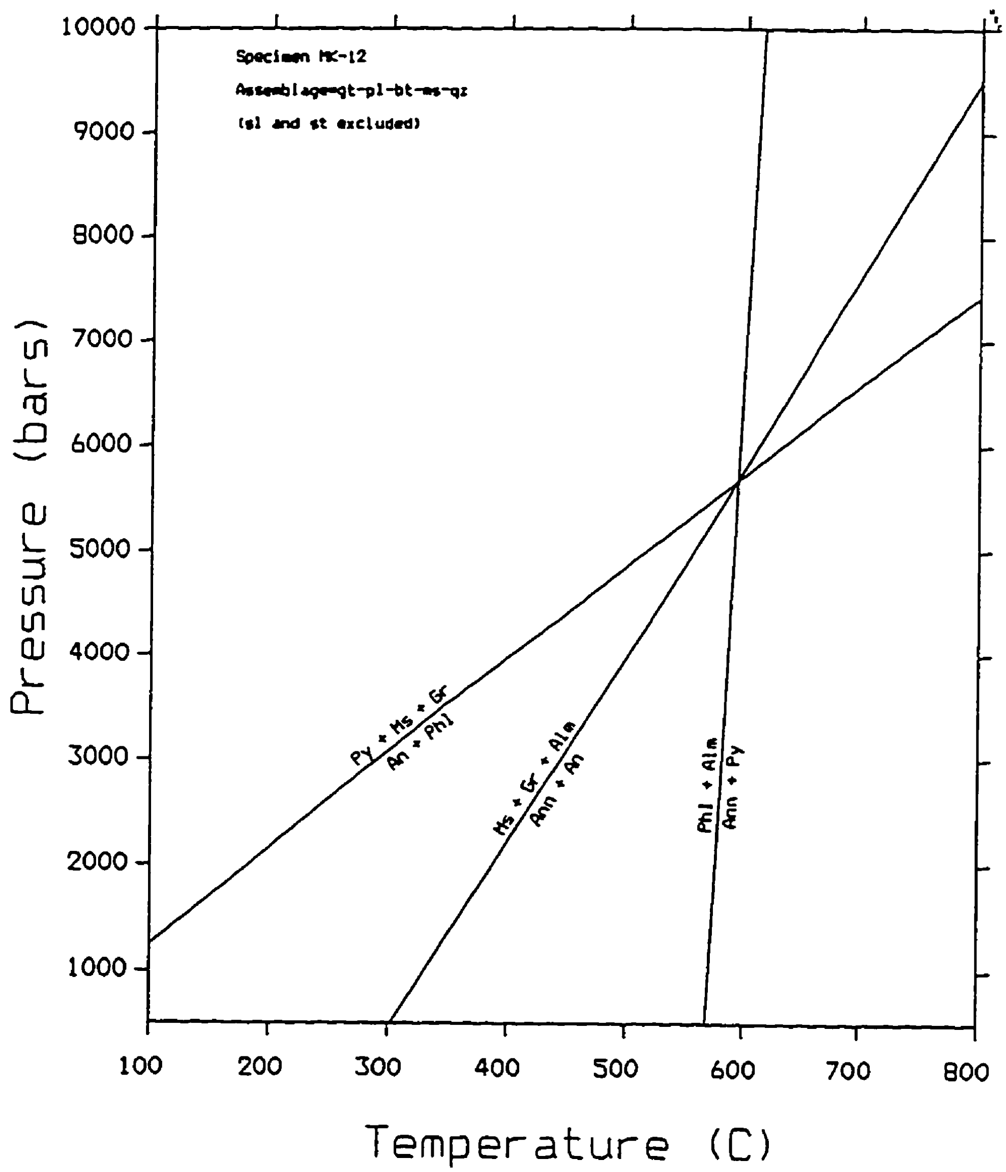

Specimen MK-J1
Assemblage=gt-pl-bt-ms-qz-st-ch
Pressure (bars)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Temperature (C)


Specimen MK-J1
Assemblage=gt-pl-bt-ms-qz-st
(ch excluded)
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Py + Ms + Gr
An + Phl
Phl + Alm
Ann + Py

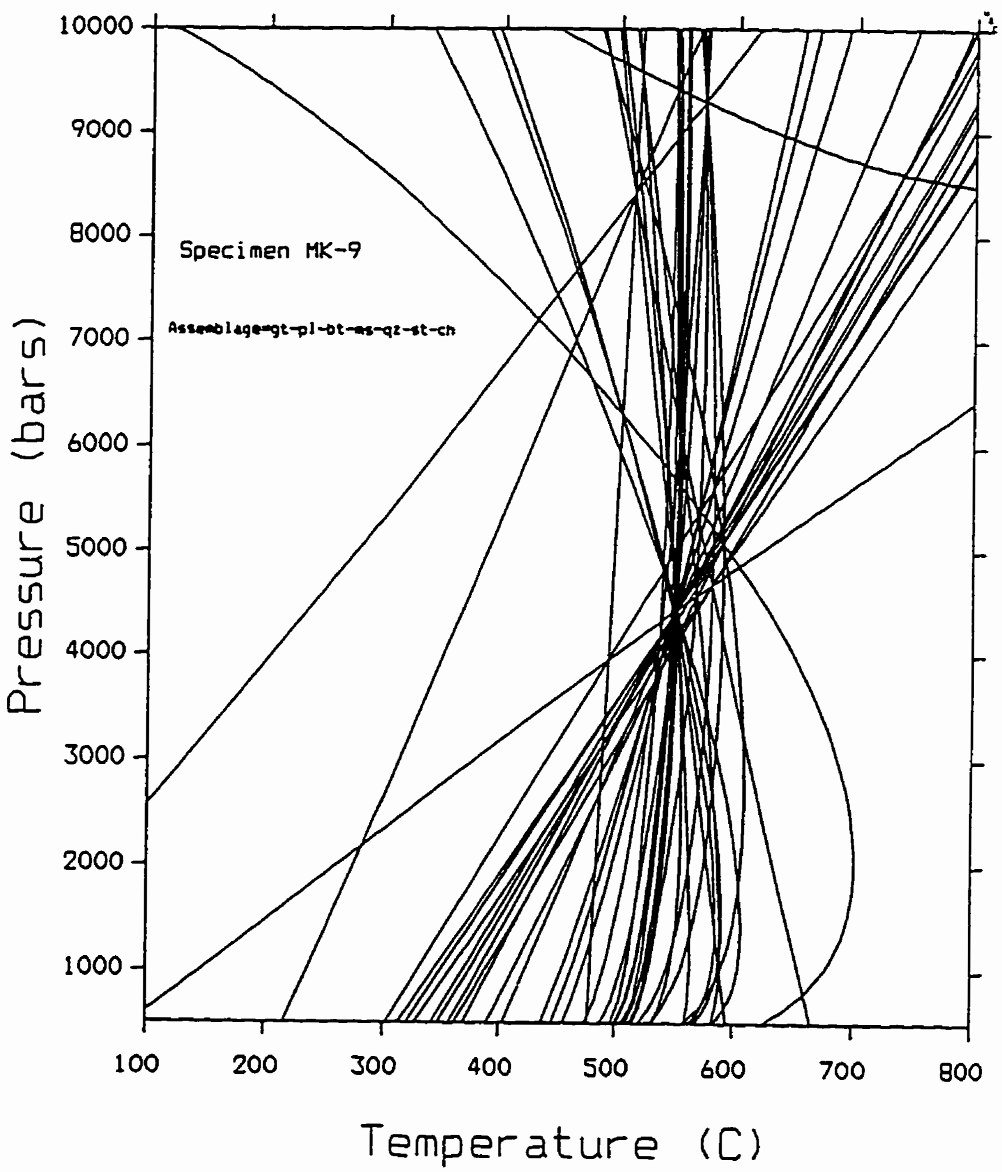

Specimen MK-J1
Assemblage=gt-pl-bt-ms-qz
(ch and st excluded)
Py + Ms + Gr
An + Phl
Ms + Gr + Alm
Ann + An
Phl + Alm
Ann + Py
Pressure (bars)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
Temperature (C)
100
200
300
400
500
600
700
800


Specimen MK-12
Assemblage=gt-pl-bt-ms-qz-sl-st
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800

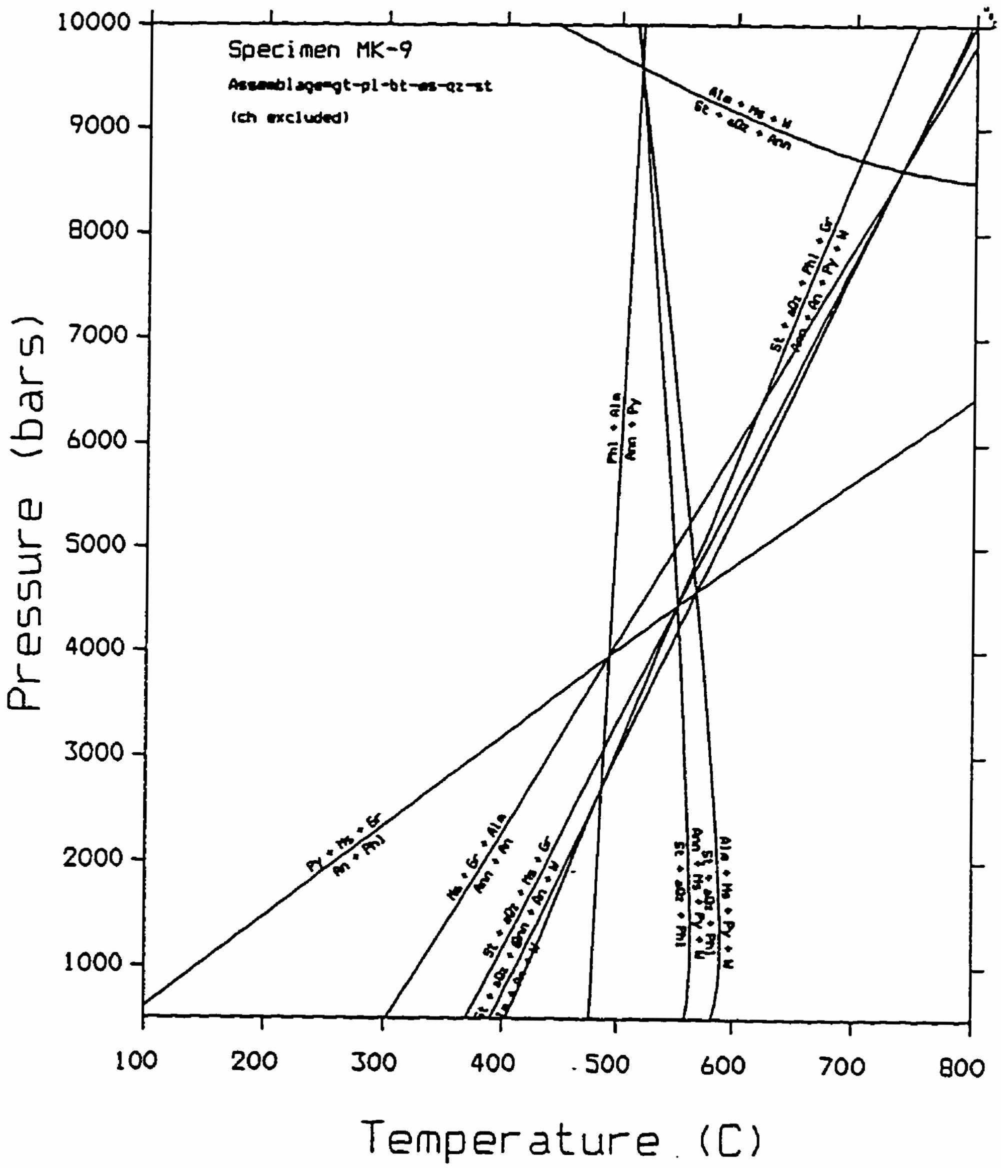

Specimen MK-12
Assemblage=gt-pl-bt-ms-qz-sl
(st excluded)
Py Ms
Phl aQz Si
Alm Ms
Si aQz Ann
Alm Gr Ms
Py Ms Gr
An Phl
Phl Alm
Ann Py
Pressure (bars)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Temperature (C)

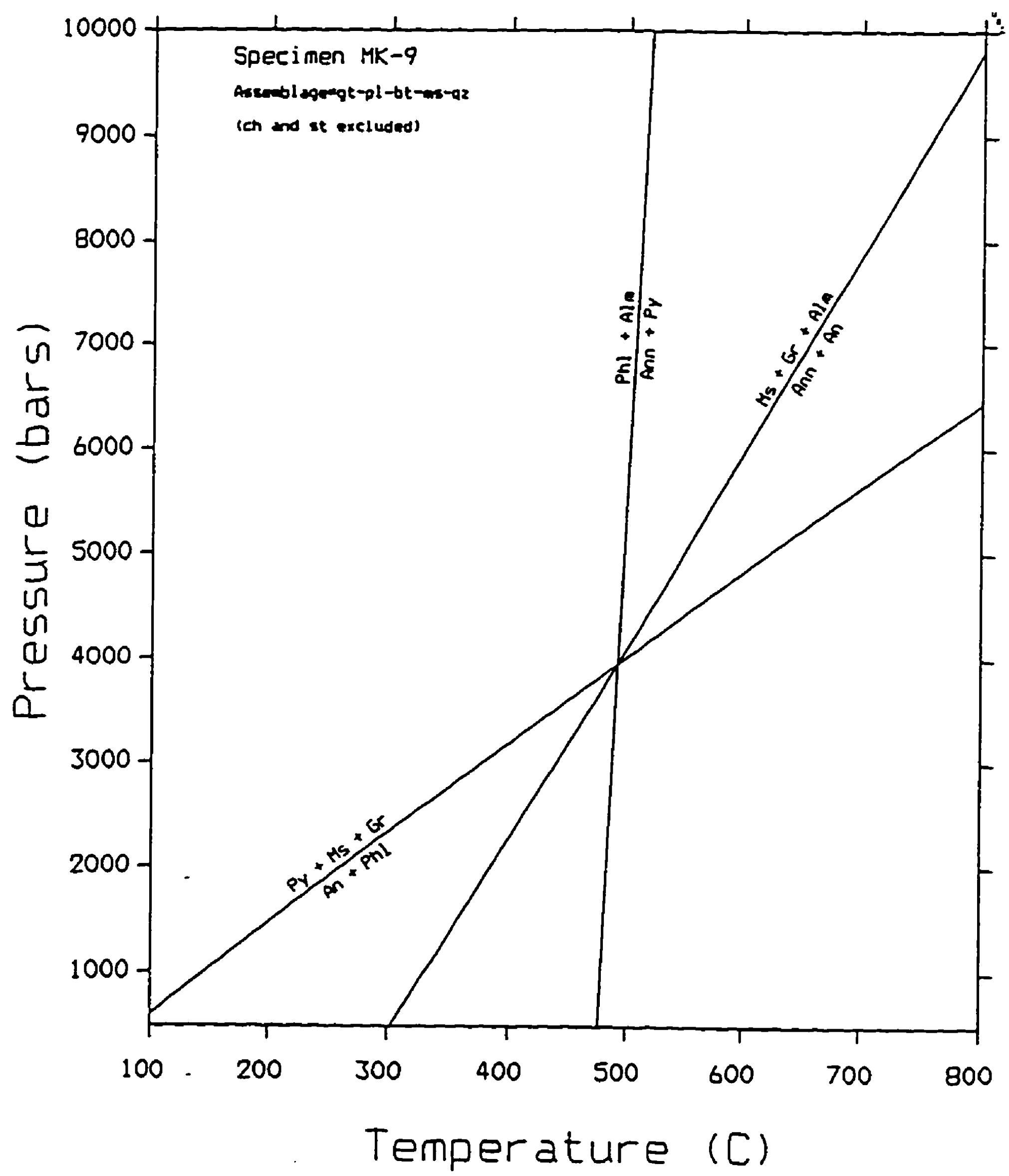


Specimen MK-12
Assemblage=gt-pl-bt-ms-qz
(sl and st excluded)
Py + Ms + Gr
An + Phl
Ms + Gr + Alm
Ann + An
Phl + Alm
Ann + Py
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
Pressure (bars)
100
200
300
400
500
600
700
800
Temperature (C)

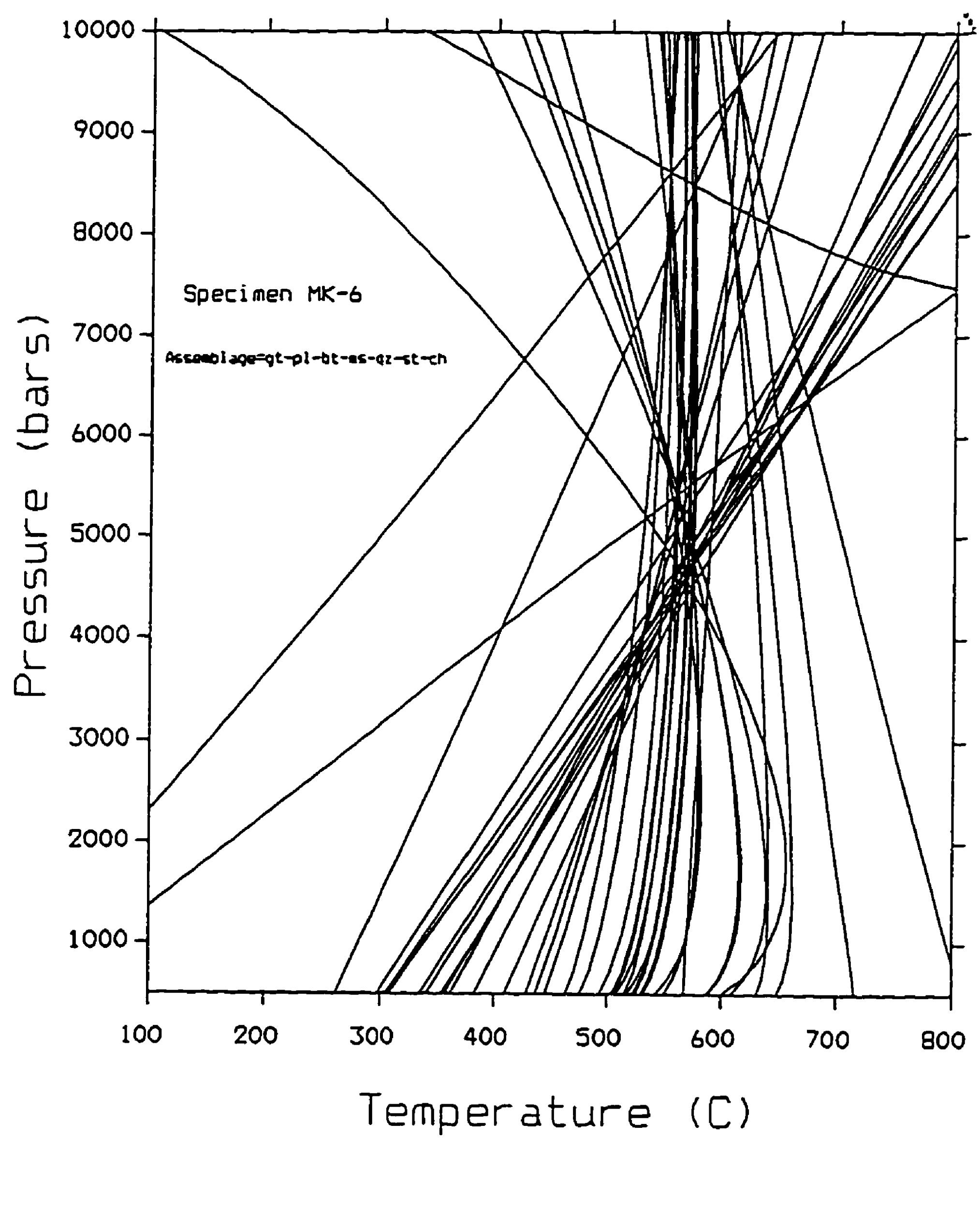


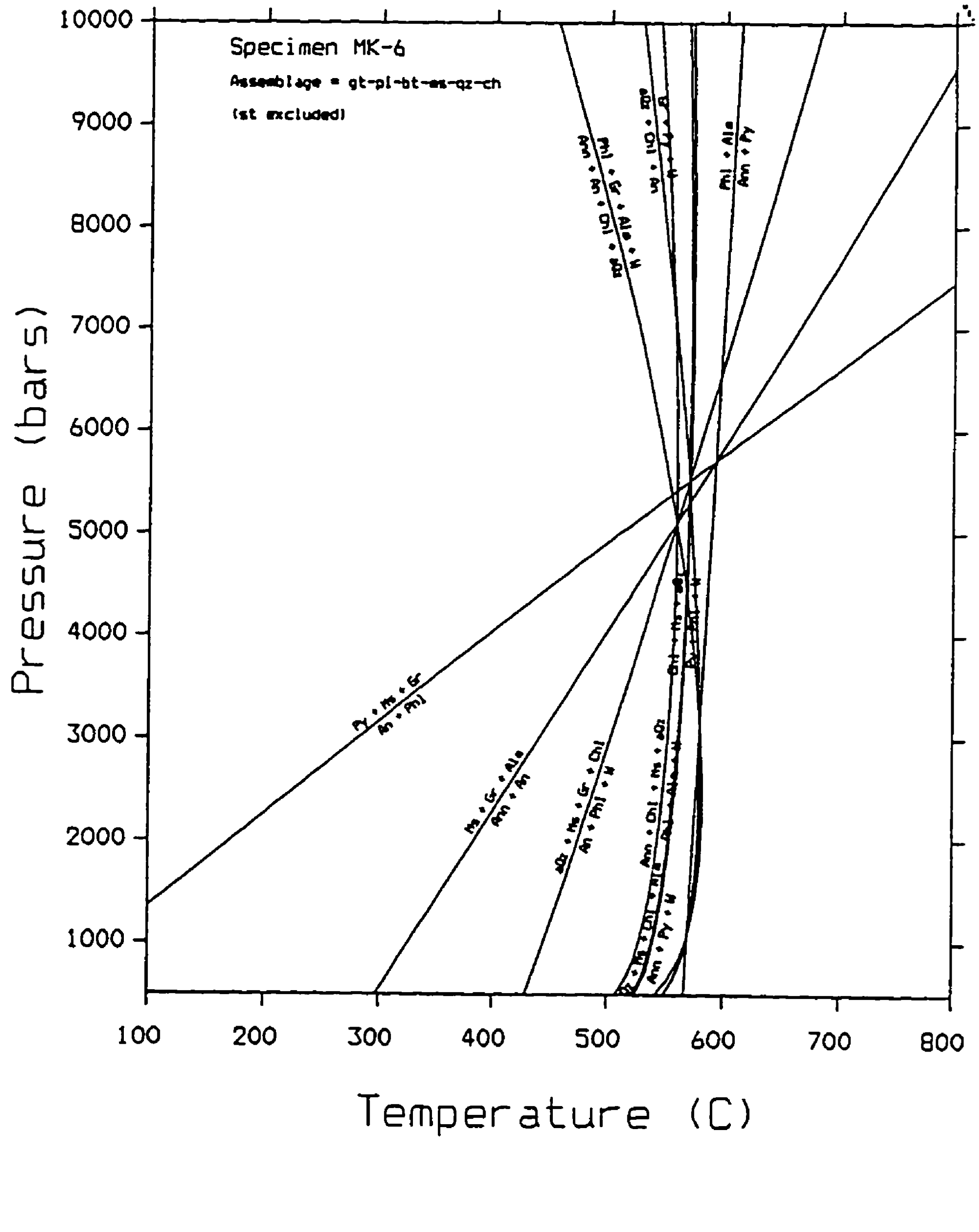
Specimen MK-9
Assemblage=gt-pl-bt-ms-qz-st-ch
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
Pressure (bars)
100
200
300
400
500
600
700
800
Temperature (C)

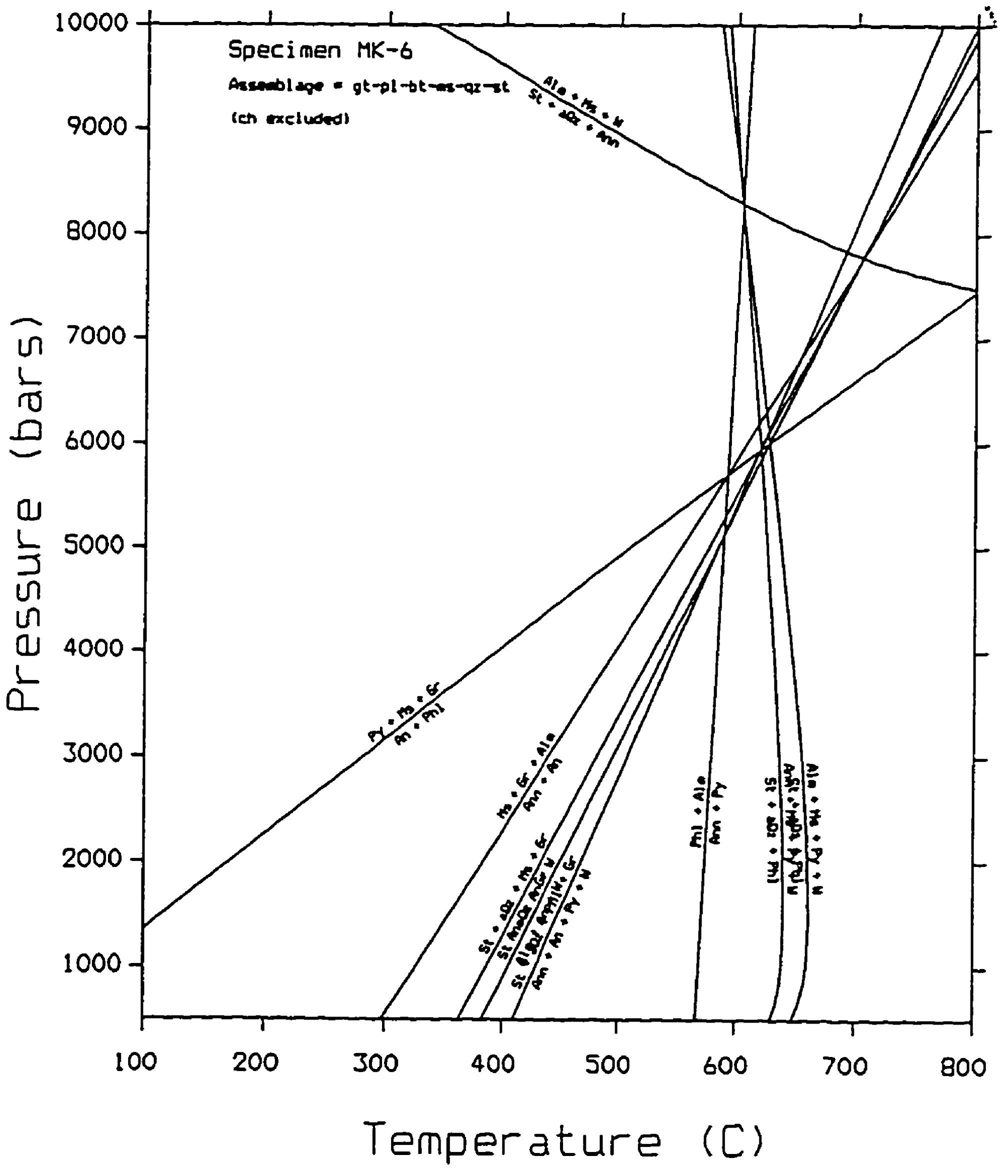

Specimen MK-9
Assemblage=gt-pl-bt-ms-qz-st
(ch excluded)
Pressure (bars)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Temperature (C)

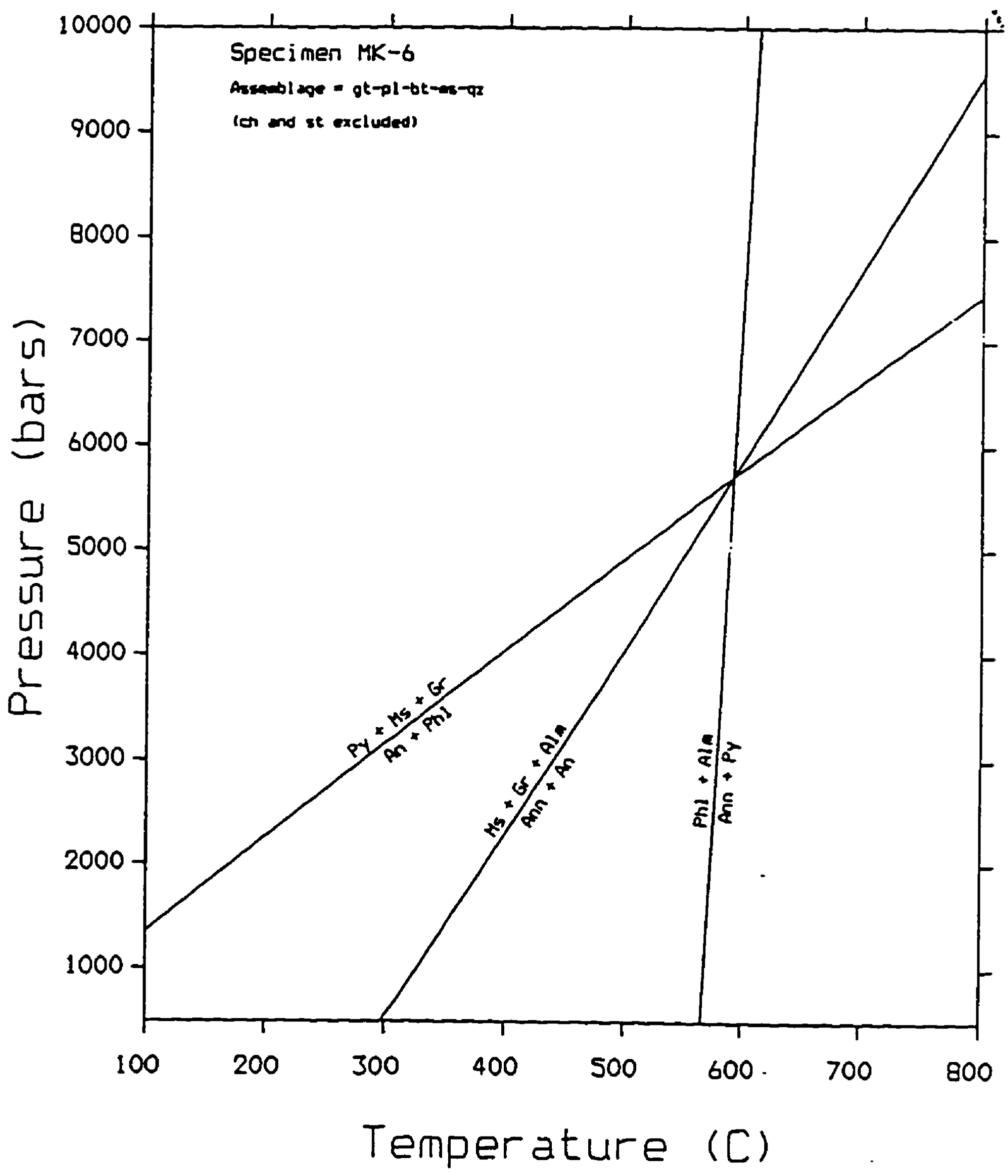


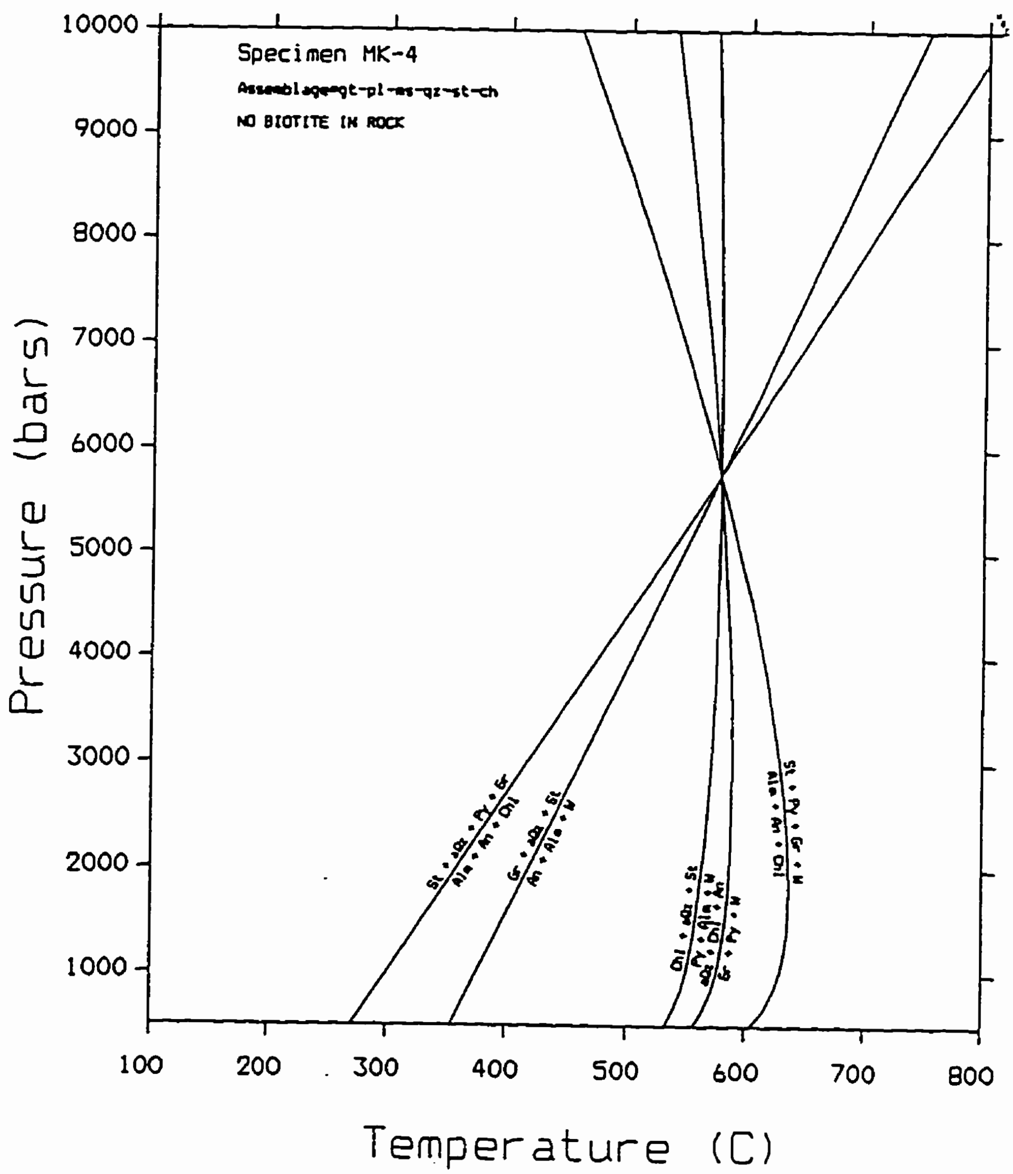
Specimen MK-9
Assemblage=gt-pl-bt-ms-qz
(ch and st excluded)
Phl + Alm
Ann + Py
Ms + Gr + Alm
Ann + An
Py + Ms + Gr
An + Phl
Pressure (bars)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Temperature (C)


Specimen MK-6
Assemblage=gt-pl-bt-ms-qz-st-ch
Pressure (bars)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Temperature (C)

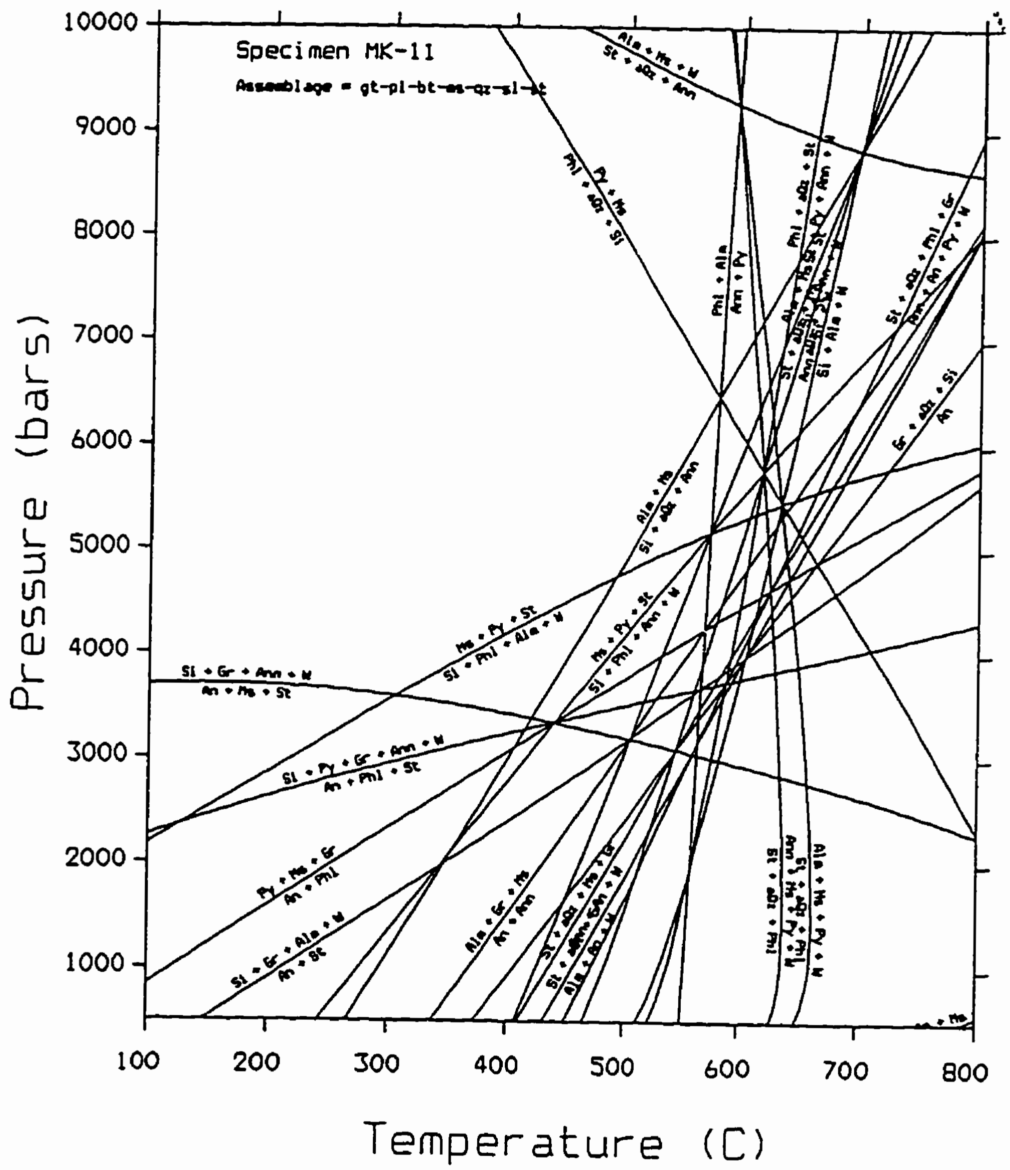

Specimen MK-6
Assemblage = gt-pl-bt-ms-qz-ch
(st excluded)
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Py + Ms + Gr
An + Phl
Ms + Gr + Als
An + An
Phl + Als
Ann + Py

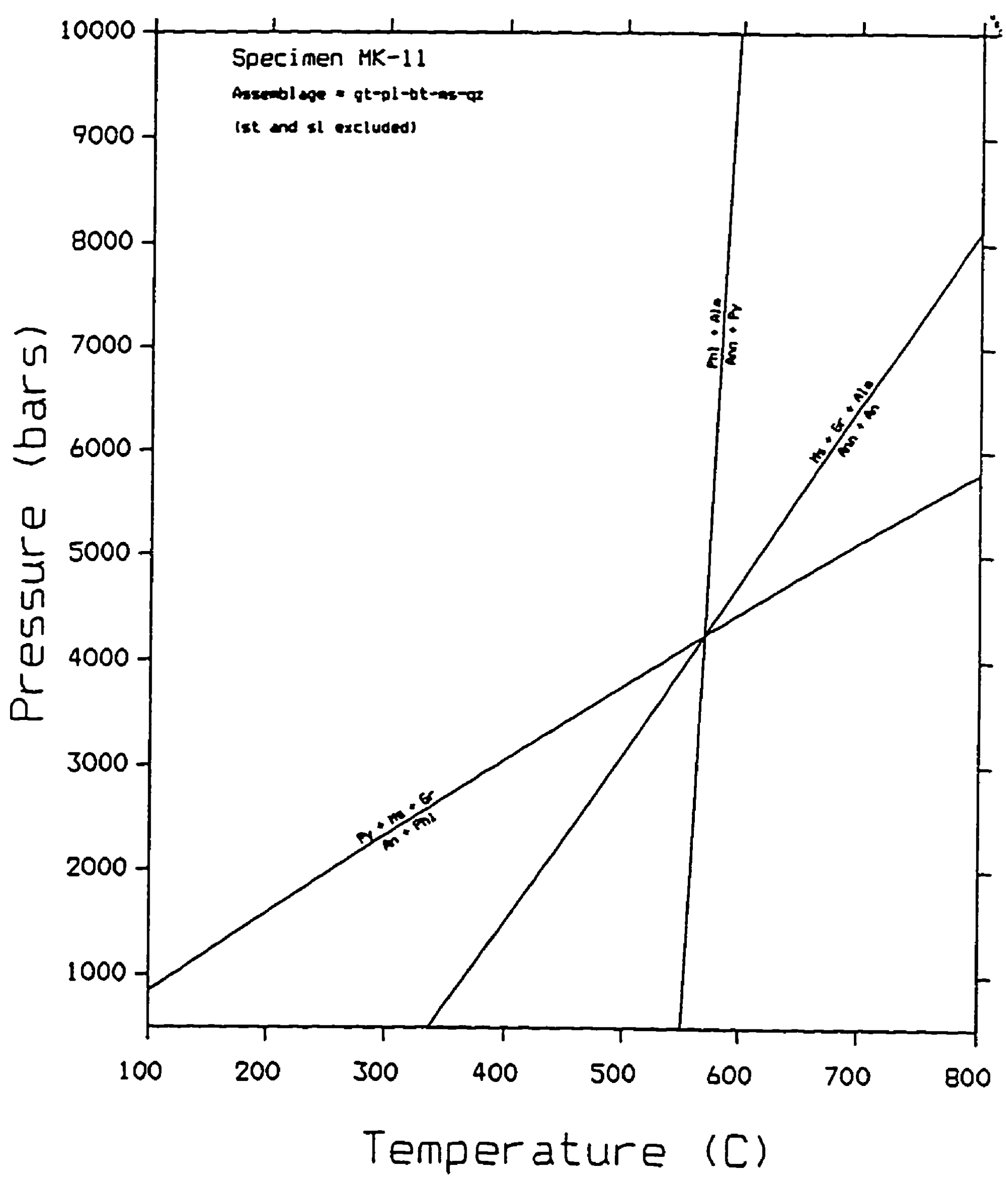

Specimen MK-6
Assemblage = gt-pl-bt-ms-qz-st
(ch excluded)
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Py + Ms + Gr
An + Phl
Ms + Gr + Alm
Ann + An
Phl + Alm
Ann + Py

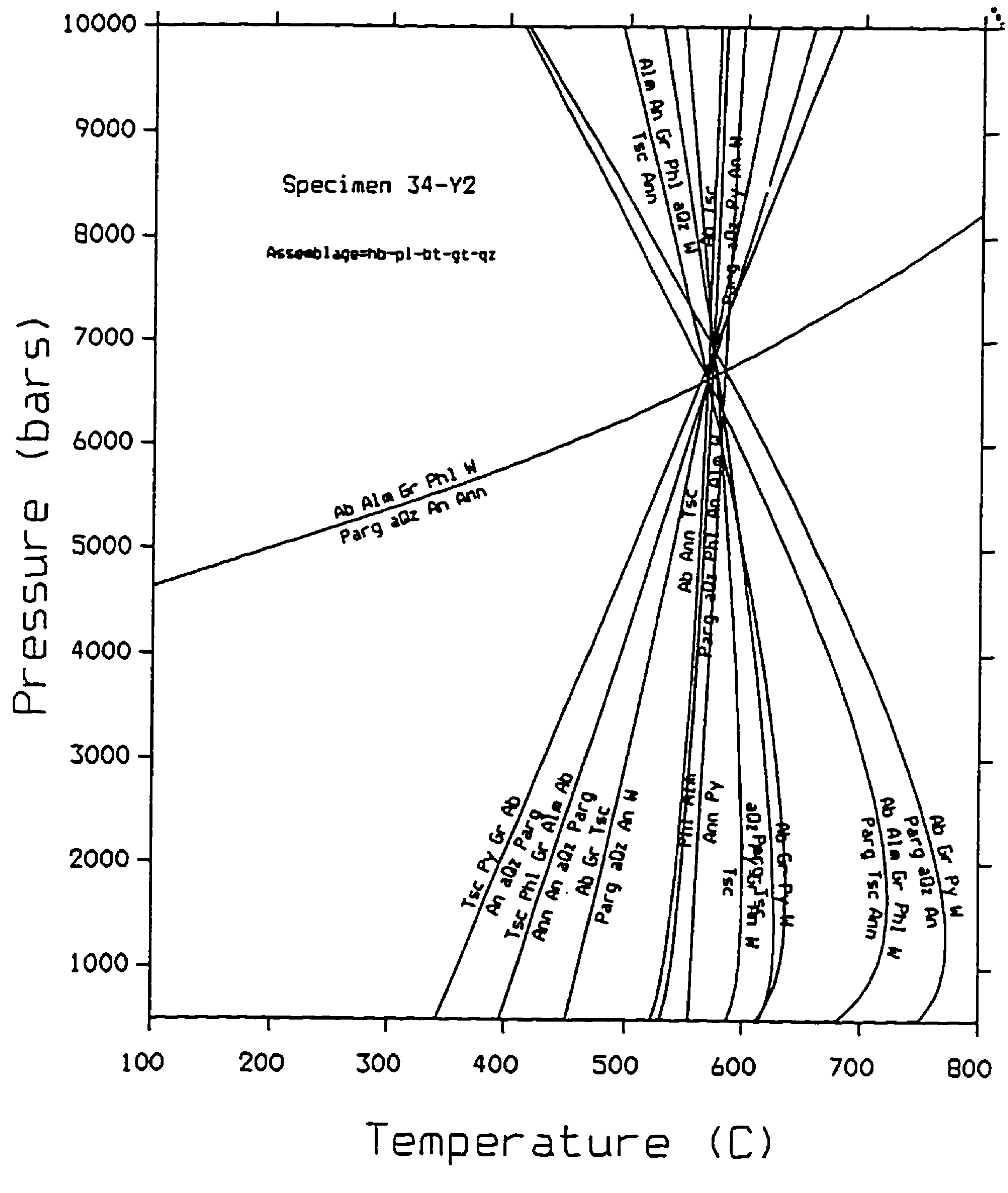

Specimen MK-6
Assemblage = gt-pl-bt-ms-qz
(ch and st excluded)
Py + Ms + Gr
An + Phl
Ms + Gr + Alm
Ann + An
Phl + Alm
Ann + Py
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800

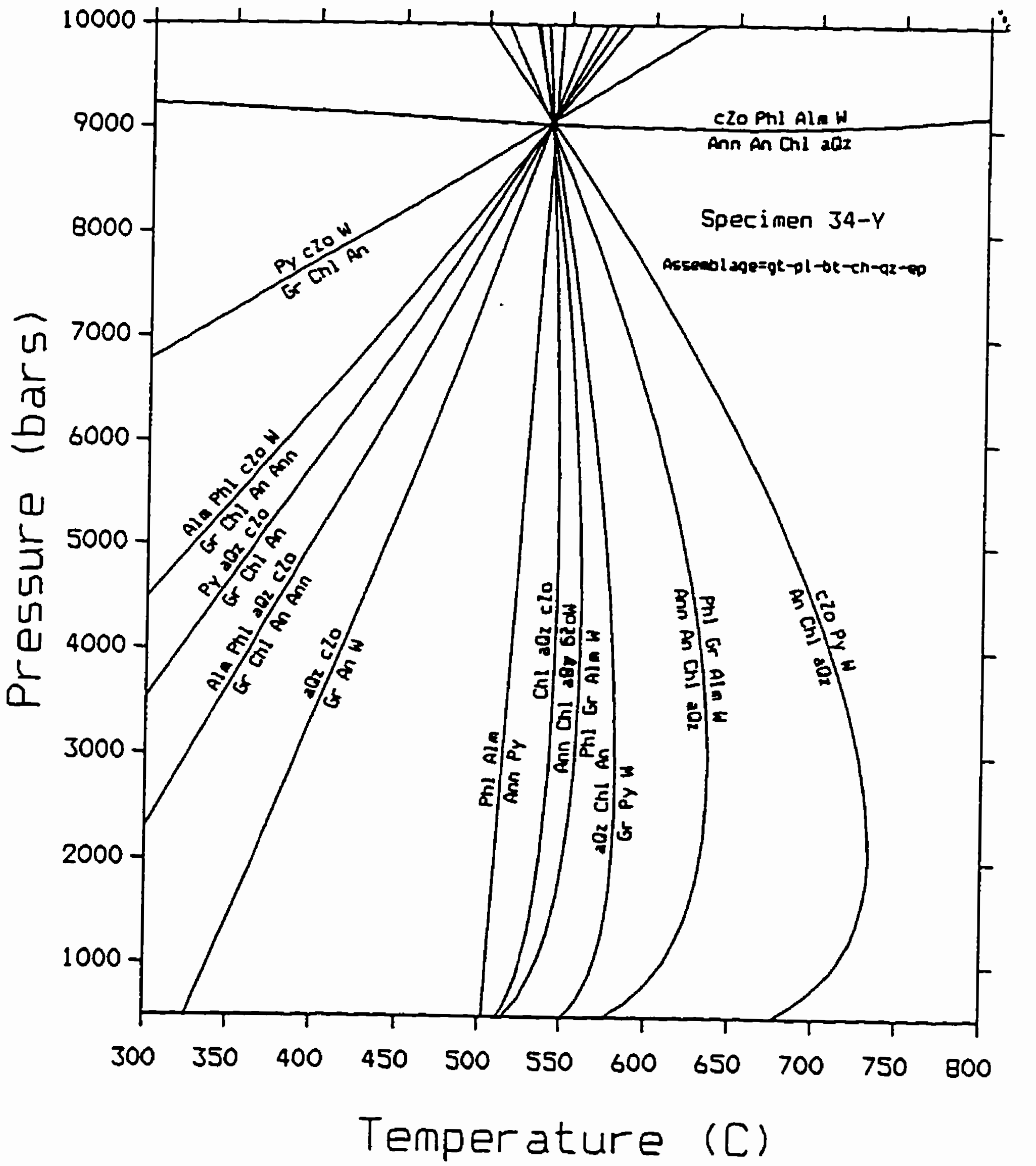


Specimen MK-4
Assemblage=gt-pl-ms-qz-st-ch
NO BIOTITE IN ROCK
Pressure (bars)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Temperature (C)

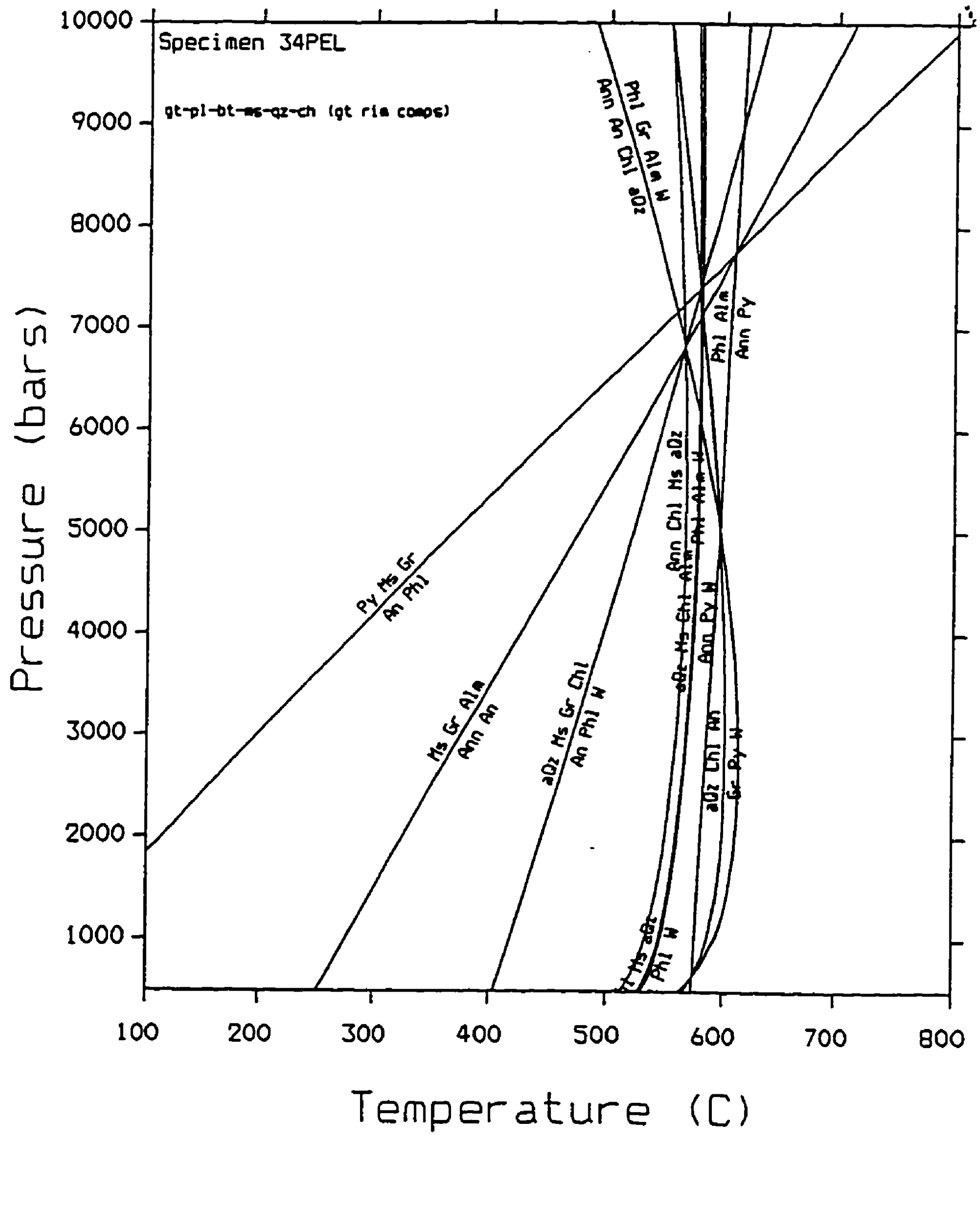

Specimen MK-11
Assemblage = gt-pl-bt-ms-qz-sl-st
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800


10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
Pressure (bars)
Specimen MK-11
Assemblage = gt-pl-bt-ms-qz
(st and sl excluded)
Phl + Alm
Ann + Py
Py + Ms + Gr
An + Phl
100
200
300
400
500
600
700
800
Temperature (C)

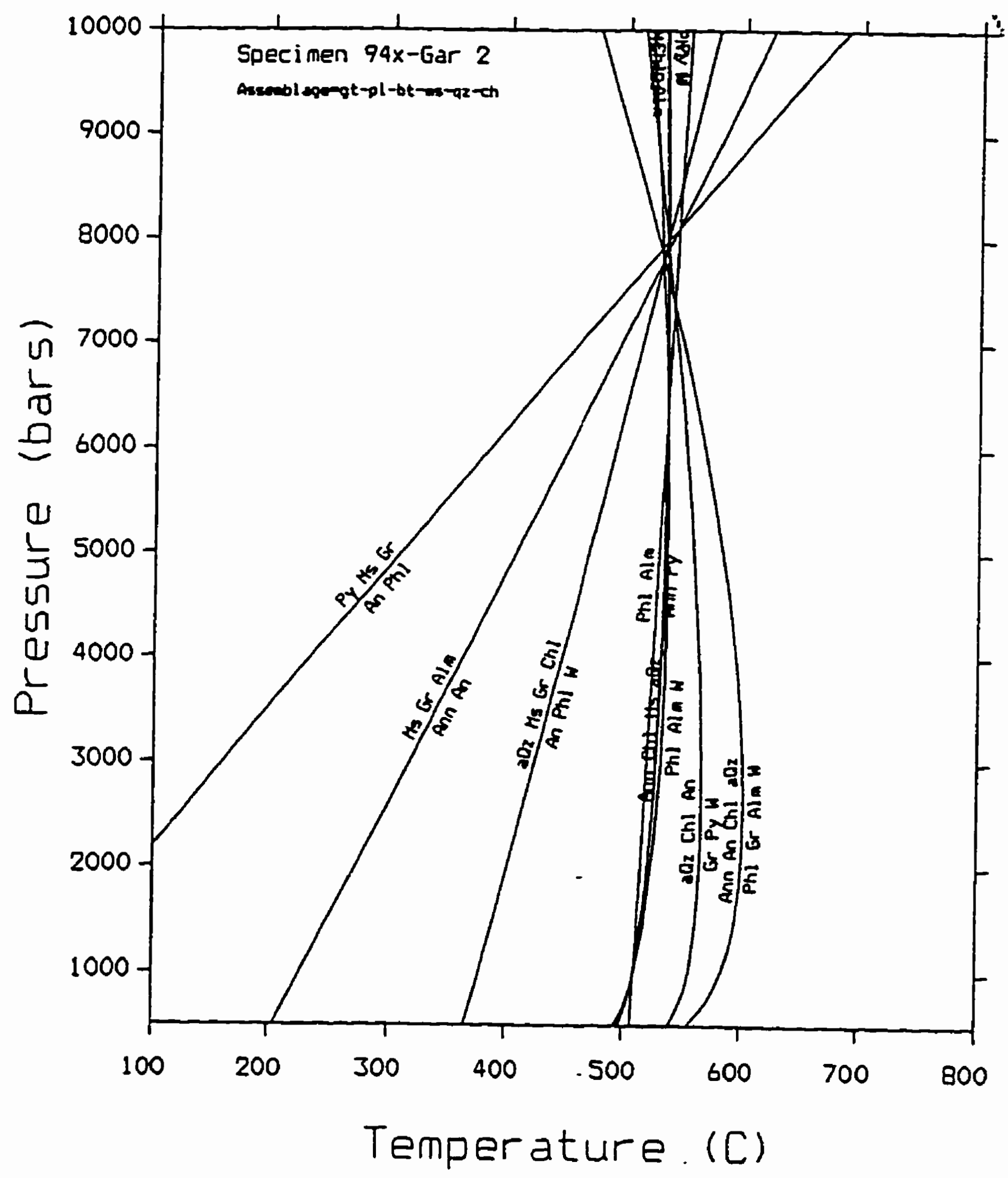

Specimen 34-Y2
Assemblage=hb-pl-bt-gt-qz
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Ab Alm Gr Phl W
Parg aQz An Ann
Alm An Gr Phl aQz W
Tsc Ann
Ab Ann Tsc
Tsc Py Gr Ab
An aQz Parg
Tsc Phl Gr Alm Ab
Ann An aQz Parg
Ab Gr Tsc
Parg aQz An W
Ann Py
Ab Gr Py W
Parg aQz An
Ab Alm Gr Phl W
Parg Tsc Ann

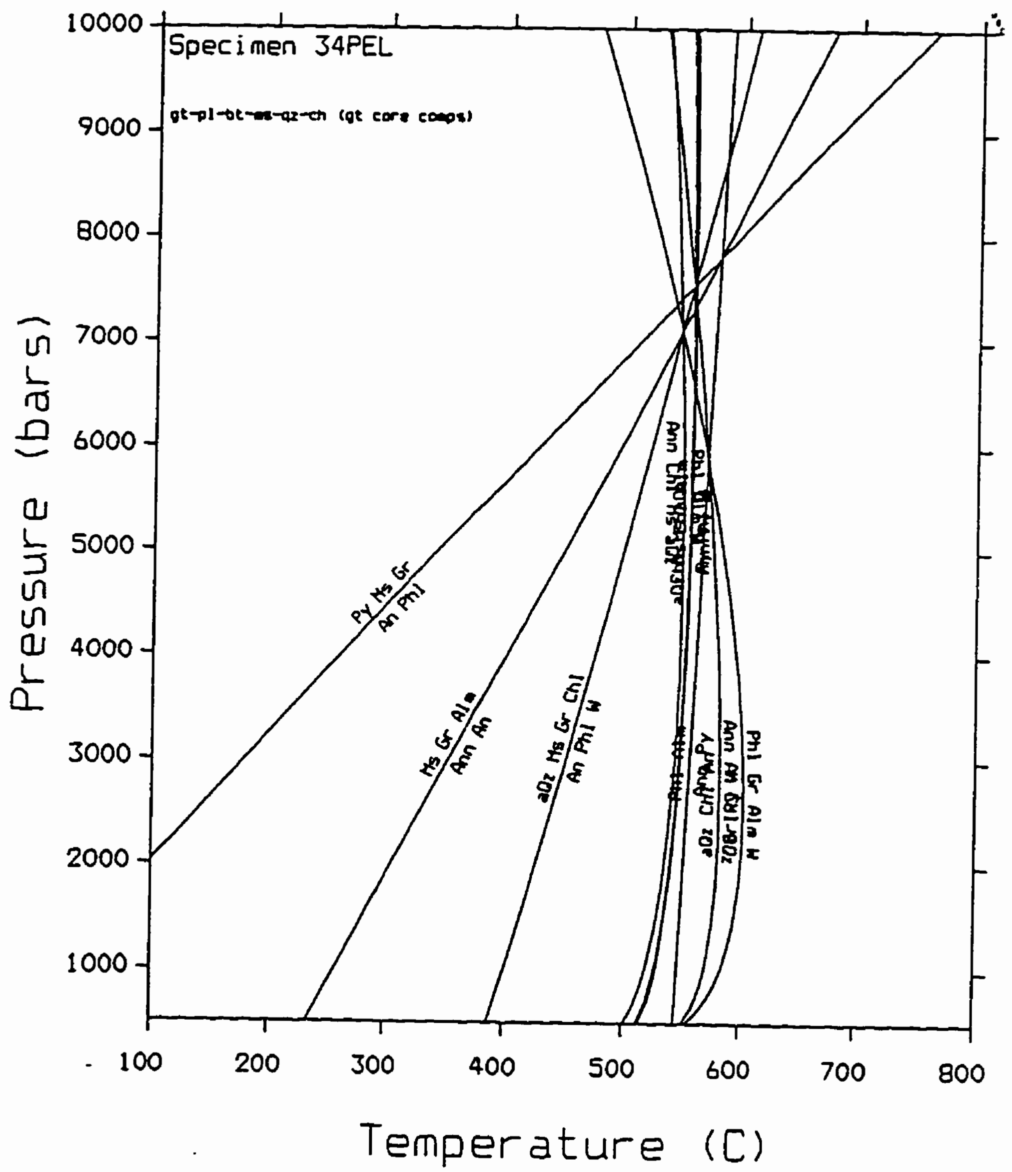


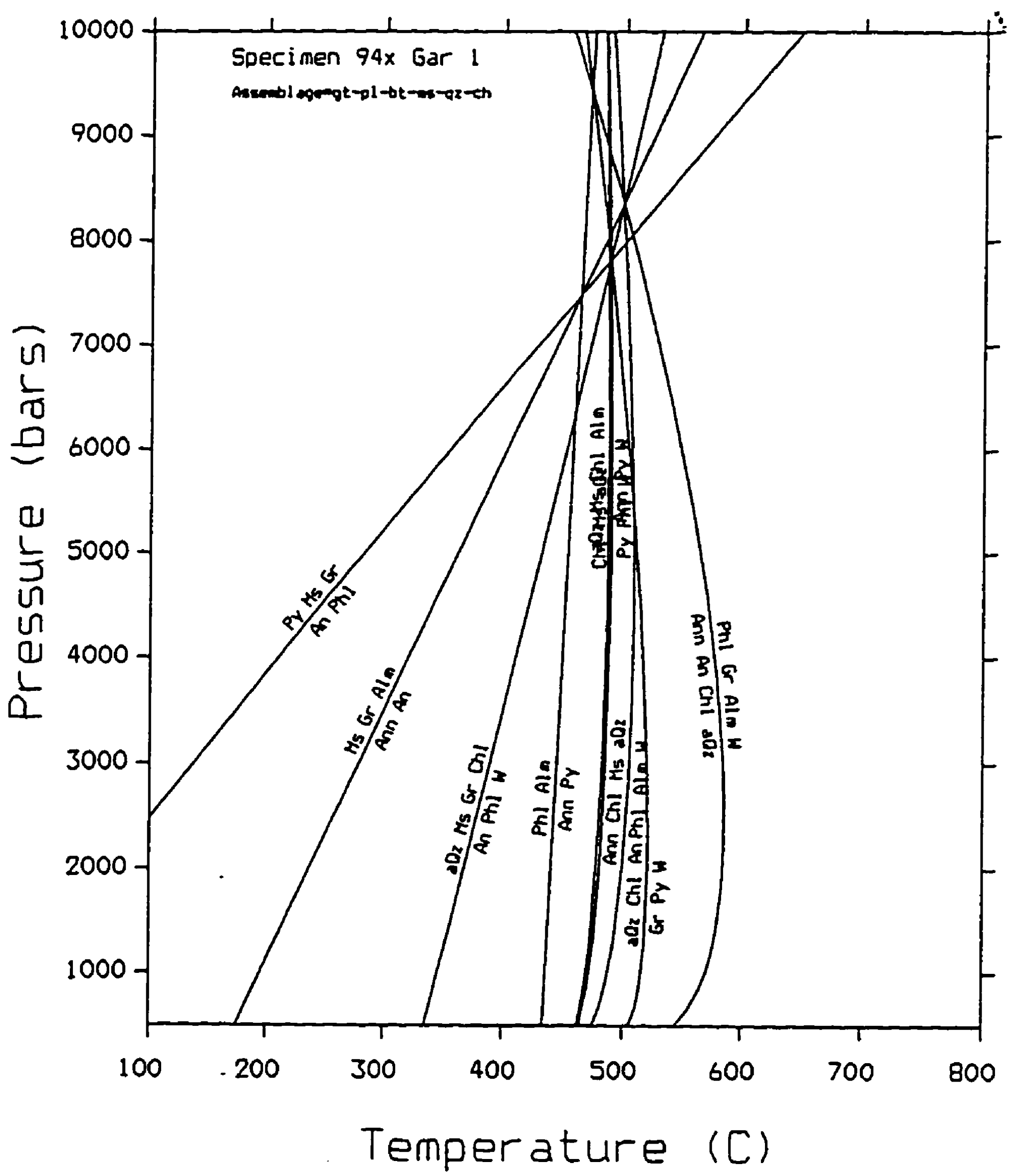
Pressure (bars)
Temperature (C)
Specimen 34-Y
Assemblage=gt-pl-bt-ch-qz-ep
cZo Phl Alm W
Ann An Chl aQz
Py cZo W
Gr Chl An
Alm Phl cZo W
Gr Chl An Ann
Py aQz cZo
Gr Chl An
Alm Phl aQz cZo
Gr Chl An Ann
aQz cZo
Gr An W
Phl Alm
Ann Py
Chl aQz cZo
Ann Chl aGy
Phl Gr Alm W
aQz Chl An
Gr Py W
Phl Gr Alm W
Ann An Chl aQz
cZo Py W
An Chl aQz
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
300
350
400
450
500
550
600
650
700
750
800



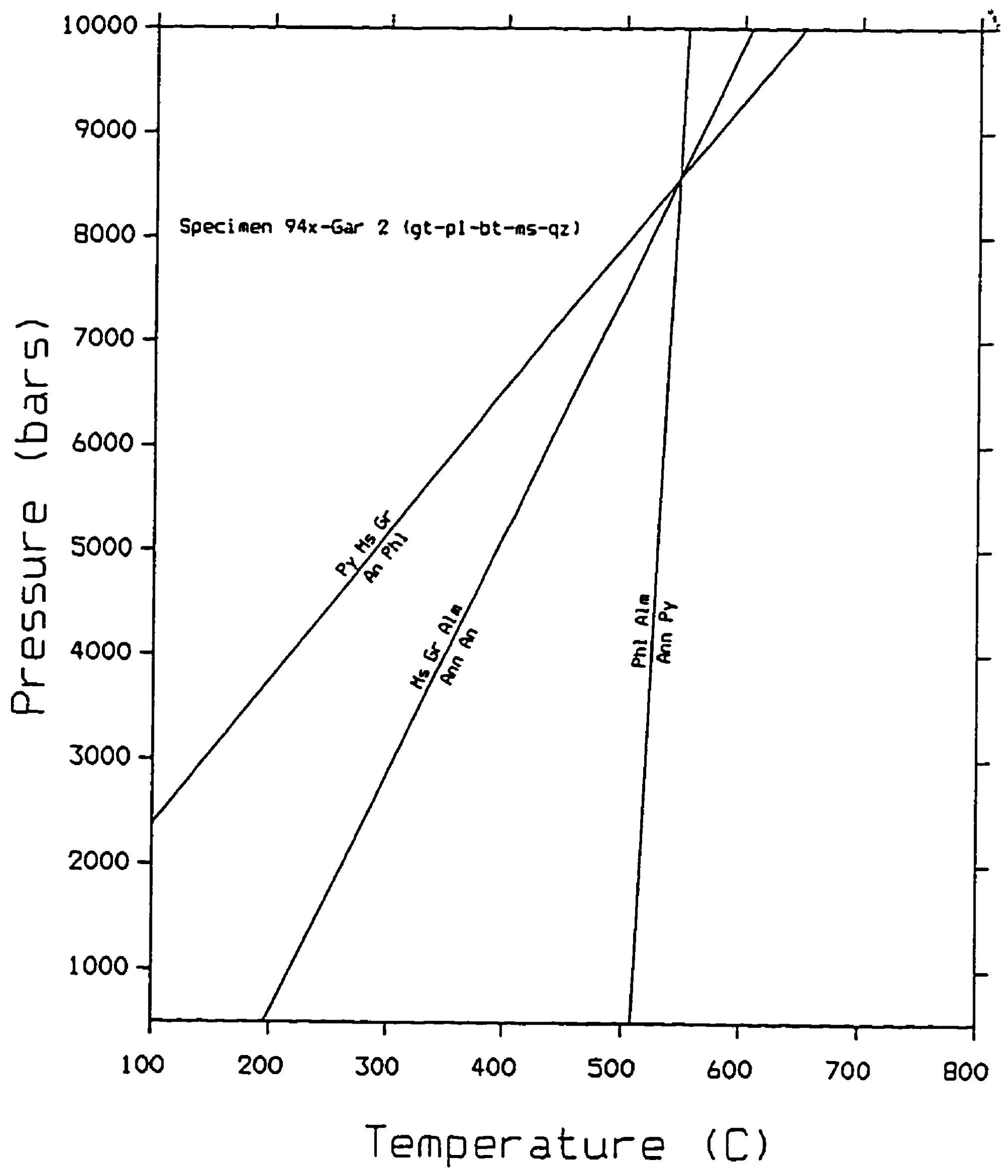
Specimen 34PEL
gt-pl-bt-ms-qz-ch (gt rim comps)
Pressure (bars)
Temperature (C)
Py Ms Gr
An Phl
Ms Gr Alm
Ann An
aQz Ms Gr Chl
An Phl W
Phl Gr Alm W
Ann An Chl aQz
Phl Alm
Ann Py
Ann Chl Ms aQz
aQz Ms Chl Alm
Phl Alm W
Ann Py W
aQz Chl An
Gr Py W
Phl W


Specimen 94x-Gar 2
Assemblage=gt-pl-bt-ms-qz-ch
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Py Ms Gr
An Phl
Ms Gr Alm
Ann An
aQz Ms Gr Chl
An Phl W
Phl Alm
Phl Alm W
aQz Chl An
Gr Py W
Ann An Chl aQz
Phl Gr Alm W

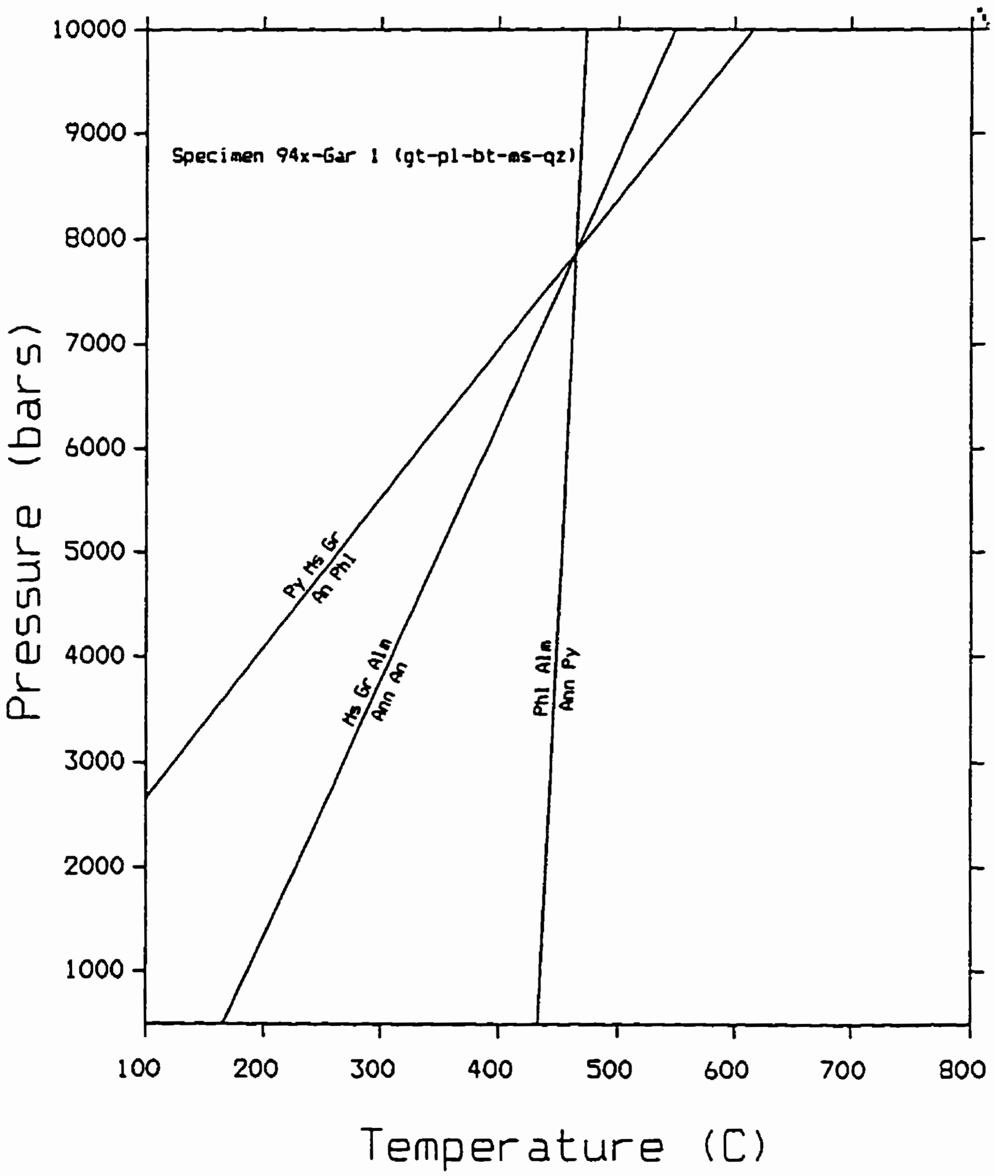

Specimen 34PEL
gt-pl-bt-ms-qz-ch (gt core comps)
Pressure (bars)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
Temperature (C)
100
200
300
400
500
600
700
800
Py Ms Gr
An Phl
Ms Gr Alm
Ann An
aQz Ms Gr Chl
An Phl W
Phl Gr Alm W

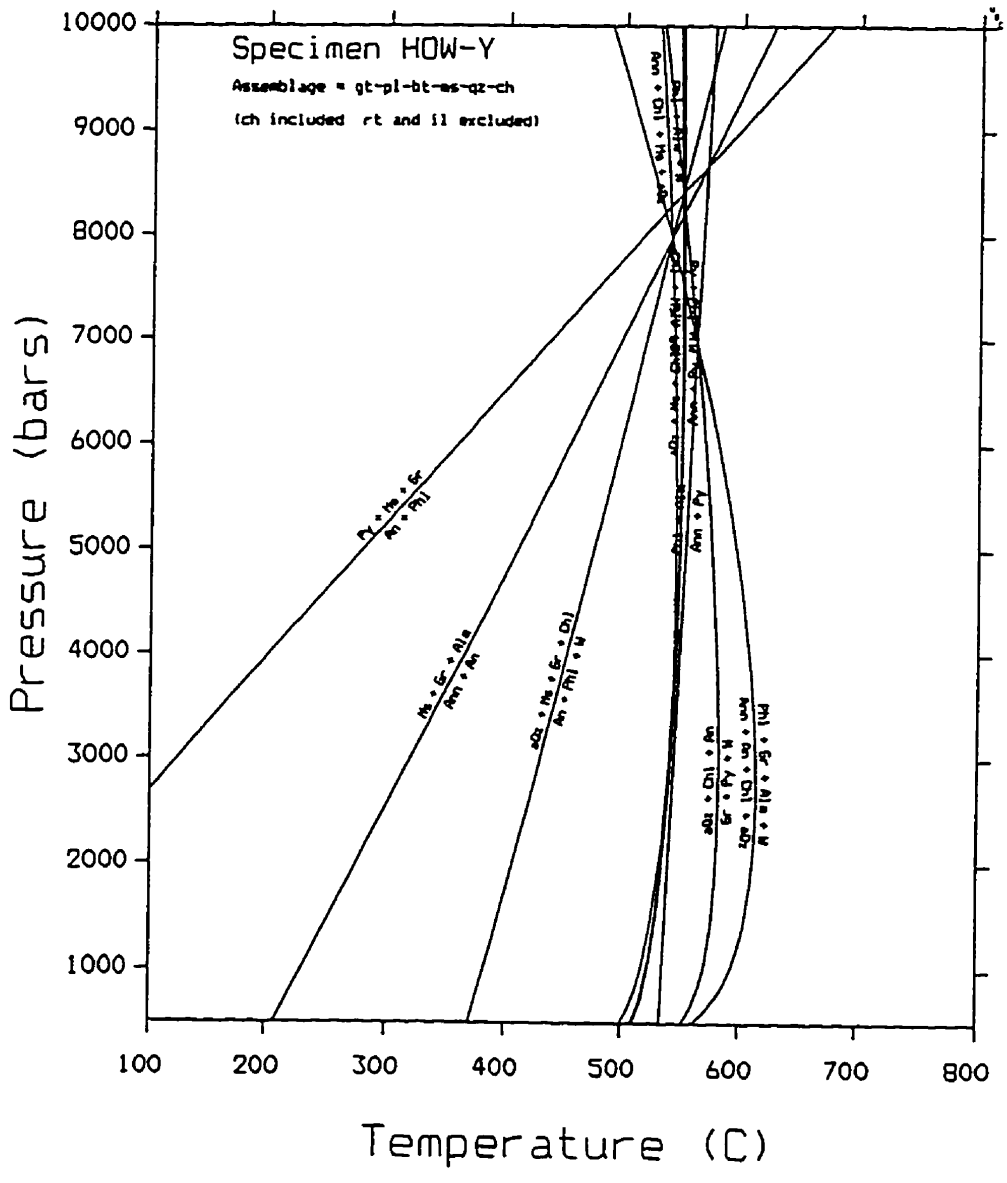


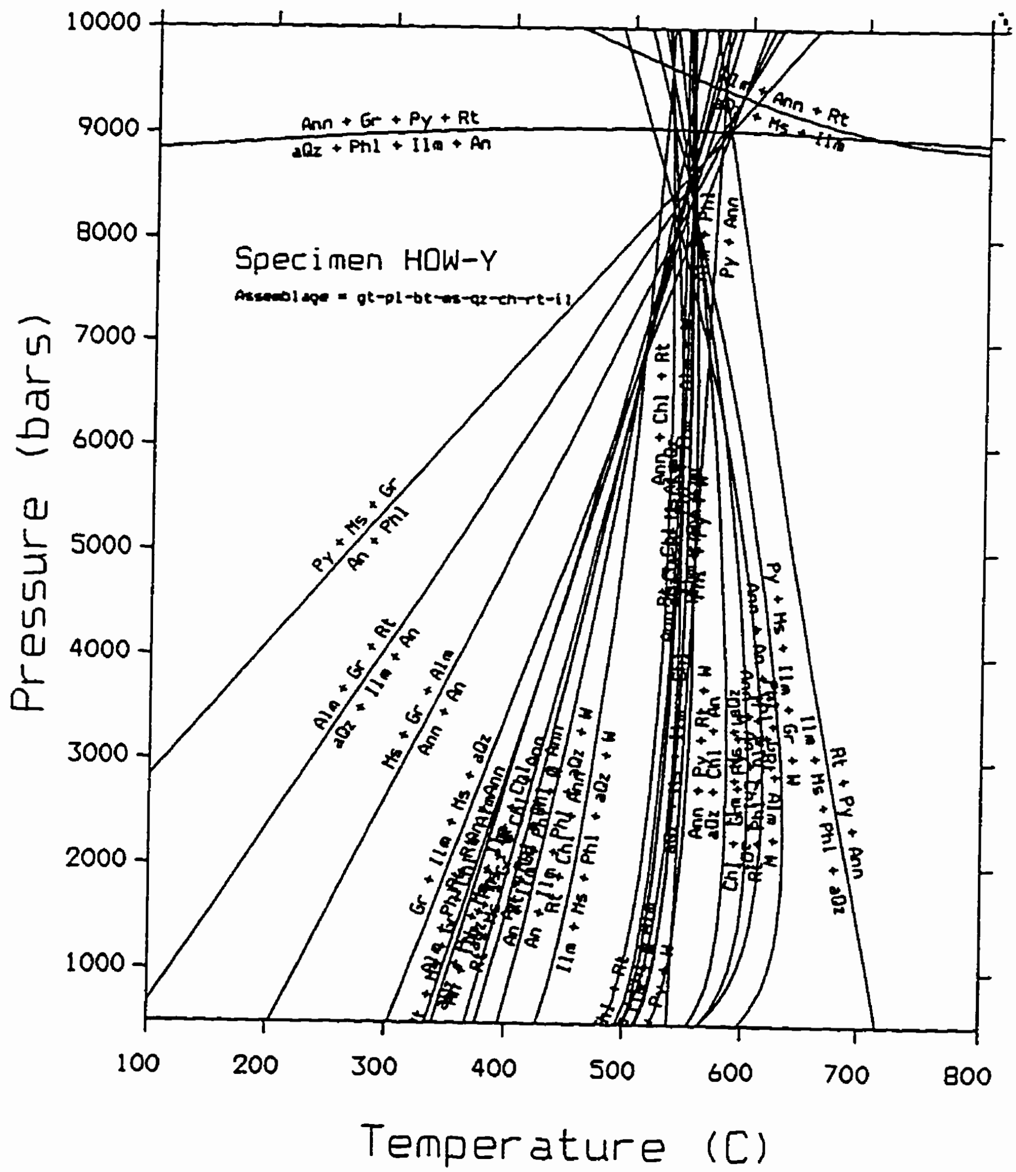
Specimen 94x Gar 1
Assemblage=gt-pl-bt-ms-qz-ch
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Py Ms Gr
An Phl
Ms Gr Alm
Ann An
aQz Ms Gr Chl
An Phl W
Phl Alm
Ann Py
Ann Chl Ms aQz
Chl Alm
Py W
Ann Chl Alm W
aQz Chl An Phl Alm W
Gr Py W
Phl Gr Alm W
Ann An Chl aQz

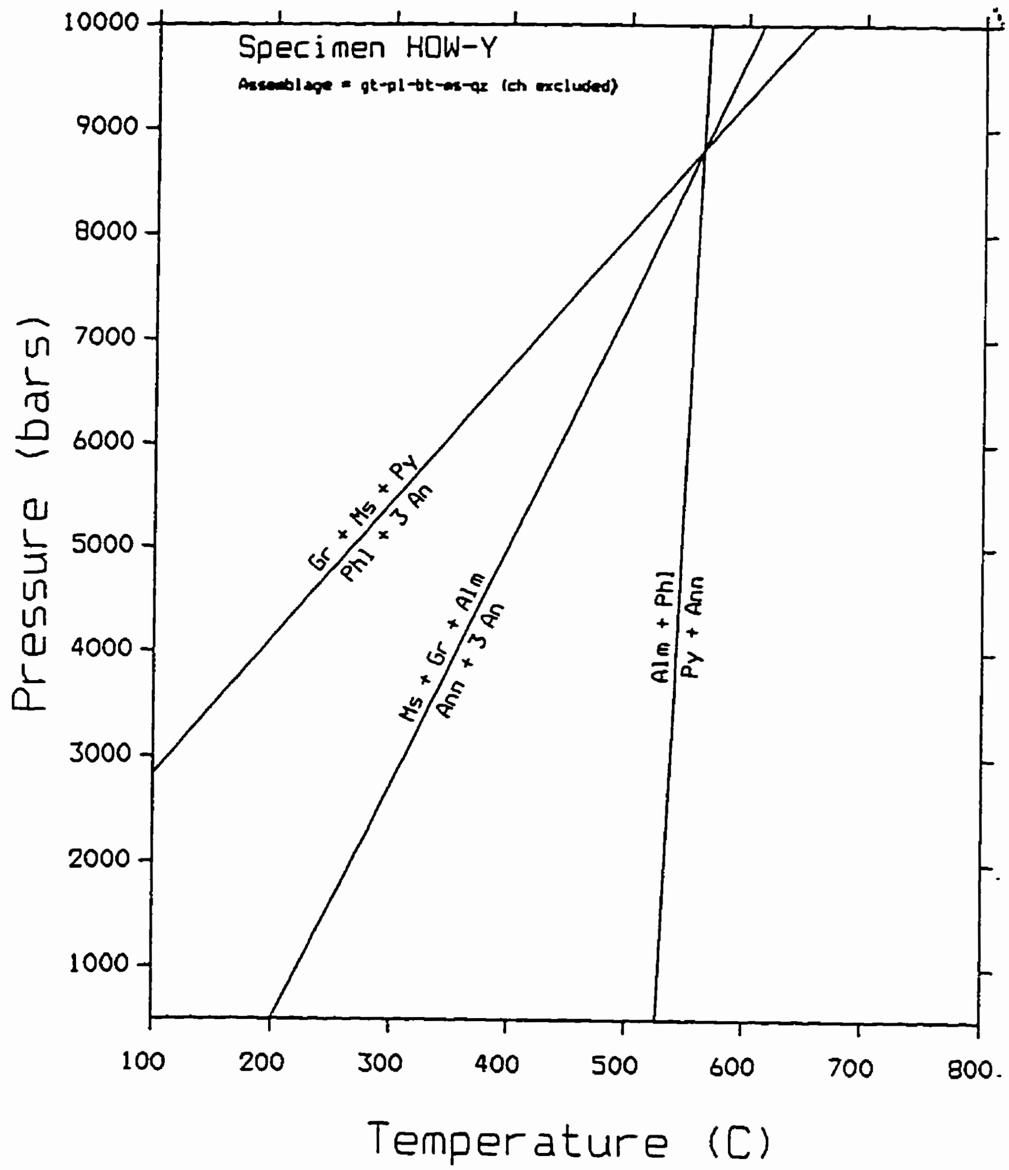


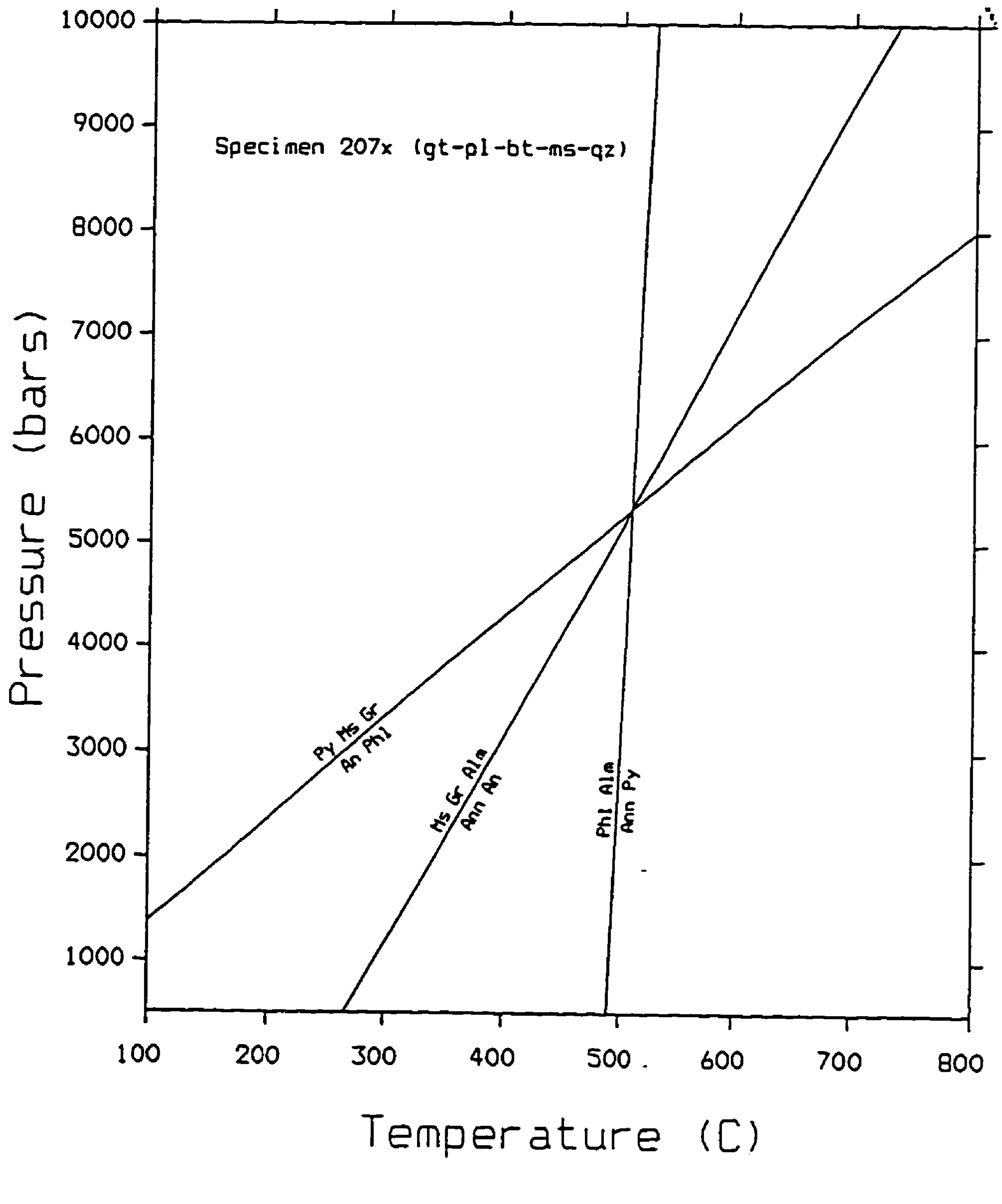
Specimen 94x-Gar 2 (gt-pl-bt-ms-qz)
Py Ms Gr
An Phl
Ms Gr Alm
Ann An
Phl Alm
Ann Py
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
Pressure (bars)
100
200
300
400
500
600
700
800
Temperature (C)

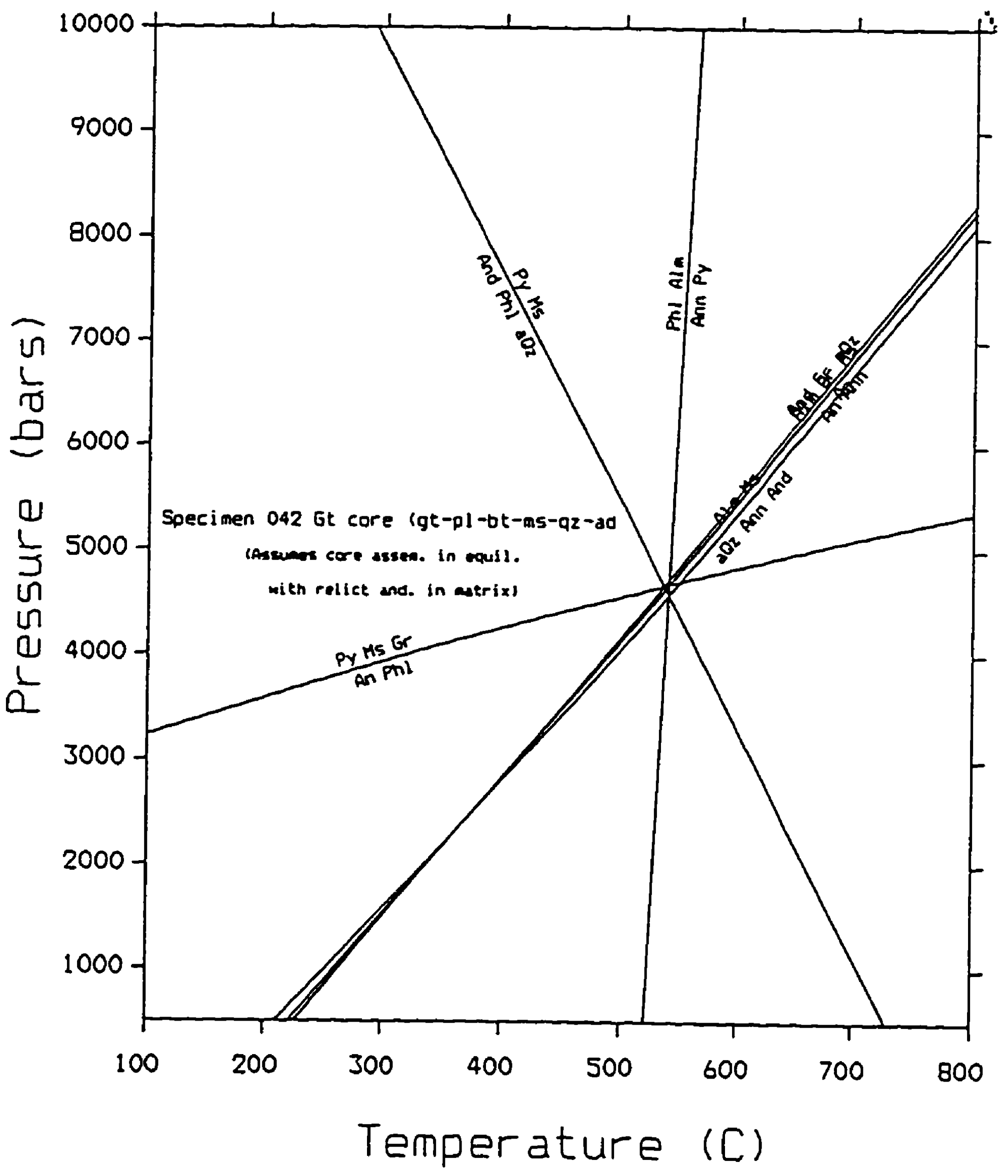

Specimen 94x-Gar 1 (gt-pl-bt-ms-qz)
Pressure (bars)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
Py Ms Gr
An Phl
Ms Gr Alm
Ann An
Phl Alm
Ann Py
100
200
300
400
500
600
700
800
Temperature (C)

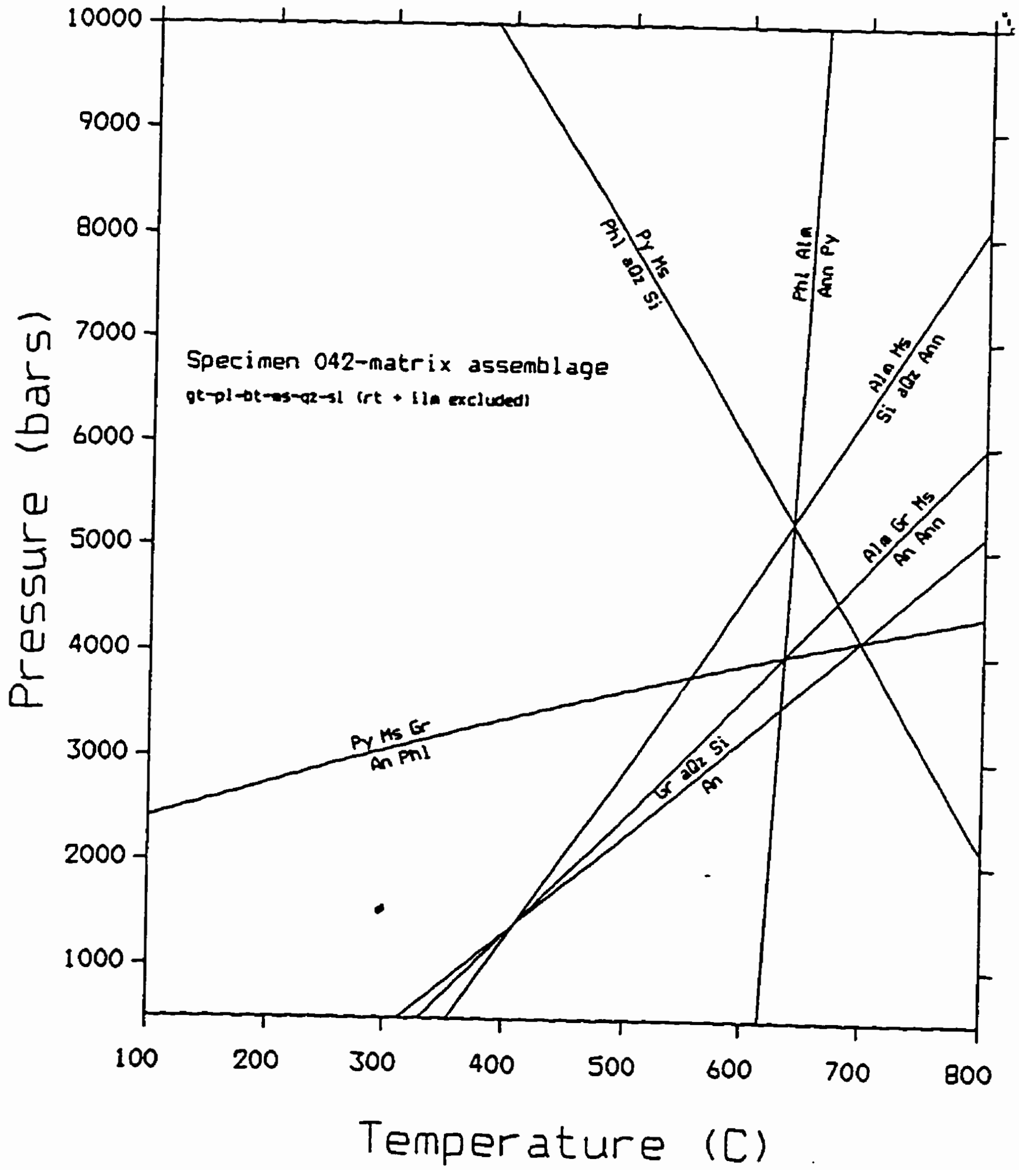

Specimen HOW-Y
Assemblage = gt-pl-bt-ms-qz-ch
(ch included rt and il excluded)
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Py + Ms + Gr
An + Phl
Ms + Gr + Alm
Ann + An
aQz + Ms + Gr + Chl
An + Phl + W
Ann + Py
Phl + Gr + Alm


Specimen HOW-Y
Assemblage = gt-pl-bt-ms-qz-ch-rt-il
Pressure (bars)
Temperature (C)
Ann + Gr + Py + Rt
aQz + Phl + Ilm + An
Py + Ms + Gr
An + Phl
Alm + Gr + Rt
aQz + Ilm + An
Ms + Gr + Alm
Ann + An
Ilm + Ann + Rt
Py + Ann
Rt + Py + Ann
Ilm + Ms + Phl + aQz
Ann + Chl + Rt
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800

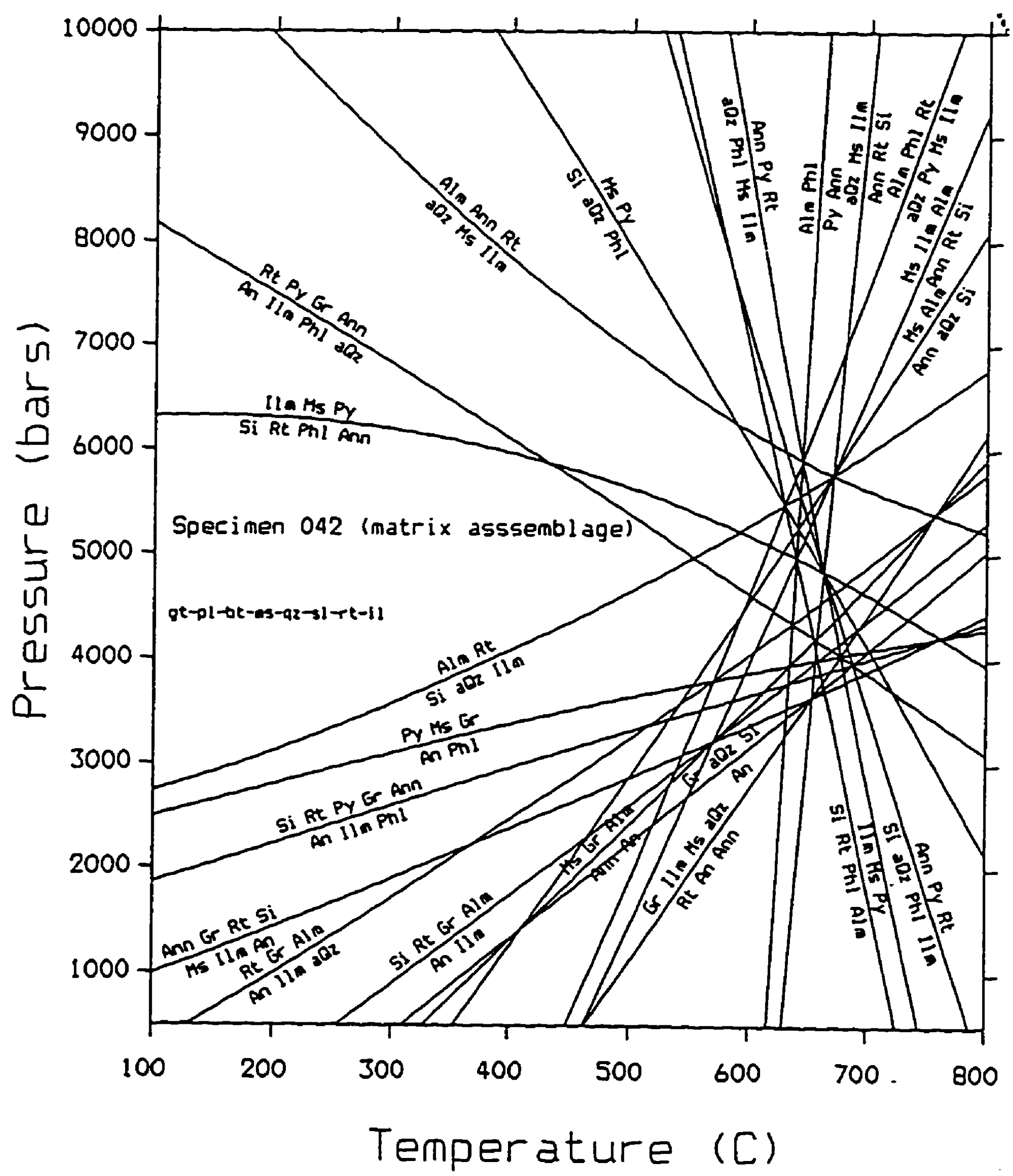

Specimen HOW-Y
Assemblage = gt-pl-bt-ms-qz (ch excluded)
Gr + Ms + Py
Phl + 3 An
Ms + Gr + Alm
Ann + 3 An
Alm + Phl
Py + Ann
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800.

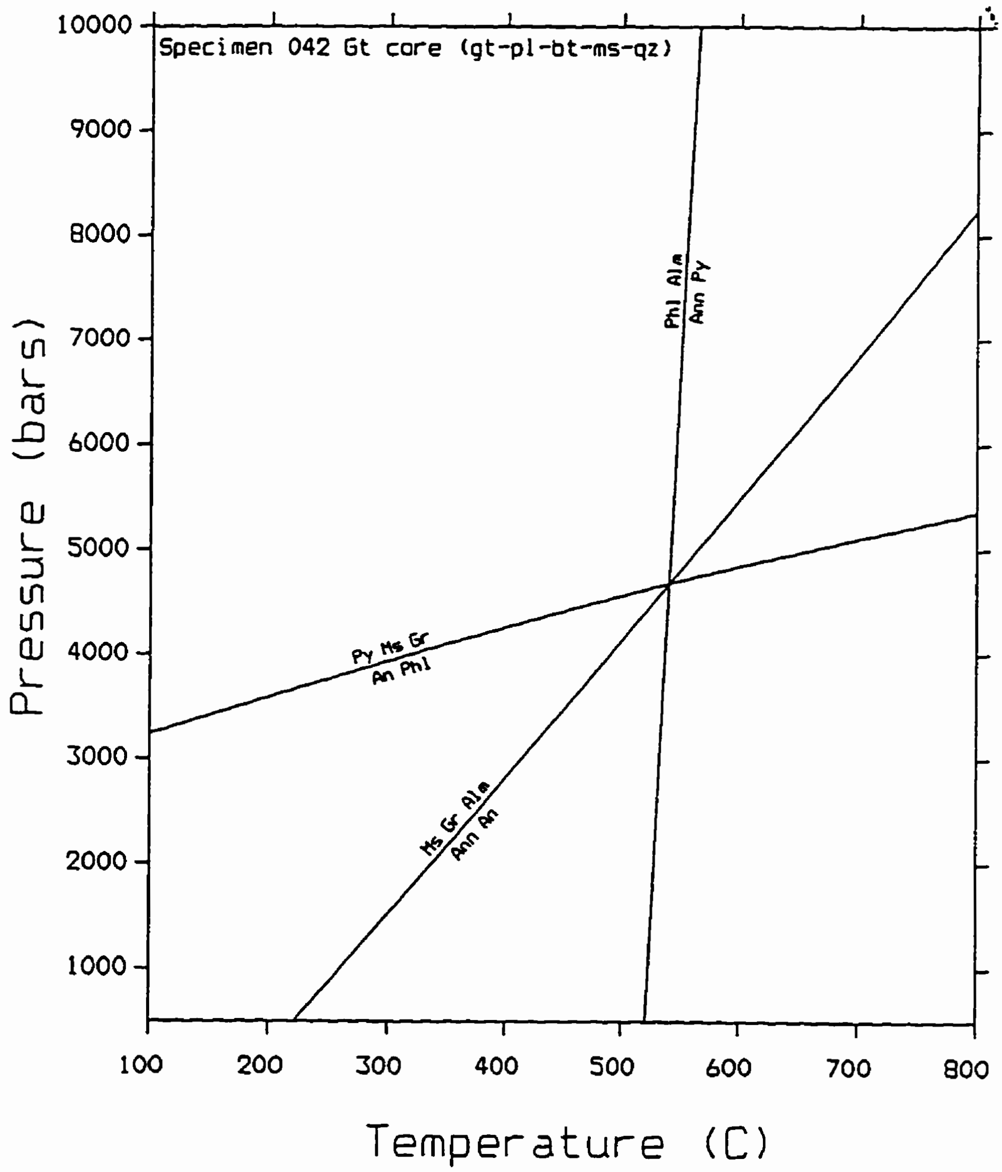

Specimen 207x (gt-pl-bt-ms-qz)
Py Ms Gr
An Phl
Ms Gr Alm
Ann An
Phl Alm
Ann Py
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800

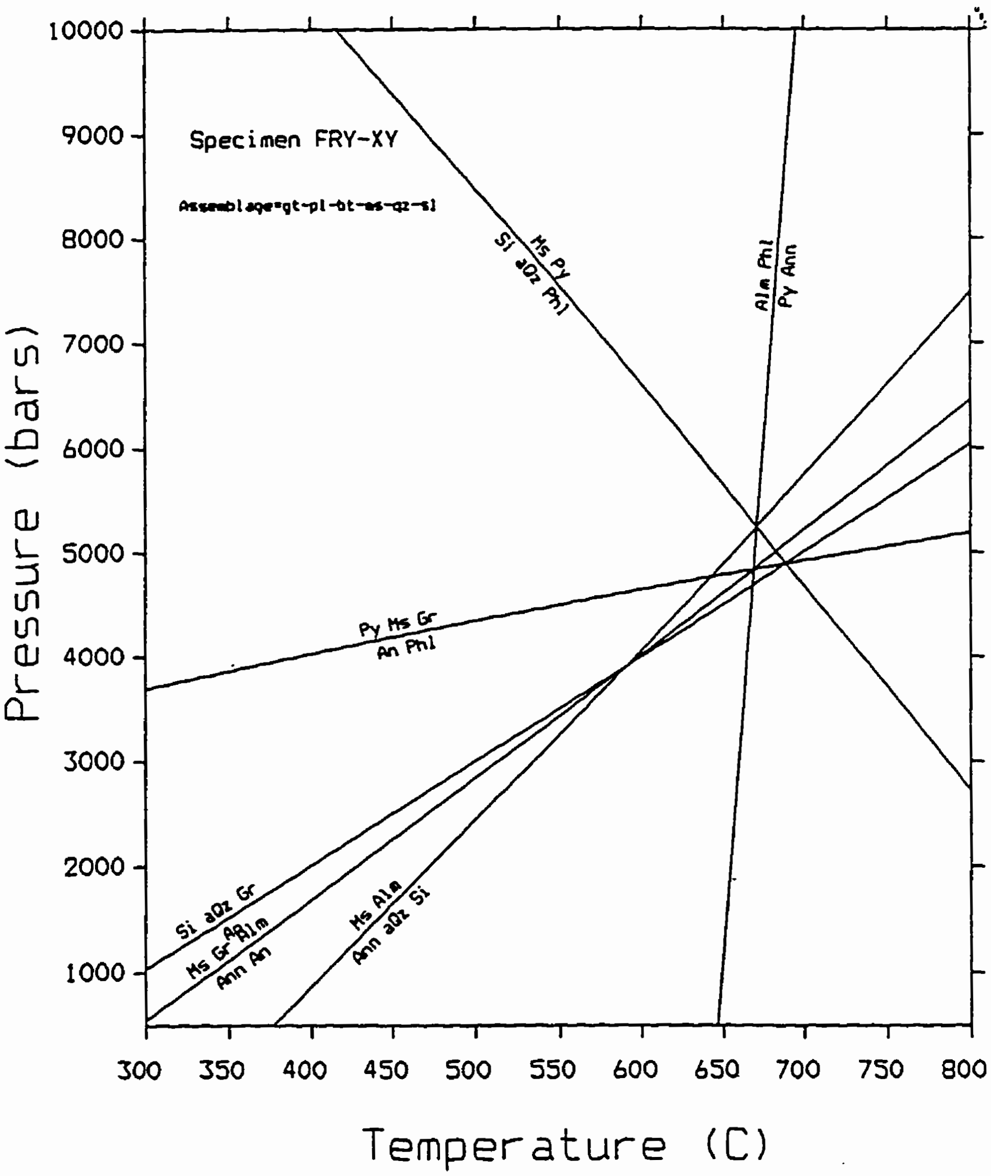

Specimen 042 Gt core (gt-pl-bt-ms-qz-ad
(Assumes core assem. in equil.
with relict and. in matrix)
Py Ms
And Phl aQz
Phl Alm
Ann Py
aQz Ann And
Py Ms Gr
An Phl
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800

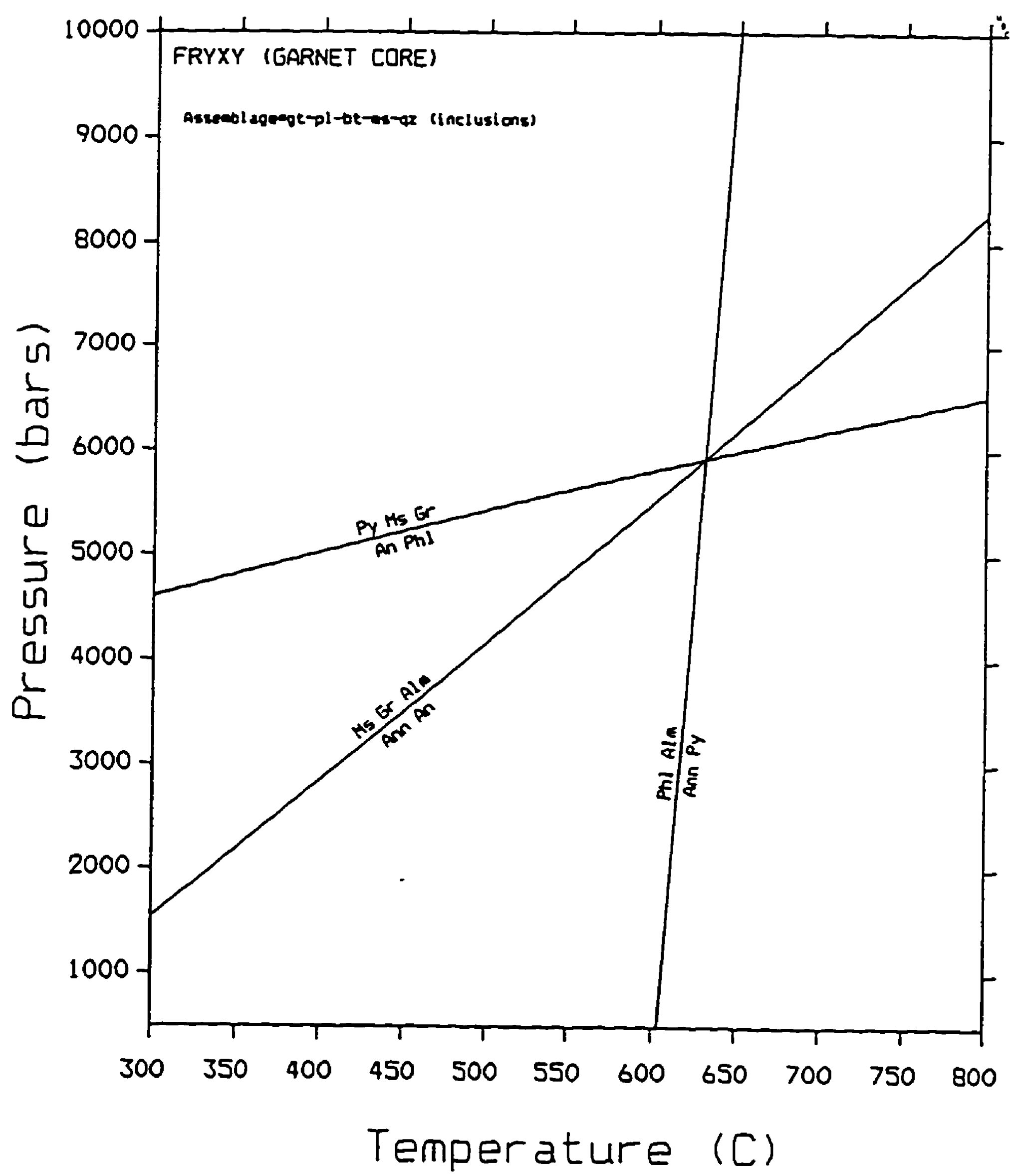


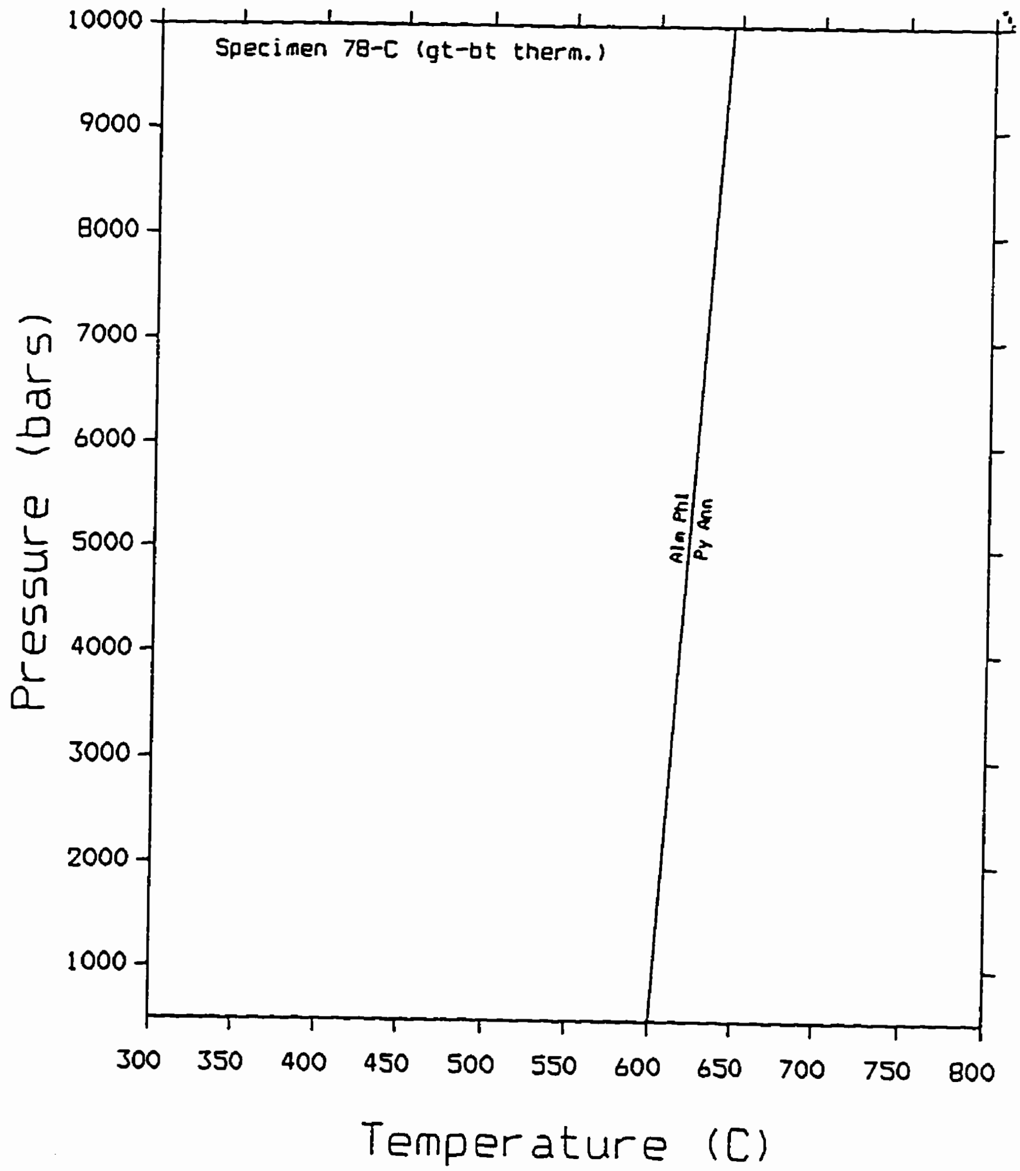
Specimen 042-matrix assemblage
gt-pl-bt-ms-qz-si (rt + ilm excluded)
Py Ms
Phl aQz Si
Phl Alm
Ann Py
Alm Ms
Si aQz Ann
Alm Gr Ms
An Ann
Py Ms Gr
An Phl
Gr aQz Si
An
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800

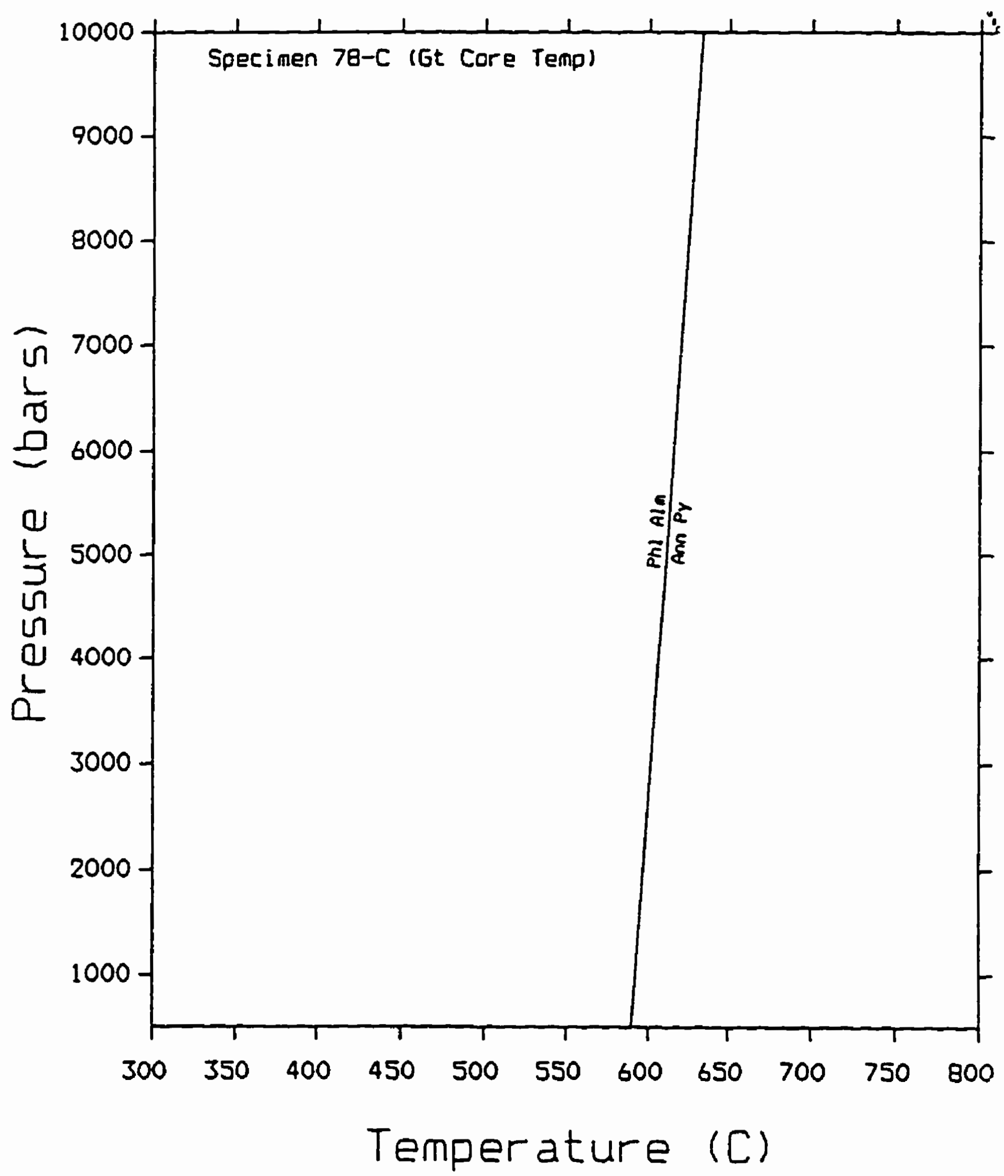

Specimen 042 (matrix asssemblage)
gt-pl-bt-ms-qz-sl-rt-il
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Alm Ann Rt
aQz Ms Ilm
Ms Py
Si aQz Phl
Rt Py Gr Ann
An Ilm Phl aQz
Ilm Ms Py
Si Rt Phl Ann
Alm Rt
Si aQz Ilm
Py Ms Gr
An Phl
Si Rt Py Gr Ann
An Ilm Phl
Ann Gr Rt Si
Ms Ilm An
Rt Gr Alm
An Ilm aQz
Si Rt Gr Alm
An Ilm
Ms Gr Alm
Ann An
Gr aQz Si
An
Gr Ilm Ms aQz
Rt An Ann
Ann Py Rt
aQz Phl Ms Ilm
Alm Phl
Py Ann
aQz Ms Ilm
Ann Rt Si
Alm Phl Rt
aQz Py Ms Ilm
Ms Ilm Alm
Ann Rt Si
Ms Alm
Ann aQz Si
Si Rt Phl Alm
Ilm Ms Py
Si aQz Phl Ilm
Ann Py Rt


Specimen 042 Gt core (gt-pl-bt-ms-qz)
Phl Alm
Ann Py
Py Ms Gr
An Phl
Ms Gr Alm
Ann An
Pressure (bars)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
100
200
300
400
500
600
700
800
Temperature (C)


Specimen FRY-XY
Assemblage=gt-pl-bt-ms-qz-sl
Ms Py
Si aQz Phl
Alm Phl
Py Ann
Py Ms Gr
An Phl
Si aQz Gr
An
Ms Gr Alm
Ann An
Ms Alm
Ann aQz Si
Pressure (bars)
Temperature (C)
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
300
350
400
450
500
550
600
650
700
750
800


FRYXY (GARNET CORE)
Assemblage=gt-pl-bt-ms-qz (inclusions)
Py Ms Gr
An Phl
Ms Gr Alm
Ann An
Phl Alm
Ann Py
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
Pressure (bars)
300
350
400
450
500
550
600
650
700
750
800
Temperature (C)


Specimen 78-C (gt-bt therm.)
Pressure (bars)
Temperature (C)
Alm Phl
Py Ann
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
300
350
400
450
500
550
600
650
700
750
800



Specimen 78-C (Gt Core Temp)
Phl Alm
Ann Py
10000
9000
8000
7000
6000
5000
4000
3000
2000
1000
Pressure (bars)
300
350
400
450
500
550
600
650
700
750
800
Temperature (C)

# Appendix 3

# GEOCHEMICAL DATA FROM METABASITES

## ANALYTICAL METHODS

Fourteen samples of greenstones or amphibolites from the Horsethief Creek Group, Hamill Group and Lardeau Group (and one sample of pyroxenite/hornblendite from the margin of the Toby stock) were selected for major- and trace-element geochemistry. The samples were crushed at Queen's University. Further sample preparation and analysis was performed at the Analytical Sciences Laboratory of the British Columbia Ministry of Energy, Mines and Petroleum Resources, Victoria, B. C. The crushed samples were pulverized in a tungsten-carbide mill, and geochemical analyses were obtained by X-ray fluorescence (major oxides and Cr, Ba, Sr, Rb, Zr, Y, Nb, Sn, Th, Ta, U, V) or by neutron activation (Ce, Cs, La, Sc). The following summary of analytical methods comprises excerpts from the Schedule of Services, Analytical Sciences Laboratory, British Columbia Ministry of Energy, Mines and Petroleum Resources, Information Circular 1991-10:

Major oxides ($SiO_2$, $Al_2O_3$, $Fe_2O_3$, MgO, CaO, $Na_2O$, $K_2O$, $TiO_2$, $P_2O_5$) are determined by X-ray fluorescence (XRF on following data tables) with a Philips PW1404 automated spectrometer. A 2 gram aliquot of the sample pulp is mixed with 10 grams of a flux comprising 80% lithium tetraborate and 20% lithium metaborate and fused in a platinum crucible to prepare a glass disc. A primary X-ray beam, directed onto the disc, excites elements to emit secondary X-ray fluorescence. Discrimination between elements based on their characterisitic frequency is performed by X-ray diffraction. The intensity of the fluorescence is measured by gas-filled proportional or scintillation counters.

Calibration of the X-ray fluorescence spectrometer is performed using rock and mineral standards. Beacause of the different rock types commonly analyzed, it is essential to use as wide a range of different standards as possible for calibration. Matrix variations such as particle size, variable background, line overlap and absorption edge interference can influence the quality of the final data. Mathematic corrections are therefore made to compensate for these variables.

"Total" trace element contents are determined by X-ray fluorescence using a fused disc prepared by the same method as for the major oxides. Different software is used to apply corrections necessary to calculate trace metal concentrations. Some of the trace metals are determined with a chromium target while others are measured with a gold target.

For elements determined by neutron activation (INAA on following data tables), a 30 gram sample is encapsulated in a PVC vial and irradiated in a nuclear reactor. Element concentrations are measured by counting the short-lived decay products.

Detection limits for both methods (also reproduced from B. C. Ministry of Energy, Mines and Petroleum Resources, Information Circular 1991-10) are given on page 331.

## RESULTS

The data were plotted as mid-ocean ridge basalt-normalized trace element patterns after Pearce (1982) and on several discrimination diagrams according to Pearce and Cann (1973) and Winchester and Floyd (1977). They suggest that, despite differences in metamorphic grade and strain level between samples within each stratigraphic group, the metabasites of the Horsethief Creek Group, Hamill Group and Lardeau Group have distinct trace-element geochemical signatures (possible significant mobility of elements during regional metamorphism or shearing is not discussed here). These data, if considered valid, have two implications that are relevant to discussion in the body of this thesis:

1) metabasites in the upper clastic sequence of the Horsethief Creek Group (Chapter 2), whose contact relationships with metasedimentary strata are uncertain, were not feeder dikes

related to mafic flows in the Hamill and Lardeau Groups. These data and rare mafic clasts in the grits of the upper Horsethief Creek Group support the interpretation that there was volcanism during the deposition of the upper clastic sequence.

2) the metabasites from each of the stratigraphic sequences plot primarily as within-plate basalts on the Ti/100 - Zr - Yx3 discrimination diagram of Pearce and Cann (1973) and the Ti/Y - Nb/Y discrimination diagram of Pearce (1982), implying that each of these sequences, including the Lardeau Group, was deposited on attenuated North American continental crust. These data thus support stratigraphic and structural arguments presented in Colpron and Price (1995) and in Chapters 3 and 4 of this thesis that "Kootenay terrane" is not far-travelled with respect to North American supracrustal rocks exposed in the Purcell anticlinorium.

These data are consistent with other trace-element geochemical data from equivalent rocks throughout the southern Canadian Cordillera and in adjacent Washington state. Trace-element geochemical data from mafic volcanic rocks at the base of the Windermere Supergroup are discussed by Bennett (1985), Devlin (1986), Sevigny (1988) and Pope (1989). Data from a dike that cuts the Windermere Supergroup but is truncated beneath the Gog Group in the Rocky Mountains are discussed by Devlin (1986). Smith (1990, and unpublished data), Smith and Gehrels (1992) and J. Logan (B. C. Geological Survey Branch, unpublished data) have obtained trace-element geochemical analyses from greenstones in the Lardeau Group and the Hamill Group in the Kootenay Arc (N. E. Washington and northern Selkirk Mountains).

## DETECTION LEVELS FOR TRACE ELEMENTS DETERMINED BY XRF

| GOLD TARGET | | CHROMIUM TARGET | |
|---|---|---|---|
| ELEMENT | DETECTION LIMIT(PPM) | ELEMENT | DETECTION LIMIT(PPM) |
| Cr | 10 | Ba | 10 |
| Nb | 5 | Ce | 15 |
| Rb | 10 | Cs | 5 |
| Sr | 5 | La | 15 |
| Sn | NA | Sc | 3 |
| Th | NA | Ti | 10 |
| Ta | NA | V | 5 |
| U | NA | | |
| W | NA | | |
| Y | 10 | | |
| Zr | 20 | | |

## DETECTION LEVELS FOR ELEMENTS BY NEUTRON ACTIVATION (30 g )

| ELEMENT | DETECTION LIMIT(PPM) | ELEMENT | DETECTION LIMIT(PPM) |
|---|---|---|---|
| Ag | 5 | As | 2 |
| Au | 5 ppb | Ba | 100 |
| Br | 1 | Ca | 1% |
| Ce | 5 | Co | 5 |
| Cr | 10 | Cs | 2 |
| Eu | 0.2% | Fe | 0.02% |
| Hf | 1 | Hg | 1 |
| Ir | 1 | La | 1 |
| Lu | 0.05 | Mo | 5 |
| Na | 0.05% | Nd | 5 |
| Ni | 50 | Rb | 30 |
| Sb | 0.2 | SC | 0.1 |
| Se | 5 | Sr | 0.05% |
| Sm | 0.1 | Sn | 0.01% |
| Ta | 1 | U | 0.5 |
| Th | 0.5 | W | 4 |
| Zn | 50 | Yb | 0.2 |

Geochem samples sent 19/12/92

Page 1 of 1

# ANALYTICAL SERVICES REQUEST

Analytical Sciences Laboratory
541 Superior Street, Victoria B.C., V8V 1X4
(604) 356-2134

Batch Number ______

Number of Samples 15
Responsibility Number ______

Geologist/Project T. Höy/M. Warren
Account Number ______

Date Samples Received ______
Date of Final Report ______

| LAB NUMBER | FIELD NUMBER | NTS MAP SHEET | UTM ZN. | UTM EASTING | UTM NORTHING | MINFILE CODE STRAT | MINFILE CODE ROCK | SAMPLE TYPE | ANALYTICAL PACKAGE | COMMENTS |
|---|---|---|---|---|---|---|---|---|---|---|
| | 031 | 82k/7 | 11 | 504637 | 5582529 | | | | ASL 4, ASL 5 | LARDEAU GP (INDEX FM) MAFIC DYKE |
| | 031-D | 82k/7 | 11 | 504637 | 5582529 | | | | ASL 4, ASL 5 | LARDEAU GP. (INDEX FM) MAFIC DYKE |
| | 035-2 | 82k/7 | 11 | 505831 | 5584323 | | | | ASL 4, ASL 5 | LOWER HAMILL GP - METABSLT |
| | 060 | 82k/7 | 11 | 504896 | 5585615 | | | | ASL 4, ASL 5 | LOWER HAMILL GP - METABSLT |
| | 061 | 82k/7 | 11 | 505392 | 5586008 | | | | ASL 4, ASL 5 | LOWER HAMILL GP. METABSLT |
| | 073 | 82k/7 | 11 | 507729 | 5590535 | | | | ASL 4, ASL 5 | UPPER HORSETHIEF CK GP META-BSLT |
| | 078-C | 82k/7 | 11 | 508810 | 5592816 | | | | ASL 4, ASL 5 | UPPER HORSETHIEF CK GP META-BSLT |
| | 171 | 82k/7 | 11 | 513276 | 5568752 | | | | ASL 4, ASL 5 | LOWER HAMILL GP - METABSLT |
| | 216 | 82k/2 | 11 | 504936 | 5563224 | | | | ASL 4, ASL 5 | LARDEAU GP (INDEX FM) METABSLT |
| | 254 | 82k/2 | 11 | 492713 | 5557368 | | | | ASL 4, ASL 5 | LARDEAU GP (INDEX FM) METABSLT |
| | 318-D | 82k/11 | 11 | 498608 | 5605137 | | | | ASL 4, ASL 5 | UPPER HORSETHIEF CR GP META-BSLT |
| | 331-A | 82k/10 | 11 | 504118 | 5596326 | | | | ASL 4, ASL 5 | HORSETHIEF CR. GP META-BSLT |
| | 524-T | 82k/2 | 11 | 532821 | 5566055 | | | | ASL 4, ASL 5 | JURASSIC? PYROXENITE/HBLNDITE (ASSOC W/ TOBY STOCK) |
| | H-R1A | 82k/7 | 11 | 502368 | 5572173 | | | | ASL 4, ASL 5 | LARDEAU GP (INDEX FM?) METABSLT |
| | MM5 | 82k/3 | 11 | 498559 | 5563068 | | | | ASL 4, ASL 5 | LARDEAU GP (JOWETT FM) SHEARED METABSLT |
| | | | | | | | | | ASL 4, ASL 5 | |
| | | | | | | | | | ASL 4, ASL 5 | |
| | | | | | | | | | | |
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MINISTRY OF ENERGY, MINES AND PETROLEUM RESOURCES
GEOLOGICAL SURVEY BRANCH LABORATORY

ANALYTICAL SERVICES REPORT

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Batch: 93/010
Sample Prep: RSL2

Submitting Geologist: T. HOY

Number of Samples: 16
Number of Requests: 27

Date Submitted: 01/20/93
Date Reported: 03/22/93

Date Completed: 03/22/93

Page 1
03/22/93

| Lab Number | Field Number | $SiO_2$ | $TiO_2$ | $Al_2O_3$ | $Fe_2O_3$ | MnO | MgO | CaO | $Na_2O$ | $K_2O$ | $P_2O_5$ | LOI | SUM |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 047155 | 031 | 46.60% | 3.40% | 13.64% | 17.45% | 0.20% | 4.71% | 7.37% | 2.31% | 0.35% | 0.36% | 3.09% | 99.48% |
| 047156 | 031-D | 47.30% | 1.51% | 15.25% | 9.91% | 0.14% | 7.25% | 9.68% | 3.56% | 0.42% | 0.13% | 3.47% | 99.30% |
| 047157 | 035-2 | 44.35% | 1.76% | 13.15% | 14.80% | 0.17% | 8.47% | 9.81% | 2.65% | 0.25% | 0.46% | 1.44% | 99.25% |
| 047158 | 060 | 47.07% | 2.13% | 7.83% | 11.62% | 0.20% | 15.57% | 10.82% | 1.34% | 0.17% | 0.23% | 2.57% | 99.55% |
| 047159 | 061 | 46.56% | 2.52% | 14.27% | 13.14% | 0.19% | 6.52% | 12.75% | 1.07% | 0.11% | 0.28% | 2.40% | 99.81% |
| 047160 | 073 | 43.19% | 1.95% | 15.32% | 17.75% | 0.25% | 5.49% | 9.68% | 1.50% | 1.11% | 0.28% | 2.87% | 99.29% |
| 047161 | STANDARD 3-4 | 65.71% | 0.36% | 8.76% | 5.92% | 0.09% | 2.18% | 3.20% | 0.21% | 1.90% | 0.15% | 7.75% | 96.23% |
| 047162 | 078-C | 52.14% | 2.13% | 12.09% | 18.51% | 0.25% | 3.85% | 6.76% | 2.61% | 0.73% | 0.21% | 0.32% | 99.63% |
| 047163 | 171 | 43.66% | 2.26% | 9.95% | 14.16% | 0.25% | 12.21% | 11.22% | 1.82% | 0.32% | 0.26% | 3.21% | 99.32% |
| 047164 | 216 | 39.95% | 2.81% | 14.45% | 14.63% | 0.18% | 7.94% | 8.71% | 2.61% | 0.55% | 0.34% | 6.20% | 98.37% |
| 047165 | 254 | 44.40% | 1.29% | 10.34% | 9.93% | 0.14% | 14.55% | 11.07% | 1.61% | 0.13% | 0.25% | 5.10% | 98.81% |
| 047166 | 318-D | 47.76% | 2.29% | 13.49% | 15.27% | 0.22% | 6.05% | 10.25% | 2.31% | 0.27% | 0.24% | 1.35% | 99.50% |
| 047167 | 331-A | 48.53% | 2.72% | 12.63% | 16.34% | 0.22% | 5.92% | 9.57% | 2.11% | 0.57% | 0.27% | 0.68% | 99.56% |
| 047168 | 524-T | 44.73% | 1.47% | 2.61% | 18.83% | 0.24% | 12.86% | 18.10% | 0.31% | 0.11% | 0.06% | -0.06% | 99.25% |
| 047169 | E-811 | 46.26% | 2.69% | 13.89% | 13.96% | 0.16% | 6.43% | 8.66% | 3.06% | 0.41% | 0.24% | 3.31% | 99.07% |
| 047170 | KKS | 46.38% | 1.81% | 11.67% | 12.35% | 0.16% | 10.98% | 10.64% | 1.71% | 0.31% | 0.20% | 3.47% | 99.68% |

| Lab Number | Field Number | Cr Result | Cr +/- | Cr Method | Cr Instr | Ba Result | Ba +/- Method | Ba Instr | Sr Result | Sr +/- | Sr Method | Sr Instr | Rb Result | Rb +/- | Rb Method | Rb Instr |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 047155 | 031 | 22m | 10 | XRF | YY | 356m | XRF | COK | 295m | 4 | XRF | YY | <10m | 3 | XRF | YY |
| 047156 | 031-B | 175m | 15 | XRF | YY | 1848m | XRF | COK | 553m | 9 | XRF | YY | <10m | 3 | XRF | YY |
| 047157 | 035-2 | 438m | 20 | XRF | YY | 97m | XRF | COK | 399m | 5 | XRF | YY | <10m | 2 | XRF | YY |
| 047158 | 060 | 914m | 45 | XRF | YY | 36m | XRF | COK | 125m | 3 | XRF | YY | <10m | 3 | XRF | YY |
| 047159 | 061 | 62m | 10 | XRF | YY | 35m | XRF | COK | 885m | 11 | XRF | YY | <10m | 3 | XRF | YY |
| 047160 | 073 | 59m | 10 | XRF | YY | 105m | XRF | COK | 250m | 4 | XRF | YY | 38m | 3 | XRF | YY |
| 047161 | STANDARD B-4 | 139m | 15 | XRF | YY | 10725m | XRF | COK | 158m | 3 | XRF | YY | 21m | 3 | XRF | YY |
| 047162 | 078-C | 29m | 10 | XRF | YY | 167m | XRF | COK | 120m | 3 | XRF | YY | 22m | 3 | XRF | YY |
| 047163 | 171 | 1270m | 50 | XRF | YY | 46m | XRF | COK | 260m | 4 | XRF | YY | <10m | 3 | XRF | YY |
| 047164 | 216 | 189m | 15 | XRF | YY | 408m | XRF | COK | 337m | 5 | XRF | YY | <10m | 3 | XRF | YY |
| 047165 | 254 | 949m | 45 | XRF | YY | 28m | XRF | COK | 175m | 3 | XRF | YY | <10m | 3 | XRF | YY |
| 047166 | 318-B | 197m | 15 | XRF | YY | 41m | XRF | COK | 155m | 3 | XRF | YY | <10m | 3 | XRF | YY |
| 047167 | 331-A | 129m | 15 | XRF | YY | 89m | XRF | COK | 175m | 3 | XRF | YY | <10m | 4 | XRF | YY |
| 047168 | 524-T | 1014m | 50 | XRF | YY | 25m | XRF | COK | 54m | 2 | XRF | YY | <10m | 4 | XRF | YY |
| 047169 | B-RIA | 104m | 15 | XRF | YY | 195m | XRF | COK | 264m | 4 | XRF | YY | <10m | 3 | XRF | YY |
| 047170 | RR5 | 782m | 35 | XRF | YY | 274m | XRF | COK | 414m | 6 | XRF | YY | <10m | 3 | XRF | YY |

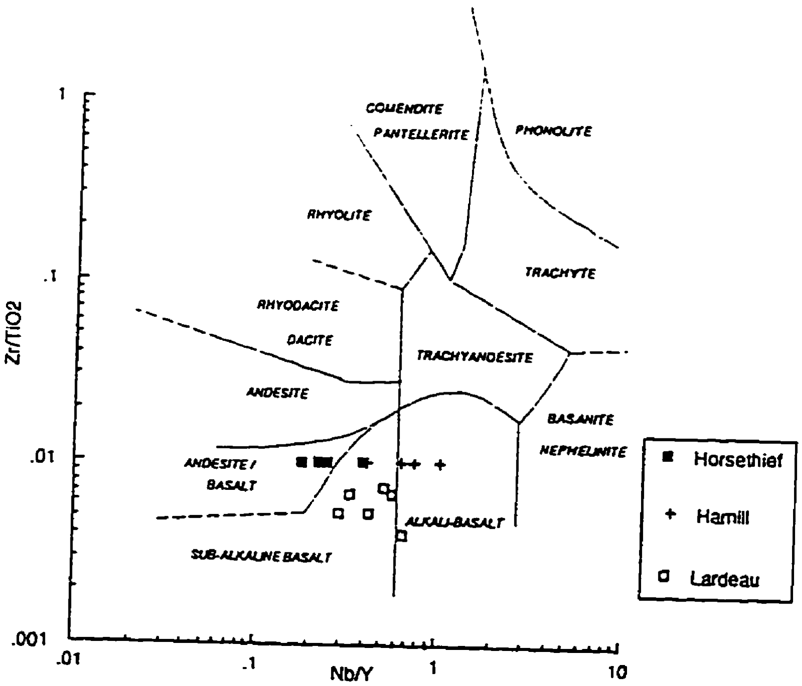
| Lab Number | Field Number | Zr Result | Zr +/- | Zr Method | Zr Anal | Y Result | Y +/- | Y Method | Y Anal | Rb Result | Rb +/- | Rb Method | Rb Anal | Sr Result | Sr +/- | Sr Method | Sr Anal |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 047155 | 031 | 139m | 3 | XRF | YY | 29m | 2 | XRF | YY | 33m | 3 | XRF | YY | 18m | 5 | XRF | YY |
| 047156 | 031-D | 55m | 2 | XRF | YY | 22m | 2 | XRF | YY | 15m | 3 | XRF | YY | 15m | 5 | XRF | YY |
| 047157 | 035-2 | 233m | 3 | XRF | YY | 31m | 2 | XRF | YY | 53m | 3 | XRF | YY | 22m | 5 | XRF | YY |
| 047158 | 050 | 138m | 3 | XRF | YY | 21m | 2 | XRF | YY | 32m | 3 | XRF | YY | 21m | 5 | XRF | YY |
| 047159 | 061 | 183m | 3 | XRF | YY | 33m | 2 | XRF | YY | 21m | 3 | XRF | YY | 28m | 5 | XRF | YY |
| 047160 | 073 | 141m | 3 | XRF | YY | 41m | 2 | XRF | YY | 9m | 3 | XRF | YY | 22m | 5 | XRF | YY |
| 047161 | STANDARD B-4 | 67m | 2 | XRF | YY | 11m | 2 | XRF | YY | <5m | 3 | XRF | YY | <= 15m | 4 | XRF | YY |
| 047162 | 078-C | 150m | 3 | XRF | YY | 40m | 2 | XRF | YY | 11m | 3 | XRF | YY | 26m | 5 | XRF | YY |
| 047163 | 171 | 148m | 3 | XRF | YY | 21m | 2 | XRF | YY | 38m | 3 | XRF | YY | 24m | 5 | XRF | YY |
| 047164 | 216 | 192m | 3 | XRF | YY | 28m | 2 | XRF | YY | 29m | 3 | XRF | YY | 23m | 5 | XRF | YY |
| 047165 | 254 | 95m | 2 | XRF | YY | 14m | 2 | XRF | YY | 26m | 3 | XRF | YY | <15m | 5 | XRF | YY |
| 047166 | 318-D | 148m | 3 | XRF | YY | 39m | 2 | XRF | YY | 13m | 3 | XRF | YY | 23m | 5 | XRF | YY |
| 047167 | 331-A | 168m | 3 | XRF | YY | 38m | 2 | XRF | YY | 20m | 3 | XRF | YY | 22m | 5 | XRF | YY |
| 047168 | 524-T | 33m | 3 | XRF | YY | 22m | 2 | XRF | YY | <5m | 3 | XRF | YY | 21m | 6 | XRF | YY |
| 047169 | B-81A | 144m | 3 | XRF | YY | 24m | 2 | XRF | YY | 22m | 3 | XRF | YY | 18m | 5 | XRF | YY |
| 047170 | XK5 | 122m | 2 | XRF | YY | 21m | 2 | XRF | YY | 17m | 3 | XRF | YY | 23m | 5 | XRF | YY |

Units Are Indicated As Follows: %=percent m=ppm b=ppb

| Lab Number | Field Number | Th Result | Th +/- Method | Th Inst | Ta Result | Ta +/- Method | Ta Inst | U Result | U +/- Method | U Inst | Ce Result | Ce +/- Method | Ce Inst |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 047155 | 031 | <15m | 4 XRF | YY | <15m | 5 XRF | YY | <15m | 3 XRF | YY | 44m | INAA | ACT |
| 047156 | 031-G | <15m | 4 XRF | YY | <15m | 4 XRF | YY | <15m | 3 XRF | YY | 31m | INAA | ACT |
| 047157 | 035-C | <15m | 4 XRF | YY | <15m | 4 XRF | YY | <15m | 3 XRF | YY | 25m | INAA | ACT |
| 047158 | 050 | <15m | 4 XRF | YY | <15m | 4 XRF | YY | <15m | 3 XRF | YY | 43m | INAA | ACT |
| 047159 | 051 | <15m | 4 XRF | YY | <15m | 4 XRF | YY | <15m | 3 XRF | YY | 47m | INAA | ACT |
| 047160 | 073 | <15m | 5 XRF | YY | <15m | 5 XRF | YY | <15m | 3 XRF | YY | 33m | INAA | ACT |
| 047161 | STANDARD S-4 | <15m | 4 XRF | YY | 17m | 4 XRF | YY | 20m | 3 XRF | YY | 52m | INAA | ACT |
| 047162 | 073-C | <15m | 5 XRF | YY | <15m | 5 XRF | YY | <15m | 3 XRF | YY | 29m | INAA | ACT |
| 047163 | 171 | <15m | 4 XRF | YY | <15m | 4 XRF | YY | <15m | 3 XRF | YY | 52m | INAA | ACT |
| 047164 | 216 | <15m | 4 XRF | YY | <15m | 5 XRF | YY | <15m | 3 XRF | YY | 52m | INAA | ACT |
| 047165 | 254 | <15m | 4 XRF | YY | <15m | 4 XRF | YY | <15m | 3 XRF | YY | 46m | INAA | ACT |
| 047166 | 318-D | <15m | 4 XRF | YY | <15m | 4 XRF | YY | <15m | 3 XRF | YY | 28m | INAA | ACT |
| 047167 | 331-A | <15m | 5 XRF | YY | <15m | 5 XRF | YY | <15m | 3 XRF | YY | 34m | INAA | ACT |
| 047168 | 524-T | <15m | 5 XRF | YY | <15m | 5 XRF | YY | <15m | 4 XRF | YY | 32m | INAA | ACT |
| 047169 | B-RIA | <15m | 4 XRF | YY | <15m | 4 XRF | YY | <15m | 3 XRF | YY | 42m | INAA | ACT |
| 047170 | BX5 | <15m | 4 XRF | YY | <15m | 4 XRF | YY | <15m | 3 XRF | YY | 31m | INAA | ACT |

Units Are Indicated As Follows: %=percent m=ppm b=ppb

| Lab Number | Field Number | Cs Result | Cs +/- | Cs Method | Cs Lab | La Result | La +/- | La Method | La Lab | Sc Result | Sc +/- | Sc Method | Sc Lab | Z Result | Z +/- | Z Method | Z Lab |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 047155 | 031 | 2m | | INAA | ACT | 21m | | INAA | ACT | 25m | | INAA | ACT | 345m | | XRF | COX |
| 047156 | 031-B | 6m | | INAA | ACT | 12m | | INAA | ACT | 44m | | INAA | ACT | 252m | | XRF | COX |
| 047157 | 035-2 | 2m | | INAA | ACT | 34m | | INAA | ACT | 31m | | INAA | ACT | 353m | | XRF | COX |
| 047158 | 060 | <1m | | INAA | ACT | 39m | | INAA | ACT | 43m | | INAA | ACT | 221m | | XRF | COX |
| 047159 | 061 | <1m | | INAA | ACT | 22m | | INAA | ACT | 29m | | INAA | ACT | 280m | | XRF | COX |
| 047160 | 073 | 13m | | INAA | ACT | 14m | | INAA | ACT | 54m | | INAA | ACT | 530m | | XRF | COX |
| 047161 | STANDARD B-4 | 2m | | INAA | ACT | 33m | | INAA | ACT | 11m | | INAA | ACT | 130m | | XRF | COX |
| 047162 | 078-C | 15m | | INAA | ACT | 13m | | INAA | ACT | 45m | | INAA | ACT | 462m | | XRF | COX |
| 047163 | 171 | <1m | | INAA | ACT | 25m | | INAA | ACT | 39m | | INAA | ACT | 265m | | XRF | COX |
| 047164 | 216 | <1m | | INAA | ACT | 24m | | INAA | ACT | 29m | | INAA | ACT | 356m | | XRF | COX |
| 047165 | 254 | <1m | | INAA | ACT | 24m | | INAA | ACT | 32m | | INAA | ACT | 197m | | XRF | COX |
| 047166 | 318-D | 5m | | INAA | ACT | 11m | | INAA | ACT | 48m | | INAA | ACT | 423m | | XRF | COX |
| 047167 | 331-A | 4m | | INAA | ACT | 17m | | INAA | ACT | 43m | | INAA | ACT | 425m | | XRF | COX |
| 047168 | 524-T | <1m | | INAA | ACT | 1.5m | | INAA | ACT | 67m | | INAA | ACT | 620m | | XRF | COX |
| 047169 | B-RIA | <1m | | INAA | ACT | 20m | | INAA | ACT | 37m | | INAA | ACT | 420m | | XRF | COX |
| 047170 | XRS | <1m | | INAA | ACT | 15m | | INAA | ACT | 35m | | INAA | ACT | 258m | | XRF | COX |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PHONOLITE
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TRACHYTE
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DACITE
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ANDESITE
BASANITE
NEPHELINITE
ANDESITE / BASALT
ALKALI-BASALT
SUB-ALKALINE BASALT
Zr/TiO2
Nb/Y
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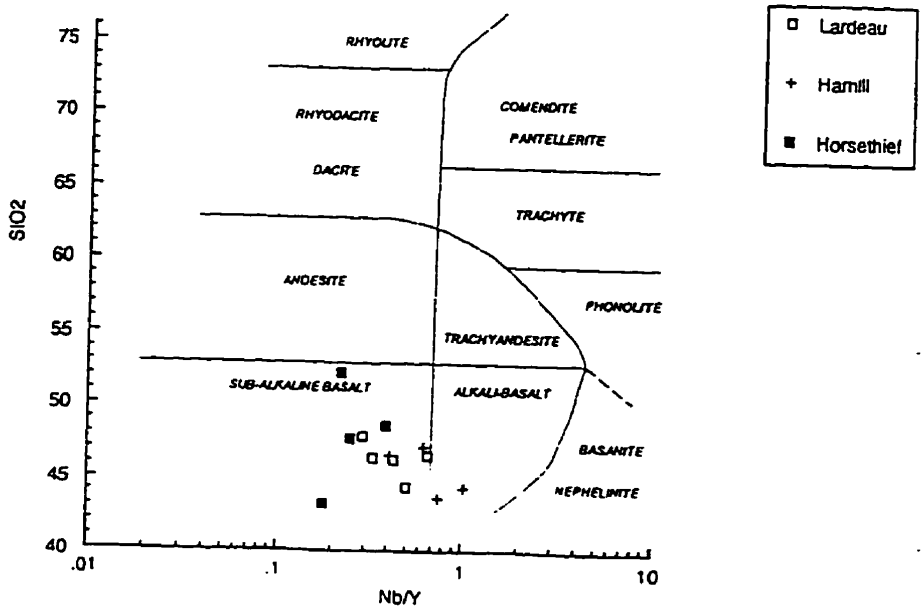
.01
.001
.01
.1
1
10
Horsethief
Hamill
Lardeau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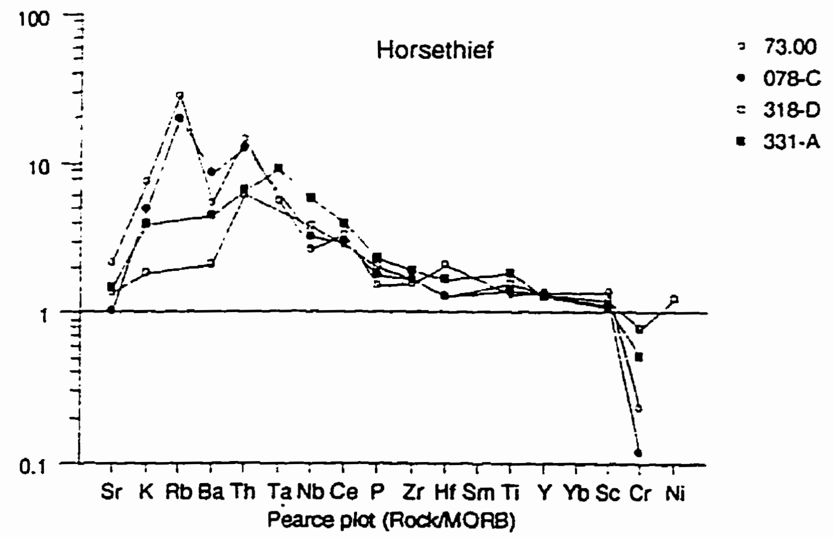
Horsethief


Horsethief
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318-D
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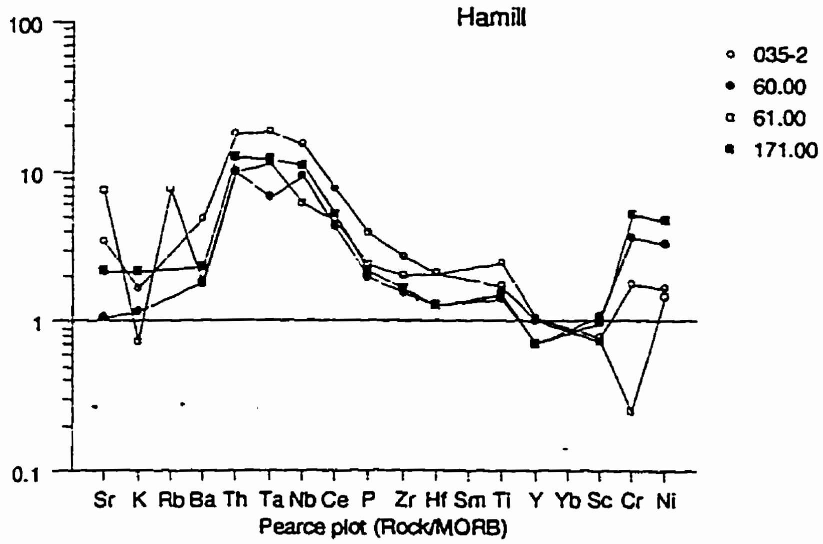
Pearce plot (Rock/MORB)


Hamill
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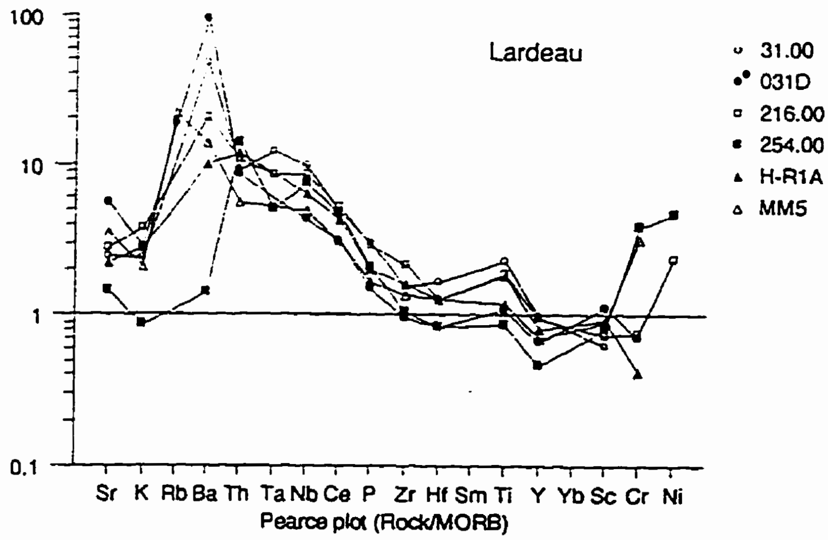
Pearce plot (Rock/MORB)



100
10
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Lardeau
31.00
031D
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MM5
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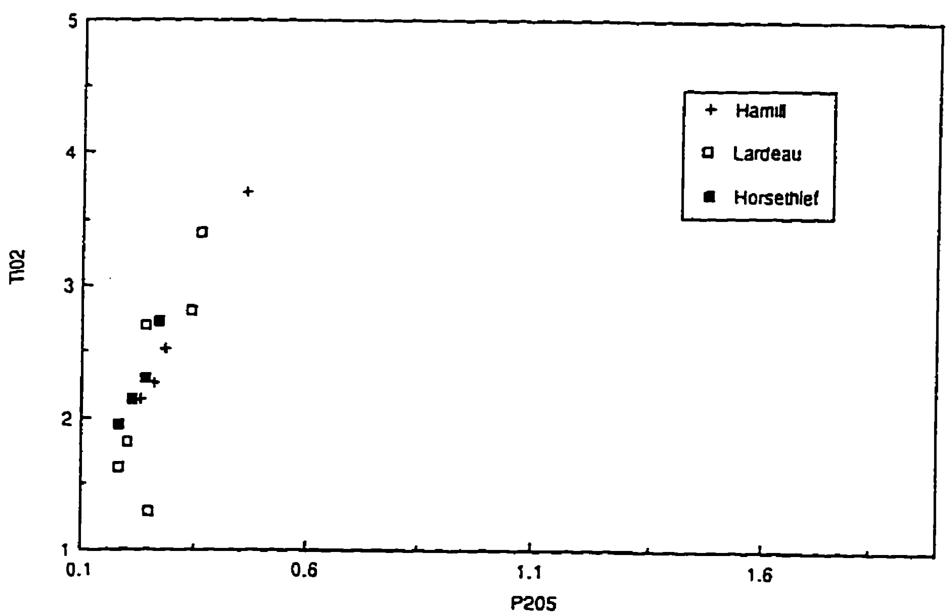
Pearce plot (Rock/MORB)



Hamill
Lardeau
Horsethief
TiO2
P2O5
5
4
3
2
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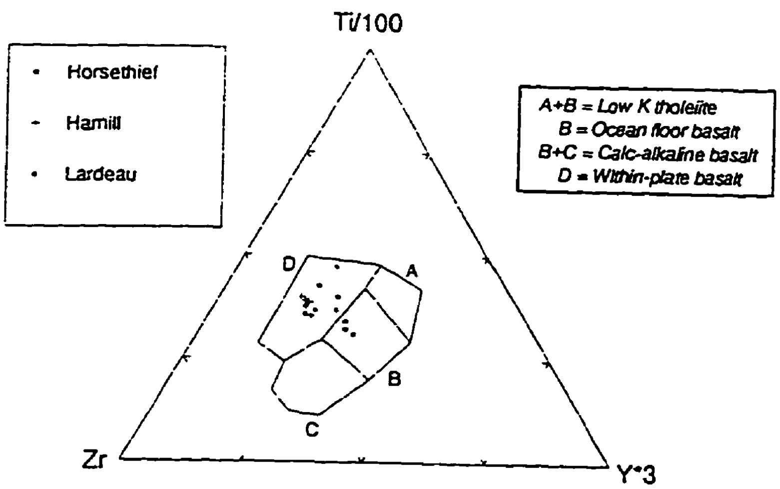
0.1
0.6
1.1
1.6


Horsethief
Hamill
Lardeau
Ti/100
A+B = Low K tholeiite
B = Ocean floor basalt
B+C = Calc-alkaline basalt
D = Within-plate basalt
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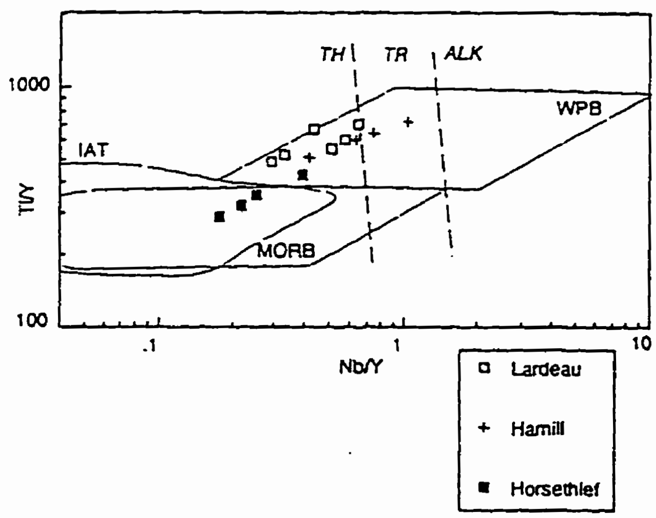
A
B
C
Zr
Y*3


TH
TR
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IAT
WPB
MORB
Ti/Y
.1
1
10
Nb/Y
Lardeau
Hamill
Horsethief

# Appendix 4

## PALINSPASTIC RESTORATION OF THE PURCELL ANTICLINORIUM AND KOOTENAY ARC

Palinspastic restoration (Plate 4; in pocket) of a regional cross section through the Purcell anticlinorium and the Kootenay Arc (Plate 4a) illustrates their Mesozoic tectonic evolution (Chapter 4) and shows their configuration following Neoproterozoic and Early Paleozoic crustal attenuation. The restored sections were constructed using palinspastic techniques coupled with boundary conditions provided by new and published geochronological and geobarometric data. The sections are presented below from youngest to oldest, in order to discuss the assumptions used in the restoration and the critical relationships revealed by each step.

### 1) PRESENT

The present section is a composite regional cross section through the Kootenay Arc and the Purcell anticlinorium from northern Kootenay Lake to the Rocky Mountain Trench near Invermere, B. C. The location of the line of section is shown in Figure 4 - 2. The Kootenay Arc segment is modified from a section presented by J. T. Fyles and T. Hōy in Gabrielse (1991), constructed from Fyles (1964), Fyles and Eastwood (1962), Read (1973), Klepacki (1985) and Klepacki and Wheeler (1985). Western Purcell anticlinorium segment is Section H of Warren (1996a; Plate 2). The eastern Purcell anticlinorium section is modified from Root (1987). Pressure data and plutons that provide critical age constraints are projected into line of section. Projected plutons are shown by dotted lines. Pressure data without ages are constrained between 178 and 164 Ma. The regional garnet and biotite isograds are also shown but are poorly constrained in the line of section. The Cordilleran basal décollement is constrained beneath the eastern Purcell anticlinorium by seismic data as discussed by Cook et al. (1988). Thrust displacement of less than 3 km on the Kootenay Arc boundary fault is based on stratigraphic and geobarometric data. The closure of the easternmost isoclinal anticline in the Kootenay Arc is not constrained and the structure need not be as large as shown.

### 2) RESTORED TO ABOUT 135 Ma (Constraints: 157 Ma - 93 Ma)

The eastern Purcell anticlinorium is restored prior to flat, out-of-sequence thrusts, which include the Mount Forster thrust and a parallel fault system at depth. Both cut the Mount Forster syncline and other folds. Motion on the Purcell thrust also must have occurred with the other out of sequence thrusts, but this motion is not included in the restoration. The faults pre-date the cross-cutting Horsethief Creek batholith, indicating that they occurred prior to 93 Ma (D. A. Archibald, unpublished $^{40}Ar/^{39}Ar$ data). See discussion of Section 3 (≈155 Ma) for further age constraints on the deformation in the eastern Purcell anticlinorium. The displacement ( 6 km) on the Mount Forster thrust used in this restoration is read from hangingwall and footwall cutoffs of the axial trace of the Mount Forster syncline and of the top of the Toby Formation as shown on Reesor's (1973) map. However, stratigraphy beneath the Toby Formation does not match across the Mount Forster thrust due to Neoproterozoic normal faulting (Root, 1987; Pope, 1990). This reconstruction assumes that motion on the Mount Forster thrust was parallel to the section (about N70E). The Cordilleran basal décollement extended beneath the entire Purcell anticlinorium and perhaps beneath the Western Main Ranges at this stage.

The Kootenay Arc is restored prior to the two late, steeply west-dipping, post-metamorphic normal faults. The configuration of the western normal fault has been modified from that shown in Klepacki (1985) and Fyles (1964) so that it is not overturned at high structural levels and so that hangingwall and footwall cutoffs are more nearly compatible. Late folds with west-dipping axial surfaces are shown as the youngest structures in the Kootenay Arc by Klepacki

(1985), but the cross-cutting relationships were not observed directly in the field. These normal faults are restored in this interpretation prior to the late folds in order to avoid significant space problems encountered by restoring the folds first. These post-metamorphic faults could be the expression of Jurassic syn-orogenic extension inferred by Colpron et al. (1996), of Cretaceous syn-orogenic extension inferred by Scammell (1993) or of Eocene extension and exhumation connected with the Purcell Trench fault system. Possible regional Eocene tilting and exhumation on the western flank of the Purcell anticlinorium (Archibald et al., 1984; Jansen, 1991; Doughty, 1996; and this study) is not considered in this step of the restoration. This restoration assumes that there was no normal reactivation of the Kootenay Arc boundary fault from post-Middle Jurassic through Eocene time.

### 3) RESTORED TO ABOUT 155 Ma (Constraints: 157 Ma - 93 Ma)

The eastern Purcell anticlinorium is restored prior to regional, upright folding. The age of these folds is poorly constrained, but they are congruent with upright folds in the western Purcell anticlinorium that are inferred to have formed before 159 - 157 Ma, and probably were part of the same episode of deformation. The Kootenay Group in the foreland basin was deposited between about 155 and 135 Ma, and it contains clastic material that records uplift and erosion of strata now exposed in the Purcell anticlinorium and Western Main Ranges. Thus, it is reasonable to propose that both the early thrusts (see Section 4; ≈170 Ma) and folds in the eastern Purcell anticlinorium occurred between about 155 and 135 Ma. Root (1987) proposed that the early thrust and the upright folds were partly synchronous. They have been restored separately, but the early thrusts must have occurred immediately prior to the time shown in this section. These age relationships imply that the upright folding was diachronous across the Purcell anticlinorium from west (pre-157 Ma) to east (post-155 Ma). The basal décollement is inferred to have propagated part way beneath the eastern Purcell anticlinorium at this stage, because the early thrusts are presumably splays from it.

The Kootenay Arc is restored prior to development of the folds with shallow, west-dipping axial surfaces. These are most conspicuous in the immediate hangingwall of the Kootenay Arc boundary fault, and they are interpreted to have developed during hangingwall rotation and sub-vertical flattening in response to increasingly steep motion on the fault (Klepacki, 1985). Part of this thrust motion is restored here, in order to illustrate this interpretation. Thrust motion on the Kootenay Arc boundary fault post-dates the west-verging folds and felsic dikes in the Kootenay Arc (173 +/- 5 Ma) and was partly synchronous with the upright folding in both the Kootenay Arc and western Purcell anticlinorium. Therefore, the thrust motion is no older than 178 Ma and probably not much younger than 159 - 157 Ma.

## 4) RESTORED TO ABOUT 170 Ma

The eastern Purcell anticlinorium is restored prior to early, east-verging thrust faults (Root, 1987). Displacement is constrained by matching hangingwall and footwall cutoffs in the repeated sequence of Dutch Creek Formation. These thrust faults root into the same décollement used by the younger Mount Forster thrust. Stratigraphic relationships across the Mount Forster thrust indicate that this surface must have significant pre-Mesozoic normal motion. Normal faulting and eastwrd tilting in the are recorded by overstepping unconformities in Lower Paleozoic strata in the eastern part of section associated with the emergence of the "Windermere High" (Walker, 1926; Reesor, 1973; Root, 1987). The youngest normal motion was Middle Devonian. It resulted in deposition of a thick sequence of coarse, immature clastic and mafic volcanic rocks (Mount Forster Formation) on the western part of the Windermere High, which contrasts with a thinner, more mature and volcanic-free sequence on the eastern Windermere High (Root, 1987). Stratigraphic relationships beneath the Toby Formation indicate that the Proterozoic antecedent to the Windermere High was a graben, and that the Horsethief Creek Group immediately to the west was deposited on a relatively high-standing block.

This section is restored prior to the upright folding in the Kootenay Arc and prior to most of the upright folding in the western Purcell anticlinorium. This is based on the interpretation that the episode of upright folding in the western Purcell anticlinorium was related to both phases of folding in the Kootenay Arc. Toby and Glacier Creek stocks were emplaced during this episode of folding in the western Purcell anticlnorium. They are no older than 185 Ma and no younger than 159 Ma, and they are assumed to have been emplaced at about the same time as the 170 Ma Mine stock that intrudes the western Purcell anticlinorium to the south (Archibald et al., 1983). This section shows the western Purcell anticlinorium immediately prior to emplacement of the stocks. The stocks' contact aureoles overprint the regional assemblages that are associated with the upright folding, but the contact assemblages are also deformed by the latter stages of upright folding. The pressures recorded by the contact aureoles are significantly lower than the regional assemblages that they overprint, indicating that there was significant, rapid exhumation between regional peak metamorphic conditions and emplacement of the syn-kinematic plutons. The upright folding in the Kootenay Arc is younger than 173 +/- 5 Ma and had ceased by 165 Ma, the K-Ar hornblende age of a post-kinematic phase of the Nelson batholith (Mt. Carlyle stock; Wanless et al., 1968).

Restoration of the upright folds in the Kootenay Arc shows that the early isoclinal folds were west-verging structures and not back-rotated east-verging structures as suggested by many previous workers (Fyles, 1964; Hōy, 1980; Archibald et al., 1983; Varsek and Cook, 1991). The thrust motion on the Kootenay Arc boundary fault is also restored, because it was synchronous with the upright folds and their axial planar schistosity. The steep part of the Kootenay Arc boundary fault is interpreted to follow a stratigraphic discontinuity that corresponds to an older normal fault, one of several growth faults that controlled the deposition of the discontinuous lower Hamill Group (Warren, 1996b, and Chapter 3). A similar reactivated normal fault occurs in the western Purcell anticlnorium and may have partly controlled the intrusion of Toby stock. The Kootenay Arc boundary fault is interpreted as a steep splay from the floor thrust of an east-verging tectonic wedge (see Price, 1986) of North American strata that has been driven beneath the rocks of the Kootenay Arc. The stratigraphic and structural discontinuity between the Kootenay Arc and the Purcell anticlinorium corresponds to the eastern limit of progress of the tectonic wedge, which impinged upon the eastern edge of the lower Hamill Group basin. Further horizontal motion was impeded, and the wedge, with the strata above it and in the adjacent Purcell anticlinorium, were subsequently shortened by upright folding, until the Cordilleran basal décollement propagated beneath the Purcell anticlinorium from the floor thrust of the wedge, carrying the Kootenay Arc and the deformed wedge in its hangingwall. The Kootenay Arc boundary fault probably developed as a splay from the tip of the wedge at the same time.

A west-verging thrust fault that juxtaposes the Duncan anticline against the Meadow Creek anticline is shown above the wedge as a splay from its roof thrust. This fault is required by repetition of the Jowett Formation shown in the composite section by Klepacki (1985).

The western part of the section is shown significantly subsided and tilted to the west, due to 20 - 27 km of tectonic loading recorded by regional metamorphic assemblages in the Horsethief Creek and Hamill Groups in the Kootenay Arc and western Purcell anticlinorium. Possible strike-slip motion on the Kootenay Arc boundary fault and the effects of non-cylindrical folding and non-plane strain in the Kootenay Arc are not considered in this restoration. Shortening during the upright phase of folding was calculated perpendicular to axial surfaces, and the effects of simple shear during folding were not considered. The datum (line length) used was the top of the Hamill Group, because it is well constrained in the present cross-section and because the upper Hamill Group is the most competent unit in the section and thus least likely to have been affected by tectonic thinning and thickening or volume loss during folding.

### 5) RESTORED TO ABOUT 173 Ma (Constraints: 173+/-5 Ma)

The Kootenay Arc is restored to the time of emplacement of the felsic dikes (173+/-5 Ma), prior to the development of the west-verging thrust that developed above the tectonic wedge. The section illustrates an early stage in the eastward displacement of the tectonic wedge beneath the rocks of the Kootenay Arc and shows the relationship between the wedge and the west-verging folds. The west-verging, ductile structures developed because rocks above the wedge became progressively coupled to it as it propagated eastward on the Cordilleran basal décollement and on its floor thrust. The pelitic unit at the top of the Windermere Supergroup and the overlying upper Hamill Group and gently east-dipping base of the lower Hamill Group provided a favorably-oriented anisotropy for insertion of the wedge until the wedge impinged on the eastern edge of the lower Hamill Group basin.

The Meadow Creek anticline is shown as a composite structure comprising three sheared-out isoclinal anticlines, so that it is in part a west-verging ductile thrust duplex. The felsic dikes were emplaced at 173 +/- 5 Ma during the development of the strong axial planar foliation in the Meadow Creek anticline. The region of diking is shown as a flat-lying lens of sill-like intrusions (cross-hatching) above the upper detachment, but the dikes could extend into the unexposed wedge below. The dikes were perhaps the result of melting initiated by fluids channelled along the roof thrust of the wedge. The Kootenay Arc and perhaps the western Purcell anticlinorium would have been tectonically loaded by Slide Mountain and Quesnel terranes, which were probably 12 - 15 km thick (Cook et al., 1988), and perhaps by local foreland basin sediments that are not preserved, although there may not have been significant emergent topography at this stage.

East-verging thrust faults within the Milford Group to the west of the Meadow Creek anticline (Klepacki, 1985) represent 187 - 178 Ma telescoping of the North American marginal basin and obduction of Slide Mountain and Quesnel terranes prior to involvement of Lower Paleozoic and older North American strata in the orogeny.

## 6) RESTORED TO POST-375 Ma

The section is restored to its post-Frasnian (lower Upper Devonian), pre-Middle Jurassic configuration. The line length used for the restoration of the west-verging folds in the Kootenay

Arc is the top of the upper Hamill Group. Several km of thrust motion in addition to isoclinal folding are assumed on each of the "horses" in the Meadow Creek "duplex," prior to the emplacement of the felsic dikes. The incipient tectonic wedge is shown schematically at the far left of section. The restored volume (area) of the Kootenay Arc is significantly greater than its present volume (area); this is attributed to significant volume loss due to pressure solution, for which there is ample field evidence. The section shows that normal faulting that controlled the deposition of the lower Hamill Group had ceased before deposition of the upper Hamill Group and Mohican and Badshot Formations. This restoration does not show subsequent Early Paleozoic normal motion (post-Badshot) that reactivated the Kootenay Arc boundary fault antecedent (Chapters 3, 4; Devlin, 1989; Kubli, 1990; Colpron et al., in review), because this displacement is unconstrained (see schematic stratigraphic section in Fig. 4-3). A normal fault of lower Windermere age is shown schematically beneath the western Purcell anticlinorium to represent both along-strike and transverse syn-Windermere normal faults discussed in Chapter 2. This zone of normal faults in the wester Purcell anticlinorium marks the rifted western and northern edge of the Purcell Supergroup, which was truncated and removed to the north and west during Neoproterozoic continental rifting.

# Appendix 5

## PHOTOGRAPHIC PLATES

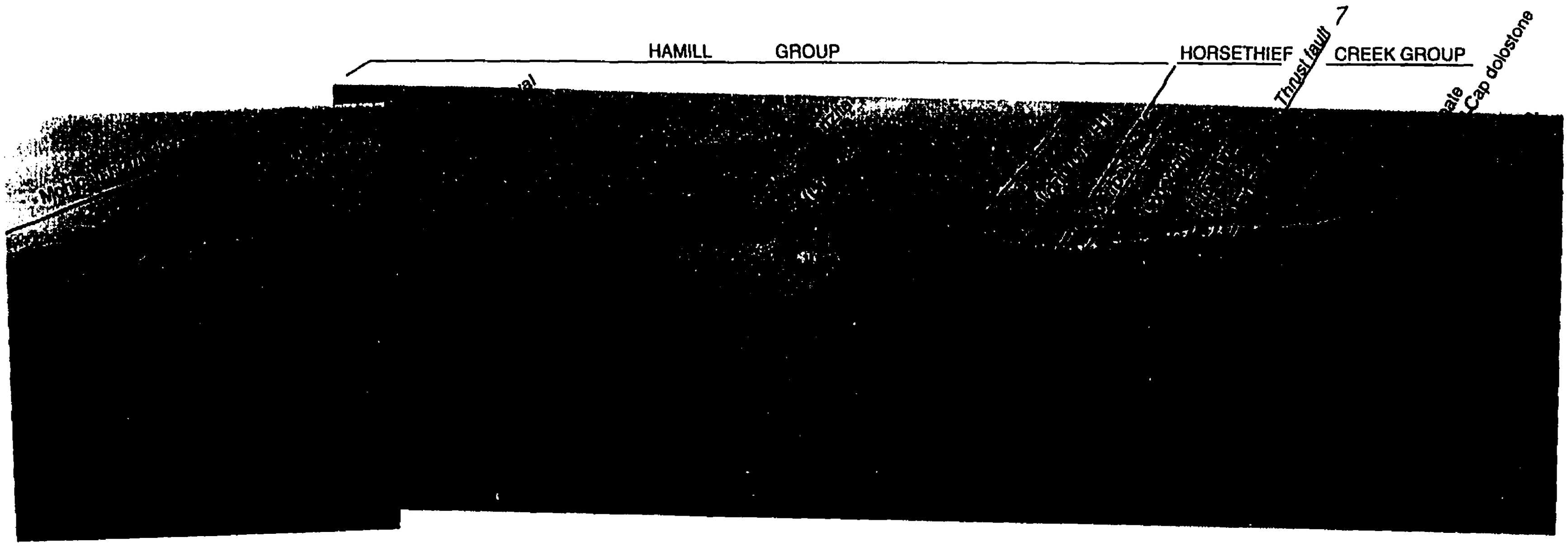
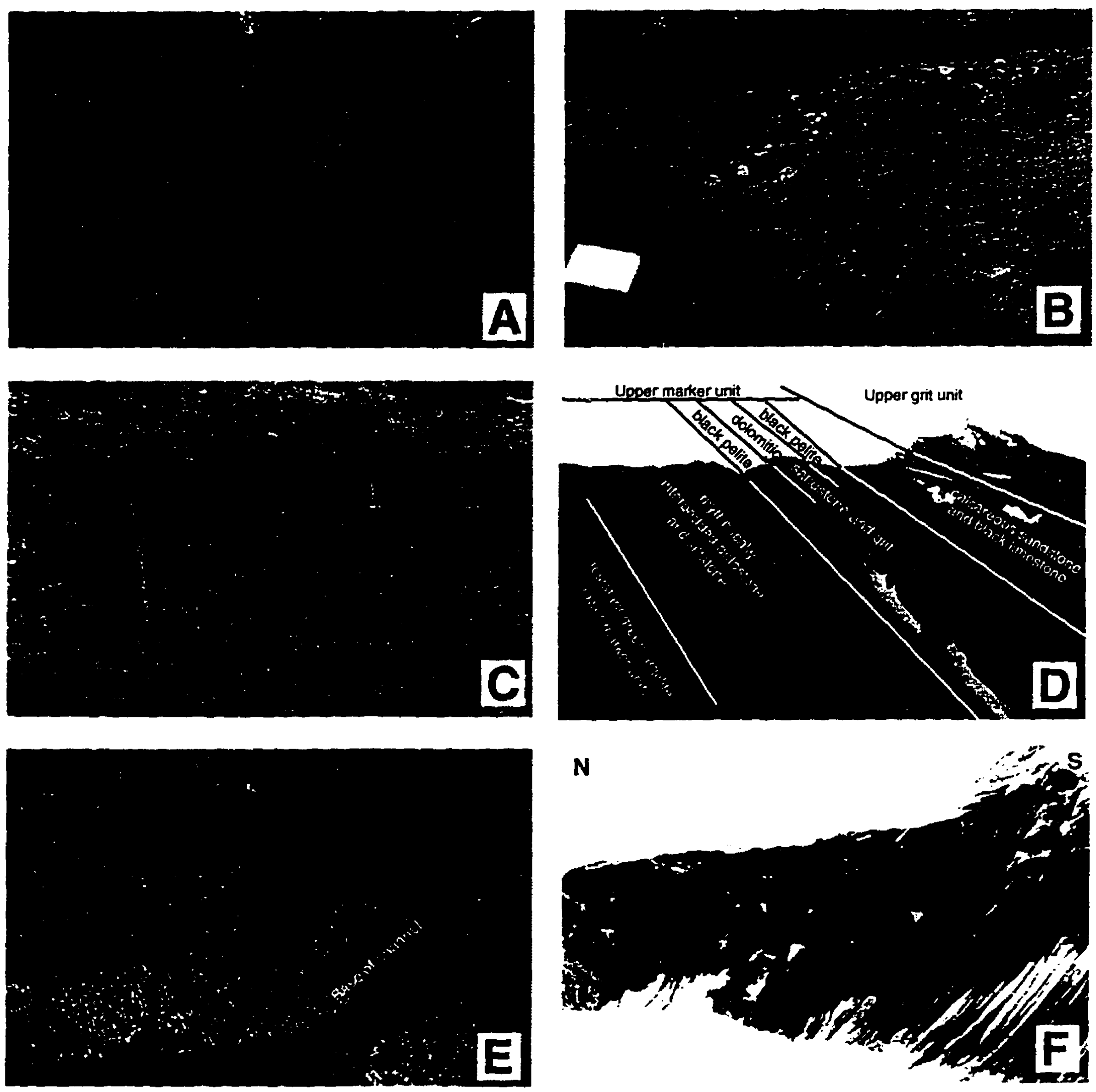



**PHOTO PLATE 1: COMPLETE SECTION THROUGH THE HORSETHIEF CREEK AND HAMILL GROUPS**

Panorama of exposures of complete sections of the Horsethief Creek and Hamill Groups in the Jumbo Creek drainage (north fork). West is at left of photo; north is at right. The section at the headwaters of Jumbo Creek (right of center) is the location of the Windermere Supergroup stratigraphic column shown in Figure 2-6A. The exposures of the Hamill Group, on the east limb of the Blockhead Mountain syncline, are the basis for the composite stratigraphic column from the Hamill Group shown in Figure 3-3. The lower quartzite unit of the Hamill Group appears abnormally thick because it is viewed for several km along strike. Photo taken from just below the top of the Toby Formation, east of Jumbo Creek. Jumbo Pass is immediately to left of photo. See Plate 1 for geological map and place names.




## PHOTO PLATE 2: THE HORSETHIEF CREEK GROUP

A) Rhythmically graded beds in turbidites of the lower pelite unit (PHClp on Plates 1-3), upper Toby Creek watershed.
B) Poorly sorted pebble to cobble conglomerate, of the lower calcareous clastic unit (PHClcg), upper Horsethief Creek watershed. Pebbles and cobbles include marble, dolostone, blue and white quartz and k-spar. Matrix is calcareous mudstone to sandstone or grit. Notebook is about 10 cm long.
C) Carbonate cobble conglomerate lithofacies of the regional carbonate and fine clastic marker unit (PHCm or PHCum), east side of Starbird Glacier near Starbird Pass. Rounded cobbles are fine-grained, recrystallized dolostone or marble. Slabs are primarily dolomitic siltstone.
D) Section through the lower (PHClm) and upper marker unit (PHCum; about 125 m thick) on the divide between upper Howser and Stockdale Creeks (north of Eyebrow Peak). Note that these strata are strongly cleaved and lithological contacts are sheared.
E) Amalgamated channeled grit and conglomerate beds of the upper grit unit (PHCug), at headwaters of the north fork of Jumbo Creek. Vertical scale of photo is about 2 m.
F) Laterally discontinuous grit channels, and intercalated pelite (levee deposits?) thin toward the south in the upper grit unit (PHCug), on the ridge west of Birthday Peak. Grit beds are more laterally continuous and thinner on this ridge to north and south of photo; grit bed geometry and pebble imbrication suggest that sediment was transported from southeast to northwest in a significant distributory axis in this area. Vertical relief on right of cliff face is about 350 m.

**All place names are located on Plate 1.**

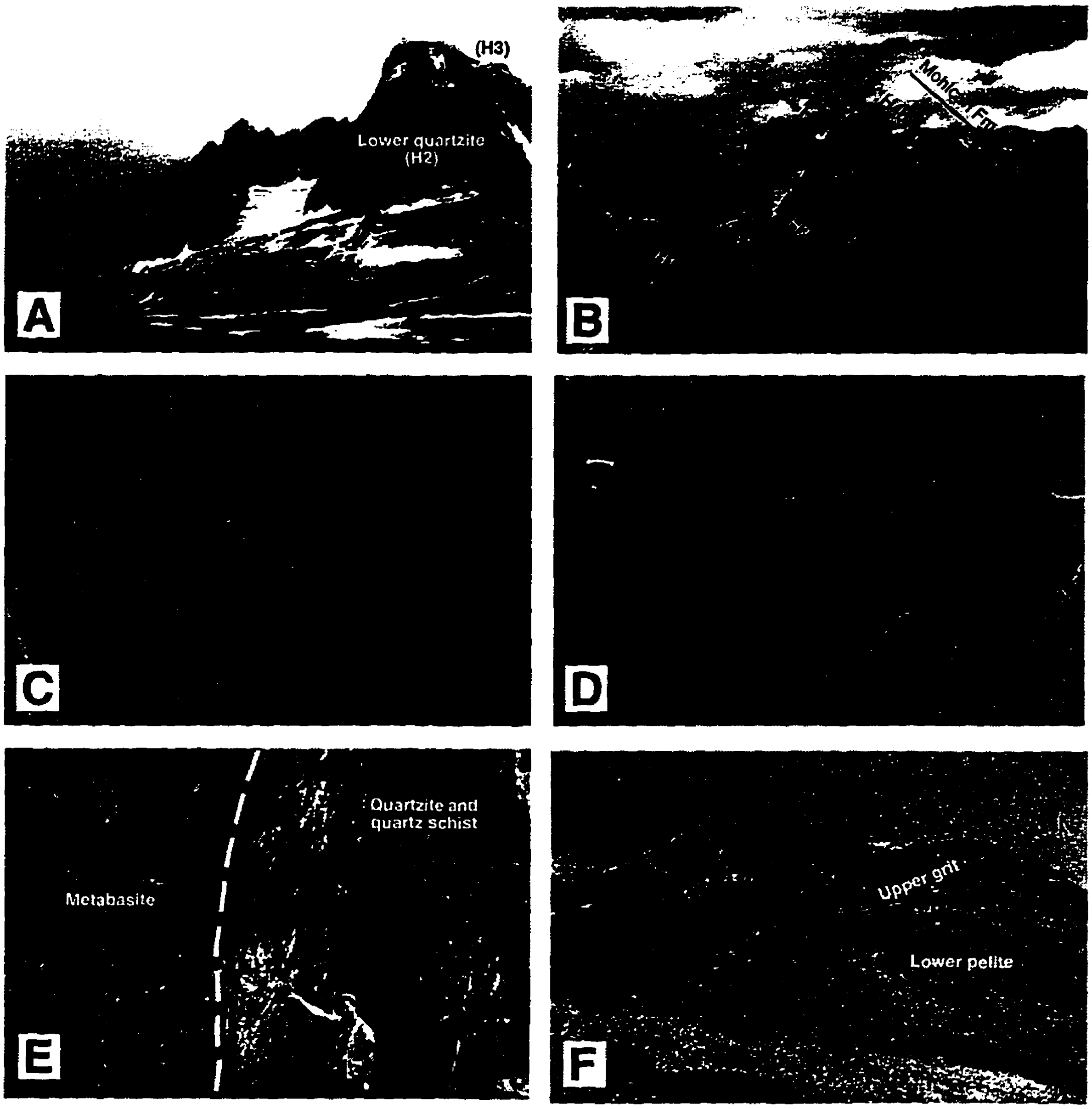



**PHOTO PLATE 3: THE HAMILL GROUP**

A) The lower Hamill Group on the east limb of the Cauldron Mountain syncline, upper Glacier Creek watershed (north fork). Prominent peak is Mt. MacDuff. Photo looks southwest. Vertical relief is about 400 m.
B) The Hamill Group on the west limb of the Blockhead Mountain syncline, upper Glacier Creek watershed (south fork). Photo looks north. Jumbo Pass is to right of photo. Vertical relief is about 500 m.
C) Large-scale uni-directional cross beds and planar beds typical of the lower quartzite unit (H2), east of Jumbo Pass.
D) Small-scale multi- or bi-directional cross beds typical of the upper quartzite unit, east of Jumbo Pass.
E) Contact between metabasite and quartzite and schist in the lower part of the Hamill Group in the Kootenay Arc (Unit HA). Outcrop is along shore of Duncan Lake, 500 m west of the Kootenay Arc boundary fault. Metabasites occur both as sills or flows and as dikes in the lower Hamill Group.
F) Exposure of the Earl Grey Pass thrust fault (heavy dashed line) about 1 km south of Earl Grey Pass. Light dashed lines outline grit beds that are truncated against the fault. Horsethief Creek Group strata are cut out along the fault, implying that it was a post-Windermere normal fault that was reactivated as a thrust fault. The Horsethief Creek Group marker unit and the lower part of the upper grit unit are missing in this exposure. The Hamill Group occurs in the hangingwall but is absent in the footwall to the east, and its deposition may have been controlled by this normal fault (see Chapter 3).
**All place names are located on Plate 1.**

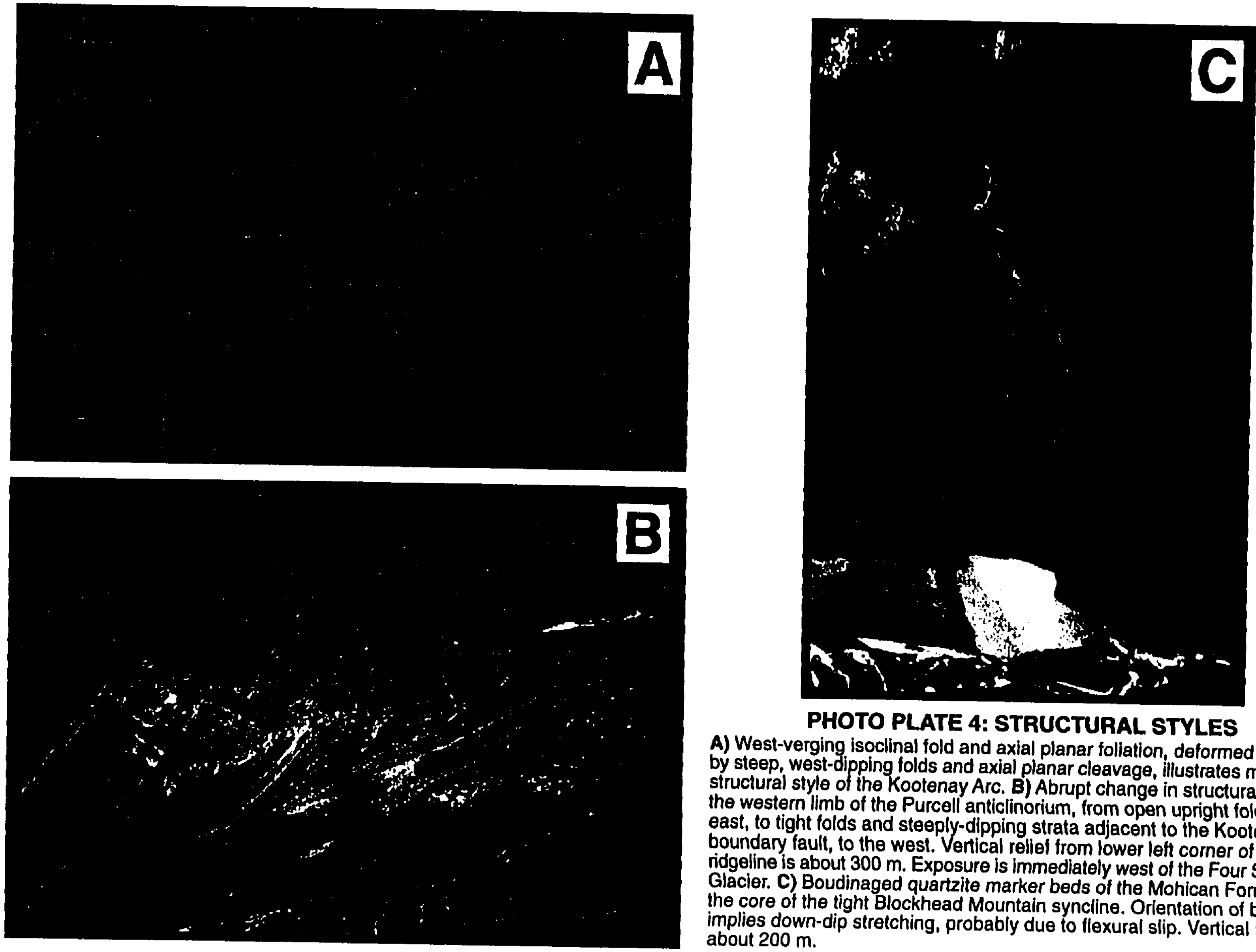



**PHOTO PLATE 4: STRUCTURAL STYLES**

**A)** West-verging isoclinal fold and axial planar foliation, deformed coaxially by steep, west-dipping folds and axial *planar* cleavage, illustrates map-scale structural style of the Kootenay Arc. **B)** Abrupt change in structural style on the western limb of the Purcell anticlinorium, from open upright folds, to the east, to tight folds and steeply-dipping strata adjacent to the Kootenay Arc boundary fault, to the west. Vertical relief from lower left *corner of* photo to ridgeline is about 300 m. Exposure is immediately west of the Four Squatters Glacier. **C)** Boudinaged *quartzite marker beds* of the Mohican Formation in the core of the tight Blockhead Mountain syncline. Orientation of boudins implies down-dip stretching, probably due to flexural slip. Vertical relief is about 200 m.

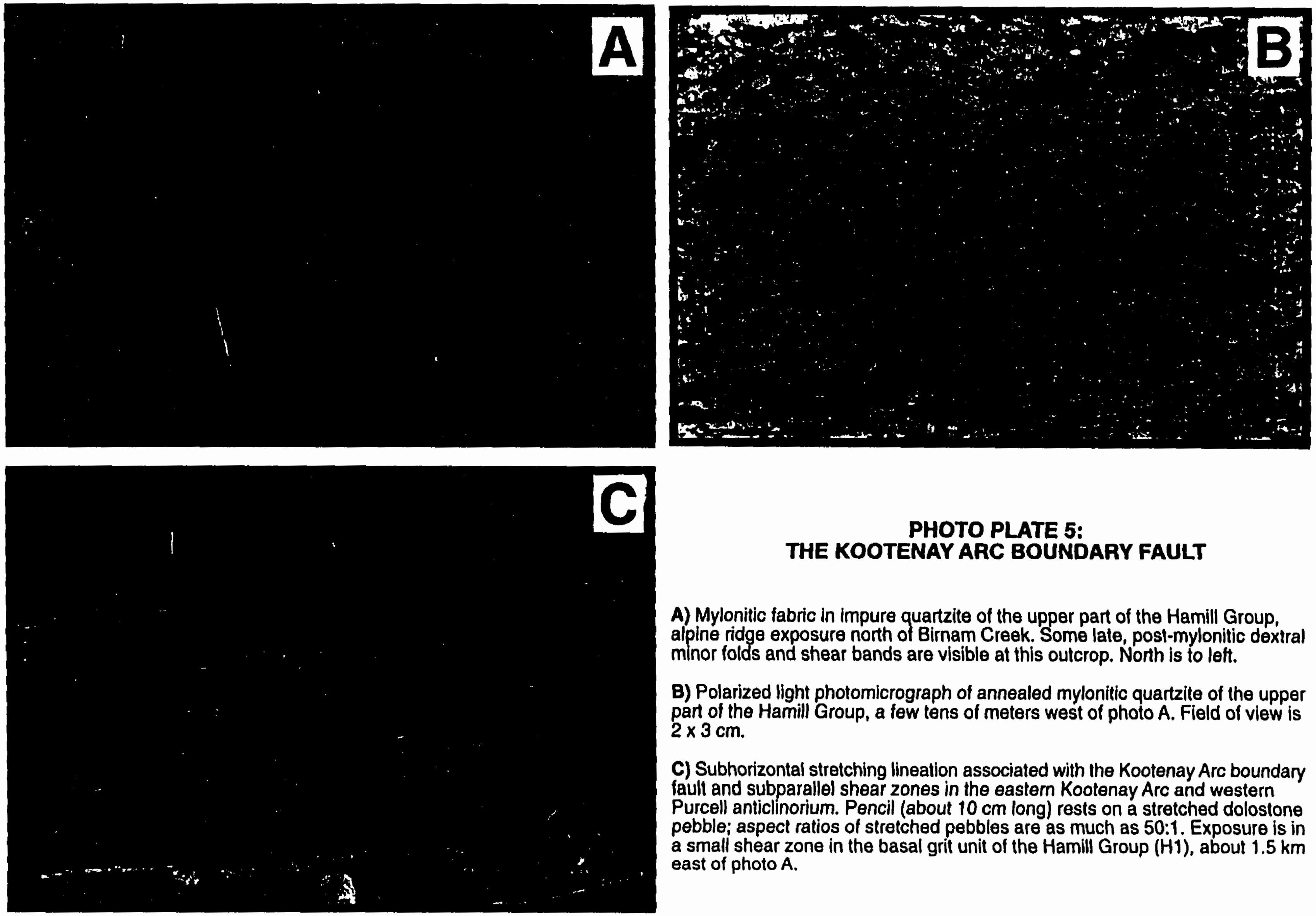



**PHOTO PLATE 5:**
**THE KOOTENAY ARC BOUNDARY FAULT**

**A)** Mylonitic fabric in impure quartzite of the upper part of the Hamill Group, alpine ridge exposure north of Birnam Creek. Some late, post-mylonitic dextral minor folds and shear bands are visible at this outcrop. North is to left.

**B)** Polarized light photomicrograph of annealed mylonitic quartzite of the upper part of the Hamill Group, a few tens of meters west of photo A. Field of view is 2 x 3 cm.

**C)** Subhorizontal stretching lineation associated with the Kootenay Arc boundary fault and subparallel shear zones in the eastern Kootenay Arc and western Purcell anticlinorium. Pencil (about 10 cm long) rests on a stretched dolostone pebble; aspect ratios of stretched pebbles are as much as 50:1. Exposure is in a small shear zone in the basal grit unit of the Hamill Group (H1), about 1.5 km east of photo A.

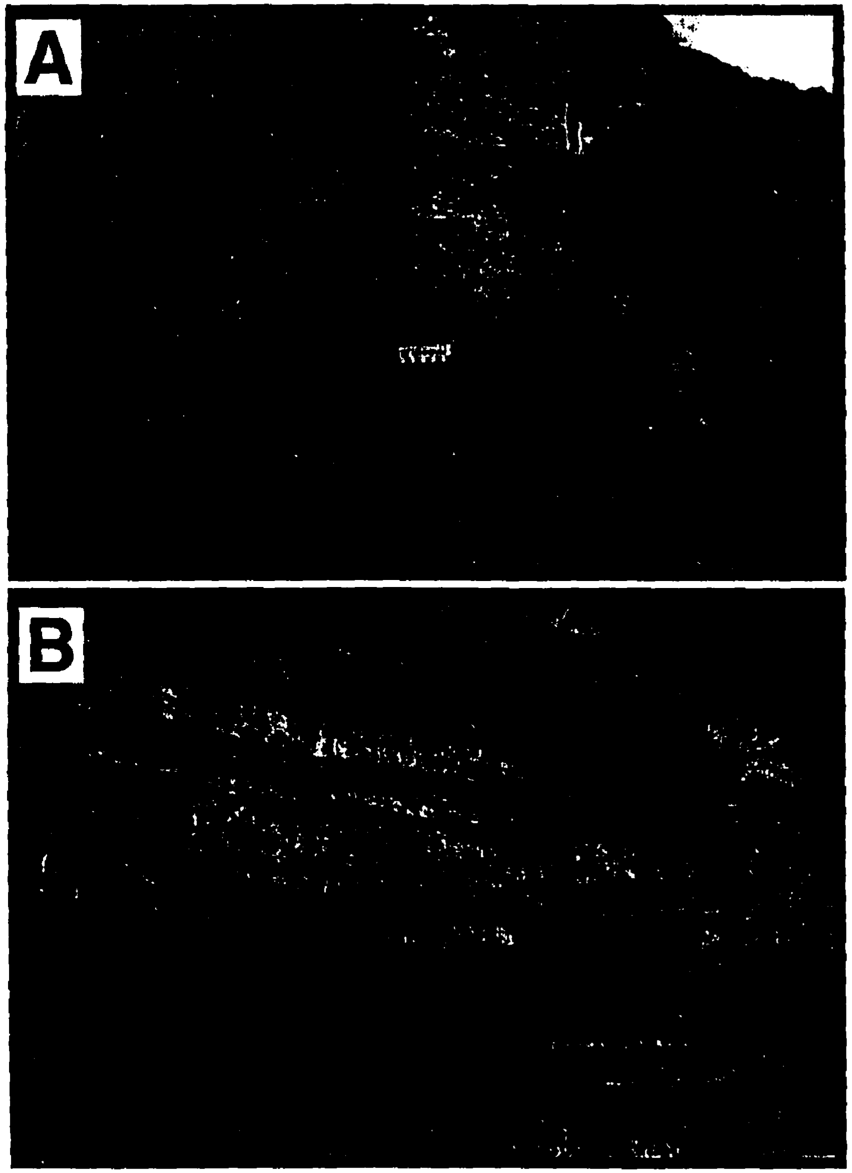



**PHOTO PLATE 6:
INTRUSIONS AND DEFORMATION**

**A)** Syn-kinematic veins associated with the Toby stock are deformed by folds that are parasitic to the Blockhead Mountain syncline, but also locally cut the axial surfaces of these folds. Scale bar in centimeters.

**B)** Syn-kinematic felsic dikes exposed in the hinge region of the recumbent F1 Meadow Creek anticline, west of Kootenay and Duncan lakes. The dikes are parallel to the dominant F1 foliation in the country rock, locally cut bedding at minor F1 hinges and locally contain the F1 foliation. The F1 foliation and the dikes are deformed by the steeper F2 structures associated with the Kootenay Lake antiform, which appear at this outcrop only as a crenulation or weak spaced cleavage in the country rock (quartzose schist of the Hamill Group). Thickest dike in photo is about 1.5 m thick.

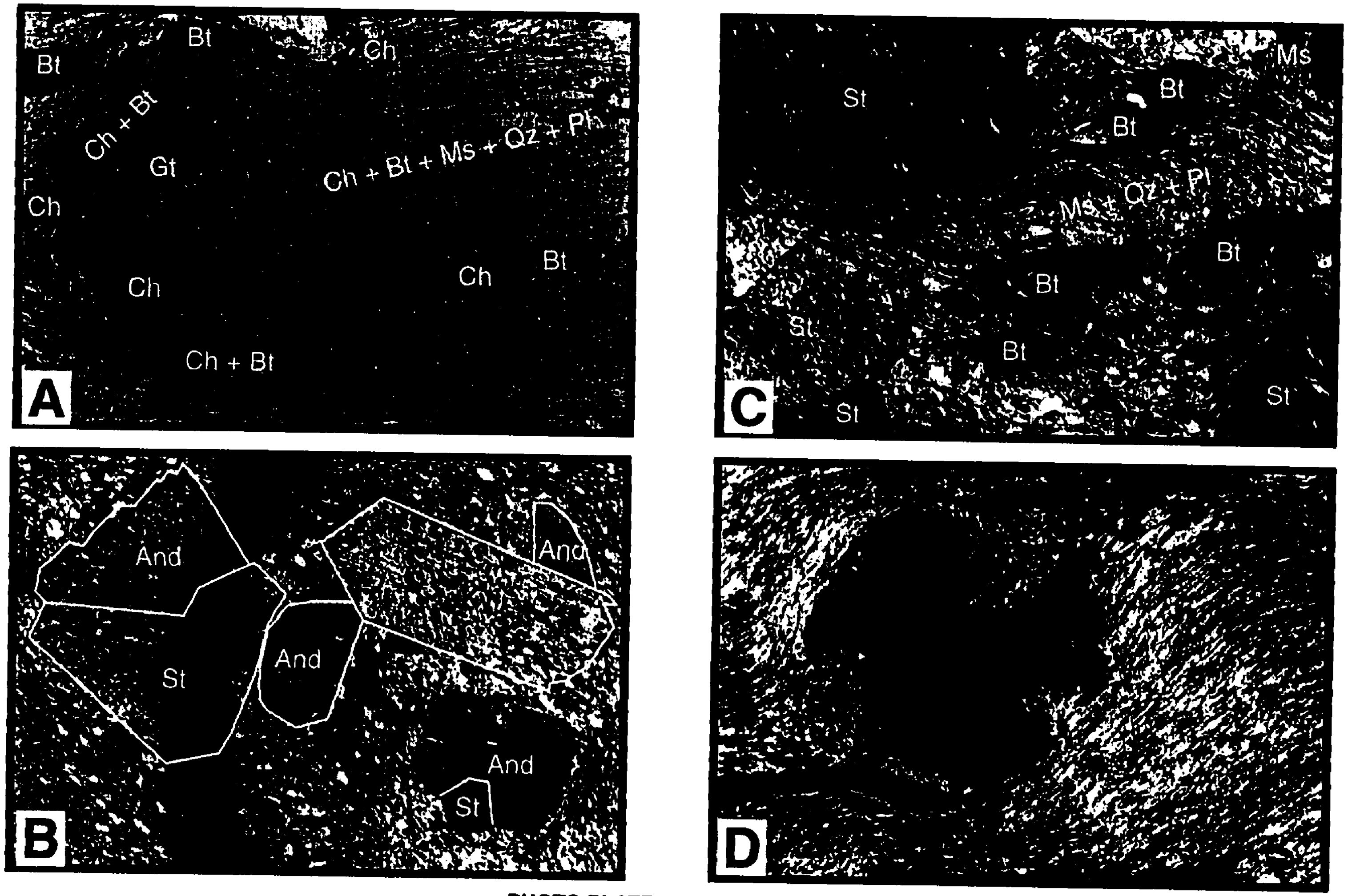



**PHOTO PLATE 7: METAMORPHISM**

**A - Specimen 34-PEL)** Plane light photomicrograph of garnet porphyroblast in regional metamorphic assemblage garnet-biotite-musovite-plagioclase-quartz-chlorite, from the hangingwall of the Kootenay Arc boundary fault. *Note that inclusions in the garnet define an older foliation* that is perpendicular to the matrix foliation. **B - Specimen 506.2)** Polarized light photomicrograph of invariant-point contact metamorphic assemblage from the aureole of the Toby stock. Andalusite, staurolite, kyanite (porphyroblasts) and biotite, muscovite, plagioclase, quartz, chlorite (matrix) indicate a pressure of 4.2 Kbars and 520°C (D. M. Carmichael, unpublished). Dark arc is mark on thin section. **C - Adjacent to specimen MK-J1)** Polarized light photomicrograph of syn-kinematic contact metamorphic assemblage from the inner part of the aureole of the Glacier Creek stock. Garnet and chlorite are also present in this rock. **D - Specimen HT-3)** Polarized light photomicrograph of rotated chloritoid porphyroblast, possibly pseudomorphed after garnet, from the outer part of the aureole of the Glacier Creek stock. A retrograde muscovite-chlorite cleavage commonly is superposed on the outer parts of the contact aureoles and on the prograde metamorphic assemblages and foliation in the immediate footwall of the Kootenay Arc boundary fault. Specimens MK-J1 and HT-3 are described also in Kells (1993). Field of view for A, B and D is 2 x 3 cm; *field of view for C is 1 x 1.5 cm.*

# VITA

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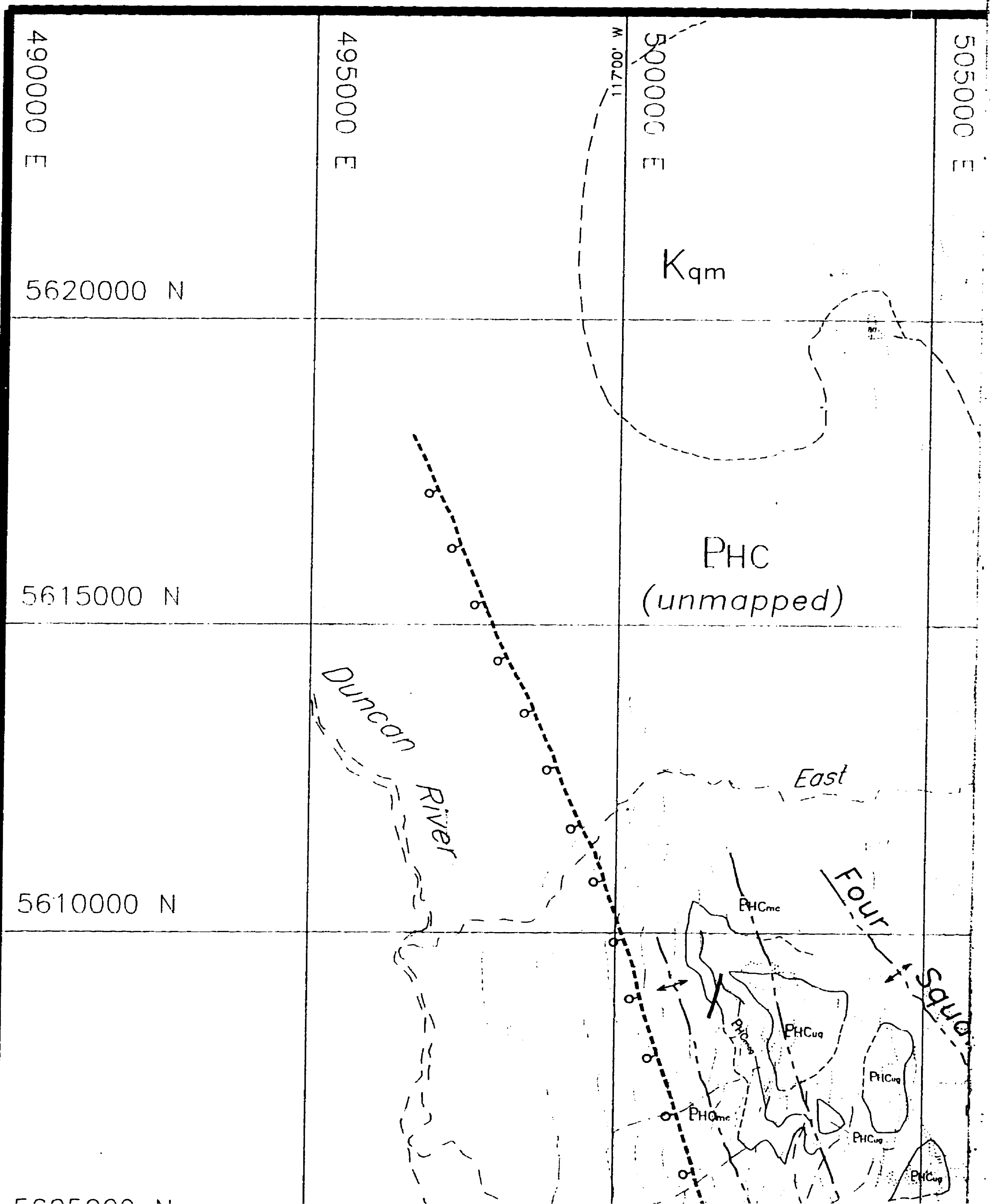
Reinhardt Scholarships, Queen's University

Carmichael Fellowship, Queen's University

Elected to membership in Sigma Xi (1986)

Most improved in women's geo-hockey (1993): *Most significant accomplishment of my Canadian graduate career*



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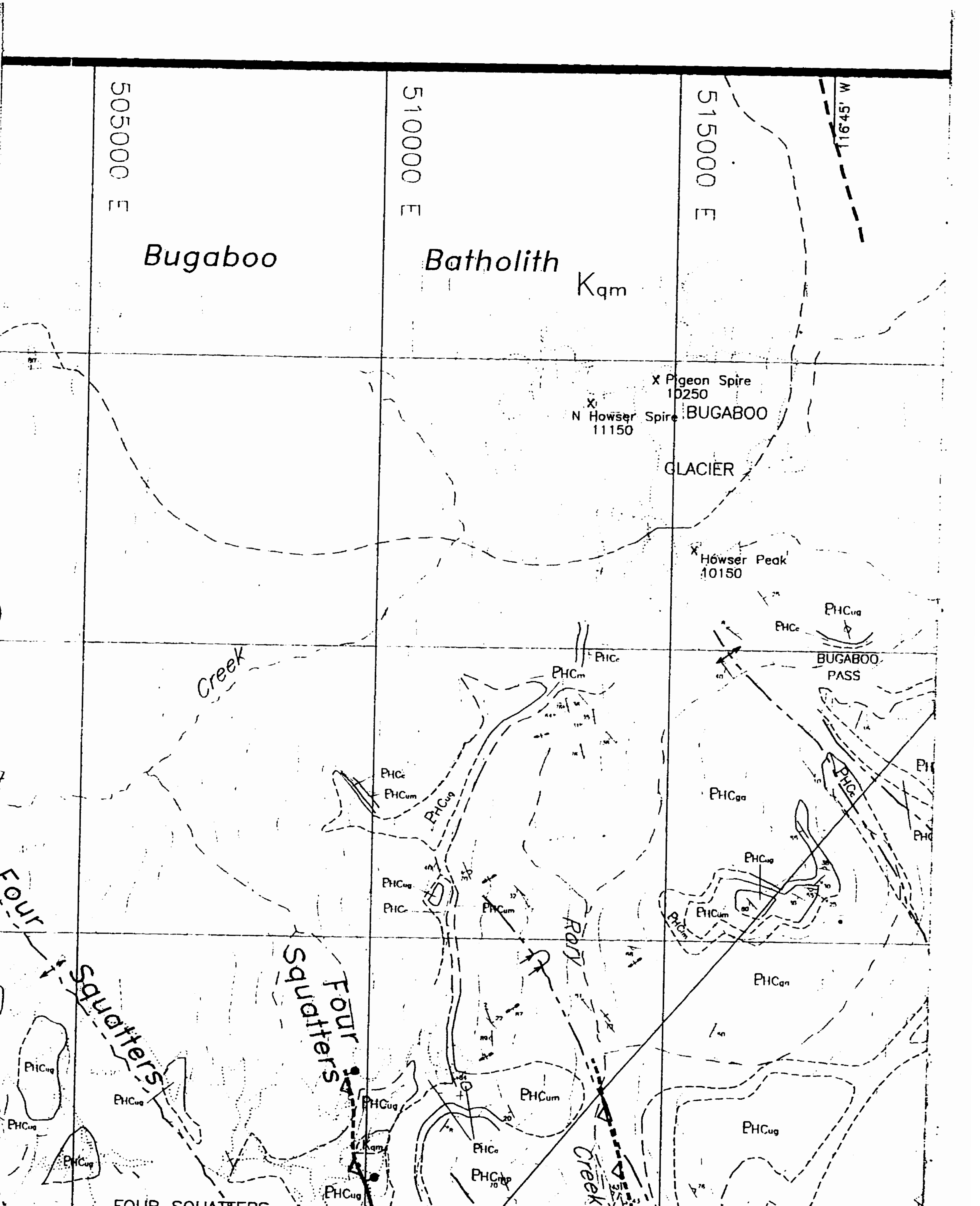
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Bugaboo Batholith
Kqm
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BUGABOO
GLACIER
Howser Peak
10150
BUGABOO
PASS
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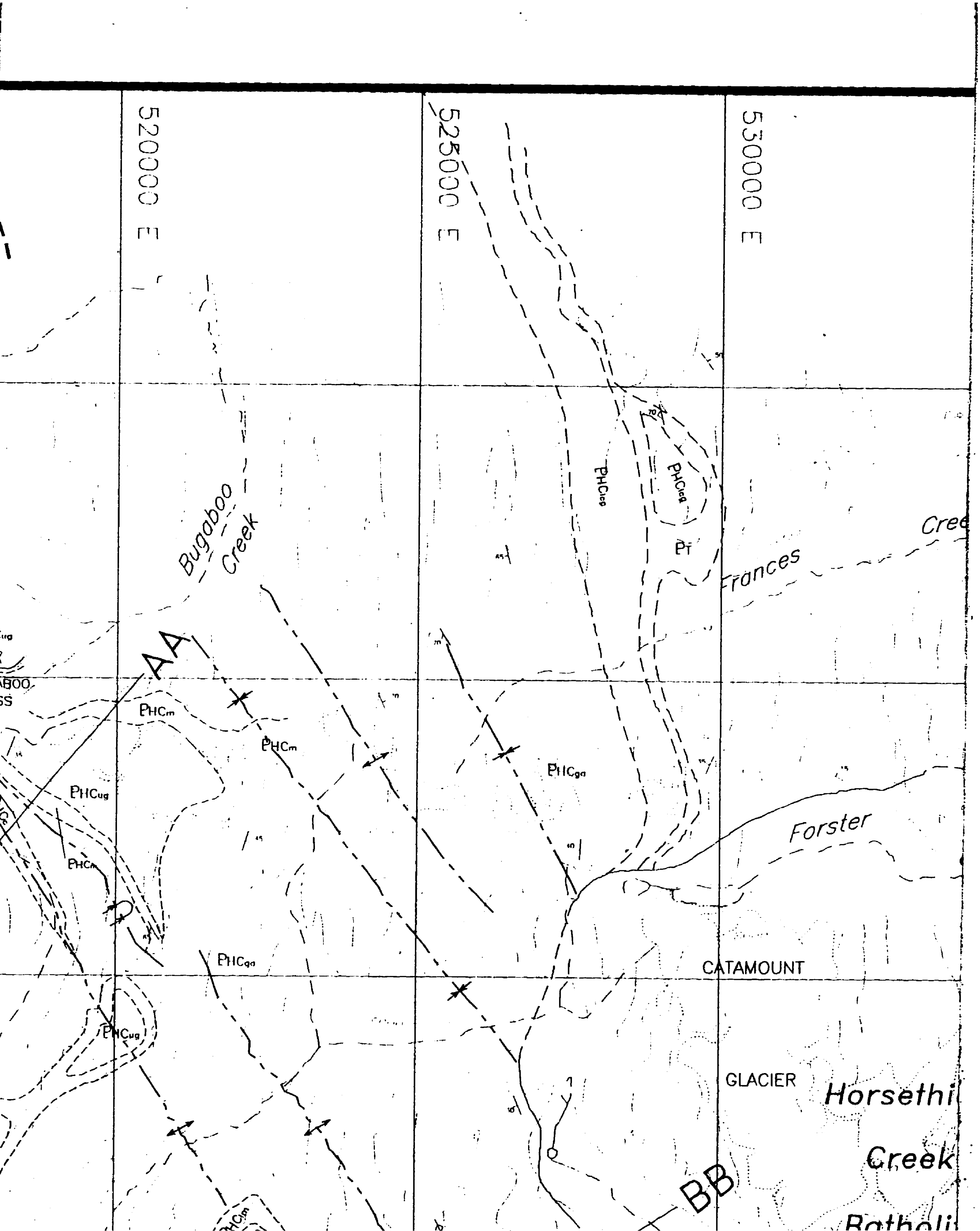

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CATAMOUNT
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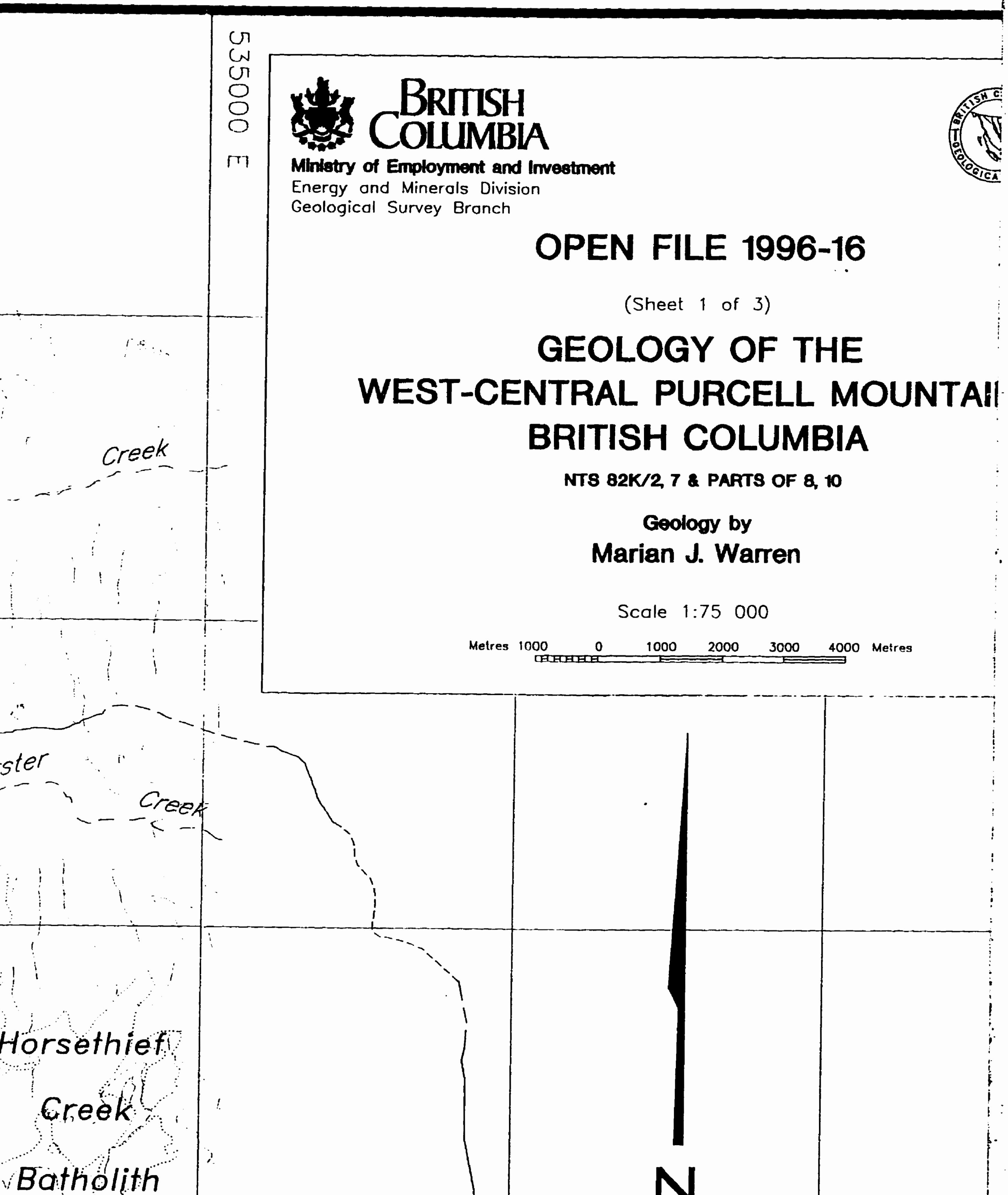
Horsethi
Creek
BB



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Energy and Minerals Division
Geological Survey Branch
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(Sheet 1 of 3)
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WEST-CENTRAL PURCELL MOUNTAI
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NTS 82K/2, 7 & PARTS OF 8, 10
Geology by
Marian J. Warren
Scale 1:75 000
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**Ministry of Employment and Investment**
Energy and Minerals Division
Geological Survey Branch

# OPEN FILE 1996-16

(Sheet 1 of 3)

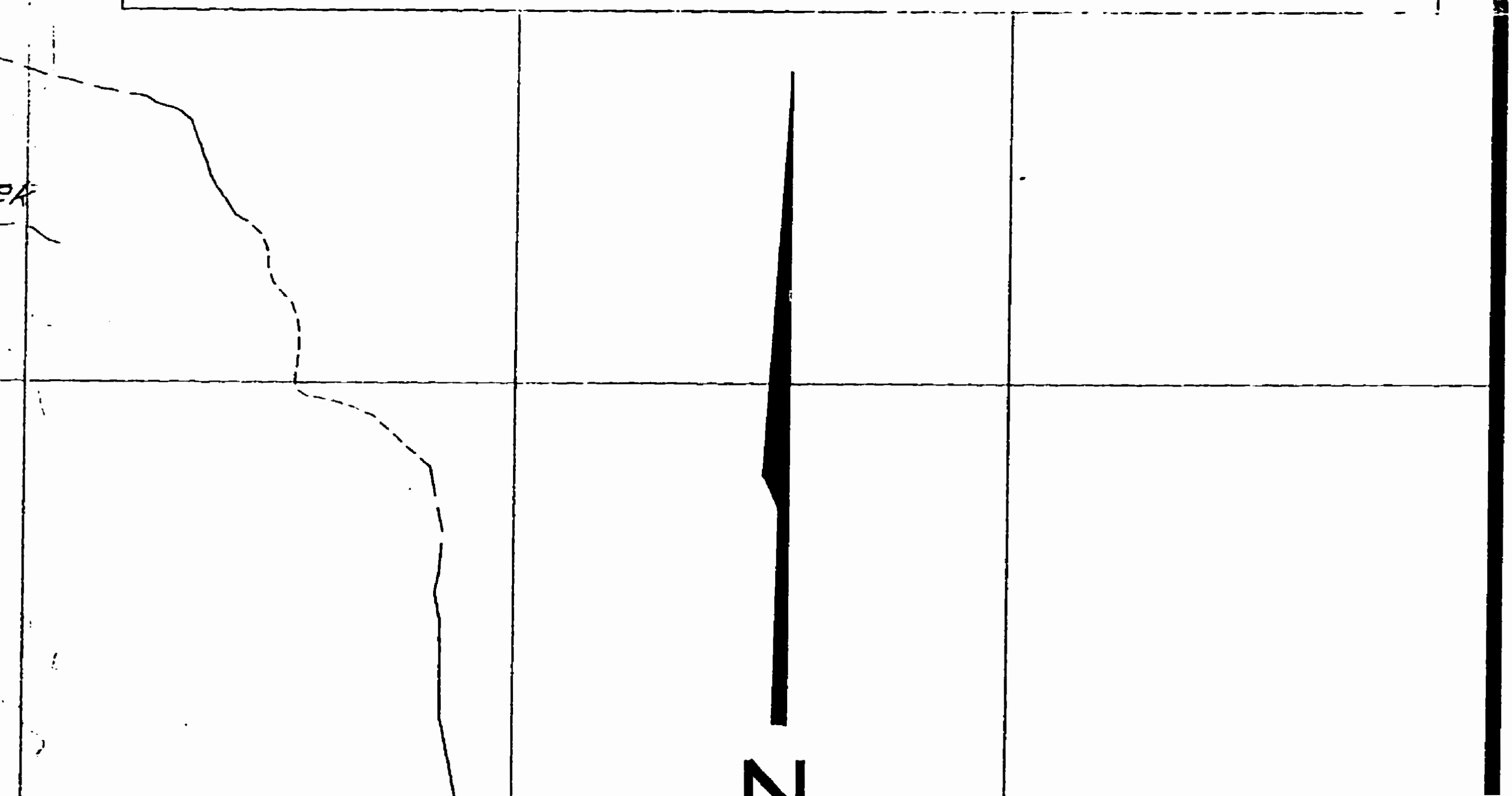
# GEOLOGY OF THE WEST-CENTRAL PURCELL MOUNTAINS BRITISH COLUMBIA

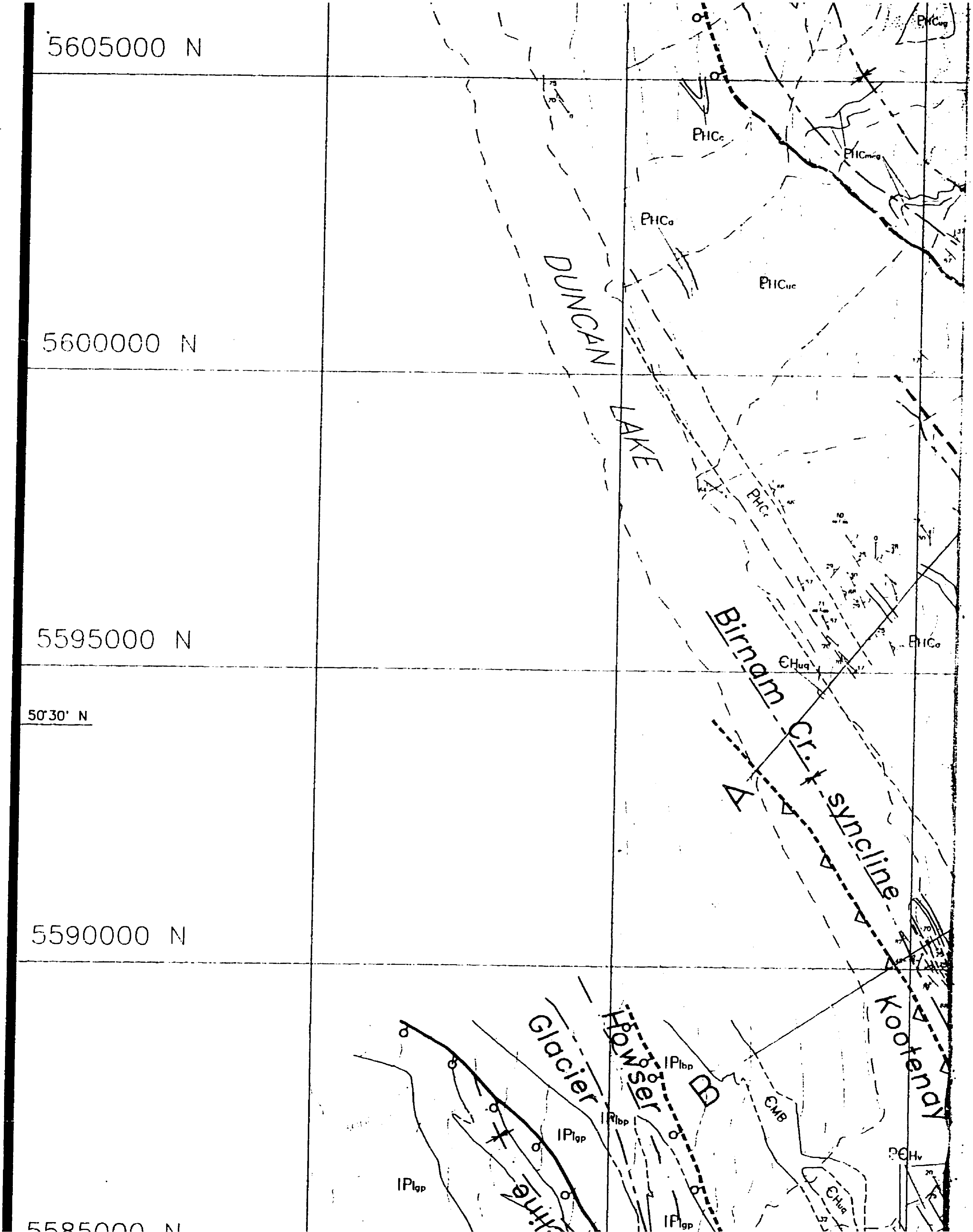
**NTS 82K/2, 7 & PARTS OF 8, 10**

**Geology by**
**Marian J. Warren**

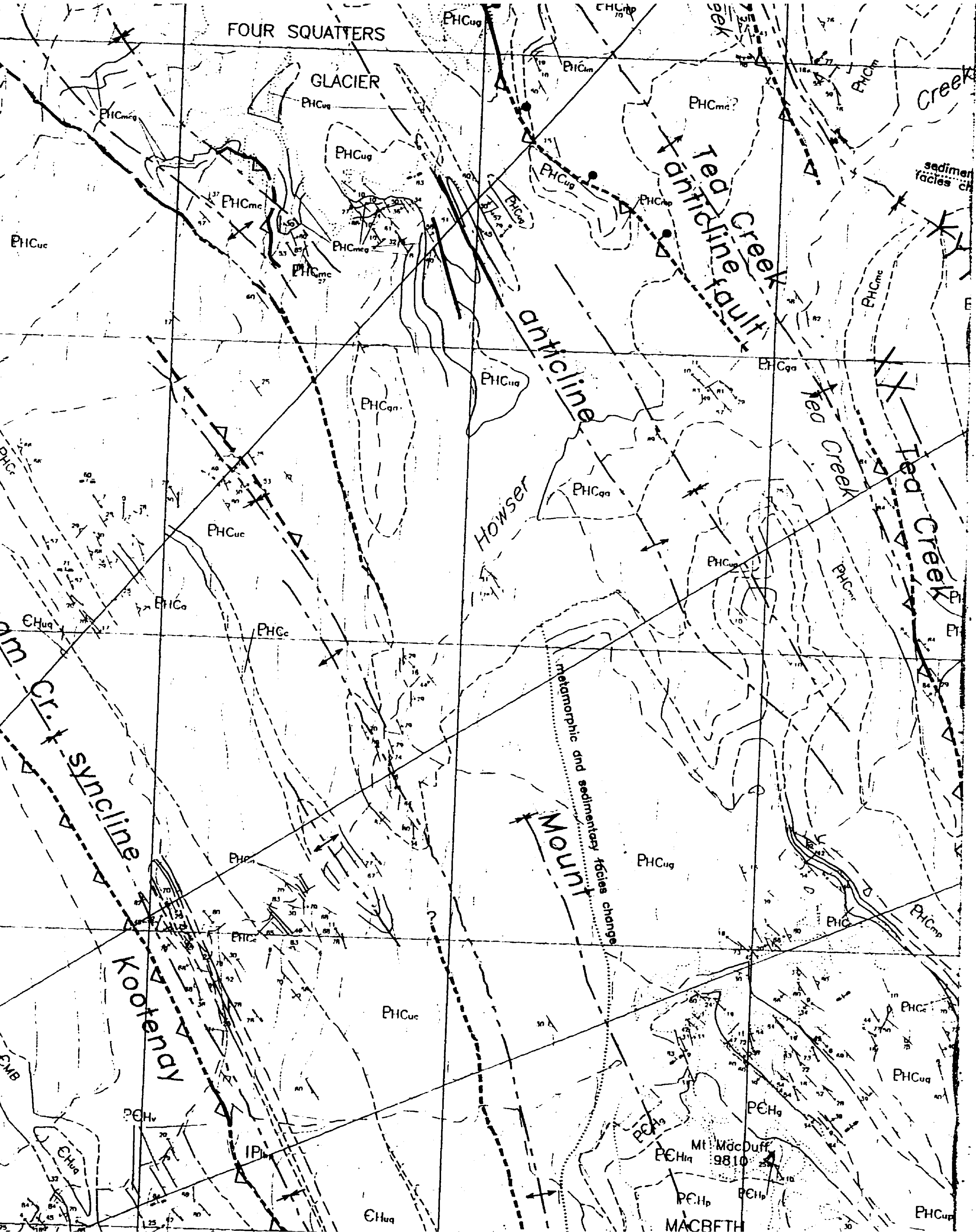
Scale 1:75 000

Metres 1000 0 1000 2000 3000 4000 Metres

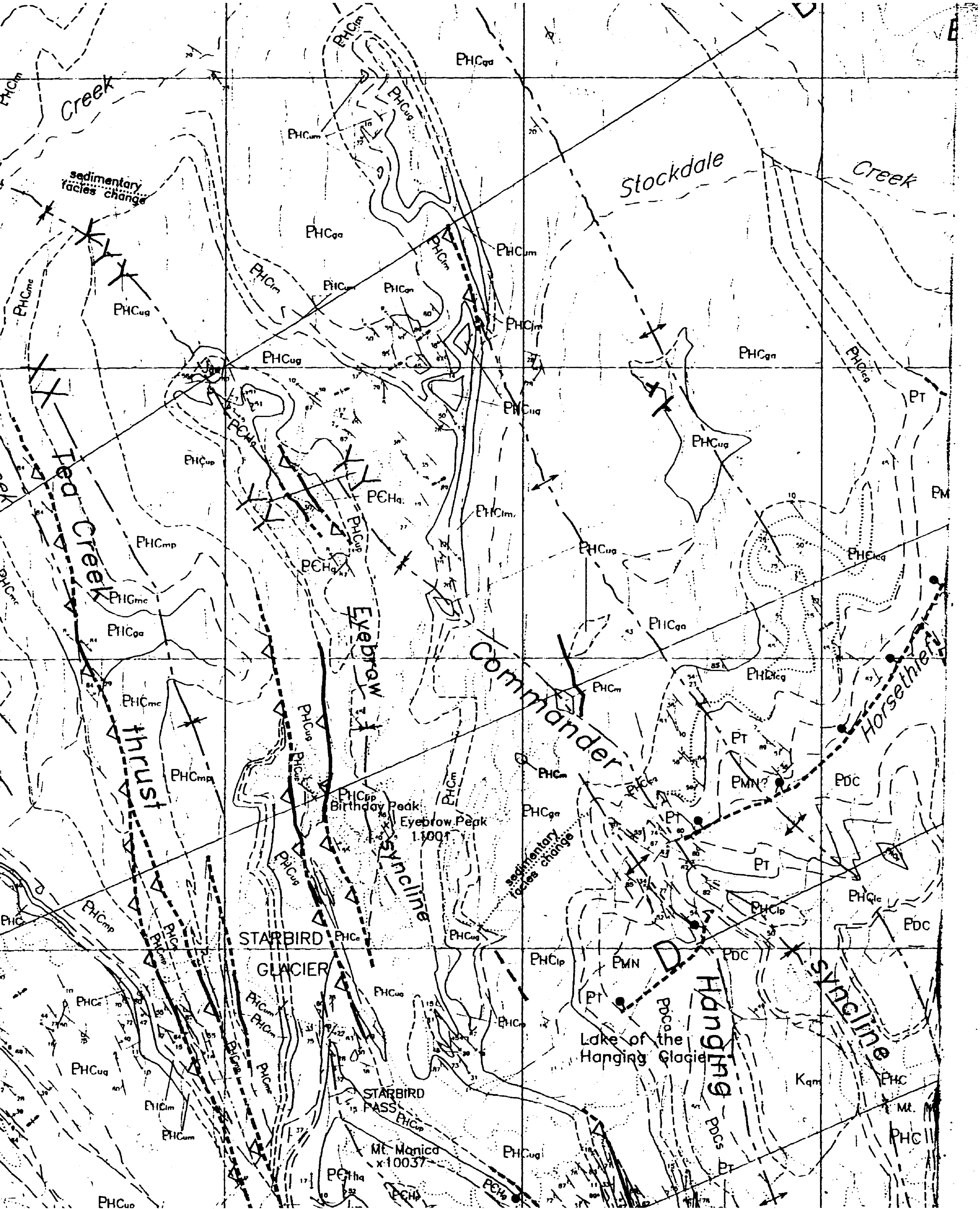


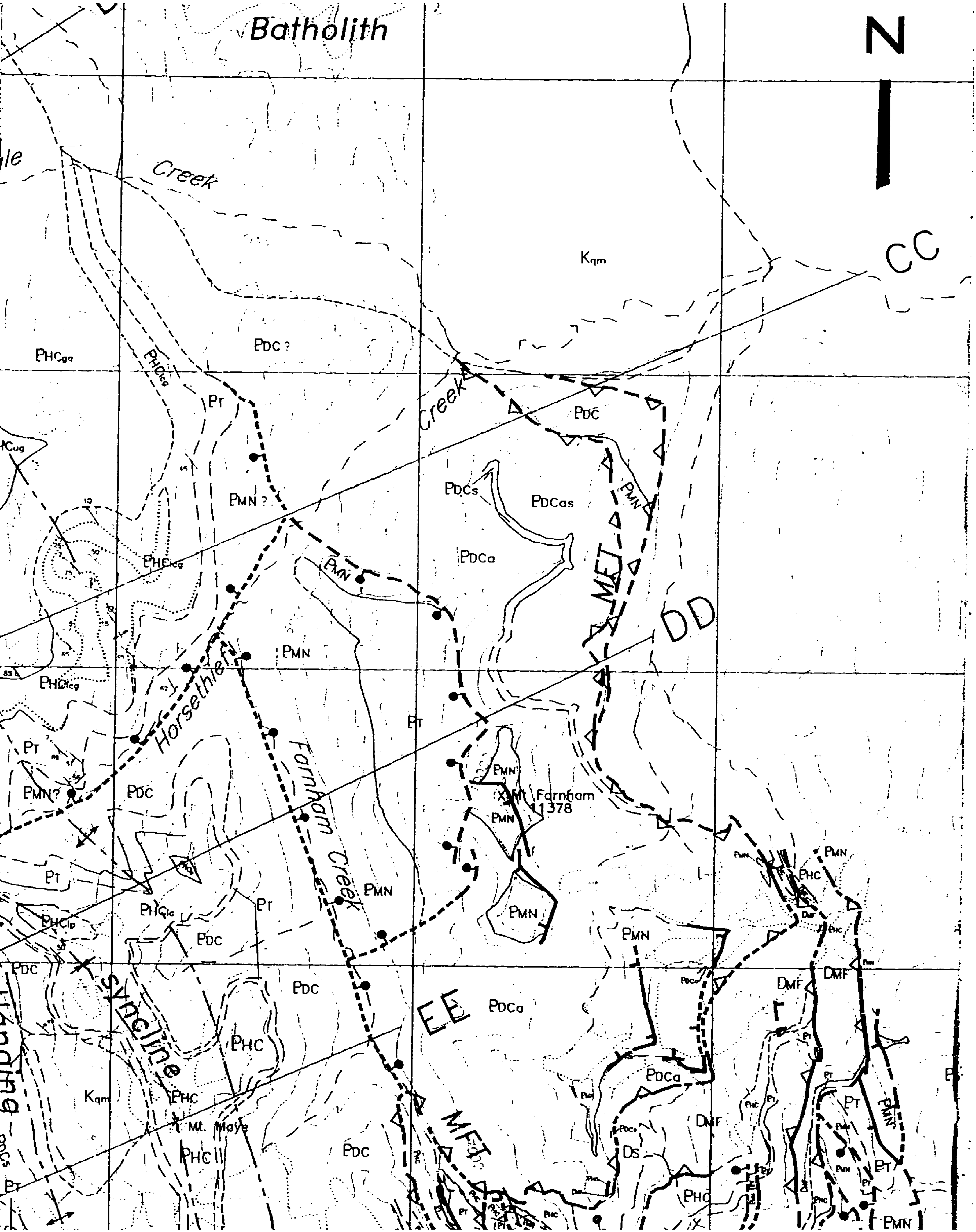


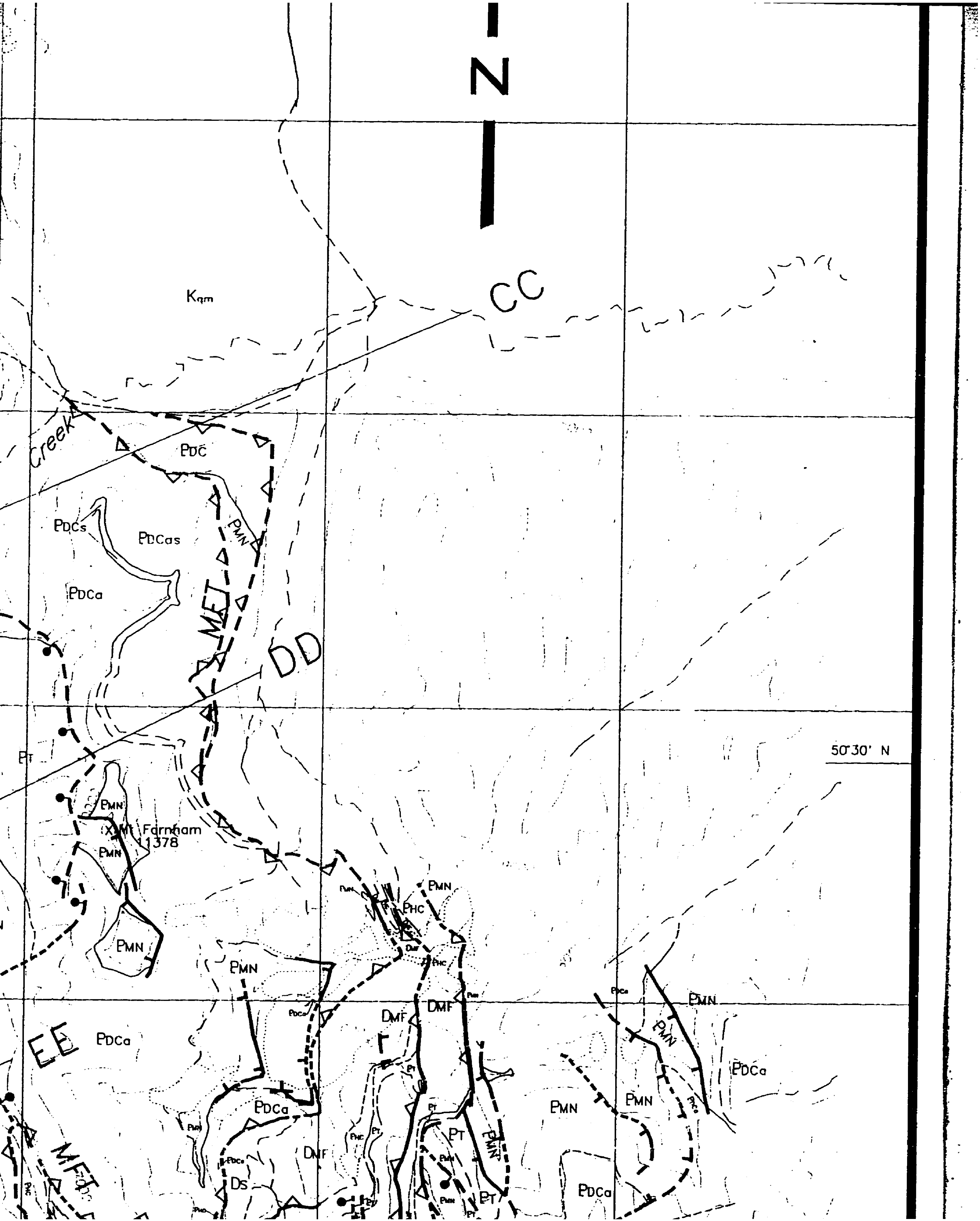
5605000 N
5600000 N
5595000 N
50°30' N
5590000 N
DUNCAN LAKE
Birnam Cr. syncline
A
B
Kootenay
Glacier
Howser
PHCa
PHCuc
EHuq
IPIgp
IPIbp

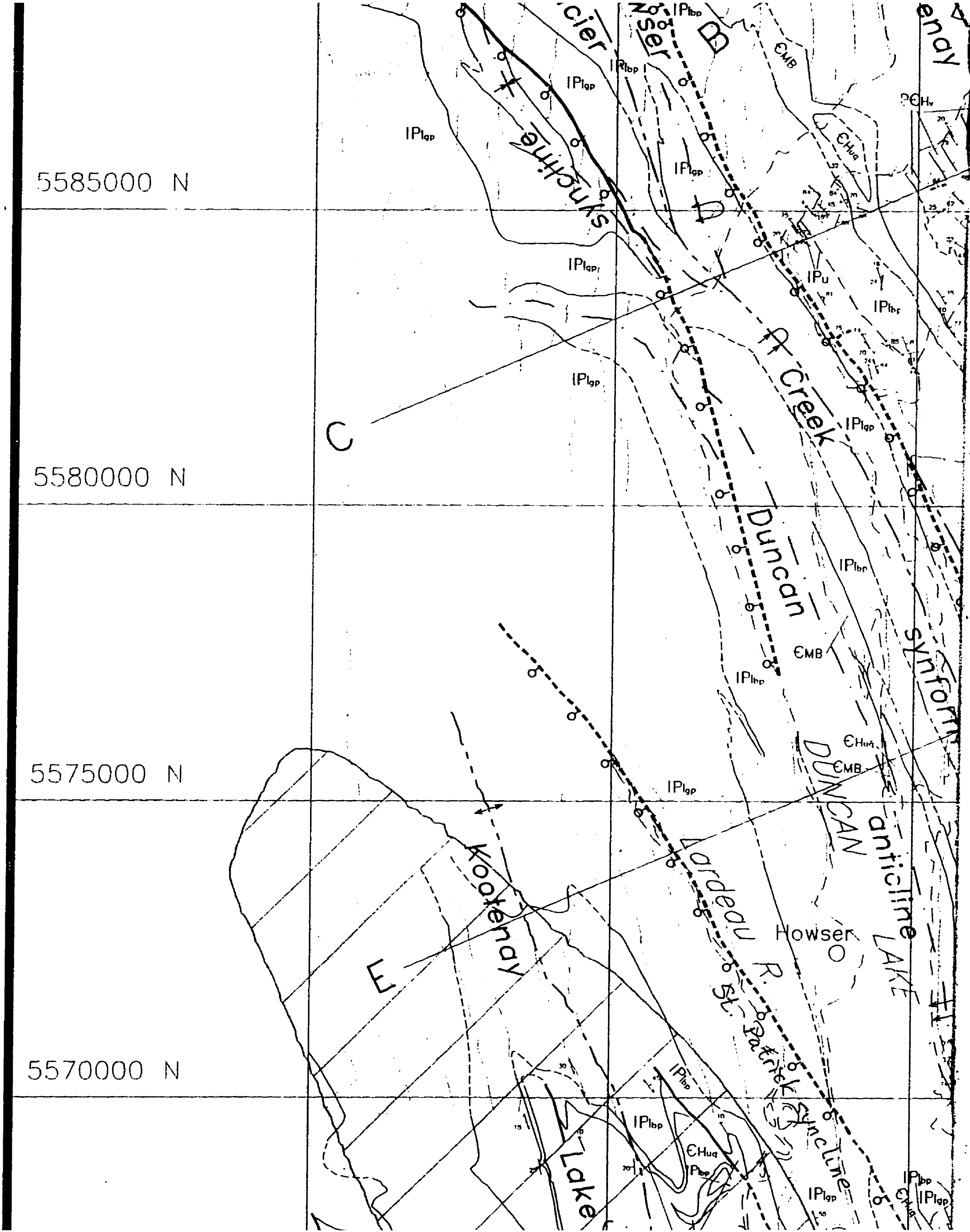


FOUR SQUATTERS
GLACIER
PHCug
PHCmcg
PHCmc
PHCuc
PHCga
PHCmp
PHCc
Ted Creek anticline fault
anticline
Ted Creek
Howser
ЄHuq
metamorphic and sedimentary facies change
Mount
Cr. syncline
Kootenay
PЄHv
IPlq
PЄHg
PЄHlq
PЄHp
Mt. MacDuff
9810
PHCup

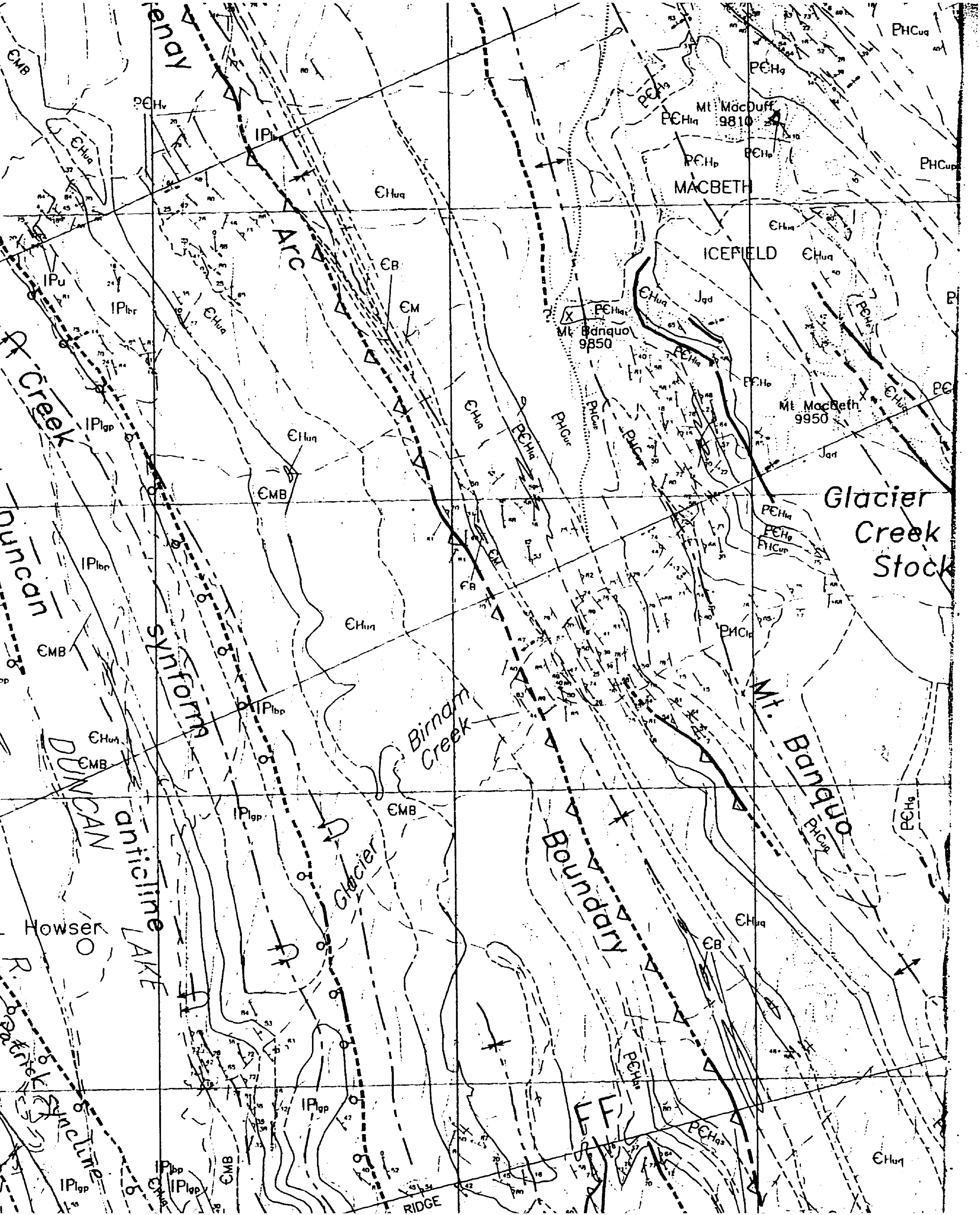

Creek
Stockdale Creek
sedimentary facies change
Ted Creek thrust
Eyebrow syncline
Commander
Horsethief
Hanging syncline
Birthday Peak
Eyebrow Peak
11,001
STARBIRD GLACIER
STARBIRD PASS
Mt. Manica
x 10037
Lake of the Hanging Glacier


Batholith
N
Creek
CC
Kqm
PHCgn
PDC ?
PT
Creek
PDC
PMN ?
PDCs
PDCas
PMN
PDCa
MFT
DD
PMN
PHCicg
Horsethief
Farnham Creek
PT
PMN
X Mt Farnham
11378
PMN
PT
PMN?
PDC
PMN
PMN
PHC
PHClc
PT
PDC
PHClp
PDC
PDC
PDCa
EE
DMF
DMF
syncline
PHC
PHC
PDCa
Kqm
Mt. Maye
PHC
PDC
MFT
DMF
Ds
PT
PMN
PT
PHC
PMN

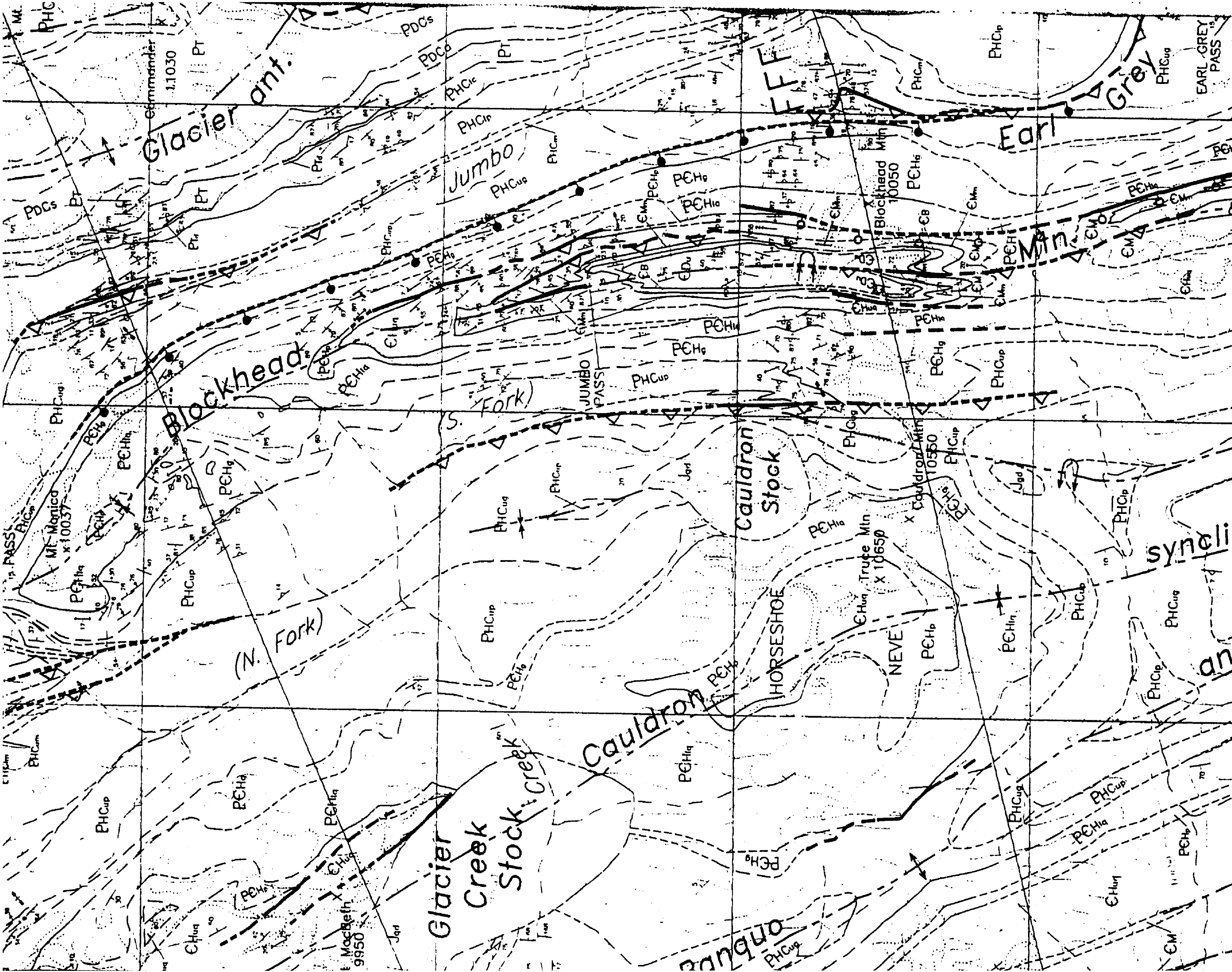

N
Kqm
CC
Creek
PDC
PDCs
PDCas
PDCa
PMN
MFT
DD
PT
Mt Farnham
11378
50°30' N
PHC
DMF
EE
Ds

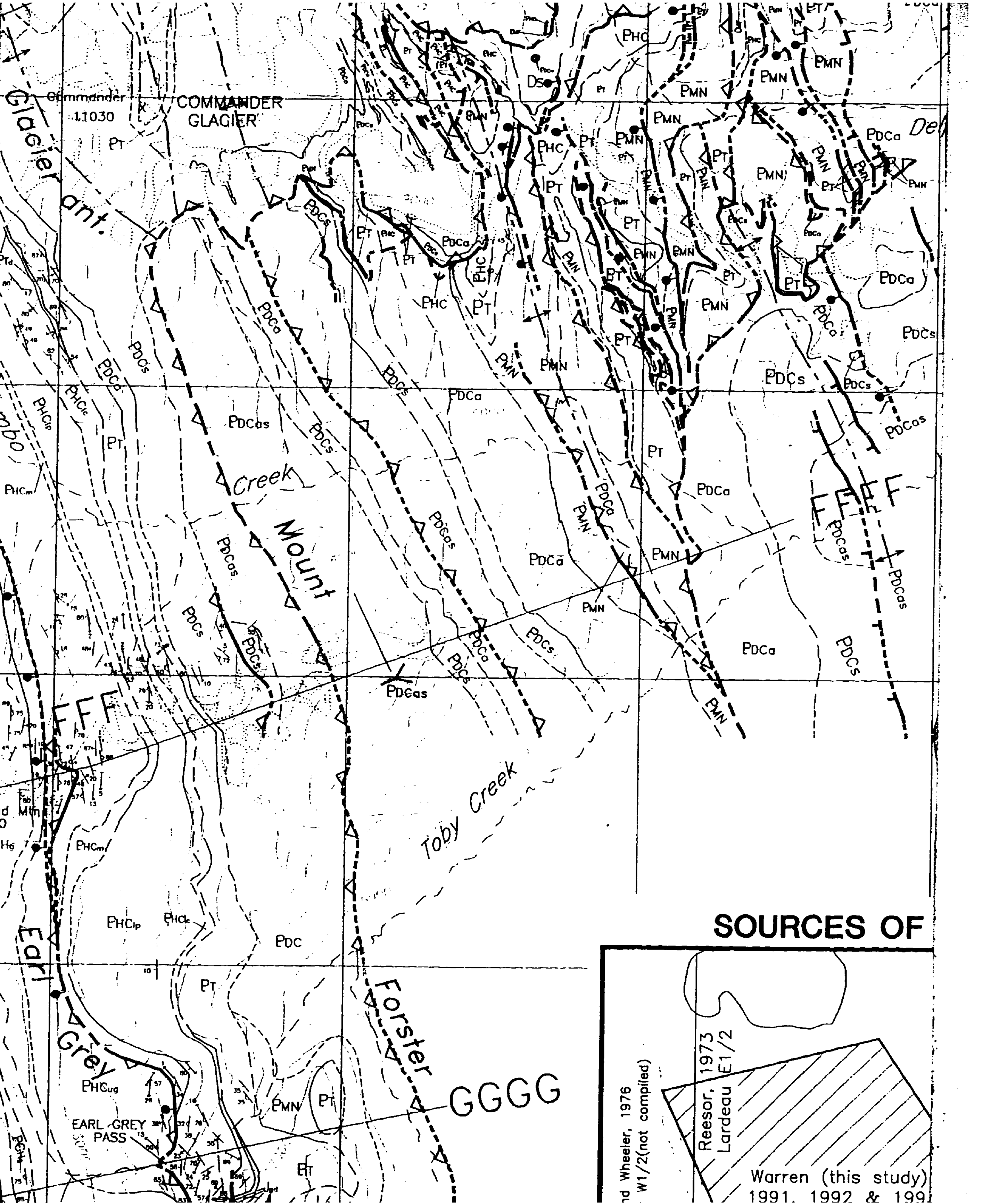


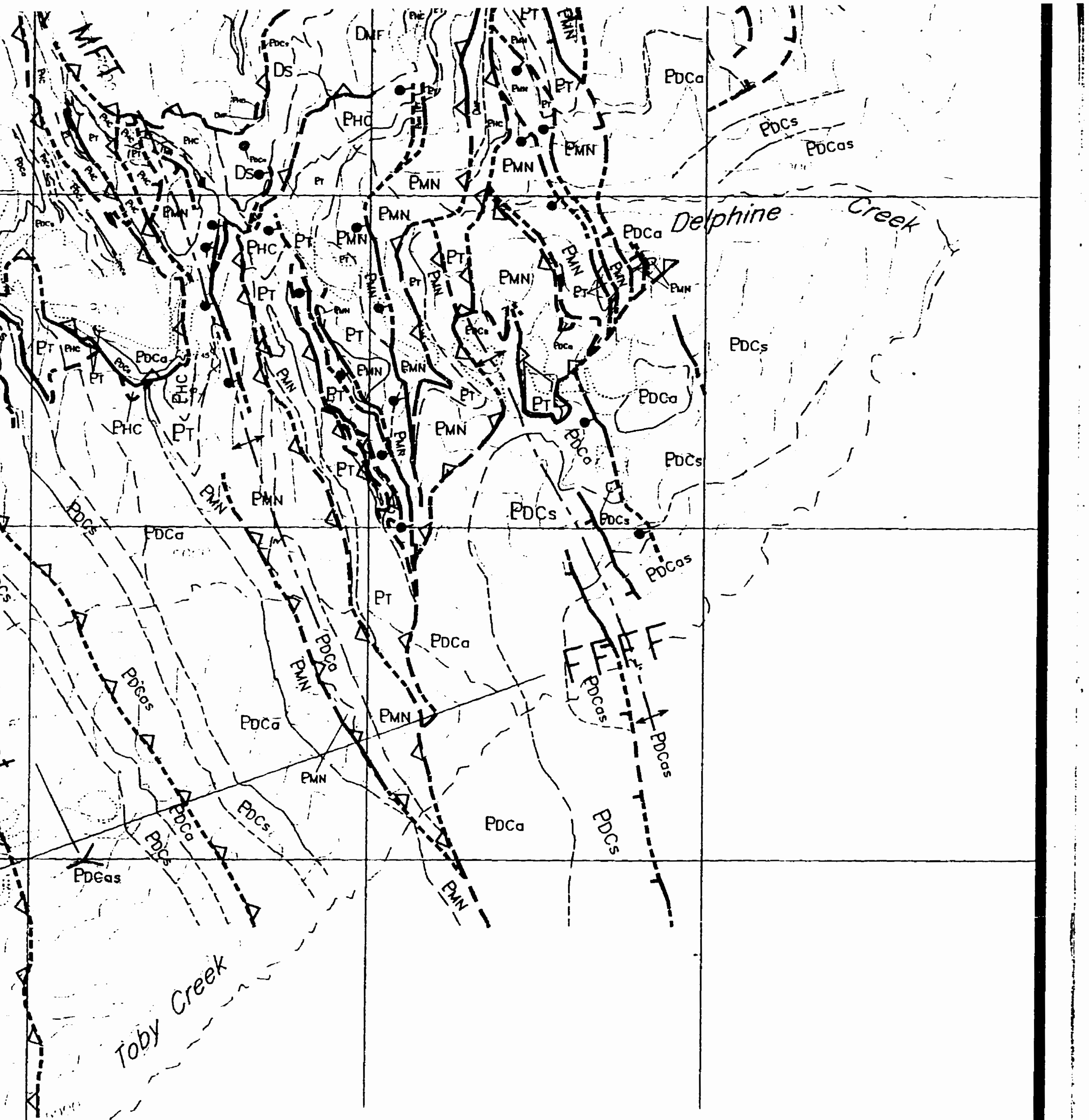
5585000 N
5580000 N
5575000 N
5570000 N
C
E
Duncan
Creek
synform
anticline
Howser
DUNCAN
LAKE
Kootenay
Lake
Lardeau R.
St. Patrick Syncline



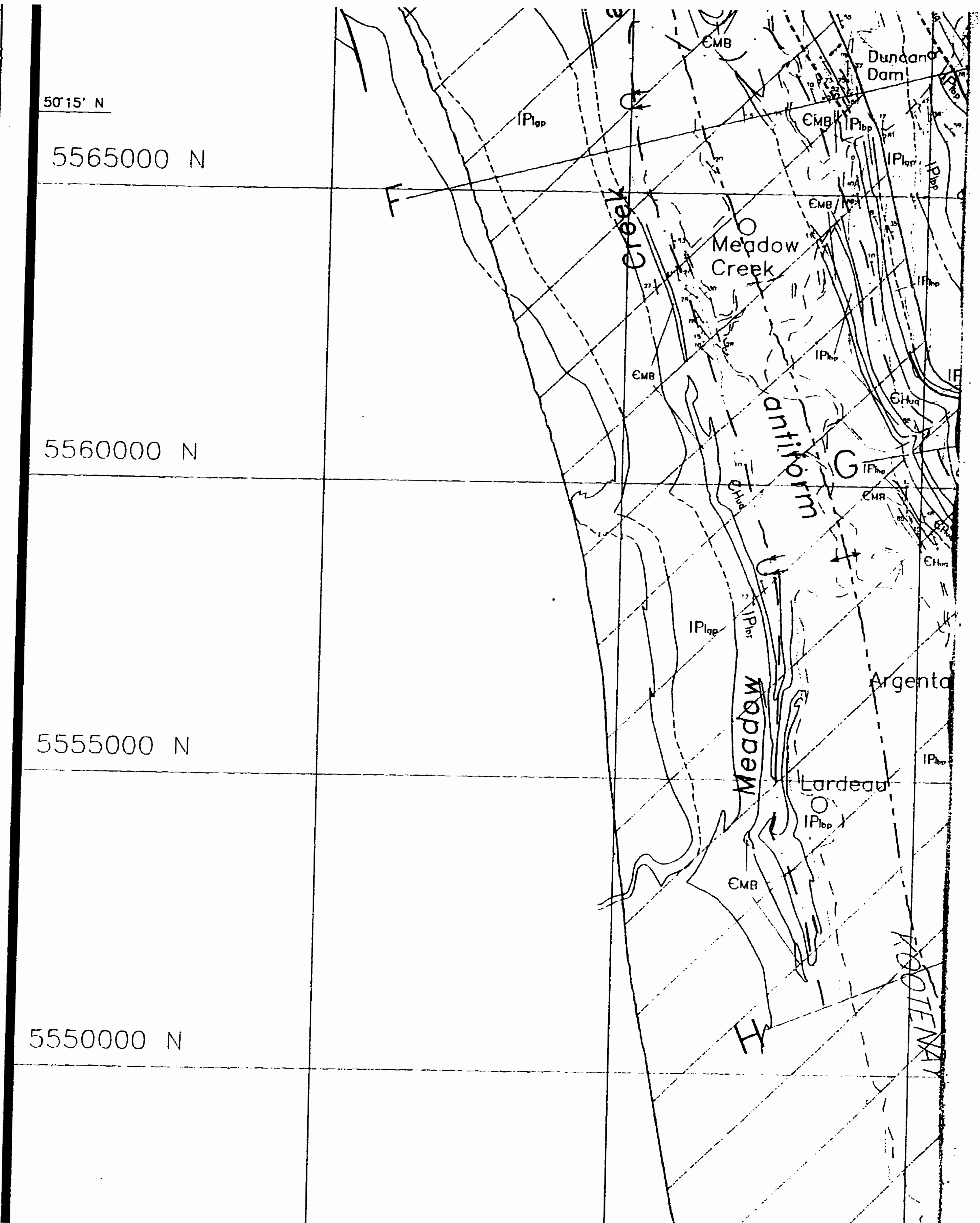
Mt. MacDuff
9810
MACBETH
ICEFIELD
Mt. Banquo
9850
Mt. MacBeth
9950
Glacier
Creek
Stock
Arc
Birnam
Creek
Mt. Banquo
Boundary
Duncan
synform
DUNCAN
LAKE
anticline
Howser
Glacier
Creek
FF
RIDGE


Glacier ant.
Commander
11030
Jumbo
Blockhead
Earl
Grey
Mtn
EARL GREY PASS
Blockhead Mtn
10050
JUMBO PASS
(S. Fork)
(N. Fork)
Mt. Monica
x 10037
PASS
Cauldron Stock
x Cauldron Mtn
10560
Truce Mtn
x 10650
HORSESHOE
NEVE
syncli
an
Cauldron
Glacier Creek Stock
Creek
MacBeth
9950
Banquo

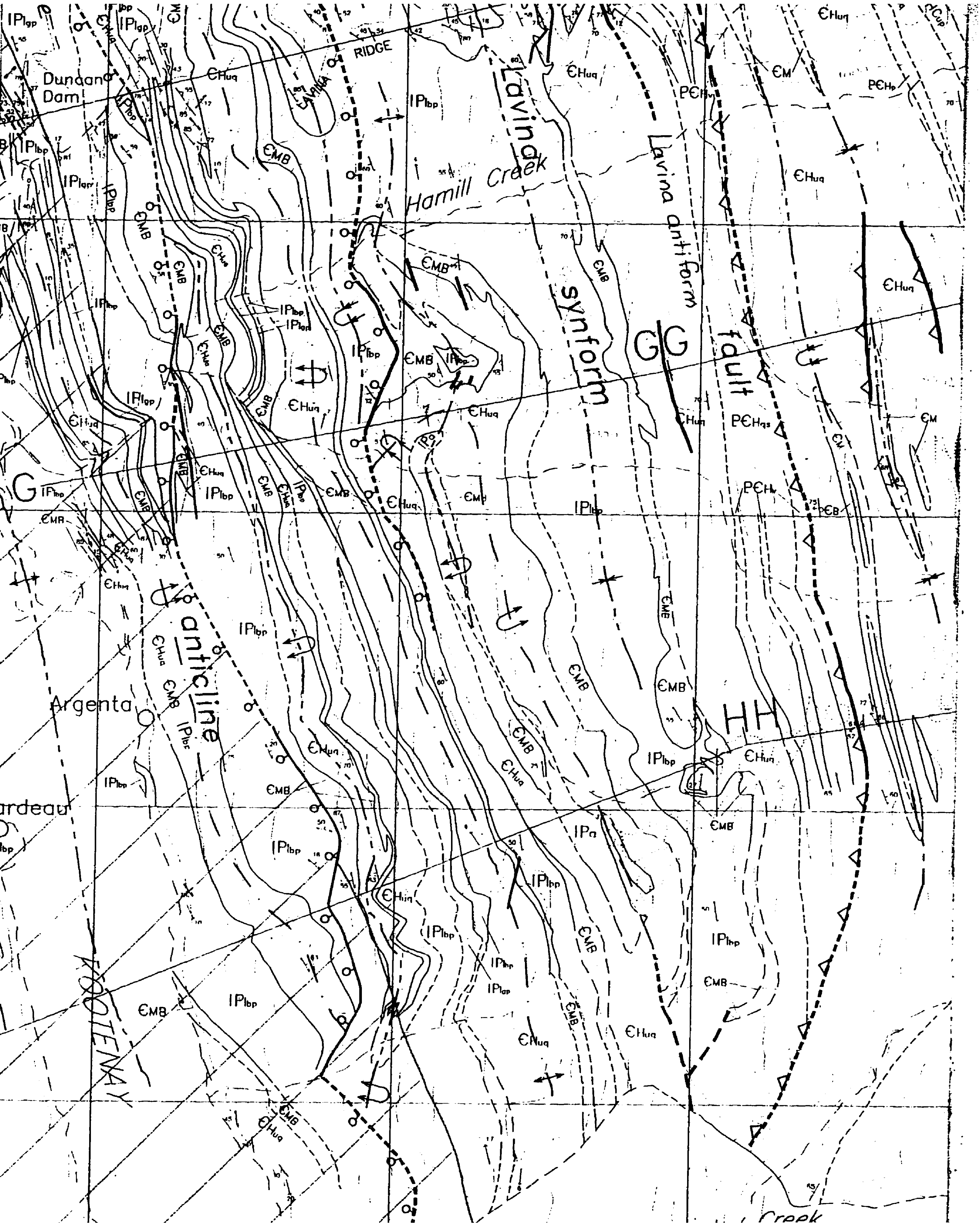

Commander
11030
COMMANDER
GLACIER
Glacier ant.
Creek
Mount
Forster
Toby Creek
Earl
Grey
EARL GREY
PASS
FFFF
FFF
GGGG
PDCa
PDCs
PDCas
PDC
PMN
PT
PHC
PHCm
PHCip
PHCug
Ds
SOURCES OF
Reesor, 1973
Lardeau E1/2
nd Wheeler, 1976
W1/2(not compiled)
Warren (this study)
1991, 1992 & 199



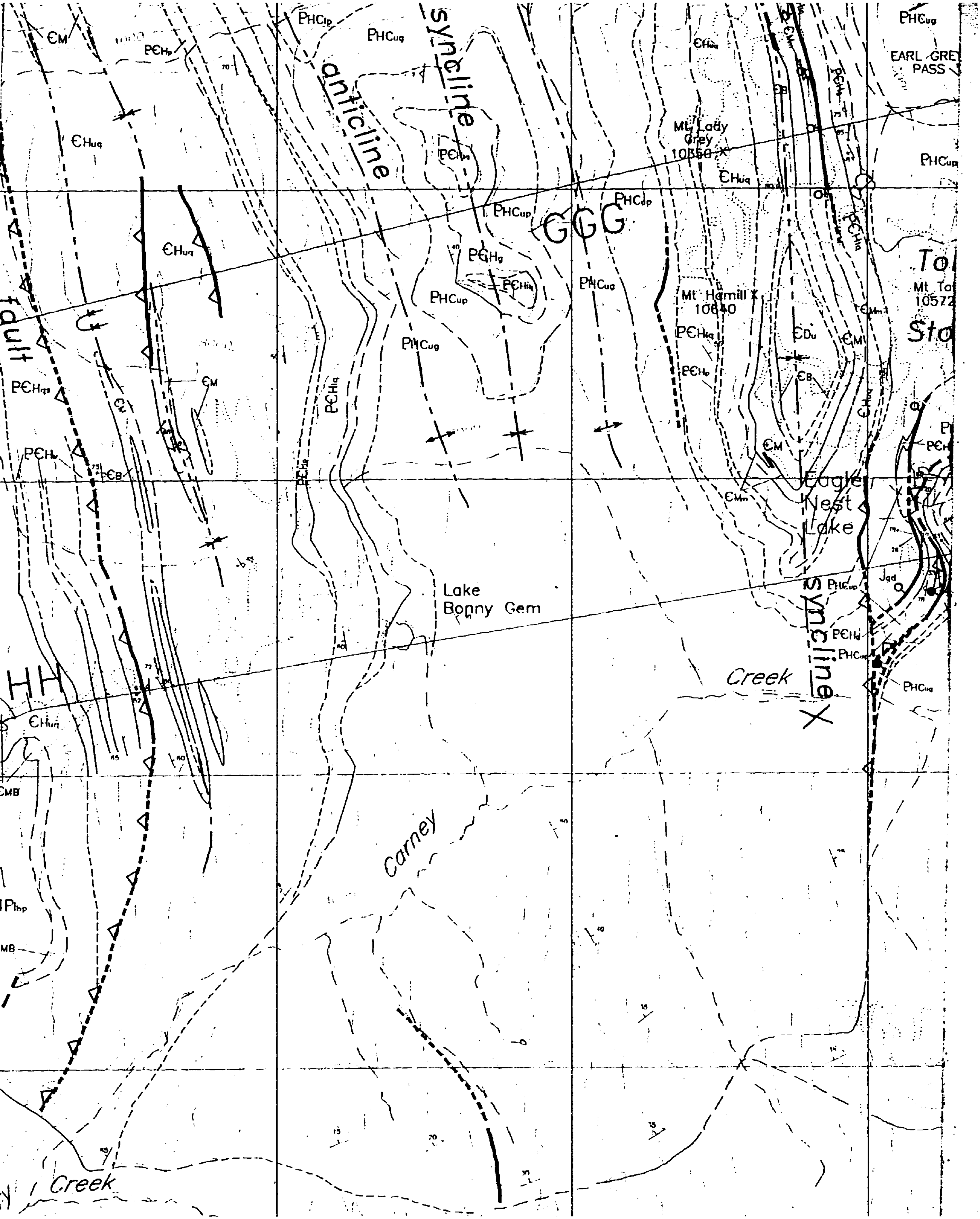

## SOURCES OF DATA



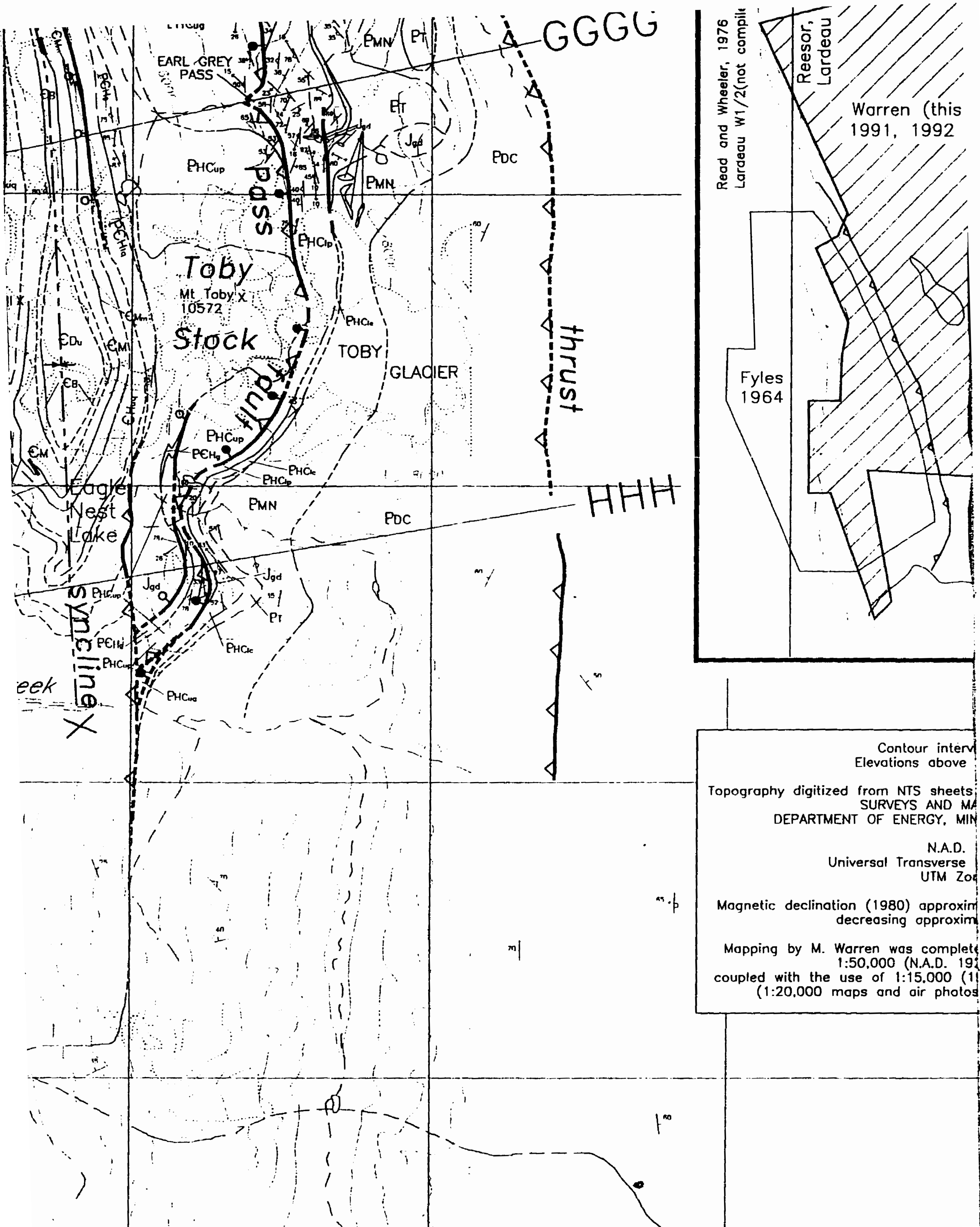
50°15' N
5565000 N
5560000 N
5555000 N
5550000 N
F
G
H
Meadow Creek
Meadow Creek antiform
Duncan Dam
Argenta
Lardeau
KOOTENAY
ЄMB
IPlgp
IPlbp
ЄHuq

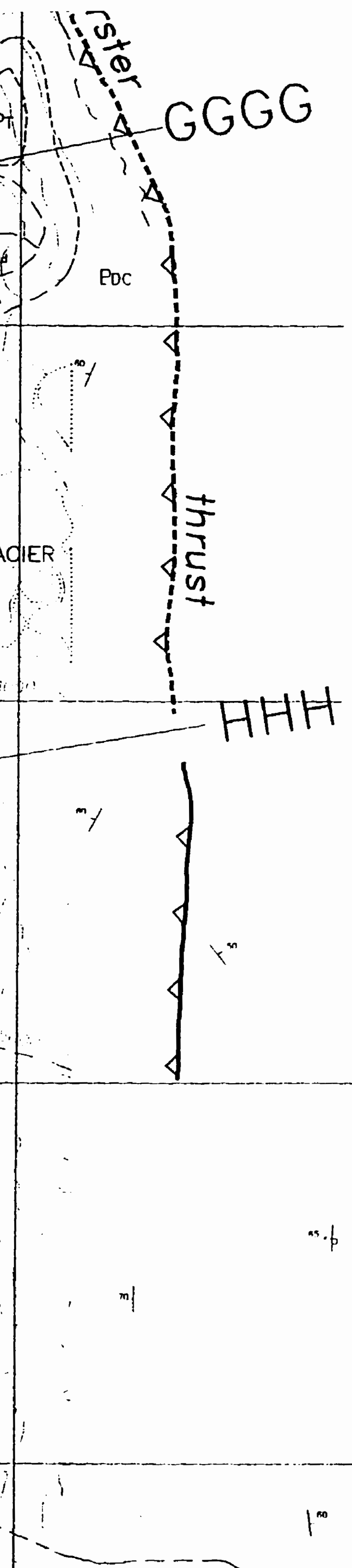


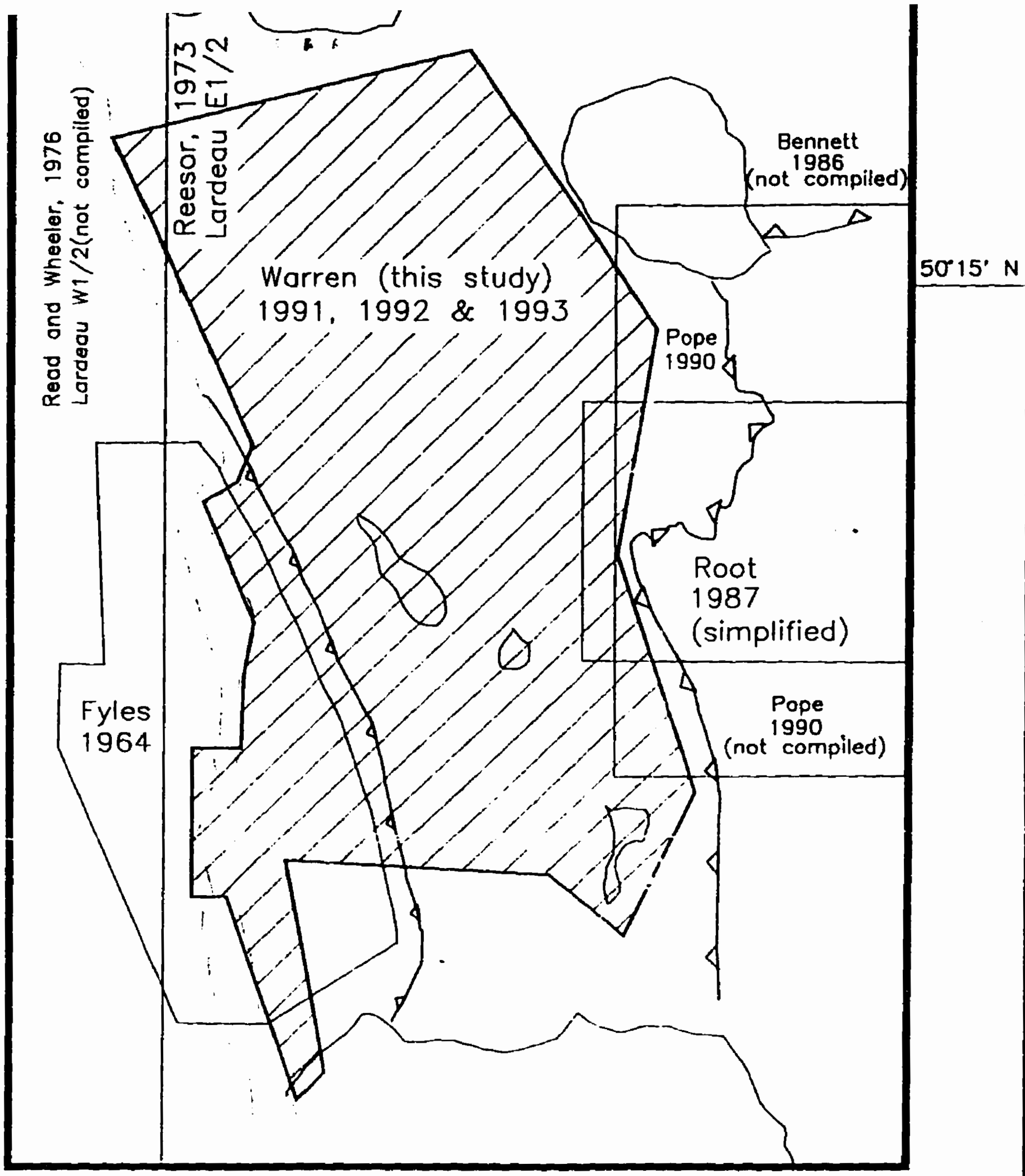
Duncan Dam
RIDGE
Lavina
Hamill Creek
Lavina synform
Lavina antiform
fault
GG
G
HH
anticline
Argenta
KOOTENAY
Creek



syncline
anticline
fault
GGG
HH
EARL GREY PASS
Mt Lady Grey 10350
Mt Hamill 10640
Mt To
10572
Eagle Nest Lake
Lake Bonny Gem
syncline
Creek
Carney
Creek


GGGG
EARL GREY PASS
Pass
Toby
Mt Toby X 10572
Stock
fault
TOBY
GLACIER
thrust
HHH
Eagle Nest Lake
syncline
eek
Read and Wheeler, 1976
Lardeau W1/2(not compil
Reesor, Lardeau
Warren (this 1991, 1992
Fyles 1964
Contour interv
Elevations above
Topography digitized from NTS sheets
SURVEYS AND MA
DEPARTMENT OF ENERGY, MIN
N.A.D.
Universal Transverse
UTM Zo
Magnetic declination (1980) approxim
decreasing approxim
Mapping by M. Warren was complet
1:50,000 (N.A.D. 19
coupled with the use of 1:15,000 (1
(1:20,000 maps and air photos



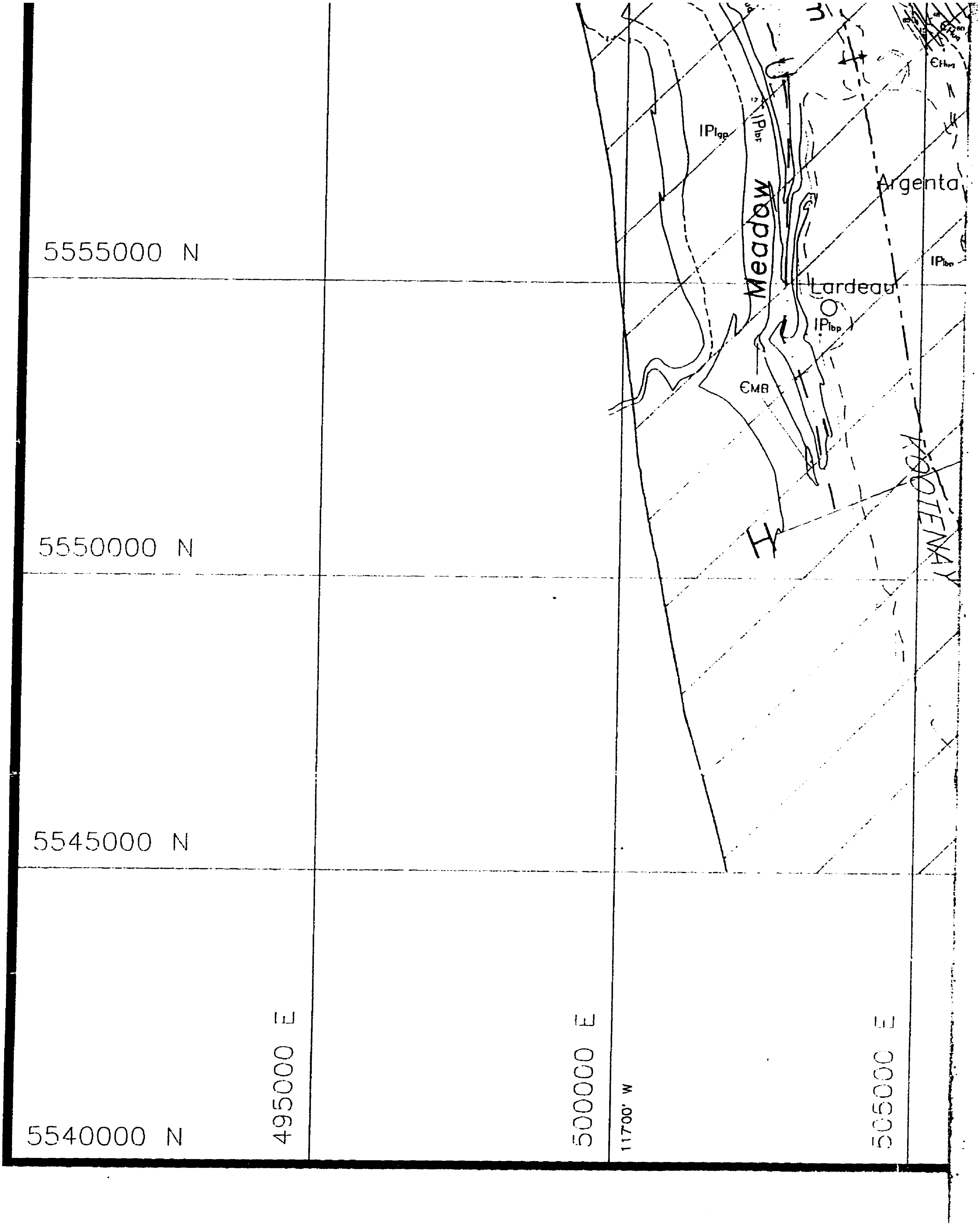

Contour interval: 1000 feet
Elevations above mean sea level

Topography digitized from NTS sheets 82K/2, 7, 8, 9 and 10 (1:50,000)
SURVEYS AND MAPPING BRANCH
DEPARTMENT OF ENERGY, MINES AND RESOURCES CANADA

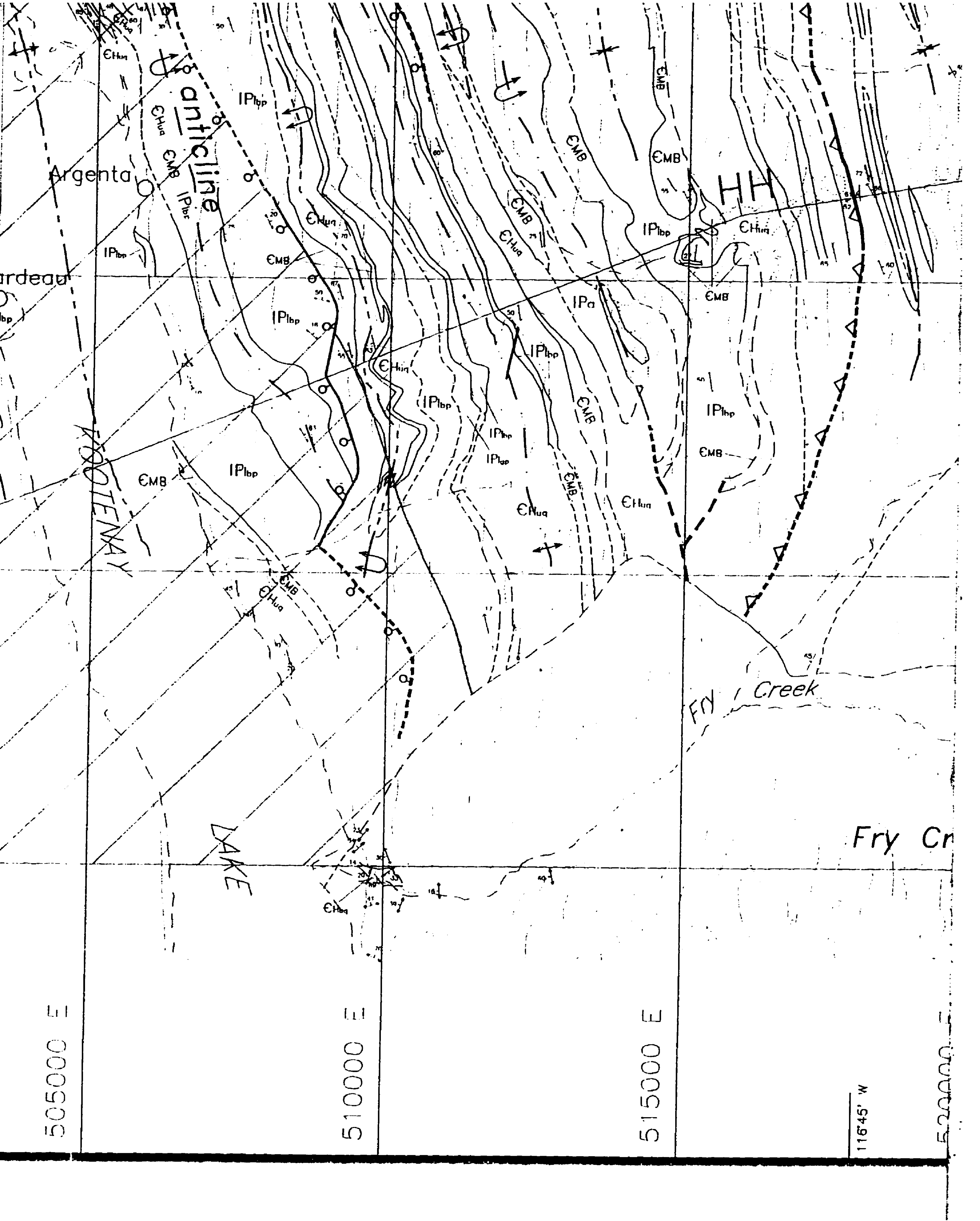
N.A.D. 1927
Universal Transverse Mercator Projection
UTM Zone 11

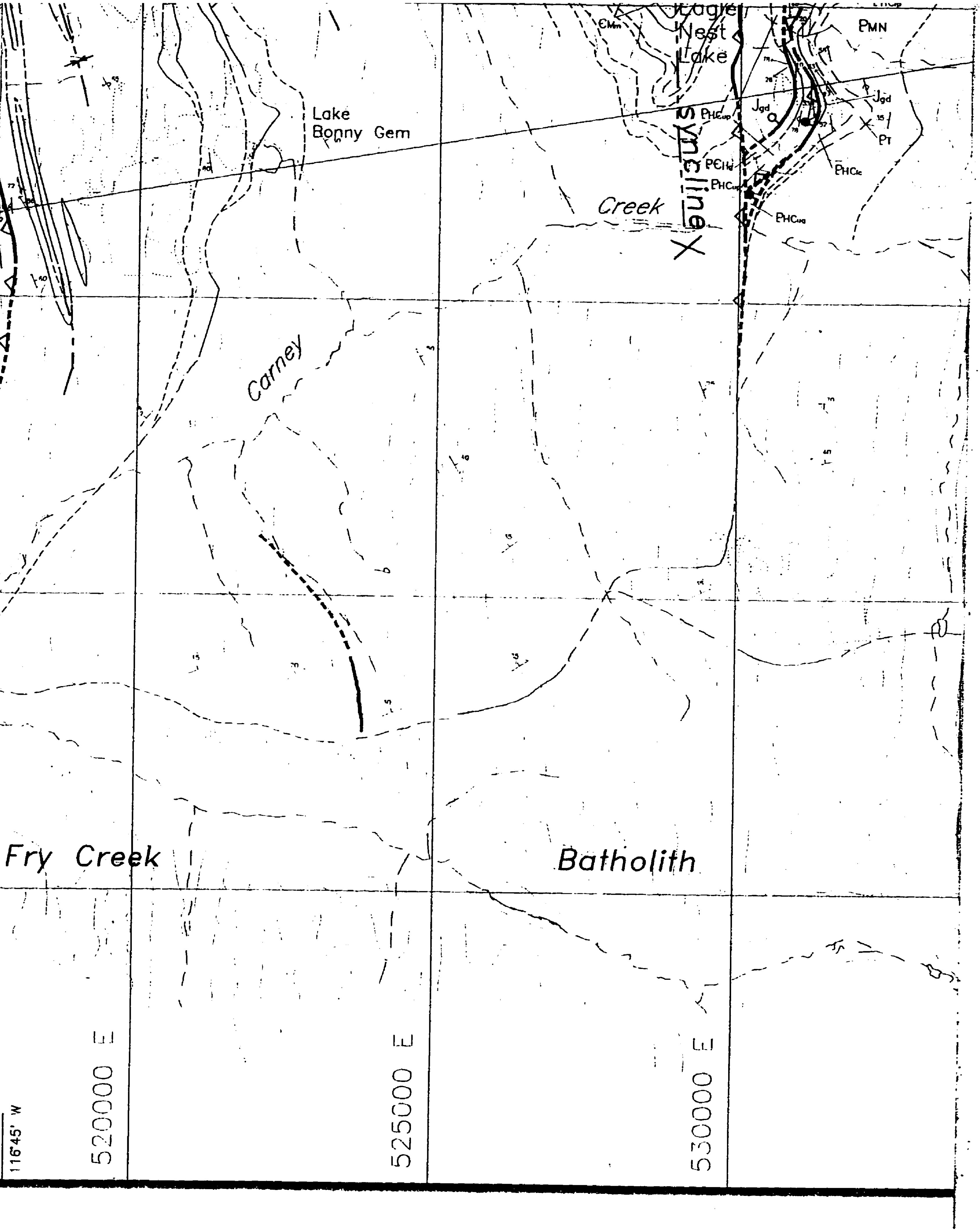
Magnetic declination (1980) approximately 21 20 E at centre of map, decreasing approximately 10' annually

Mapping by M. Warren was completed on 1:20,000 (N.A.D. 1983) and 1:50,000 (N.A.D. 1927) topographic bases, coupled with the use of 1:15,000 (1991) and older 1:60,000 air photos (1:20,000 maps and air photos available from MAPS B. C.)

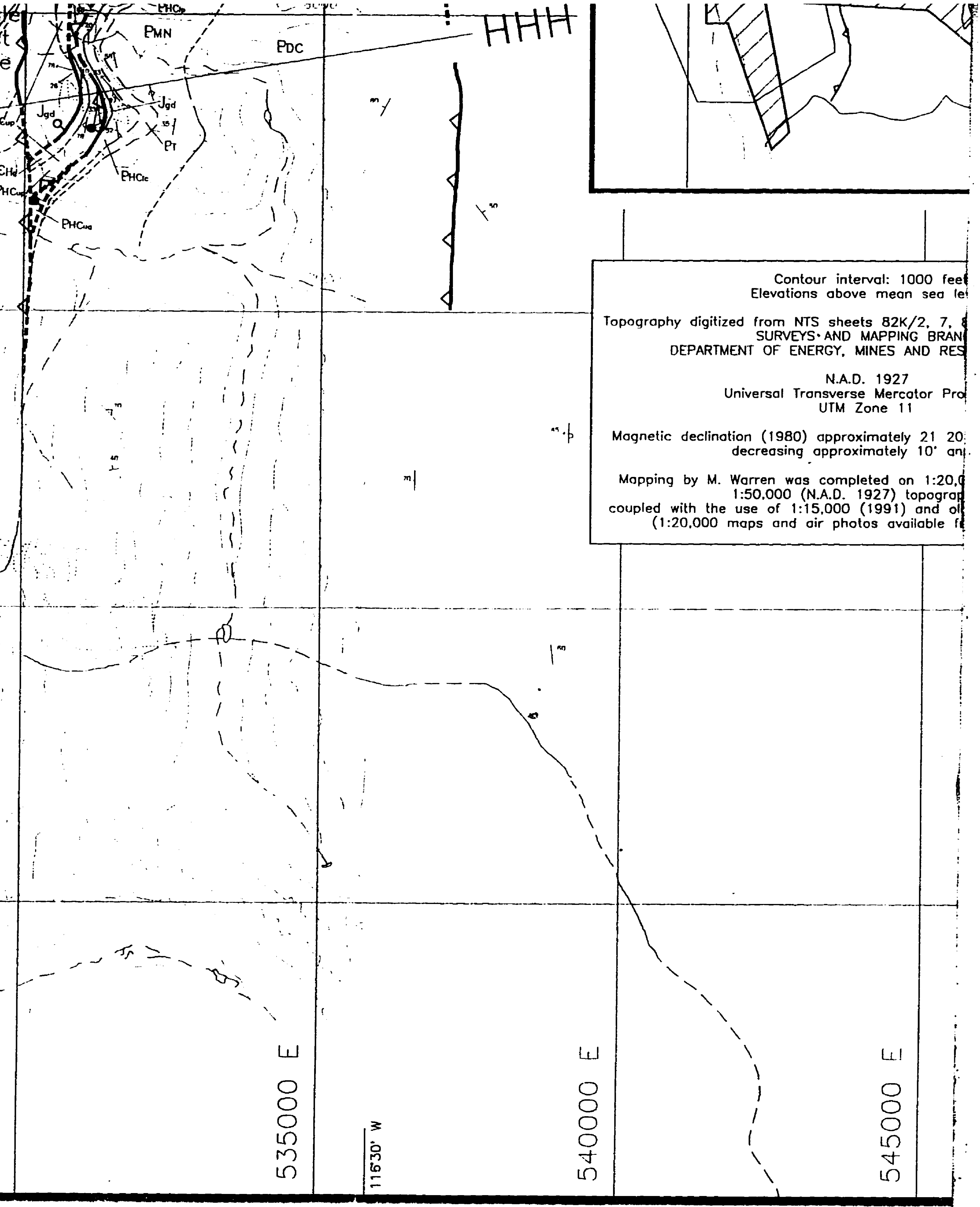


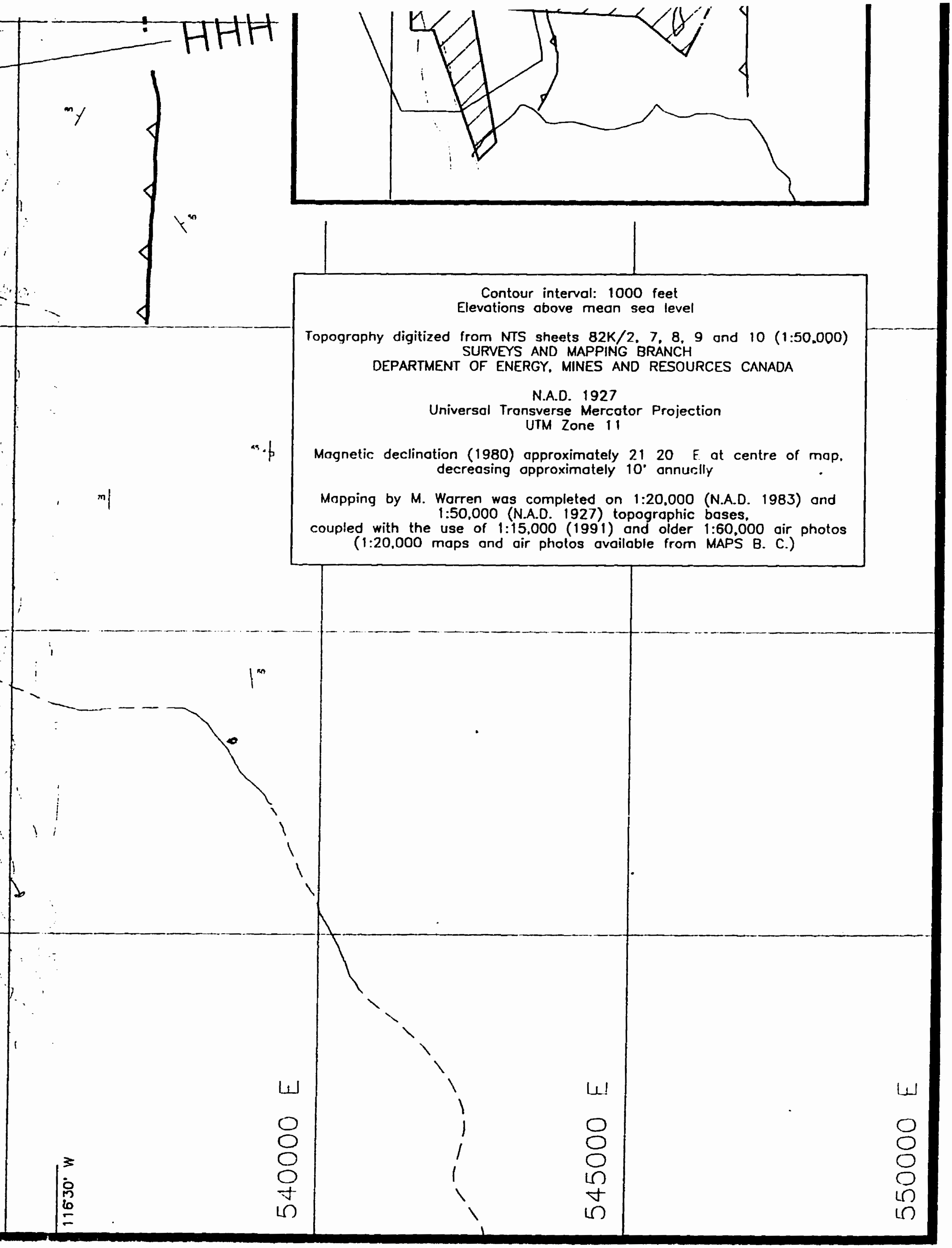
5555000 N
5550000 N
5545000 N
5540000 N
495000 E
500000 E
505000 E
117°00' W
Meadow
Argenta
Lardeau
KOOTENAY
IPlgp
IPlbr
IPlbp
IPlbp
€MB


Argenta
ardeau
anticline
HH
KOOTENAY
LAKE
Fry Creek
Fry Cr
€MB
€Hua
IPlbp
IPa
505000 E
510000 E
515000 E
116°45' W



Eagle Nest Lake
PMN
Lake Bonny Gem
Jgd
PT
PHCtc
Creek
Syncline
Carney
Fry Creek
Batholith
520000 E
525000 E
530000 E
116°45' W


PMN
PDC
Jgd
Jgd
PT
PHCtc
PHCua
Contour interval: 1000 feet
Elevations above mean sea le
Topography digitized from NTS sheets 82K/2, 7, 8
SURVEYS AND MAPPING BRAN
DEPARTMENT OF ENERGY, MINES AND RES
N.A.D. 1927
Universal Transverse Mercator Pro
UTM Zone 11
Magnetic declination (1980) approximately 21 20
decreasing approximately 10' an
Mapping by M. Warren was completed on 1:20,0
1:50,000 (N.A.D. 1927) topograp
coupled with the use of 1:15,000 (1991) and ol
(1:20,000 maps and air photos available f
535000 E
116°30' W
540000 E
545000 E


Contour interval: 1000 feet
Elevations above mean sea level
Topography digitized from NTS sheets 82K/2, 7, 8, 9 and 10 (1:50,000)
SURVEYS AND MAPPING BRANCH
DEPARTMENT OF ENERGY, MINES AND RESOURCES CANADA
N.A.D. 1927
Universal Transverse Mercator Projection
UTM Zone 11
Magnetic declination (1980) approximately 21 20 E at centre of map, decreasing approximately 10' annually
Mapping by M. Warren was completed on 1:20,000 (N.A.D. 1983) and 1:50,000 (N.A.D. 1927) topographic bases, coupled with the use of 1:15,000 (1991) and older 1:60,000 air photos (1:20,000 maps and air photos available from MAPS B. C.)
116°30' W
540000 E
545000 E
550000 E

**PLEASE NOTE:**

Oversize maps and charts are filmed in sections in the following manner:

**LEFT TO RIGHT, TOP TO BOTTOM, WITH SMALL OVERLAPS**

The following map or chart has been refilmed in its entirety at the end of this dissertation (not available on microfiche). A xerographic reproduction has been provided for paper copies and is inserted into the inside of the back cover.

Black and white photographic prints (17" x 23") are available for an additional charge.

BRITISH COLUMBIA

**Ministry of Employment and Investment**
Energy and Minerals Division
Geological Survey Branch

# OPEN FILE 1996-16

(Sheet 2 of 3)

# GEOLOGY OF THE WEST-CENTRAL PURCELL MOUNTAINS BRITISH COLUMBIA

**NTS 82K/2, 7 & PARTS OF 8, 10**

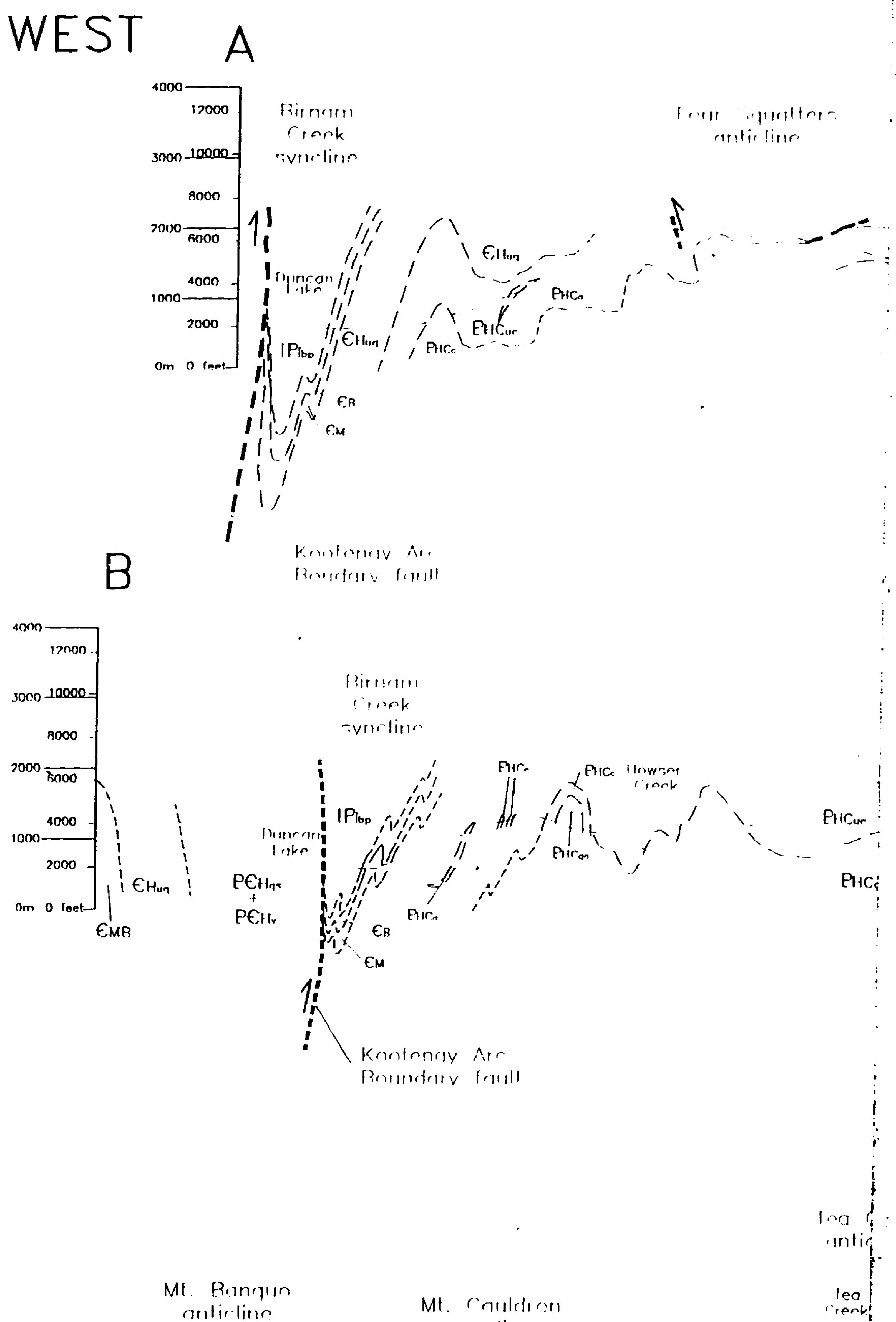
**Geology by**
**Marian J. Warren**

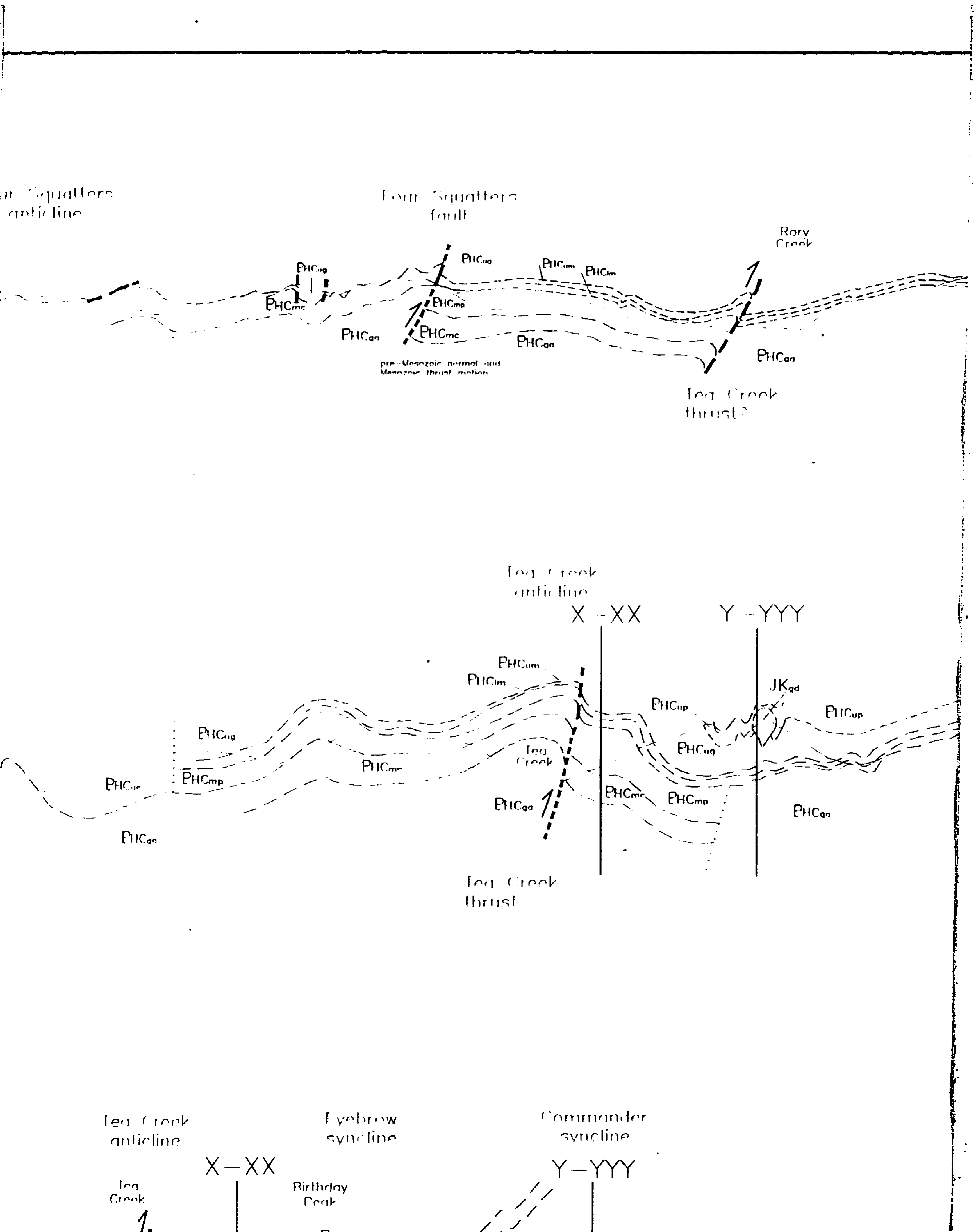
Scale 1:75 000

Metres 1000 0 1000 2000 3000 4000 Metres

Eastern segments of sections C CC, F FF, E EEEE, G GGGG and H HHHH
are after sections by Fyles, 1964
Segment EEE EEEE by Root, 1987

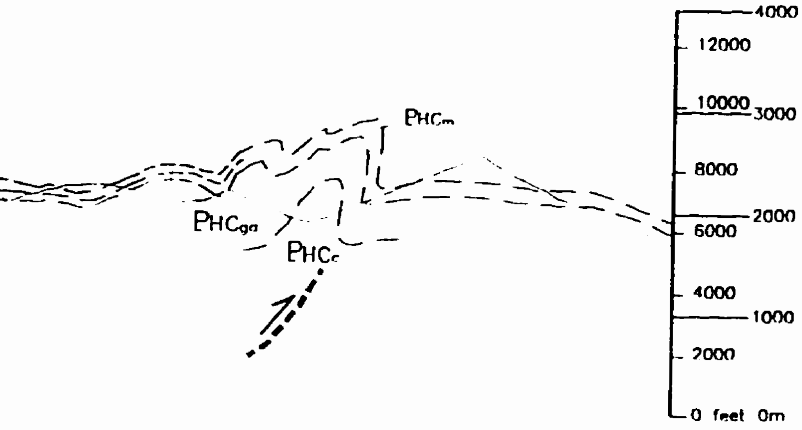
NS


WEST
A
Birnam Creek syncline
Four Squatters anticline
Duncan Lake
Kootenay Arc Boundary fault
B
Birnam Creek syncline
Howser Creek
Duncan Lake
Kootenay Arc Boundary fault
Tea Creek anticline
Mt. Banquo anticline
Mt. Cauldron

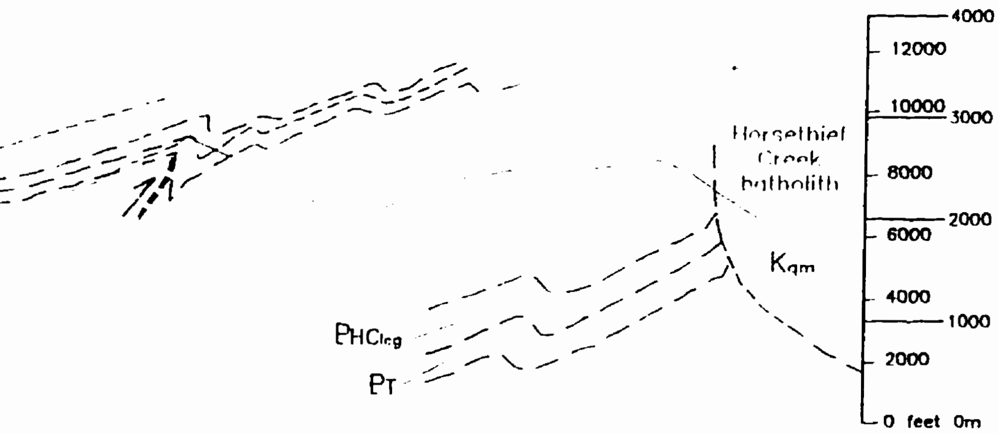


Four Squatters
fault
Rory
Creek
pre-Mesozoic normal and
Mesozoic thrust motion
Teg Creek
thrust?
Teg Creek
anticline
X-XX
Y-YYY
JKgd
Teg
Creek
Teg Creek
thrust
Teg Creek
anticline
Eyebrow
syncline
Commander
syncline
X-XX
Y-YYY
Teg
Creek
Birthday
Peak

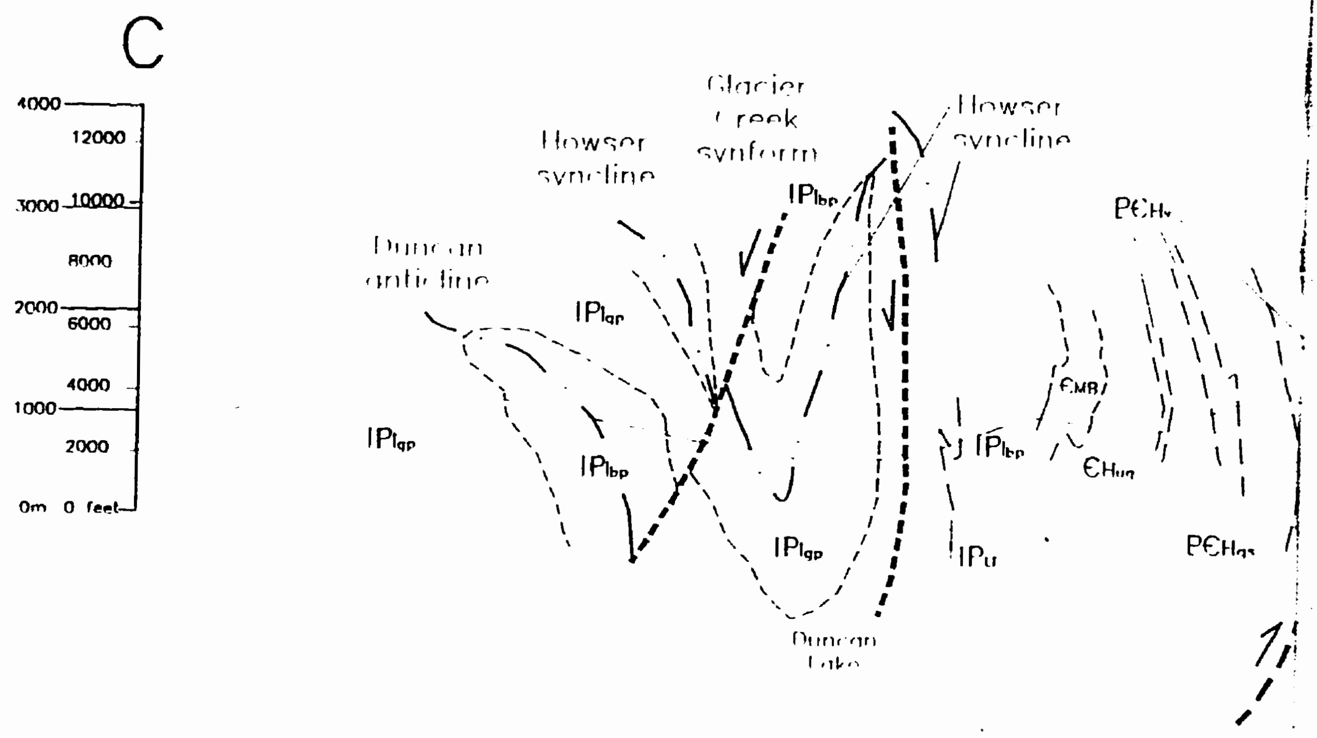
AA EAST


PHCm
PHCga
PHCc
4000
12000
10000 3000
8000
2000
6000
4000
1000
2000
0 feet 0m

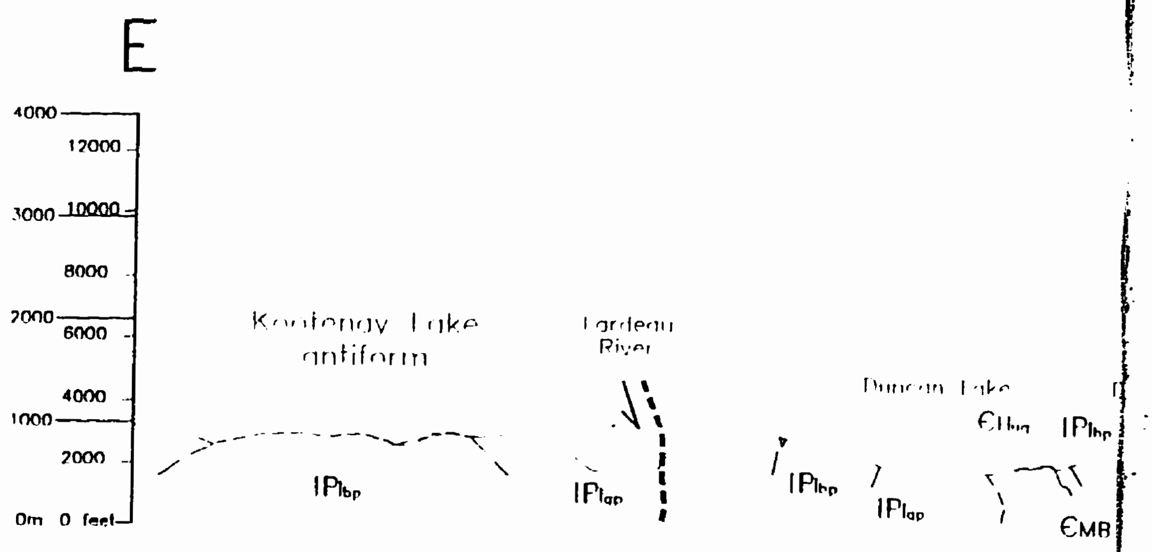
BB

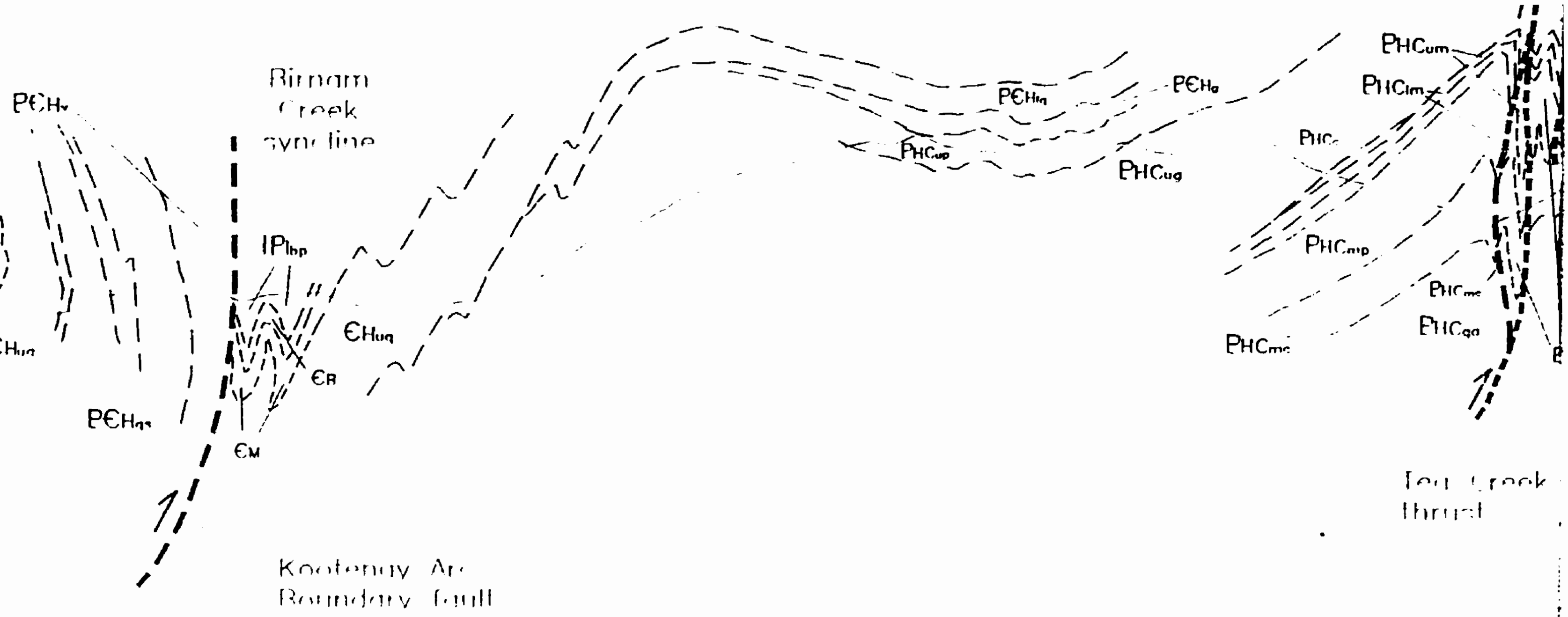

Horsethief
Creek
batholith
Kqm
PHClcg
PT
4000
12000
10000 3000
8000
2000
6000
4000
1000
2000
0 feet 0m

CC

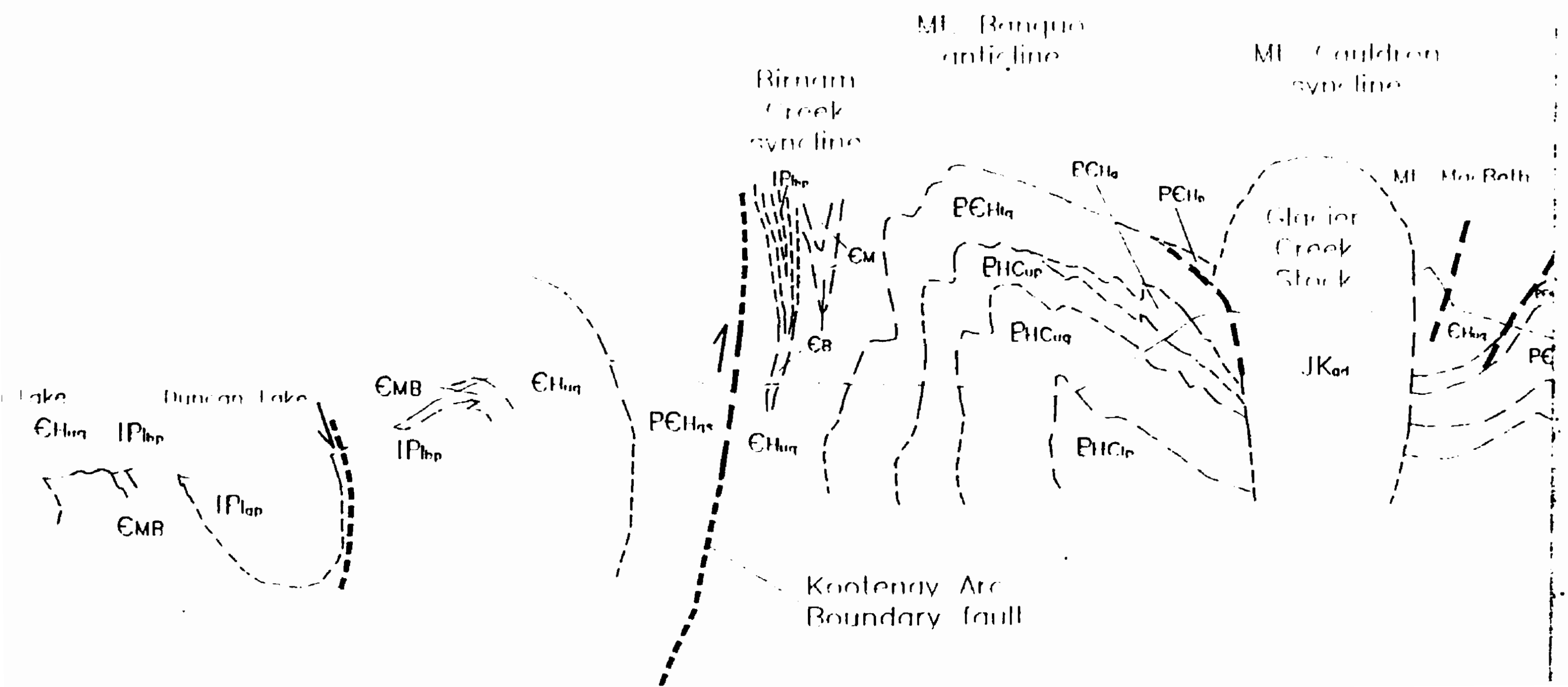


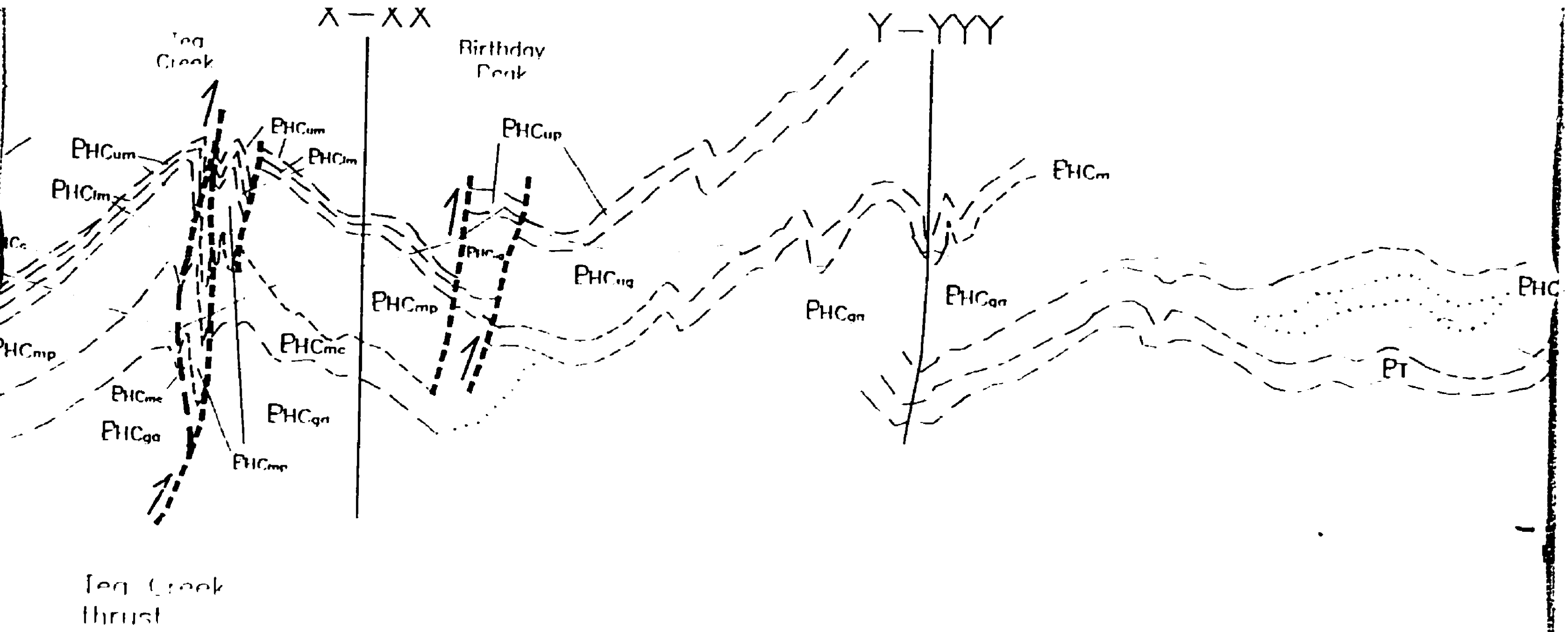

# STRUCTURAL CROSS S




Birnam Creek syncline
PЄHv
IPlbp
ЄHuq
ЄB
ЄM
PЄHqs
Kootenay Arc Boundary fault
PЄHlq
PЄHa
PHCup
PHCug
PHCum
PHClm
PHCmp
PHCms
PHCqa
PHCmc
Teu Creek thrust

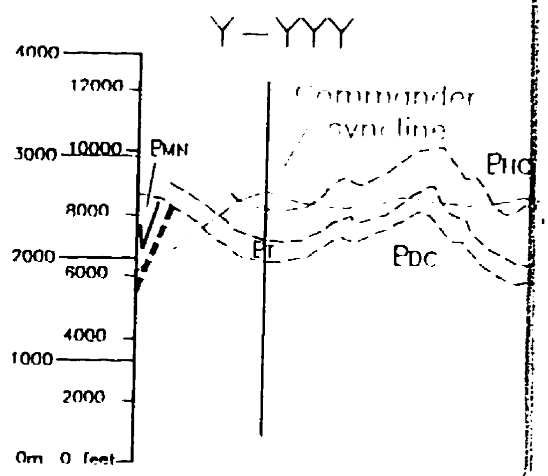
# SS SECTIONS


Mt. Banquo anticline
Mt. Cauldron syncline
Birnam Creek syncline
IPlbp
PЄHa
PЄHp
Mt. MacBeth
PЄHlq
ЄM
PHCup
Glacier Creek Stock
PHCug
ЄB
JKgd
ЄHuq
Lake
Duncan Lake
ЄMB
IPlap
PЄHqs
PHClp
Kootenay Arc Boundary fault

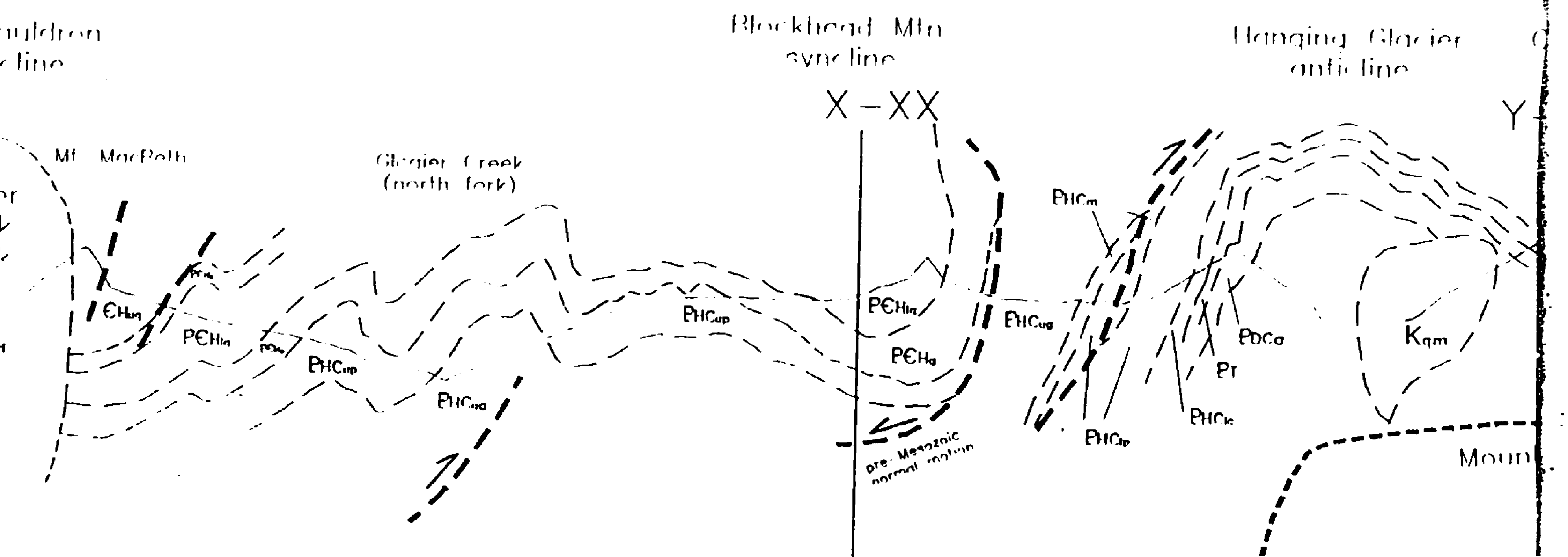


X–XX
Y–YYY
ЄHCum
ЄHClm
ЄHCup
ЄHCm
ЄHCug
ЄHCmp
ЄHCmc
ЄHCme
ЄHCga
ЄHCgo
ЄT


D

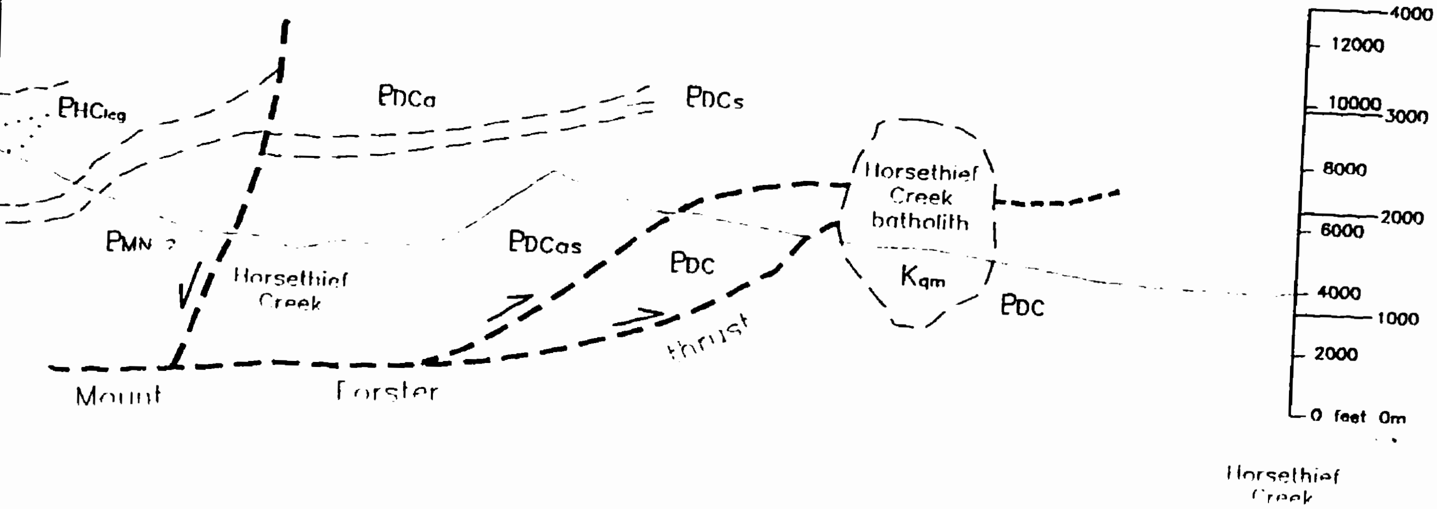


Y–YYY
4000
3000
2000
1000
0m
12000
10000
8000
6000
4000
2000
0 feet
ЄMN
ЄT
ЄDC

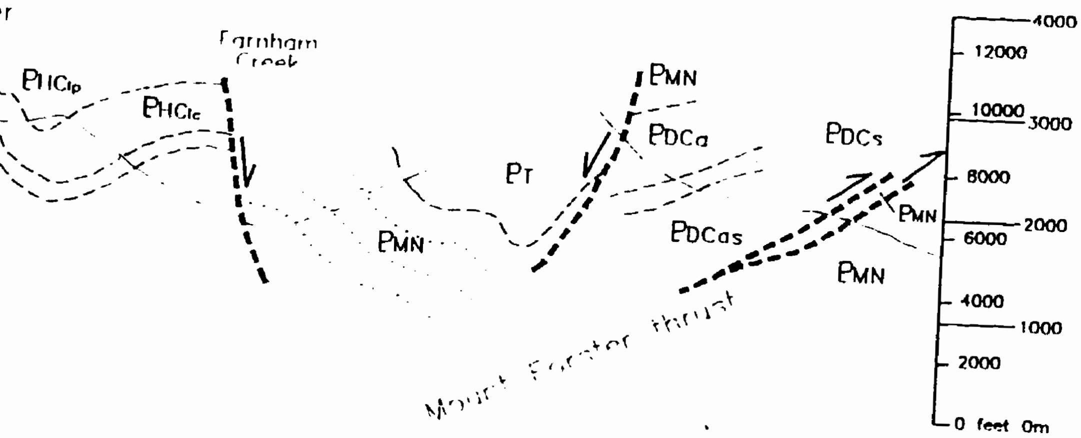



X–XX
ЄHup
PЄHlq
PЄHq
ЄHCug
ЄHCm
ЄDCa
ЄT
ЄHClc
ЄHClp
Kqm

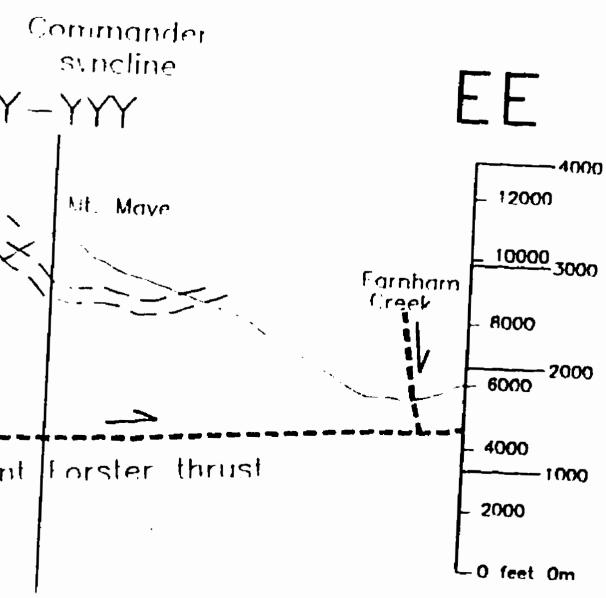
# CC

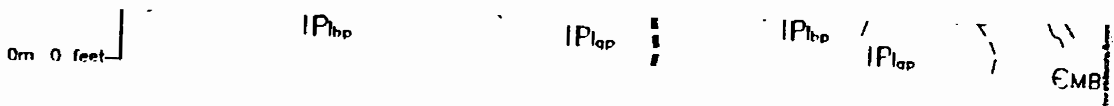

PHCicg
PDCa
PDCs
PMN ?
PDCas
PDC
Horsethief Creek
Horsethief Creek batholith
Kqm
PDC
Mount Forster thrust
4000
12000
10000
3000
8000
2000
6000
4000
1000
2000
0 feet 0m
Horsethief Creek

# DD

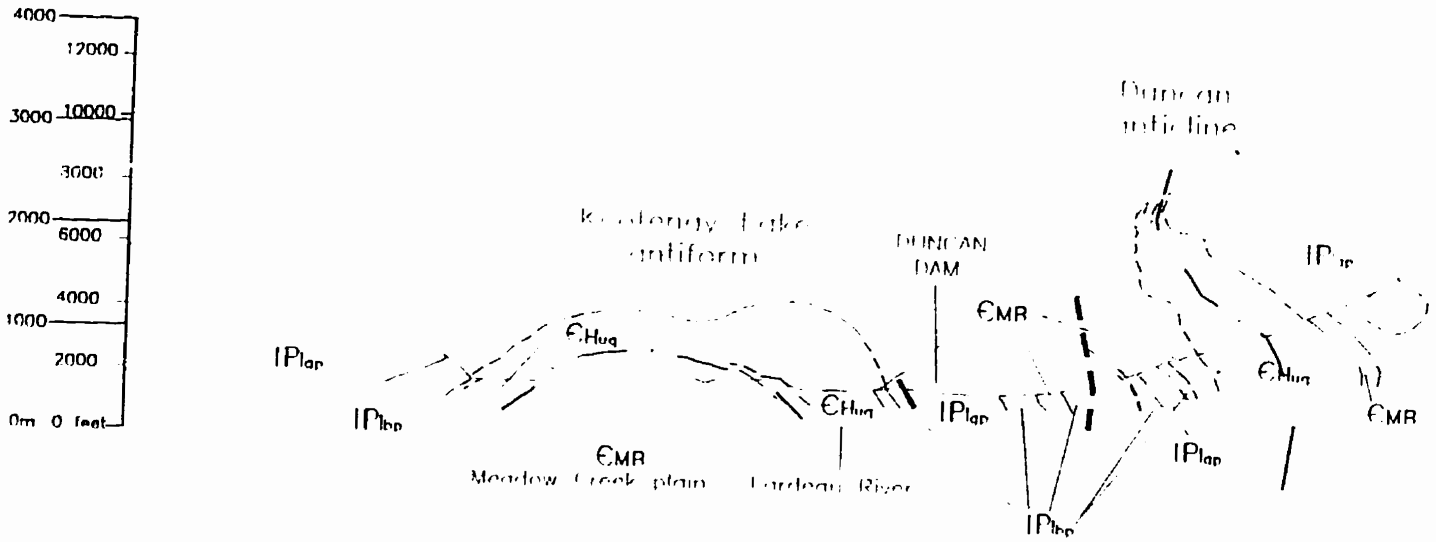

Farnham Creek
PHCip
PHCic
PMN
PDCa
PT
PDCs
PMN
PDCas
PMN
PMN
Mount Forster thrust
4000
12000
10000
3000
8000
2000
6000
4000
1000
2000
0 feet 0m

# EE

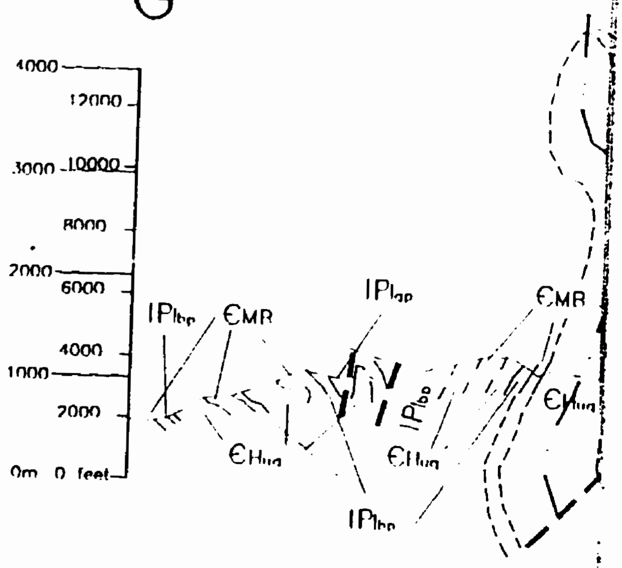

Commander syncline
Y-YYY
Mt. Maye
Farnham Creek
Forster thrust
4000
12000
10000
3000
8000
2000
6000
4000
1000
2000
0 feet 0m

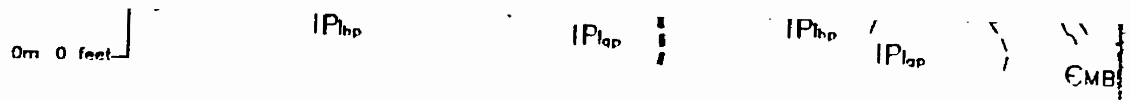

0m 0 feet
IPlbp
IPlap
IPlbp
IPlap
ЄMB

F


4000
12000
3000
10000
2000
6000
4000
1000
2000
0m 0 feet
IPlap
IPlbp
ЄHuq
ЄMR
DUNCAN
DAM
ЄMR
IPlap
ЄHuq
IPlap
IPlbp
ЄMR
ЄHuq
ЄMR

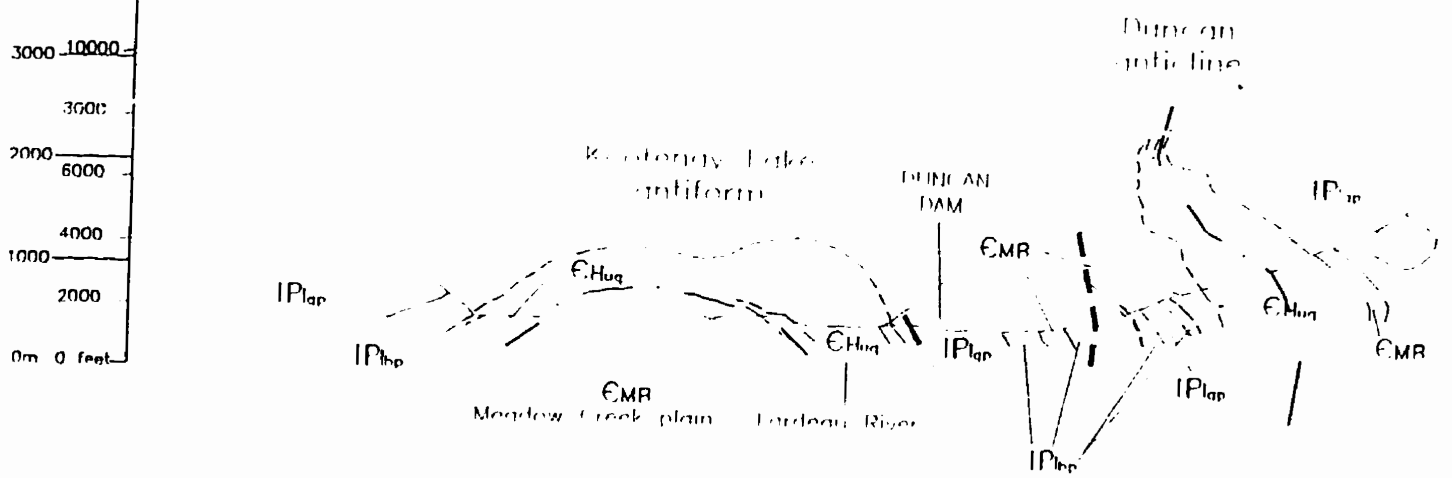
G


4000
12000
3000
10000
8000
2000
6000
4000
1000
2000
0m 0 feet
IPlbp
ЄMR
IPlap
ЄMR
ЄHuq
ЄHuq
ЄHuq
IPlbp



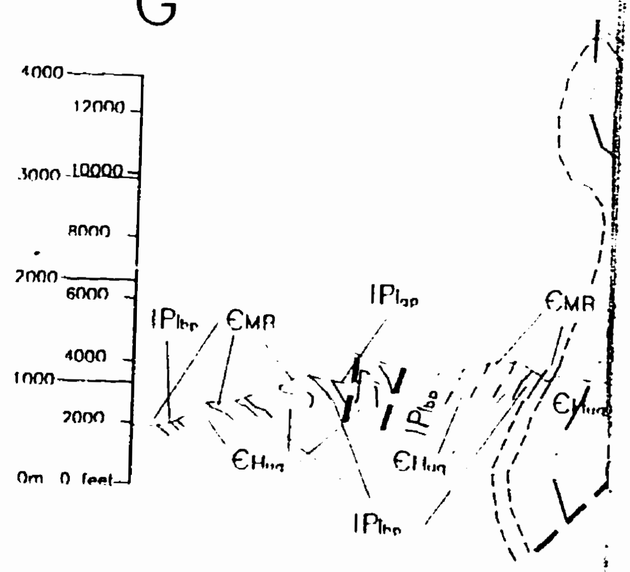
0m 0 feet
IPlbp
IPlap
IPlbp
IPlap
ЄMB


F



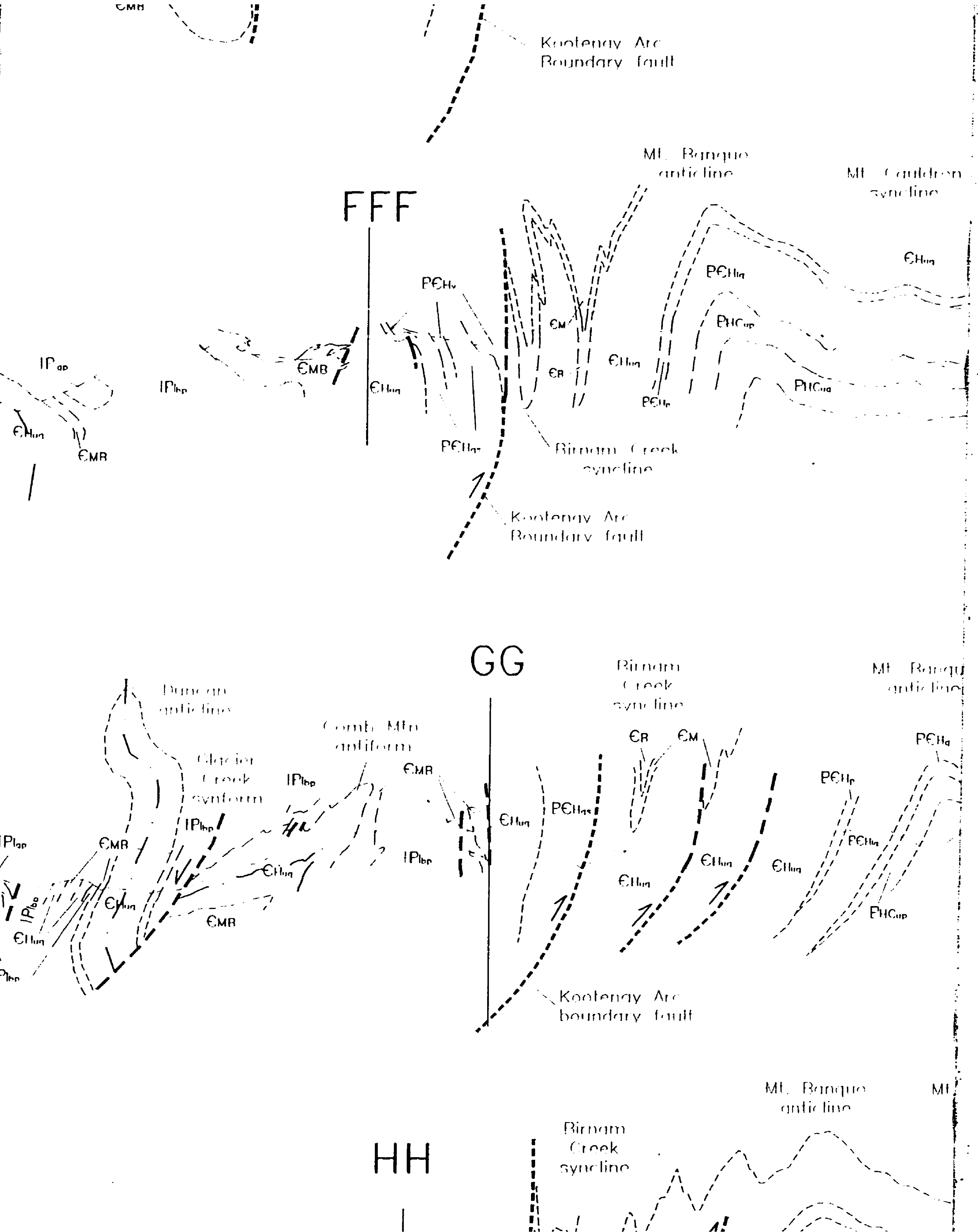
4000
12000
3000
10000
2000
6000
4000
1000
2000
0m 0 feet
IPlap
IPlbp
ЄHuq
ЄMB
DUNCAN DAM
IPlap
ЄMB
IPlbp
IPlap
ЄHuq
ЄMB

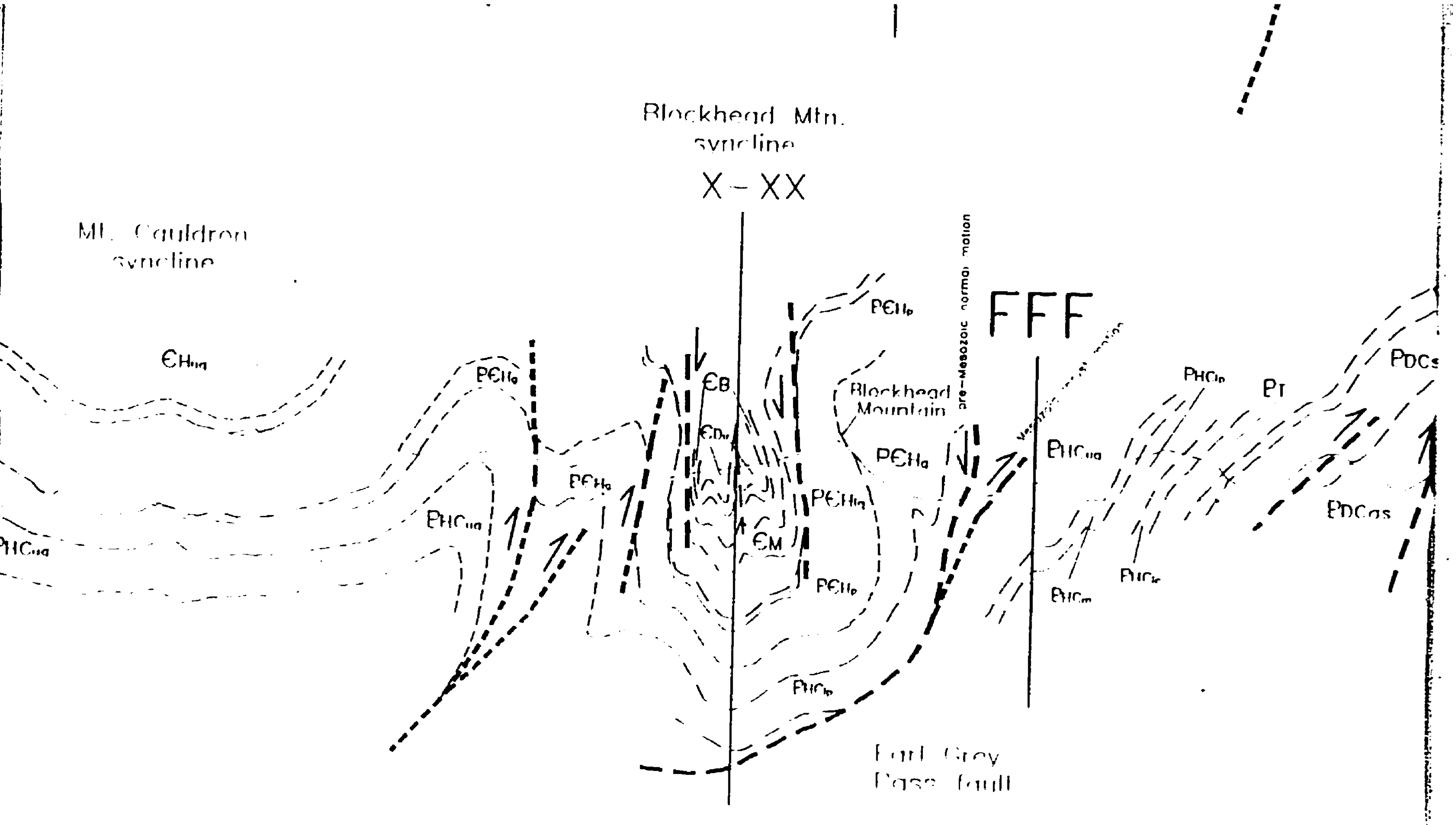

G

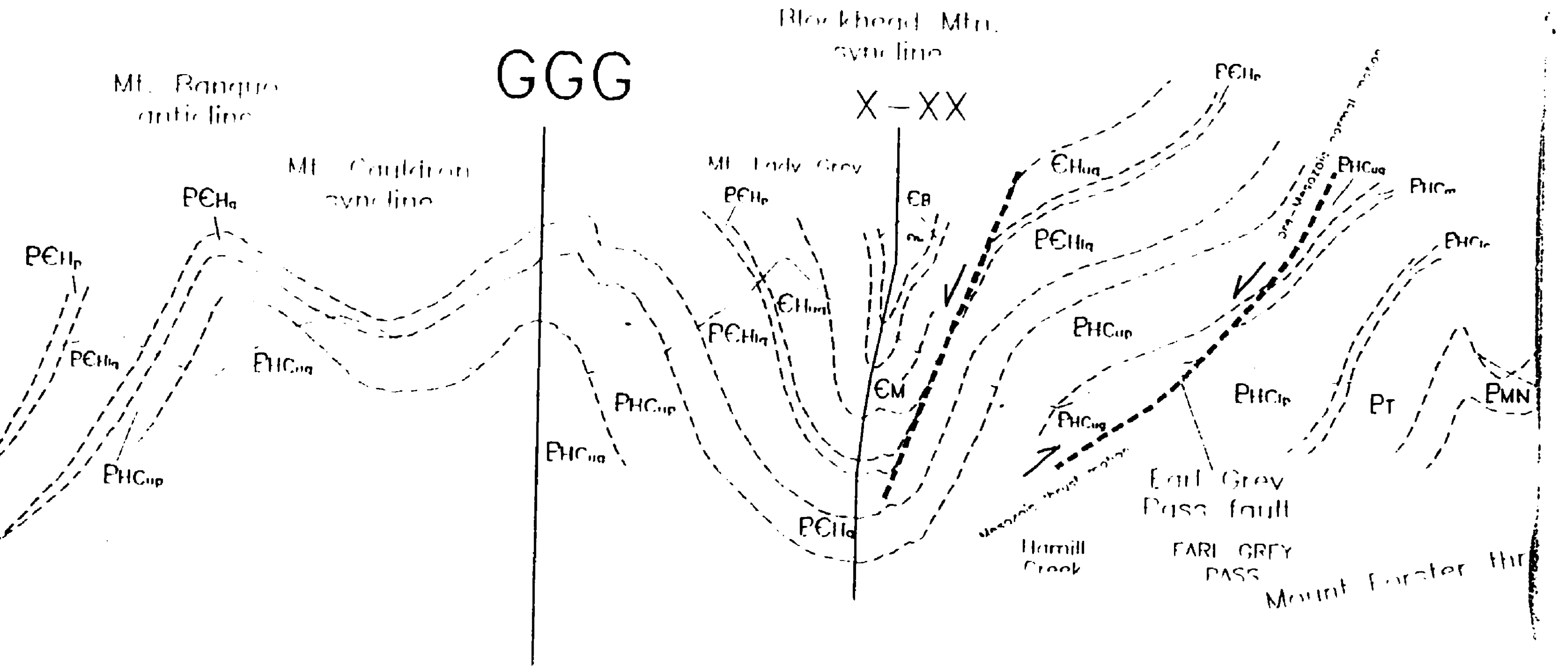


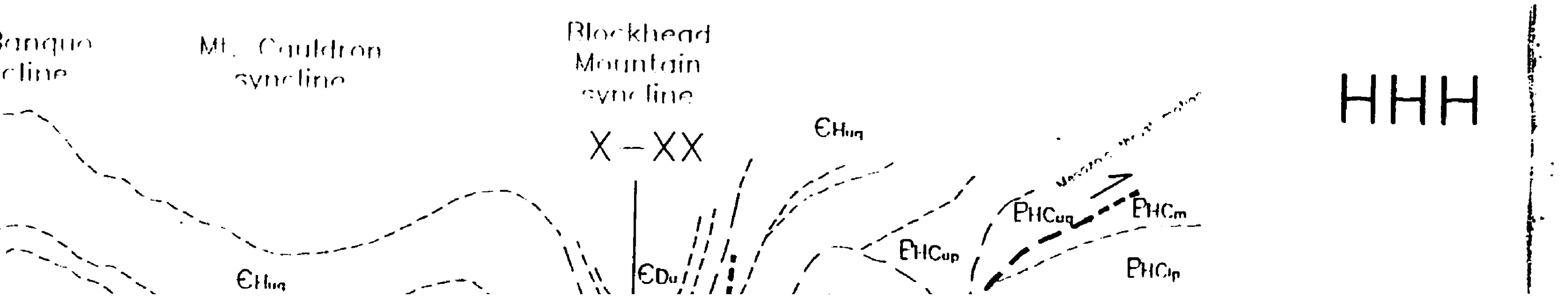
4000
12000
3000
10000
8000
2000
6000
4000
1000
2000
0m 0 feet
IPlbp
ЄMB
IPlap
ЄMB
ЄHuq
IPlbp
ЄHuq
IPlbp


H

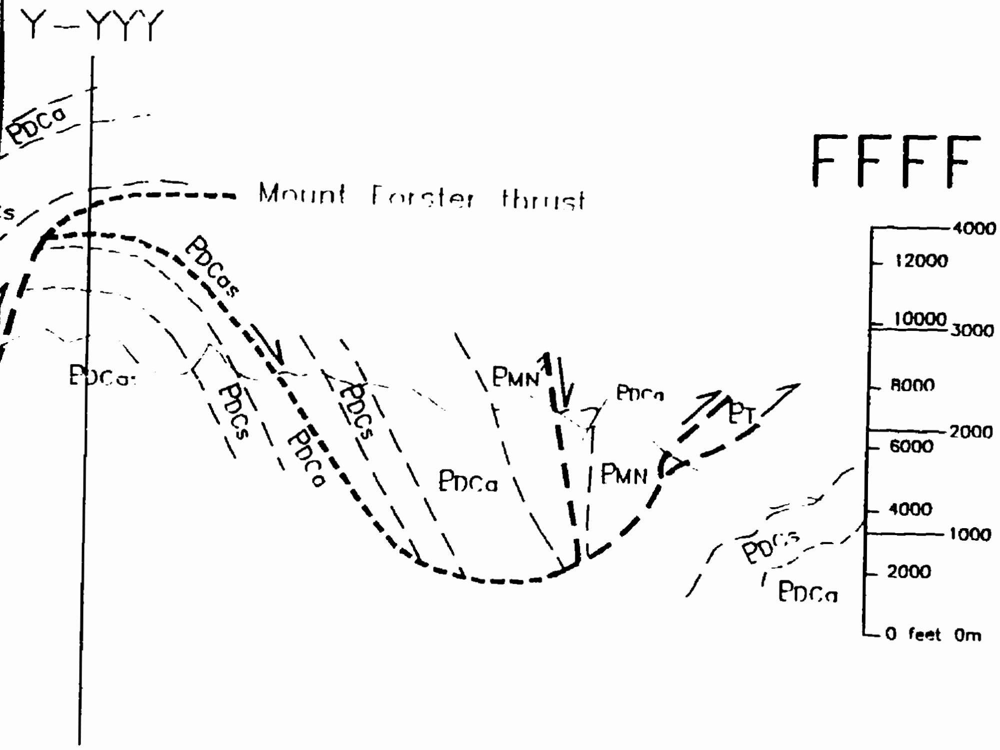

Kootenay Arc Boundary fault
FFF
Mt. Banque anticline
Mt. Cauldron syncline
Birnam Creek syncline
Kootenay Arc Boundary fault
GG
Birnam Creek syncline
Mt. Banque anticline
Duncan antiline
Glacier Creek synform
Comb Mtn antiform
Kootenay Arc boundary fault
HH
Mt. Banque anticline
Birnam Creek syncline


Blockhead Mtn.
syncline
X – XX
Mt. Cauldron
syncline
FFF
Blockhead
Mountain
pre-Mesozoic normal motion
Earl Grey
Pass fault
ЄHuq
PЄHq
PЄHp
ЄB
ЄM
PЄHq
PЄHlq
PHCuq
PHCp
PHCuq
PHCm
PHCp
PHClc
PT
PDCs
PDCas

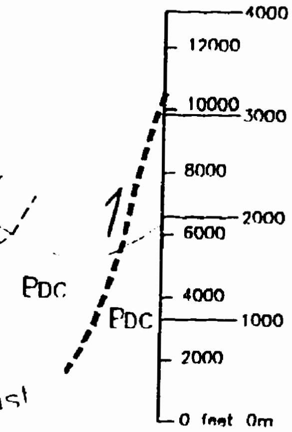

GGG
Mt. Banquo
anticline
Mt. Cauldron
syncline
Blockhead Mtn.
syncline
X – XX
Mt. Lady Grey
pre-Mesozoic normal motion
Mesozoic thrust motion
Earl Grey
Pass fault
Hamill
Creek
EARL GREY
PASS
Mount Forster thr
PЄHp
PЄHq
PЄHlq
PHCuq
PHCup
ЄB
ЄHuq
ЄHuql
ЄM
PHCm
PHClc
PHClp
PT
PMN


Banquo
cline
Mt. Cauldron
syncline
Blockhead
Mountain
syncline
X – XX
HHH
ЄHuq
ЄDu
PHCup
PHCuq
PHCm
PHClp

FFFF


Y—YYY
PDCa
Mount Forster thrust
PDCas
PDCs
PDCa
PMN
PT
PDCs
4000
12000
10000
3000
8000
2000
6000
4000
1000
2000
0 feet 0m

GGGG


4000
12000
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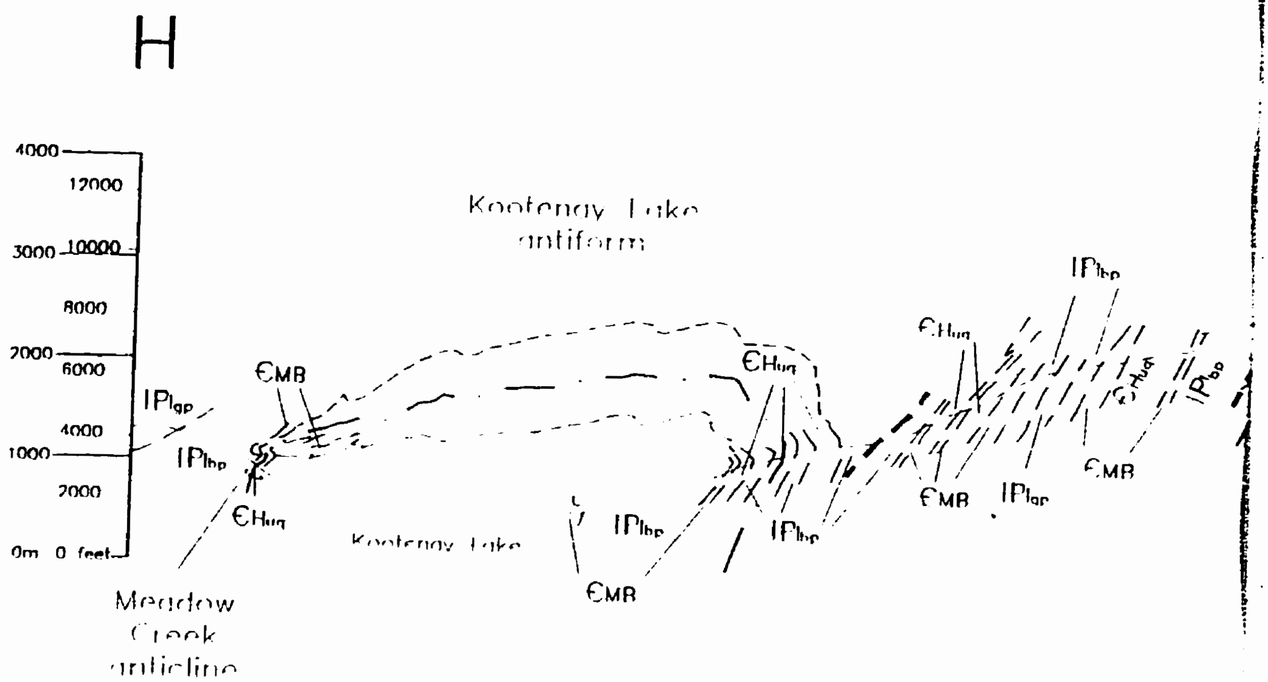
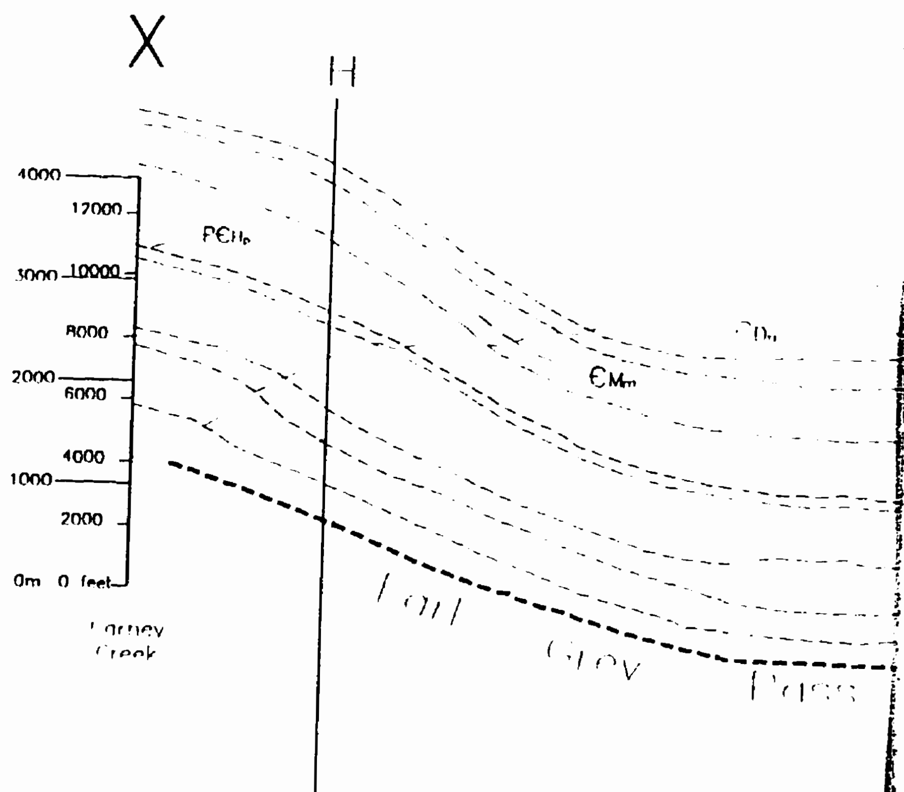
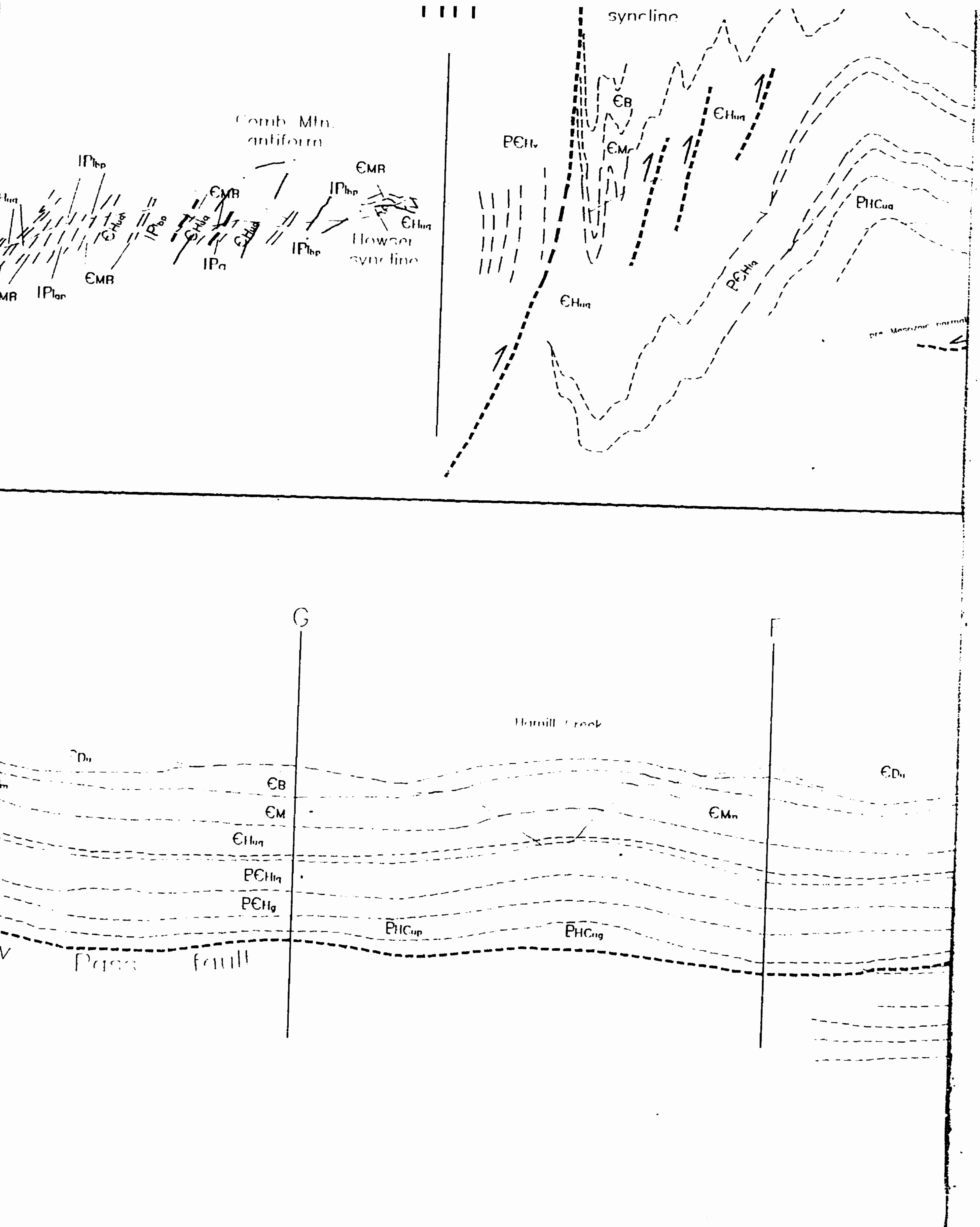
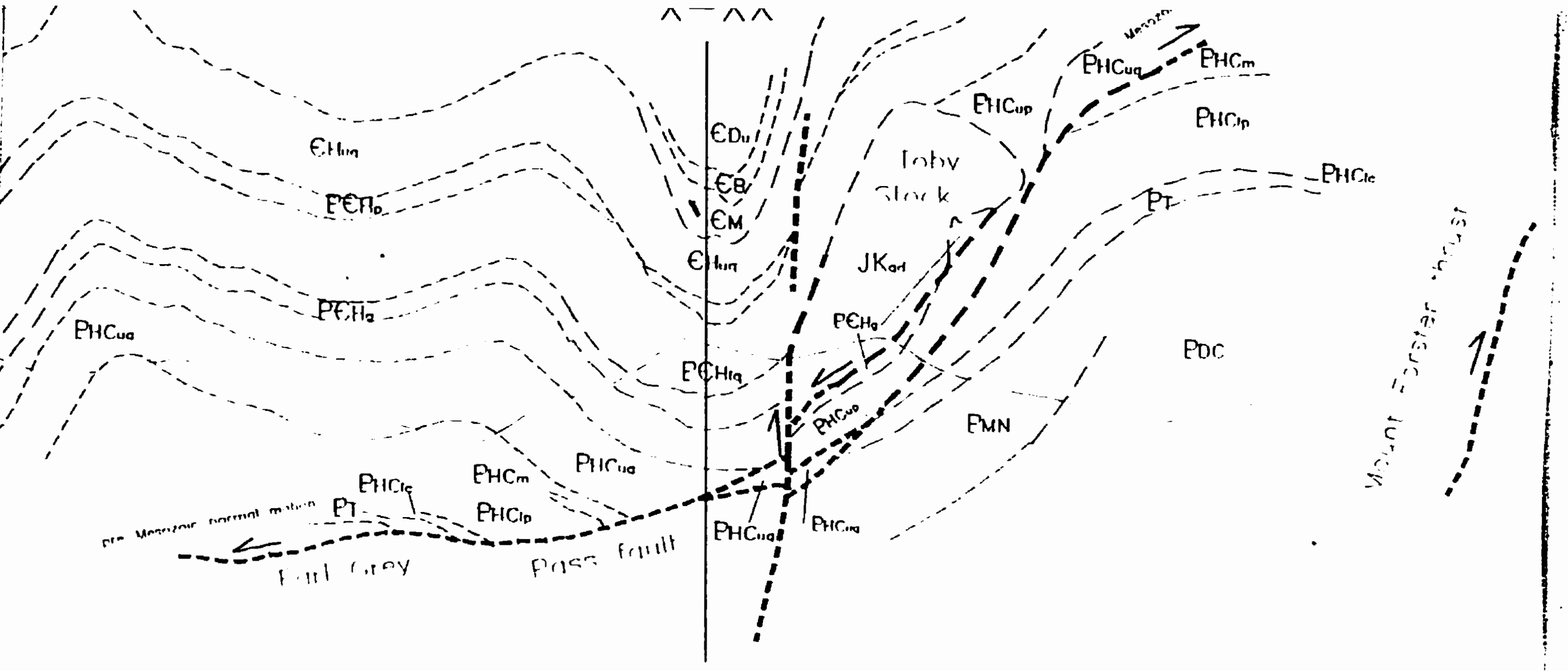
6000
4000
1000
2000
0 feet 0m
PDC

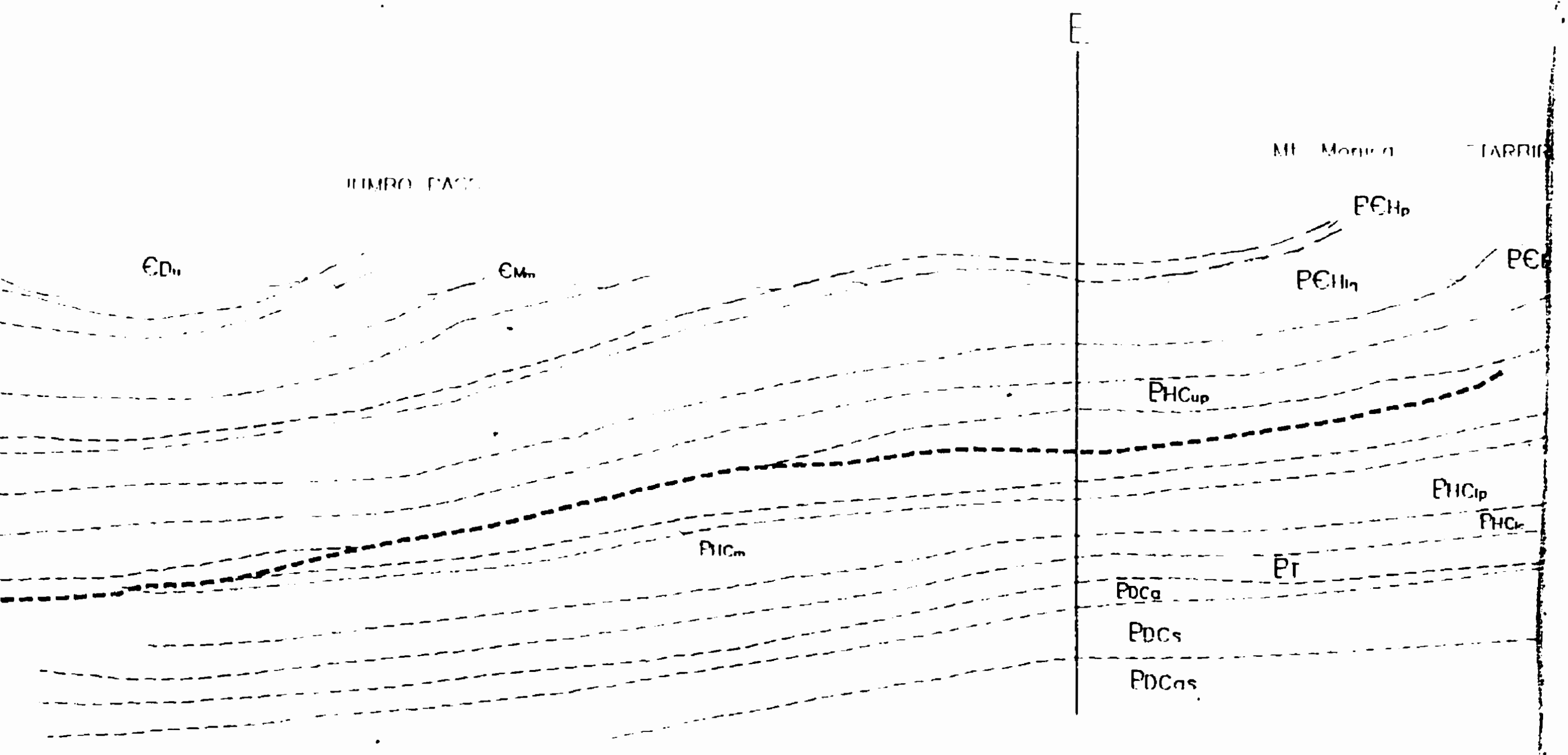

H
4000
12000
3000
10000
8000
2000
6000
4000
1000
2000
0m 0 feet
Kootenay Lake antiform
ЄMB
ЄHuq
Meadow Creek anticline
Kootenay Lake

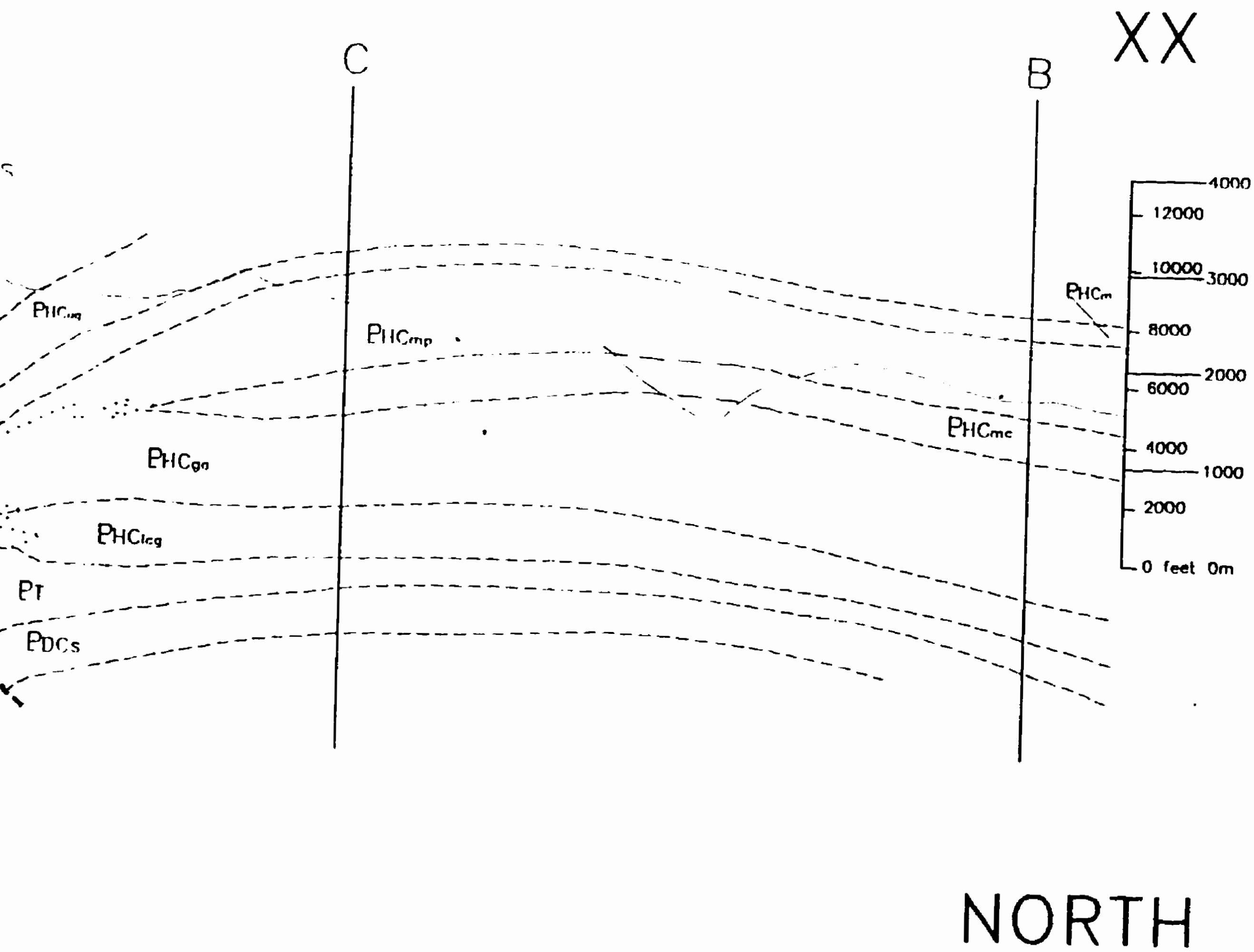

X
H
4000
12000
3000
10000
8000
2000
6000
4000
1000
2000
0m 0 feet
PЄHp
ЄMm

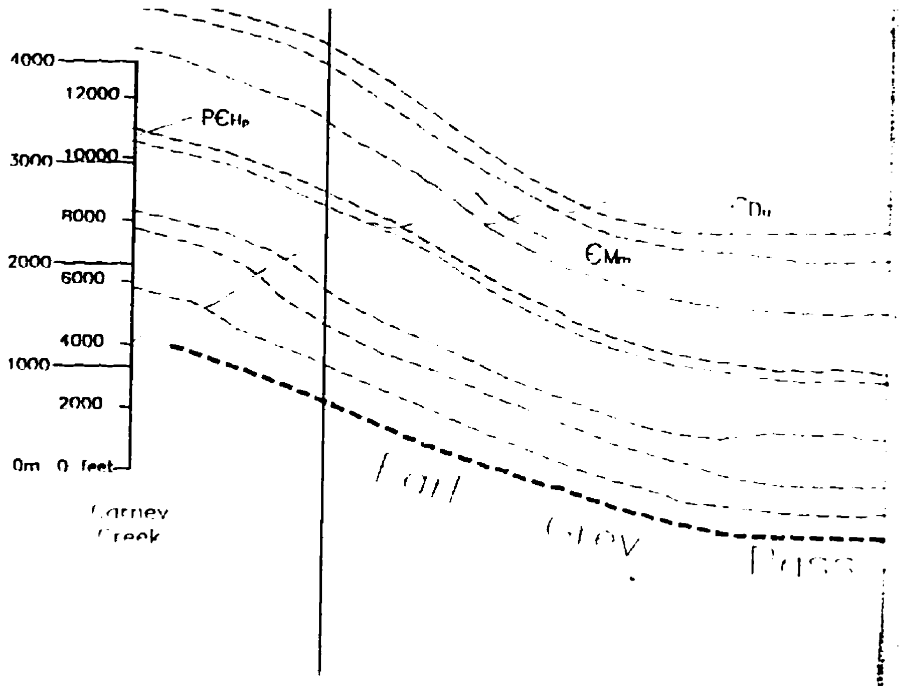
SOUTH


syncline
Comb Mtn. antiform
Howser syncline
€MB
IPlbp
€Hlq
IPq
IPlap
€B
€Mr
€Hlq
P€Hv
P€Hlq
PHCug
pre-Mesozoic normal
G
F
Hamill Creek
€B
€M
€Hlq
P€Hlq
P€Hg
PHCup
PHCug
€Mn
€Du
Pass fault


ЄHuq
PЄHp
PЄHq
PHCuq
ЄDu
ЄB
ЄM
ЄHuq
Toby
Stock
JKqd
PЄHq
PЄHtq
PHCup
PHClc
PHCm
PHCuq
PT
PHClp
pre Mesozoic normal motion
Earl Grey
Pass fault
PHCuq
PHCuq
PMN
PHCuq
PHCm
PHCup
PHClp
PHClc
PT
PDC
Mount Forster thrust


JUMBO PASS
E
Mt. Monica
PЄHp
ЄDu
ЄMn
PЄHtq
PHCup
PHClp
PHClc
PHCm
PT
PDCq
PDCs
PDCqs


C
B
XX
PHCm
PHCmp
PHCmc
PHCga
PHClcg
PT
PDCs
4000
12000
10000
3000
8000
6000
2000
4000
1000
2000
0 feet 0m
NORTH

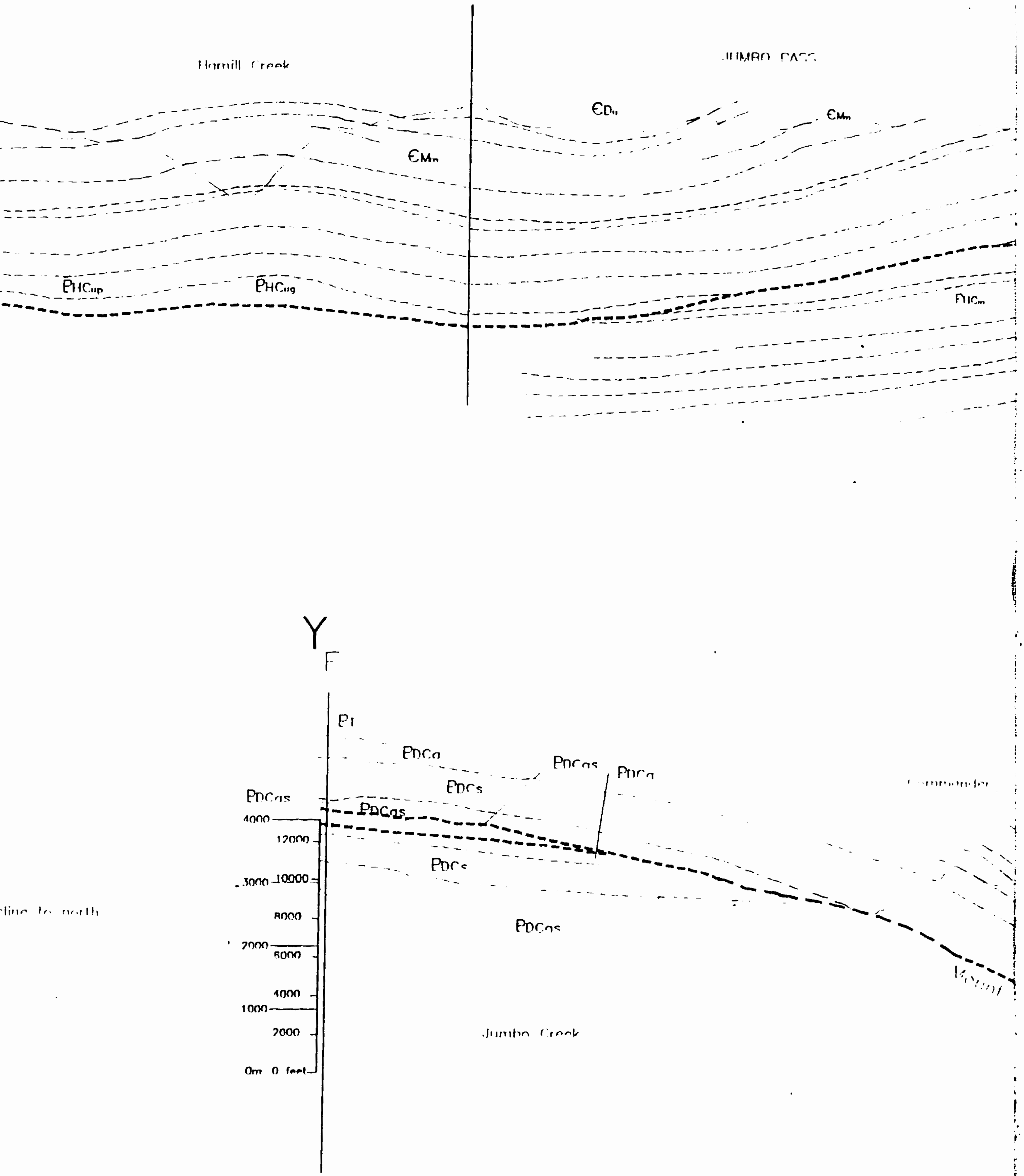



# SOUTH

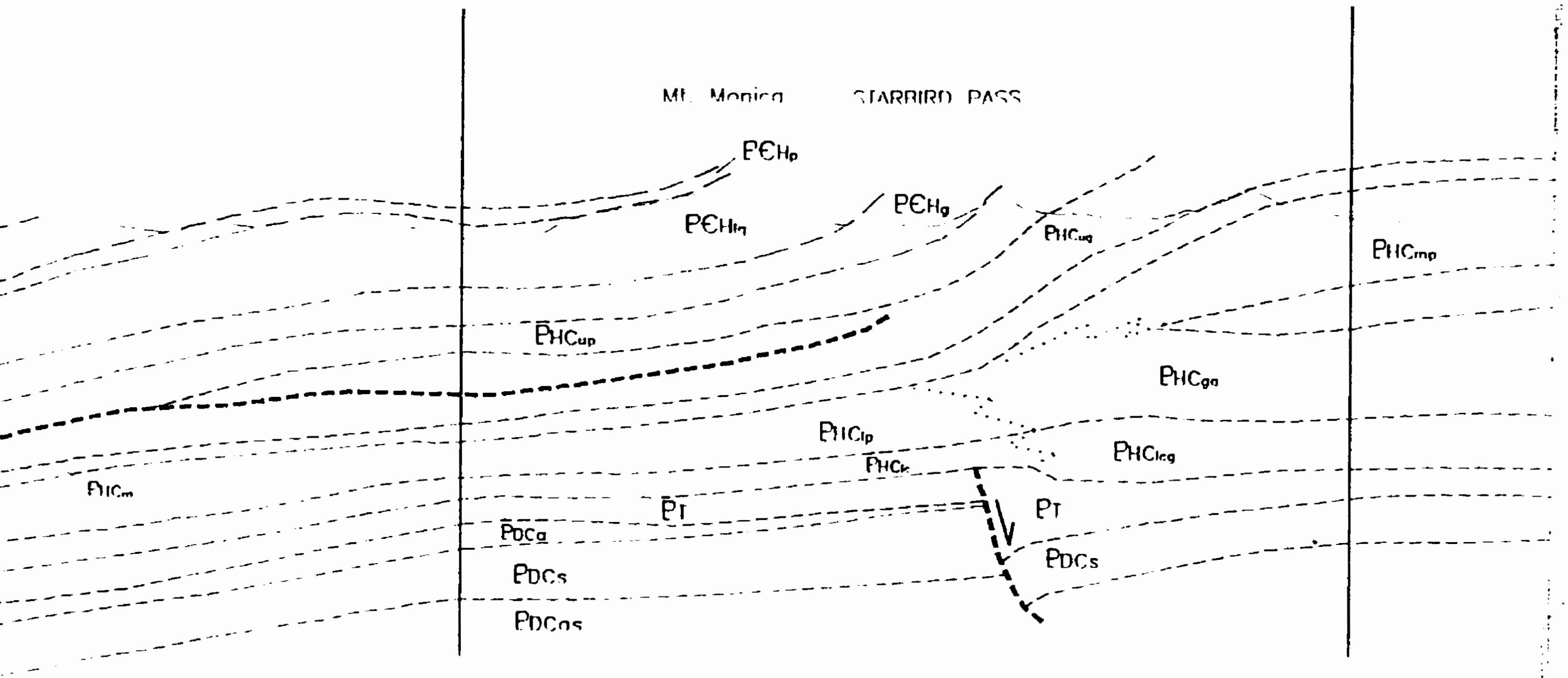
# LONGITUDINAL SECTIO

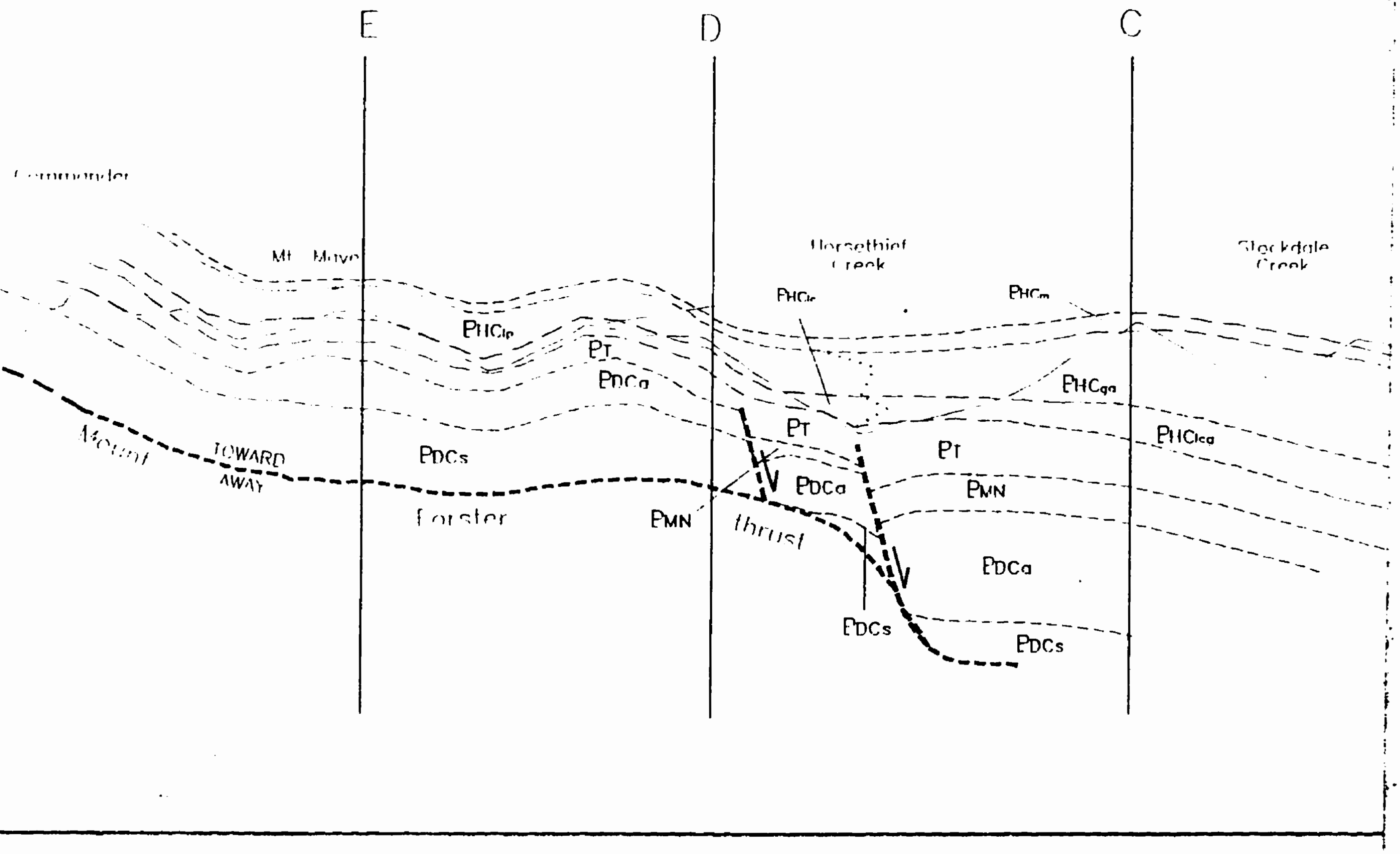
X XX – subvertical section along axial surface of Blackhead Mountain syncline from Carney Creek joins axial surface of unnamed syncline to north

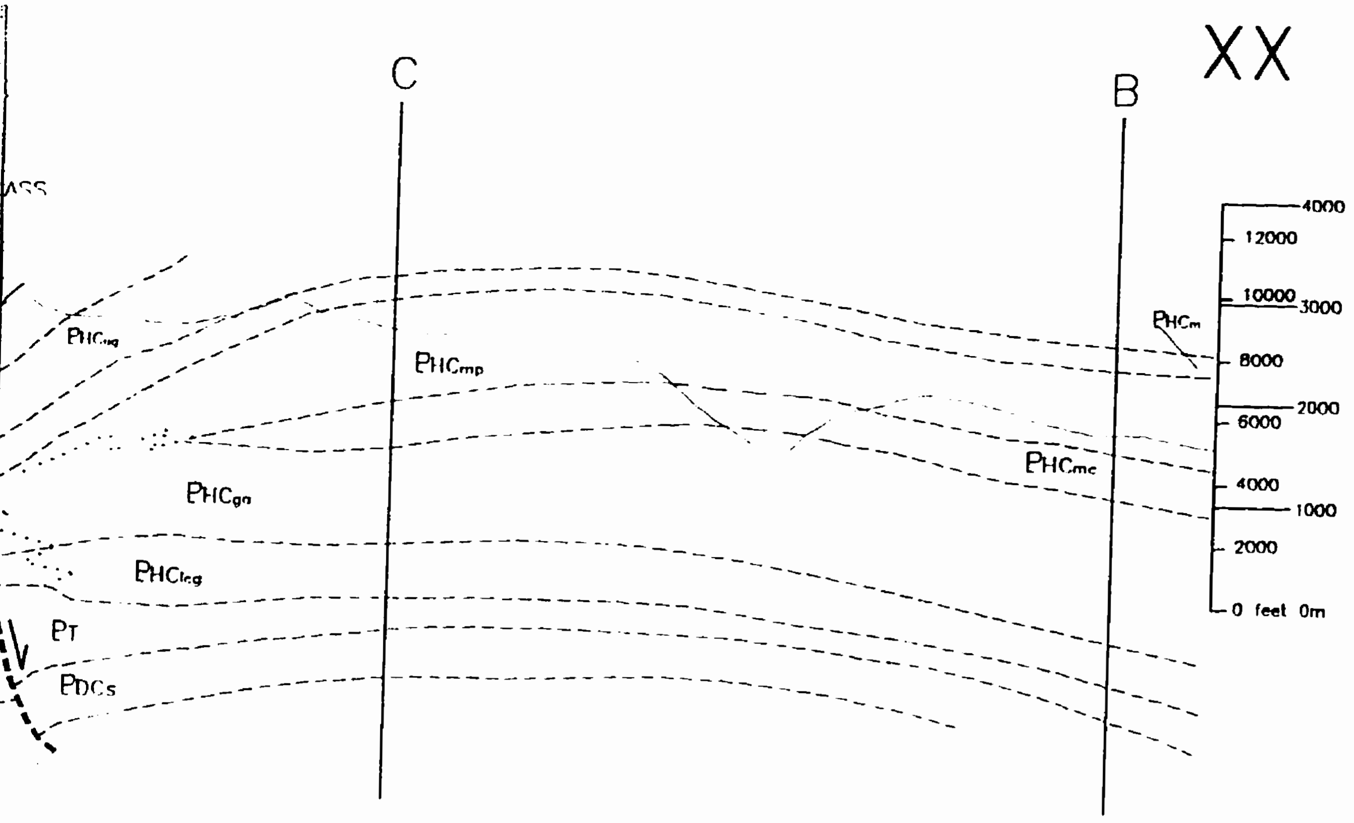
Y YYY – vertical section in footwall of Mount Forster thrust, follows axial surface of Commander



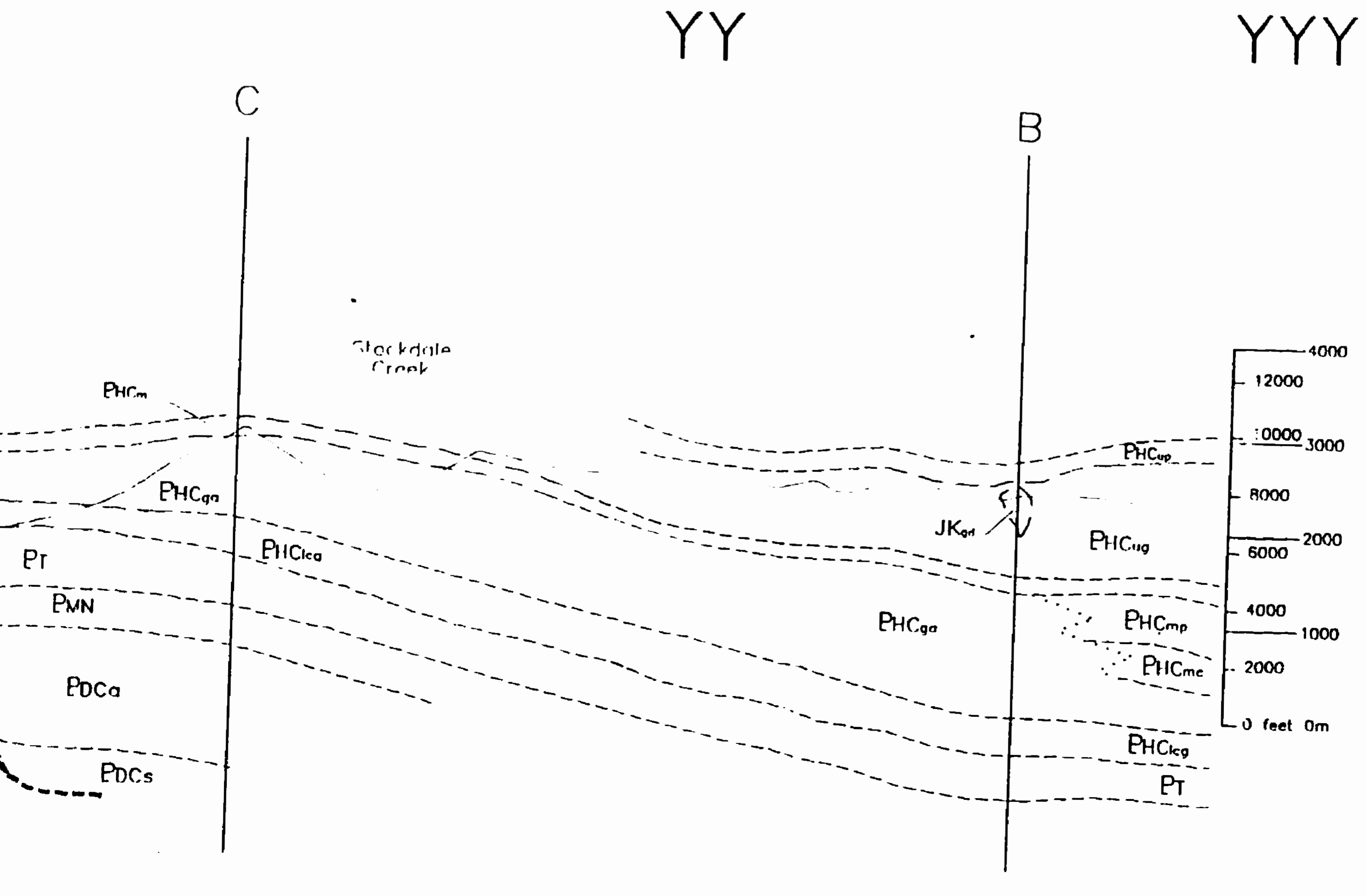
Hamill Creek
JUMBO PASS
ЄDu
ЄMn
ЄMn
PHCup
PHCug
PHCm
Y
F
PI
PDCa
PDCas
PDCa
PDCs
PDCas
PDCas
4000
12000
PDCs
3000
10000
8000
PDCas
2000
6000
4000
1000
2000
0m 0 feet
Jumbo Creek
Mount


Mt. Monica
STARBIRD PASS
PЄHp
PЄHlq
PЄHg
PHCug
PHCmp
PHCup
PHCga
PHClp
PHClcg
PHCk
PHCm
PT
PDCa
PT
PDCs
PDCs
PDCas


E
D
C
Mt. Maye
Horsethief Creek
Stockdale Creek
PHClc
PHCm
PHClp
PT
PDCa
PHCga
PHClcg
PT
PT
PDCs
PDCa
PMN
TOWARD
AWAY
Mount Forster thrust
PMN
PDCa
PDCs
PDCs


XX
C
B
PHCug
PHCmp
PHCm
PHCme
PHCga
PHClcg
PT
PDCs
4000
12000
10000
3000
8000
2000
6000
4000
1000
2000
0 feet 0m

NORTH


YY
YYY
C
B
Stockdale Creek
PHCm
PHCga
PT
PHClcg
PMN
PDCa
PDCs
JKgd
PHCup
PHCug
PHCmp
PHCme
PT
4000
12000
10000
3000
8000
2000
6000
4000
1000
2000
0 feet 0m

# LAYERED ROCKS

## EASTERN KOOTENAY ARC
## (HANGINGWALL OF KOOTENAY ARC BOUNDARY FAULT)

### LOWER PALEOZOIC

LARDEAU GROUP

INDEX FORMATION

lPI$_{gp}$ — Green phyllite and chlorite–actinolite schist, grey marble and calcareous schist, minor metabasite

lPI$_{bp}$ — Dark phyllite or schist, locally graphitic, black marble and dark quartzite

### LOWER CAMBRIAN

MOHICAN AND BADSHOT FORMATIONS (Undivided):

ЄMB — Grey marble, dolomitic marble, tan, green and grey calcareous or dolomitic schist, minor pelitic schist and micaceous quartzite

### LATEST NEOPROTEROZOIC TO LOWER CAMBRIAN

HAMILL GROUP (Informal subdivisions)

UPPER QUARTZITE

ЄH$_{uq}$ — (Designated age based on correlation with western and northern Purcell anticlinorium) Lower part: discontinuous lower white orthoquartzite (equivalent to MT. GAINER FORMATION of Fyles and Eastwood, 1962, and Read and Wheeler, 1976); upper part: intercalated grey and tan quartzite, micaceous quartzite and pelitic schist (MARSH ADAMS FORMATION of Fyles, 1964)

LOWER VOLCANIC AND CLASTIC

PЄH$_{v}$ — Metabasite and biotite–chlorite schist, minor dolostone and grit

PЄH$_{qs}$ — Laterally discontinuous intercalated green, dark and white quartzite, dark schist, feldspathic grit and minor pebble conglomerate

## WESTERN PURCELL ANTICLINORIUM
## (HANGINGWALL OF MOUNT FORSTER THRUST)

### LOWER PALEOZOIC

INDEX FORMATION (Birnam Creek syncline)

lPI$_{bp}$ — Black phyllite and quartzite, minor dark grey marble

# LEGEND

' FAULT)

ietabasite

l of Fyles
ɔn quartzite,

iic grit and

ƽT)

EASTERN PURCELL ANTICLINORIUM
(FOOTWALL OF MOUNT FORSTER THRU

**PALEOZOIC**

**UPPER DEVONIAN**

STARBIRD FORMATION

| Ds | Fossiliferous grey dolostone and limestone, and interbedded black limestone ai siltstone, minor argillite and microcrystalline quartz |
|---|---|

**MIDDLE DEVONIAN**

MOUNT FORSTER FORMATION

| DMF | Red dolomitic sandstone, grit and minor conglomerate, metabasalt and argillite |
|---|---|

**NEOPROTEROZOIC**

HORSETHIEF CREEK GROUP (Undivided)

| PHC | Grit and pebble conglomerate, grey marble and calcareous slate, dolostone an |
|---|---|

TOBY FORMATION

| PT | Diamictite and breccia with argillaceous, sandy or calcareous matrix, grit and local mafic volcanic rocks |
|---|---|

**MESOPROTEROZOIC**

**PURCELL SUPERGROUP**

MOUNT NELSON FORMATION (Undivided)

| PMN | Well-stratified succession of white and light brown quartzite, purple and green commonly laminated dolostone, minor grit and pebble to boulder conglomerate |
|---|---|

DUTCH CREEK FORMATION

| | |
|---|---|
| PDC | Undivided Dutch Creek Formation |
| PDCa | Grey, black and green argillite, rhythmically interbedded siltstone and argillite |
| PDCs | Brown, green and grey sandstone, dolomitic sandstone, dolostone and argillite |
| PDCas | Grey and green argillite, siltstone and argillaceous sandstone, minor sandstone |

INTRUSIVE ROCKS

ID

ANTICLINORIUM
FORSTER THRUST)

erbedded black limestone and argillite, red and brown

ite, metabasalt and argillite

careous slate, dolostone and slate

alcareous matrix, grit and slate,

quartzite, purple and green argillite, siltstone,
e to boulder conglomerate

ed siltstone and argillite

ne, dolostone and argillite

ndstone, minor sandstone and argillaceous dolostone

OCKS

BRITISH COLUMBIA
Ministry of Employment and Investment
Energy and Minerals Division
Geological Survey Branch

OPEN F

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GEOLO
WEST-CENTRAL
BRITISH

NTS 82K/2

G
Maria

Scal

Metres 1000 0 10

SUMMAR

The west-central Purcell Mountains of southeastern were deposited on the western margin of ancestral Nort strata were subsequently deformed, regionally metamorph Mesozoic terrane accretion.

The map-area is divisible into three distinct stratig eastern Purcell anticlinorium in the footwall of the Moun of the Mount Forster thrust and the Kootenay Arc. New anticlinorium, in order to investigate the evolution of Pr and footwall of the Mount Forster thrust, as outlined by contrasts between the Purcell anticlinorium and the mor and Reesor (1973).

In the Purcell anticlinorium, the Mesoproterozoic Pu Windermere Supergroup, which comprises the Toby Form Horsethief Creek Group comprises immature siliciclastic submarine fan setting. The Horsethief Creek Group is la northwestern successions that can be correlated using a abrupt, NE-trending facies and thickness change in the northwest-side-down normal fault that was active durin the Horsethief Creek Group (see figure below). Other fa basin deepened to the north and west and was more d presence of apparently conformable metabasites in the

BRITISH
COLUMBIA

of Employment and Investment
ınd Minerals Division
al Survey Branch

# OPEN FILE 1996-16

(Sheet 3 of 3)

# GEOLOGY OF THE
# WEST-CENTRAL PURCELL MOUNTAINS
# BRITISH COLUMBIA

**NTS 82K/2, 7 & PARTS OF 8, 10**

**Geology by**
**Marian J. Warren**

Scale 1:75 000

Metres 1000 0 1000 2000 3000 4000 Metres

## SUMMARY OF GEOLOGY

ntral Purcell Mountains of southeastern British Columbia expose Mesoproterozoic to Lower Paleozoic rocks that ı the western margin of ancestral North America during at least two episodes of continental rifting. These quently deformed, regionally metamorphosed and intruded by at least two suites of granitic plutons during accretion.

a is divisible into three distinct stratigraphic and structural domains. These comprise, from east to west, the ıticlinorium in the footwall of the Mount Forster thrust, the western Purcell anticlinorium in the hangingwall ster thrust and the Kootenay Arc. New mapping by M. Warren has focused in the western Purcell rder to investigate the evolution of Proterozoic and Paleozoic stratigraphic contrasts between the hangingwall ıe Mount Forster thrust, as outlined by Reesor (1973) and Root (1987), and the evolution of structural the Purcell anticlinorium and the more complexly deformed Kootenay Arc, as documented by Fyles (1964) ).

l anticlinorium, the Mesoproterozoic Purcell Supergroup is unconformably overlain by the Neoproterozoic group, which comprises the Toby Formation and the previously undivided Horsethief Creek Group. The Group comprises immature siliciclastic and resedimented carbonate strata that were deposited primarily in a ting. The Horsethief Creek Group is laterally variable, but it can be divided into southeastern and essions that can be correlated using a conspicuous marker unit. The two successions are separated by an ıg facies and thickness change in the upper Horsethief Creek drainage, interpreted to record motion on a wn normal fault that was active during the deposition of the Toby Formation and perhaps of the lower part of ek Group (see figure below). Other facies changes in the northwestern part of the map-area indicate that the

## LOWER PALEOZOIC

INDEX FORMATION (Birnam Creek syncline)

| IPIbp | Black phyllite and quartzite, minor dark grey marble

### LOWER CAMBRIAN

UPPER DONALD FORMATION (Blockhead Mountain syncline)

| ЄDu | Thinly interbedded silver muscovite schist or phyllite, light green chlorite schist or phyllite, tan calcareo or dolomitic schist

BADSHOT FORMATION

| ЄB | Tan or grey marble, dolomitic marble and minor calcareous or dolomitic schist

MOHICAN FORMATION

| ЄM | Thinly interbedded tan dolomitic schist and dolostone, minor impure quartzite, light green phyllite and quartzose schist

| ЄMm | Three laterally continuous, thickly cross-bedded coarse quartz arenite intervals, up to 10 m thick each; separated by thinner intervals of rusty-weathering schist (Blockhead Mountain syncline)

## LATEST NEOPROTEROZOIC TO LOWER CAMBRIAN

Designated ages based on correlation with partly fossiliferous strata in the Dogtooth Range (Evans, 1933; Kubli and Simony, 1992)

HAMILL GROUP (Informal subdivisions)

| ЄH | Undivided Hamill Group

UPPER QUARTZITE

| ЄHuq | Thinly to medium bedded, commonly cross-bedded white quartzite, minor dark quartzite and pelite (low part); interbedded dark, pink and green quartzite and dark pelite (upper part). Lower and upper parts separated by distinct rusty-weathering pelitic interval in Blockhead Mountain syncline

MIDDLE PELITE

| PЄHp | Rusty-weathering dolomitic schist, black pelite, dolostone and blue quartz pebble conglomerate, minor dolostone breccia and cobble conglomerate, cross-bedded orthoquartzite (Blockhead Mountain syncline); fine-grained rusty-weathering pelitic schist, green chlorite schist and minor dolomitic schist (Mt. Cauldron syncline)

LOWER QUARTZITE

| PЄHlq | Thickly bedded to massive orthoquartzite, coarse quartz arenite and grit, locally feldspathic grit and pebble conglomerate. Tabular cross beds common

BASAL GRIT

| PЄHg | White and light grey tabular and trough cross-bedded quartz and arkosic sandstone, grit and conglome containing abundant blue and purple quartz, and minor interbedded tan dolostone or dolarenite, black pelite and dark green chlorite schist (Blockhead Mountain and Mt. Cauldron synclines); light grey micaceous quartzite, grey phyllite or schist and tan dolostone (Eyebrow syncline)

## NEOPROTEROZOIC

## WINDERMERE SUPERGROUP

HORSETHIEF CREEK GROUP (Informal subdivisions)

| PHC | Undivided Horsethief Creek Group

SOUTHERN AND EASTERN SUCCESSION

UPPER PELITE

| PHCup | Thinly interbedded graded micaceous quartzite and grey slate, phyllite or muscovite schist, minor grit or pebble conglomerate, minor

or phyllite, tan calcareous

it green phyllite and

up to 10 m thick each,
ncline)

nge (Evans, 1933;

uartzite and pelite (lower
ower and upper parts are
line

conglomerate, minor
ad Mountain syncline);
nitic schist (Mt.

eldspathic grit and

ne, grit and conglomerate
or dolarenite, black
ies); light grey

schist, minor grit and

grey and green argillite, siltstone and argillaceous sandstone, minor s

# INTRUSIVE ROCKS

**CRETACEOUS** (Reesor, 1973; D. A. Archibald, unpublished 40Ar/39Ar data, Qu

Kqm — Primarily quartz monzonite, some granodiorite (Bugaboo batholith), gra

**MIDDLE JURASSIC TO EARLY CRETACEOUS** (Reesor, 1973; Warren, 1996

JKgd — Primarily granodiorite, some quartz monzonite, quartz diorite, with horn

**MIDDLE JURASSIC** (Smith et al., 1992)

Area intruded by biotite- and hornblende-bearing granitic dykes and s

**LOWER PALEOZOIC ?** (Fyles, 1964)

lPa — Coarse-grained amphibolite (intrudes Hamill Group and Mohican Format

lPu — Ultramafic to mafic dikes or sills (intrude lower part of Index Formatio

# MAP SYMBOLS

Stratigraphic contact (defined, approximate, assumed)

Approximate location of lateral facies changes in Horsethief Creek Group

Thrust fault (defined, approximate, assumed)

Neoproterozoic to Early Paleozoic normal fault (defined, approximate, assumed)

Mesozoic normal fault (defined, approximate, assumed)

Normal fault; age unconstrained (defined, approximate, assumed)

Fault; motion sense undetermined (defined, approximate, assumed)

Axial trace of upright anticline or antiform, syncline or synform [F2]

Axial trace of overturned anticline or antiform, syncline or synform [F2]

Axial trace of early refolded anticline, syncline in Kootenay Arc [F1]

Location of structural cross-section

Location of longitudinal section along axial trace of fold

Glacier

## ...IVE ROCKS

...ublished 40Ar/39Ar data, Queen's University)

...ite (Bugaboo batholith), granite, quartz diorite

...(Reesor, 1973; Warren, 1996; unpublished 40Ar/39Ar data)

...ite, quartz diorite, with hornblende- and pyroxene-rich rims

...earing granitic dykes and sills

... Group and Mohican Formation in Kootenay Arc)

...ower part of Index Formation in Kootenay Arc)

## ... SYMBOLS

...d)

...orsethief Creek Group

...ned, approximate, assumed)

...ed)

...te, assumed)

...nate, assumed)

... or synform [F2]

...ine or synform [F2]

...otenay Arc [F1]

A — A'

... fold X— —X

... anticlinorium, the Mesoproter...
Windermere Supergroup, which comprises the Tot...
Horsethief Creek Group comprises immature silic...
submarine fan setting. The Horsethief Creek Grou...
northwestern successions that can be correlated ...
abrupt, NE-trending facies and thickness change...
northwest-side-down normal fault that was activ...
the Horsethief Creek Group (see figure below). C...
basin deepened to the north and west and was ...
presence of apparently conformable metabasites ...
stratigraphic succession beneath these strata ac...
deposition of the upper part of the Windermere ...
trending, have been documented in the footwall ...

The Windermere Supergroup is overlain by a ...
Cambrian Hamill Group, in the hangingwall of th...
shallow marine or fluvial units, more feldspathic ...
continuous and more mature shallow marine qua...
Horsethief Creek Group. In the Kootenay Arc, the...
Purcell anticlinorium. The lower part, whose base...
The upper shallow marine quartzite is considered ...
of the Mount Forster thrust, the Windermere Su...
(Reesor, 1973; Root, 1987; Pope, 1990). These ...
"Windermere High" during Early Paleozoic time.

The lateral variations in the Hamill Group in...
lower units of the Hamill Group were deposited ...
faults and were separated by an uplifted and ed...
exposed on the eastern limb of the Blockhead N...
thrust fault that cuts the regional cleavage, but ...
down normal fault. It marks the eastern limit of ...
the eastern Hamill basin. The fault that bounded ...
fault, a Mesozoic structure that separates the K...
Group and the overlying Mohican and Badshot F...
them after normal faulting had ceased. Regional ...
and the Rocky Mountains (Warren, 1996) suggest...
Cambrian (Placentian) and that the upper quartz...

Normal motion on the fault that bounded th...
resulting in the deposition of shallow-water silici...
Formation) and deeper-water, immature clastic ...
development of this fault and the basin that it ...
and other Lower Paleozoic carbonate strata, whic...
Purcell anticlinorium.

The boundary between the Kootenay Arc and...
Arc boundary fault) that juxtaposes a domain of ...
against a domain of upright, more open folds. S...
coupled with geobarometric data (Warren, 1996), ...
component of late dextral strike-slip motion as ...
foliations and dykes and sills dated at 173 Ma (...
that both the F1 and F2 folds in the eastern K...
blind detachment that propagated beneath the K...
location of the Kootenay Arc boundary fault, and...
location of one of the syn-Hamill normal faults.


Bennett, S. (1986): Geology of the Mount... Petroleum Resources, Preliminary Map...

Evans, C. S. (1933): Brisco-Dogtooth Ma... Summary Report 1932, Part A II, p. ...

Fyles, J. T. (1964): Geology of the Dunca... of Energy, Mines and Petroleum Reso...

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... Creek Group is laterally variable, but it can be divided into southeastern and
uccessions that can be correlated using a conspicuous marker unit. The two successions are separated by an
nding facies and thickness change in the upper Horsethief Creek drainage, interpreted to record motion on a
-down normal fault that was active during the deposition of the Toby Formation and perhaps of the lower part of
Creek Group (see figure below). Other facies changes in the northwestern part of the map-area indicate that the
l to the north and west and was more distal from the sedimentary source. Abrupt E-W thickness changes and the
parently conformable metabasites in the upper part of the Horsethief Creek Group, and a change in the
ccession beneath these strata across the Four Squatters fault, suggest that normal faulting also occurred during
ne upper part of the Windermere Supergroup. Similar syn-Windermere structures, both NE-trending and N-
been documented in the footwall of the Mount Forster thrust by Root (1987) and Pope (1990).

mere Supergroup is overlain by a succession of more mature siliciclastic rocks, the Neoproterozoic to Lower
l Group, in the hangingwall of the Mount Forster thrust. The Hamill Group is divisible into three lower
or fluvial units, more feldspathic toward the base, that thin, fine and pinch out to the west and an upper,
more mature shallow marine quartzite unit that unconformably overlies the lower units or rests directly on the
k Group. In the Kootenay Arc, the succession of Hamill Group strata is distinctly different from that in the western
rium. The lower part, whose base is not exposed, includes mafic metavolcanic rocks and possibly turbidites.
ow marine quartzite is considered equivalent to the upper quartzite in the Purcell anticlinorium. In the footwall
Forster thrust, the Windermere Supergroup is unconformably overlain by Upper Cambrian to Devonian strata
Root, 1987; Pope, 1990). These overstepping unconformities record uplift and eastward tilting of the
gh" during Early Paleozoic time.

variations in the Hamill Group in both the western Purcell anticlinorium and in the Kootenay Arc imply that the
he Hamill Group were deposited in two separate basins that were bounded to the east by west-dipping normal
separated by an uplifted and eastward-tilted block of Horsethief Creek Group strata. The Earl Grey Pass fault,
eastern limb of the Blockhead Mtn. syncline in the hangingwall of the Mount Forster thrust, is a west-side-up
cuts the regional cleavage, but stratigraphic relationships across it indicate that it was previously a west-side-
ult. It marks the eastern limit of exposure of the Hamill Group, and it could be the normal fault that bounded
nill basin. The fault that bounded the western Hamill basin is closely followed by the Kootenay Arc boundary
: structure that separates the Kootenay Arc from the Purcell anticlinorium. The upper quartzite of the Hamill
overlying Mohican and Badshot Formations were deposited across these basins and the uplifted block between
al faulting had ceased. Regional correlation with fossiliferous strata in other parts of the Purcell, the Selkirk
Mountains (Warren, 1996) suggests that the lower units of the Hamill Group are Neoproterozoic to lower Lower
ntian) and that the upper quartzite is upper Lower Cambrian (Waucoban) in age.

ion on the fault that bounded the western Hamill basin may have been renewed during Early Paleozoic time,
deposition of shallow-water siliciclastic and carbonate rocks to the east (Lower Cambrian upper Donald
deeper-water, immature clastic and volcanic rocks to the west (Lower Paleozoic Lardeau Group). The
this fault and the basin that it bounded may have strongly influenced mineralisation of the Badshot Formation
Paleozoic carbonate strata, which host numerous sulphide deposits in the Kootenay Arc but not in the adjacent
um.

y between the Kootenay Arc and the Purcell anticlinorium is a steep, locally mylonitic fault zone (the Kootenay
lt) that juxtaposes a domain of complexly refolded, high-amplitude, west-verging ductile folds (Fyles, 1964)
of upright, more open folds. Stratigraphic relationships across the fault where it intersects Duncan Lake,
barometric data (Warren, 1996), imply west-side-up thrust motion, but motion sense indicators imply a
e dextral strike-slip motion as well. Re-examinantion of cross-cutting relationships between axial planar
es and sills dated at 173 Ma (Smith et al., 1992), coupled with palinspastic analysis (Warren, 1996) indicate
and F2 folds in the eastern Kootenay Arc developed in response to west-verging crustal shortening above a
that propagated beneath the Kootenay Arc, but not the Purcell anticlinorium, in mid-Jurassic time. The
ootenay Arc boundary fault, and perhaps of the eastern limit of the blind detachment, was controlled by the
f the syn-Hamill normal faults.

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oterozoic and Paleozoic Development of the Cordilleran Divergent Margin; unpublished Ph.D.

## HORSETHIEF CREEK GROUP (Informal subdivisions)

| | |
|---|---|
| PHC | Undivided Horsethief Creek Group |

### SOUTHERN AND EASTERN SUCCESSION

#### UPPER PELITE

| | |
|---|---|
| $PHC_{up}$ | Thinly interbedded graded micaceous quartzite and grey slate, phyllite or muscovite schist, minor grit pebble conglomerate, minor dolostone and coarse dolomitic sandstone toward top |

#### UPPER GRIT

| | |
|---|---|
| $PHC_{ug}$ | Thinly to thickly bedded fining– and thinning–upward succession of feldspathic grit and pebble conglomerate and interbedded pelite and micaceous quartzite |

#### MARKER UNIT (Undivided)

| | |
|---|---|
| $PHC_{m}$ | Thinly bedded tan dolomitic siltstone (upper part) and competent, homogeneous green argillite or gree micaceous quartzite (lower part) |

#### LOWER PELITE

| | |
|---|---|
| $PHC_{lp}$ | Brown–weathering thinly interbedded slate to schist, siltstone or quartzite, dolomitic siltstone or schist, minor grit lenses |

#### LOWER CARBONATE

| | |
|---|---|
| $PHC_{lc}$ | Light grey marble and dark calcareous slate or schist |

### NORTHERN AND WESTERN SUCCESSION

#### UPPER CLASTIC

| | |
|---|---|
| $PHC_{a}$ | Garnet amphibolite, apparently concordant with sedimentary contacts |
| $PHC_{uc}$ | Pelitic schist, calc–silicate schist, tourmaline–muscovite schist, graded quartzite, grit, calcareous grit o marble, locally intruded by felsic dykes or sills (west of Four Squatters anticline only) |
| $PHC_{c}$ | Dark grey to black marble and siliceous marble or dark calcareous schist (in contact with $PHC_{um}$); mi grey siliceous marble or dolostone and dolomitic coarse sandstone and pebble conglomerate (stratigrap position uncertain) |

#### UPPER MARKER UNIT

| | |
|---|---|
| $PHC_{um}$ | Rhythmically interbedded dolomitic siltstone, cream dolostone and green phyllite or slate with minor lenses of carbonate conglomerate; locally capped by black pelite and/or marble |

#### LOWER MARKER UNIT

| | |
|---|---|
| $PHC_{lm}$ | Competent homogeneous green argillite, siltstone or schist and minor dolomitic siltstone |

#### MIDDLE PELITE

| | |
|---|---|
| $PHC_{mp}$ | Brown–weathering pelite, siltstone or quartz schist, minor grit |

#### MIDDLE CARBONATE

| | |
|---|---|
| $PHC_{mc}$ | Thickly interbedded and laterally continuous intervals of light to medium grey marble, siliceous marble and dark grey calcareous grit; thickens to north and west |
| $PHC_{mcm}$ | Two tan or white marker intervals of competent dolomitic and/or feldspathic grit and pebble conglomer (west of Four Squatters anticline only) |

#### LOWER CLASTIC

| | |
|---|---|
| $PHC_{ga}$ | Thickly interbedded, laterally discontinuous intervals of light feldspathic grit or pebble conglomerate and darker calcareous grits and marble conglomerate (eastern exposures); interbedded green and grey slat phyllite or schist, minor grit and pebble conglomerate and siliceous marble (western exposures); proportion of argillite increases to west and north |

#### LOWER CALCAREOUS CLASTIC

| | |
|---|---|
| $PHC_{lcg}$ | Calcareous and dolomitic grit, conglomerate, coarse sandstone, slate and siliceous marble containing abundant blue and white quartz and feldspar (mappable arkosic grit and conglomerate lenses in Horse Creek valley shown as dotted contact) |

#### TOBY FORMATION

minor grit and

ble

ite or green

or schist,

eous grit and

PHCum ); medium
(stratigraphic

minor

s marble

conglomerate

merate and
grey slate,
es);

ntaining
in Horsethief

Axial trace of early refolded anticline, syncline in Kootenay Arc [F1]

Location of structural cross-section

Location of longitudinal section along axial trace of fold

Glacier

## STRUCTURAL DATA

(Deformation "phases" not necessarily correlative across map a

Bedding; top known (upright, overturned)

Bedding; top unknown (inclined, vertical, horizontal)

Dominant, regional penetrative schistosity or cleavage [S2] (inclined, vertical)

Bedding-parallel schistosity or cleavage [S1], older than regional penetrative foliation S2 (inclined, vertical)

Bedding-parallel schistosity [S1], locally stronger than S2 (inclined, vertical)

Composite foliation [S2/S1 or S3/S2] (inclined, vertical)

Late spaced or crenulation cleavage [east] or schistosity [west], locally developed [S3] (inclined, vertical)

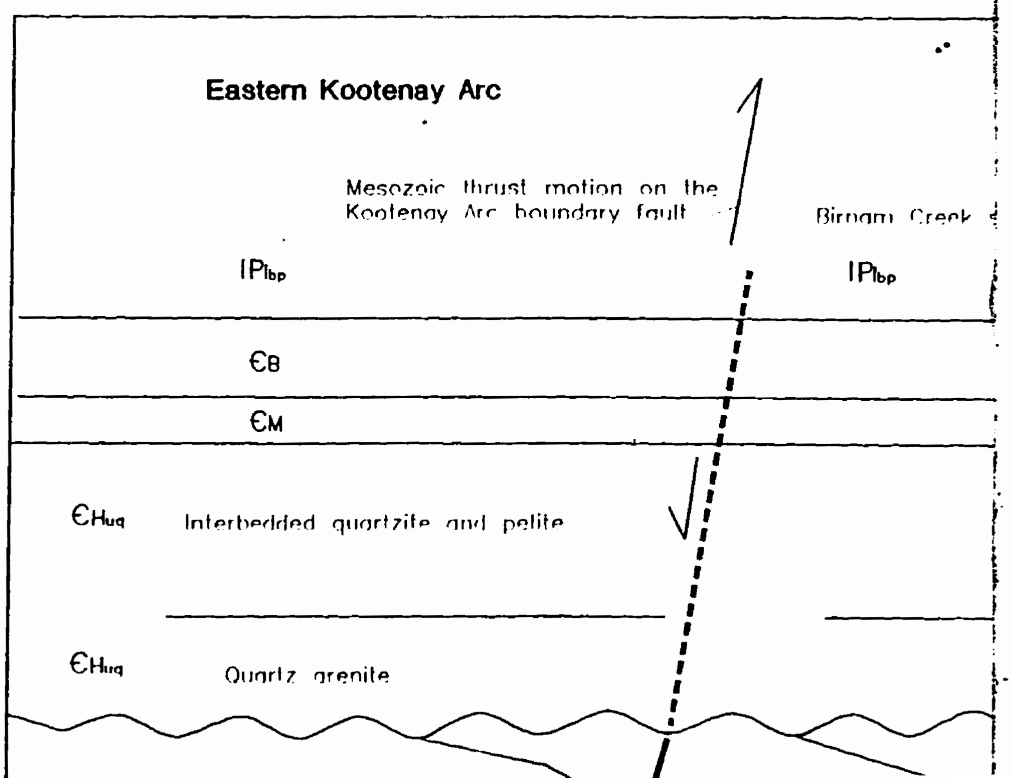
Igneous foliation, not necessarily primary (inclined, vertical)

Sinistral kink bands [S4] (inclined, vertical)

Intersection lineation, minor fold axis or pebble stretching lineation

WEST/ NORTHWEST

## SCHEMATIC S



A——AA'

'old X– –XX

## JRAL DATA

ecessarily correlative across map area)

[S2] (inclined, vertical)

ɔn regional

S2 (inclined, vertical)

ıl)

ity [west],

cal)

ng lineation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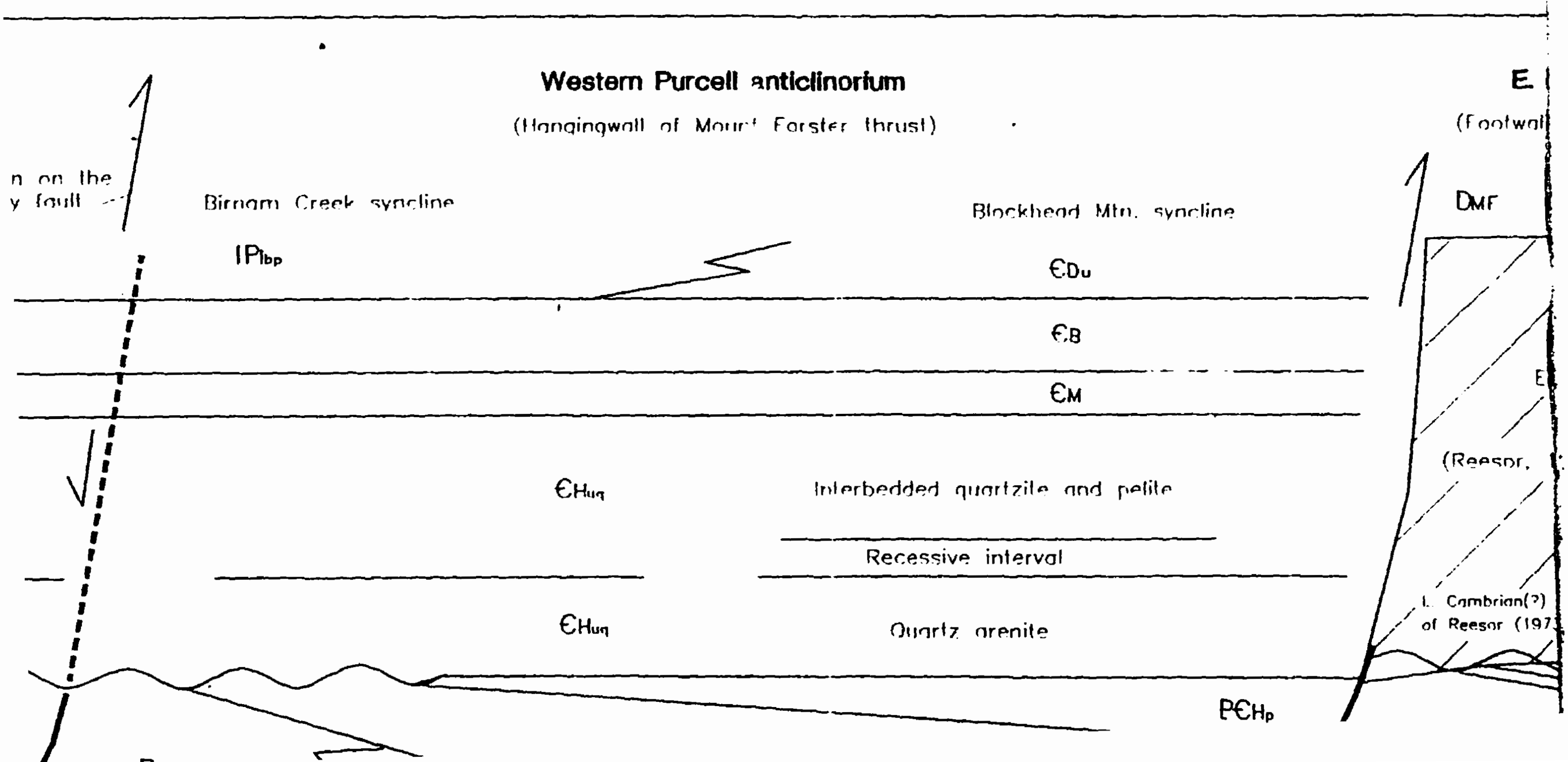
Canadian Helicopters (Nelson) and Pemberton
equipment and expertise was provided by D.
the Greenlaw family (Meadow Creek) and Kath
assistance and hospitality. The author is grat
phases of this project.


## SCHEMATIC SUMMARY OF STRATIGRAPHIC RELATIONSHIPS



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sity.

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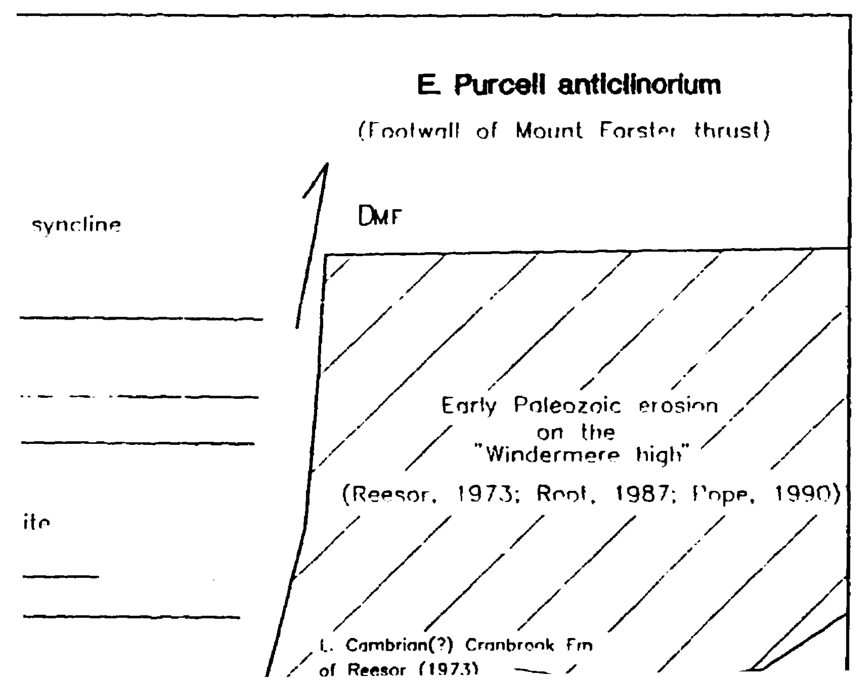
and hospitality. The author is grateful for input and assistance from Maurice Colpron during all
this project.

# ONSHIPS

EAST/ SOUTHEAST



| | |
|---|---|
| PHCum | Rhythmically interbedded dolomitic siltstone, cream dolostone and green phyllite or slate with minor lenses of carbonate conglomerate; locally capped by black pelite and/or marble |

LOWER MARKER UNIT

| | |
|---|---|
| PHCIm | Competent homogeneous green argillite, siltstone or schist and minor dolomitic siltstone |

MIDDLE PELITE

| | |
|---|---|
| PHCmp | Brown-weathering pelite, siltstone or quartz schist, minor grit |

MIDDLE CARBONATE

| | |
|---|---|
| PHCmc | Thickly interbedded and laterally continuous intervals of light to medium grey marble, siliceous marble and dark grey calcareous grit; thickens to north and west |
| PHCmcm | Two tan or white marker intervals of competent dolomitic and/or feldspathic grit and pebble conglomer (west of Four Squatters anticline only) |

LOWER CLASTIC

| | |
|---|---|
| PHCgn | Thickly interbedded, laterally discontinuous intervals of light feldspathic grit or pebble conglomerate and darker calcareous grits and marble conglomerate (eastern exposures); interbedded green and grey slate phyllite or schist, minor grit and pebble conglomerate and siliceous marble (western exposures); proportion of argillite increases to west and north |

LOWER CALCAREOUS CLASTIC

| | |
|---|---|
| PHClcg | Calcareous and dolomitic grit, conglomerate, coarse sandstone, slate and siliceous marble containing abundant blue and white quartz and feldspar (mappable arkosic grit and conglomerate lenses in Horset Creek valley shown as dotted contact) |

TOBY FORMATION

| | |
|---|---|
| PT | Homogeneous cream dolostone |
| PTd | Diamictite, dolostone and slate; diamictite comprises well-rounded to angular pebbles to boulders primarily of quartzite, marble and dolostone in red argillaceous, grey calcareous or tan sandy matrix; upper part interbedded with and capped by feldspathic grit (Horsethief Creek valley) or by homogeneous cream dolostone (Toby and Jumbo creeks) |

## MESOPROTEROZOIC

## PURCELL SUPERGROUP

MOUNT NELSON FORMATION (Lower part only)

| | |
|---|---|
| PMN | Primarily white quartzite and tan micaceous quartzite, minor brown dolostone |

DUTCH CREEK FORMATION (Subdivisions after Root, 1987)

| | |
|---|---|
| PDC | Undivided Dutch Creek Formation |
| PDCa | Grey, black and green argillite, rhythmically interbedded siltstone and argillite |
| PDCs | Brown sandstone and minor dolostone and argillite |
| PDCas | Grey and green argillite, siltstone and argillaceous sandstone, minor sandstone and argillaceous doloston |

minor

marble

conglomerate

erate and
grey slate,
s);

taining
in Horsethief

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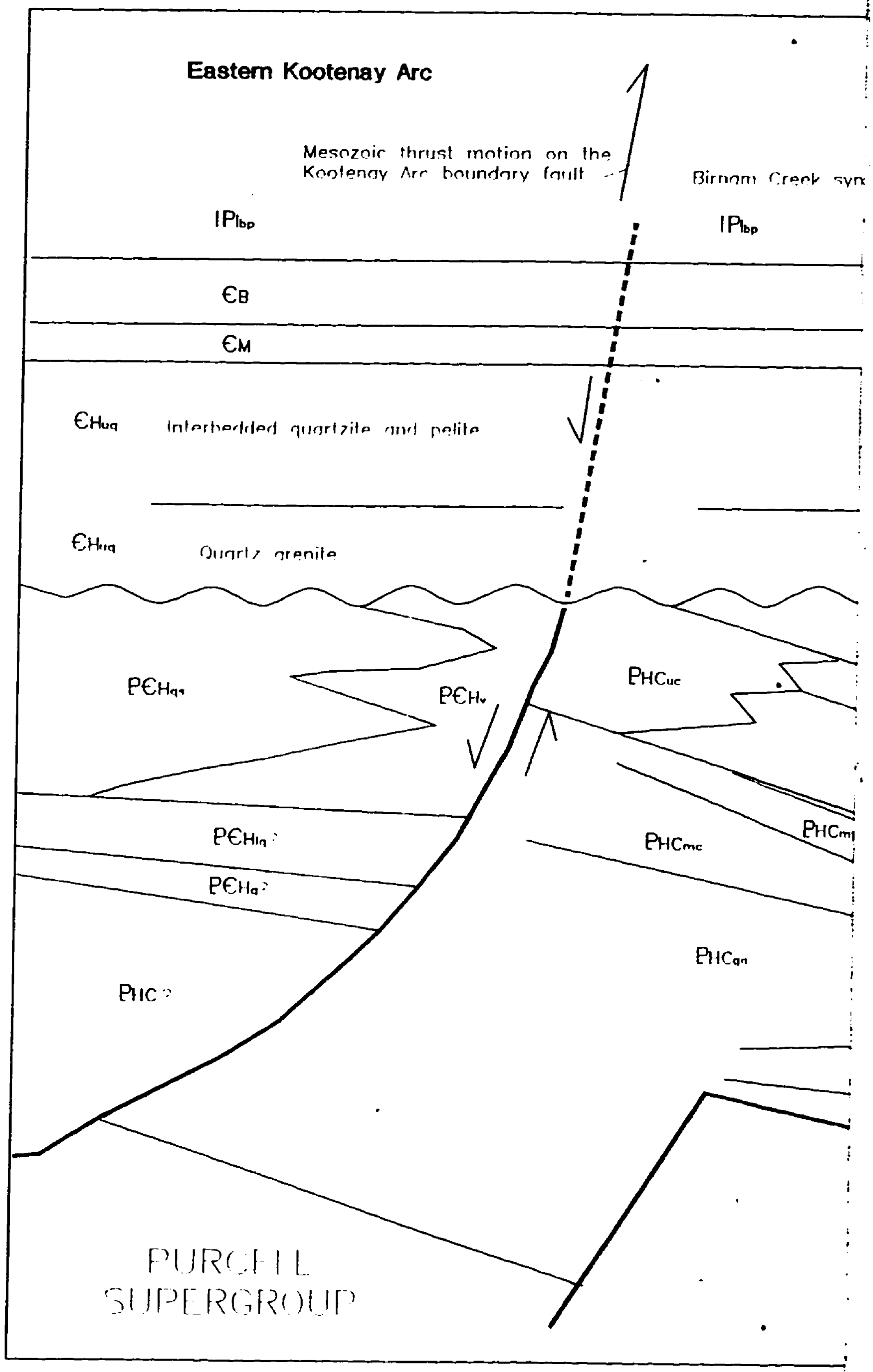
matrix;
nogeneous

dolostone

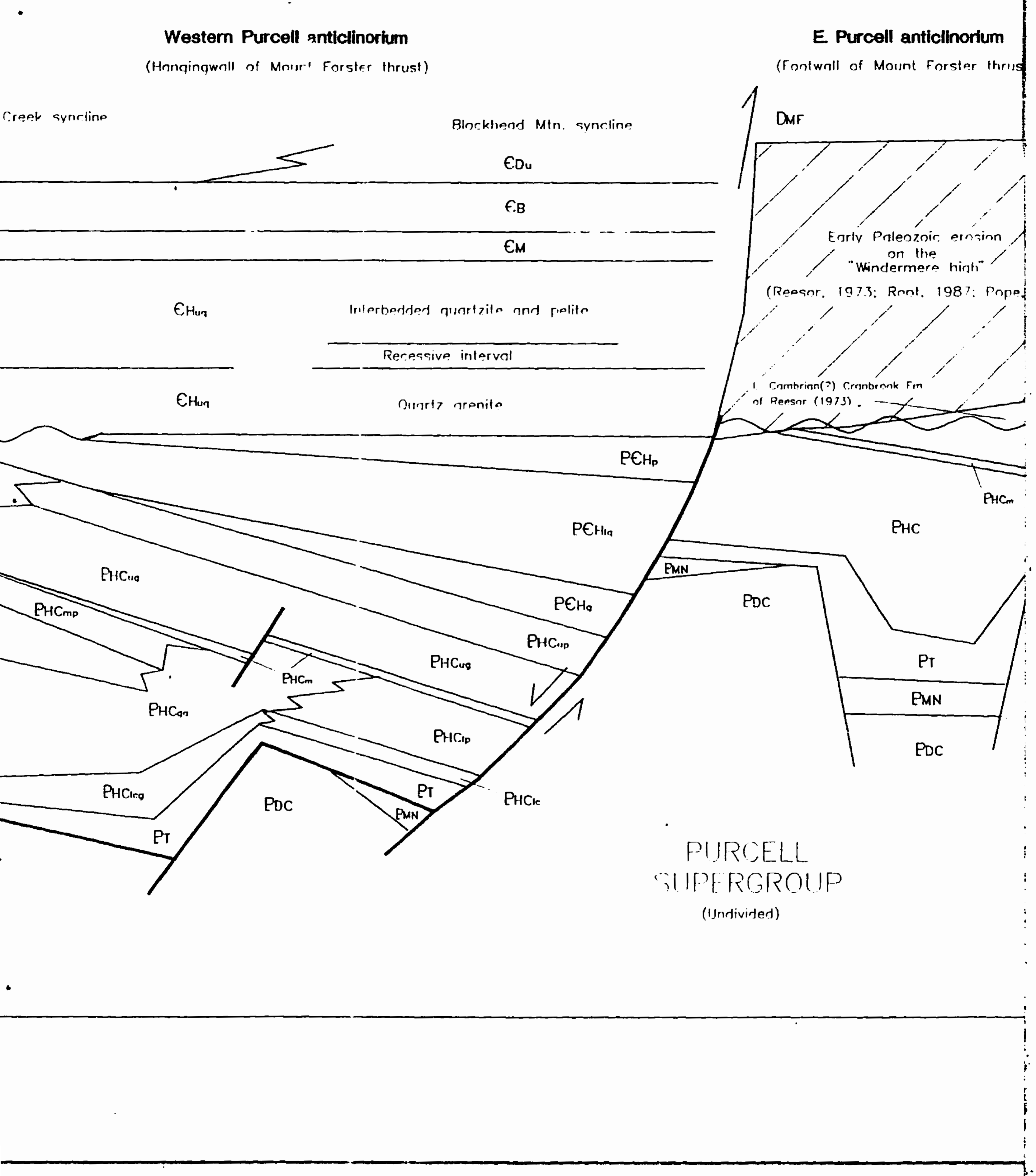
WEST/ NORTHWEST

**SCHEMATIC SU**



## C SUMMARY OF STRATIGRAPHIC RELATIONSHIPS

EAST/ SOUTHE



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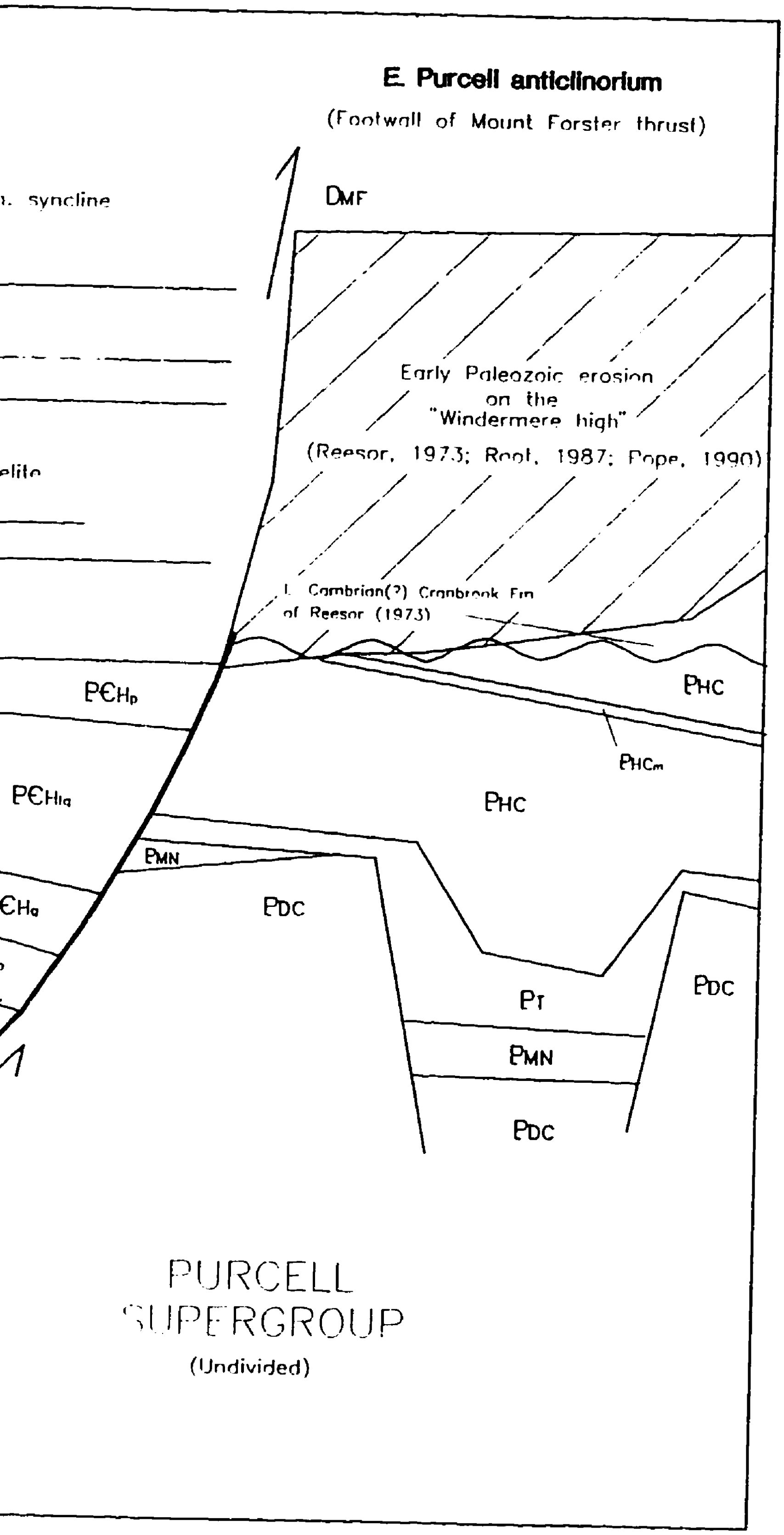

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# PLATE 4

## RESTORED REGIONAL CROSS-SE
## THE KOOTENAY ARC AND PURCE

M. J. Warren, 1

See Chapter 4 for discussi

# LEGEND

MID–CRETACEOUS

| | |
|---|---|
| Kqm | Granite and quartz monzonite |

ZOIC

MIDDLE JURASSIC

50

Scale for restored

No vertic

TE 4

OSS–SECTION THROUGH

PURCELL ANTICLINORIUM

ren, 1996

for discussion

2) RE
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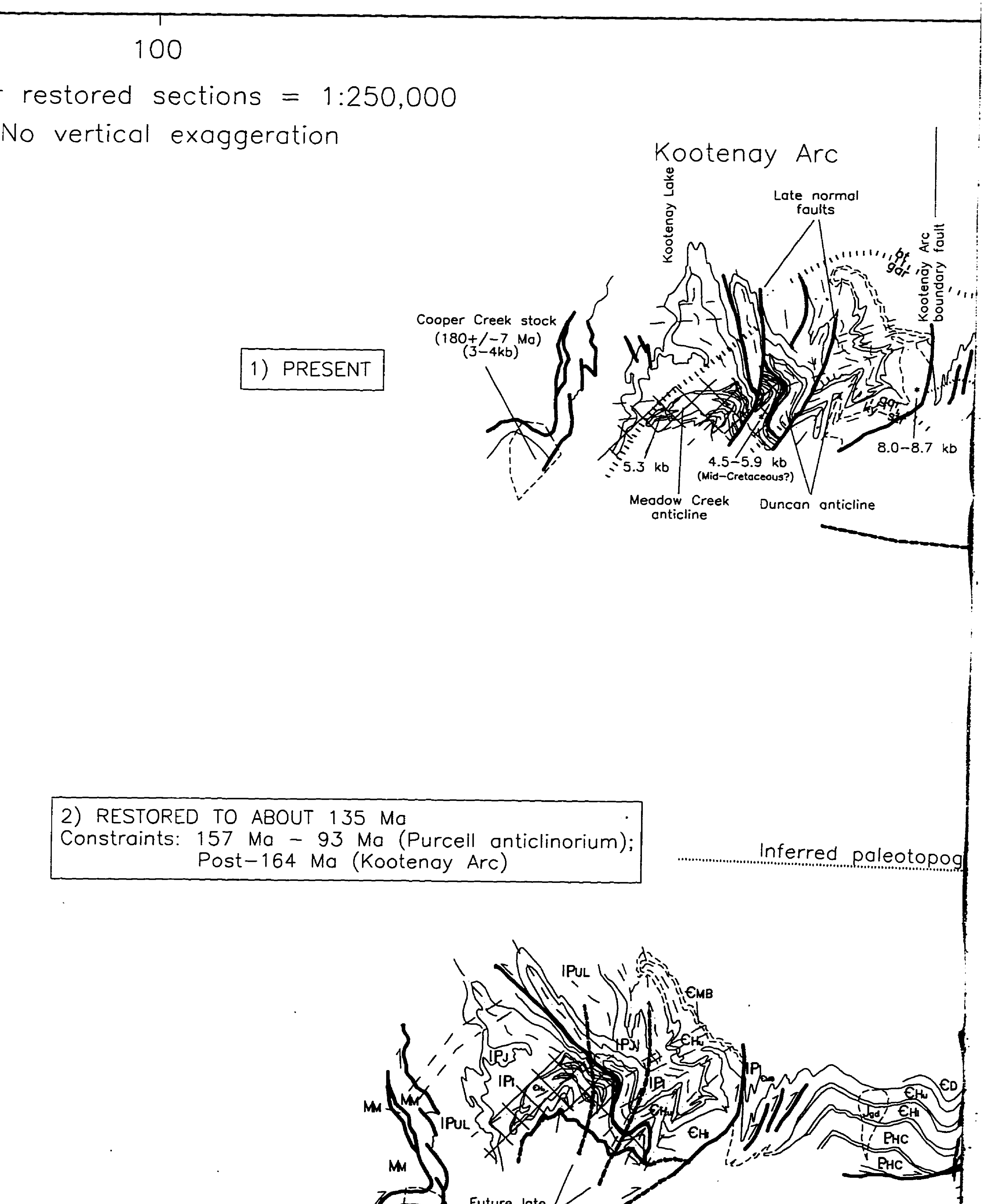
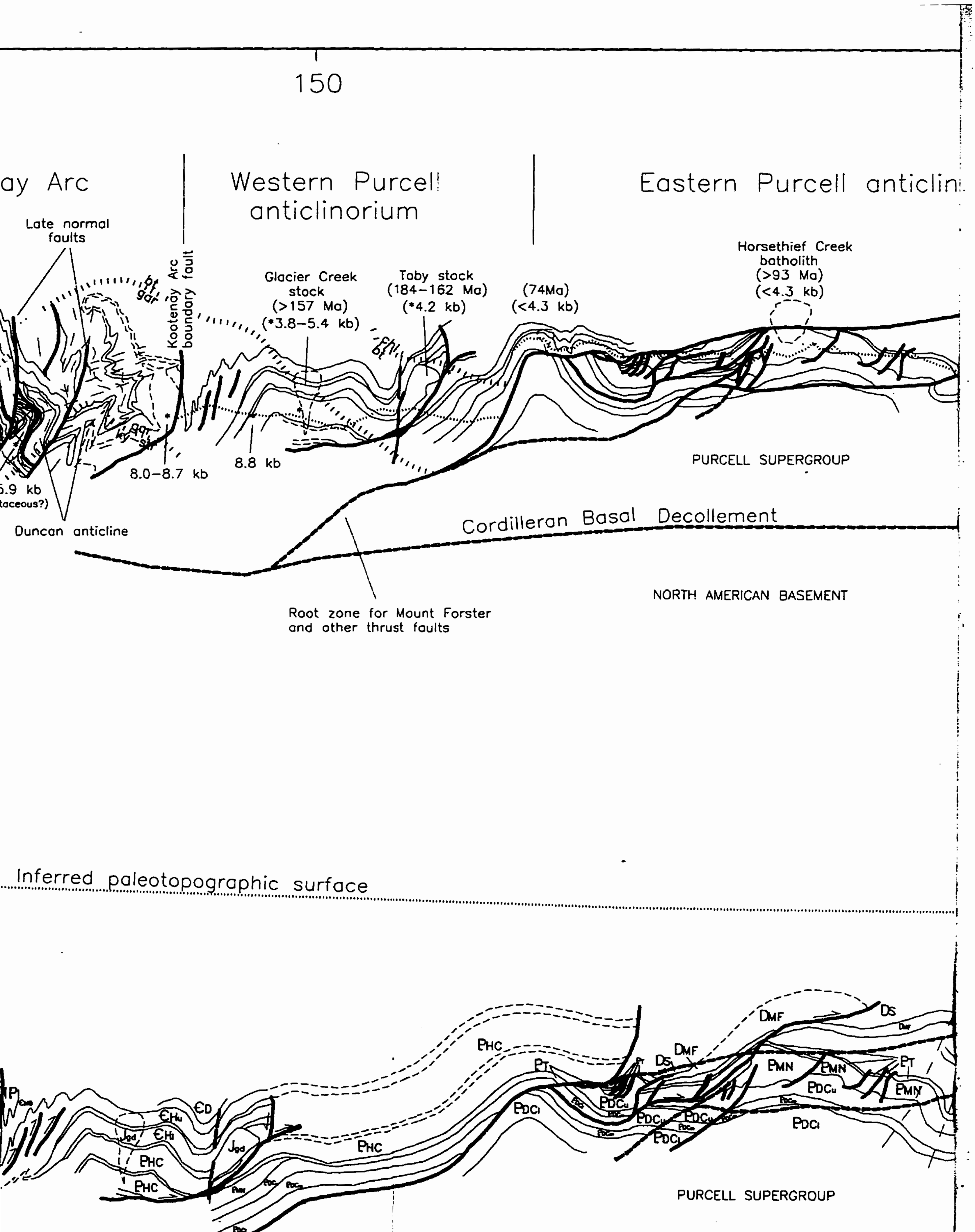
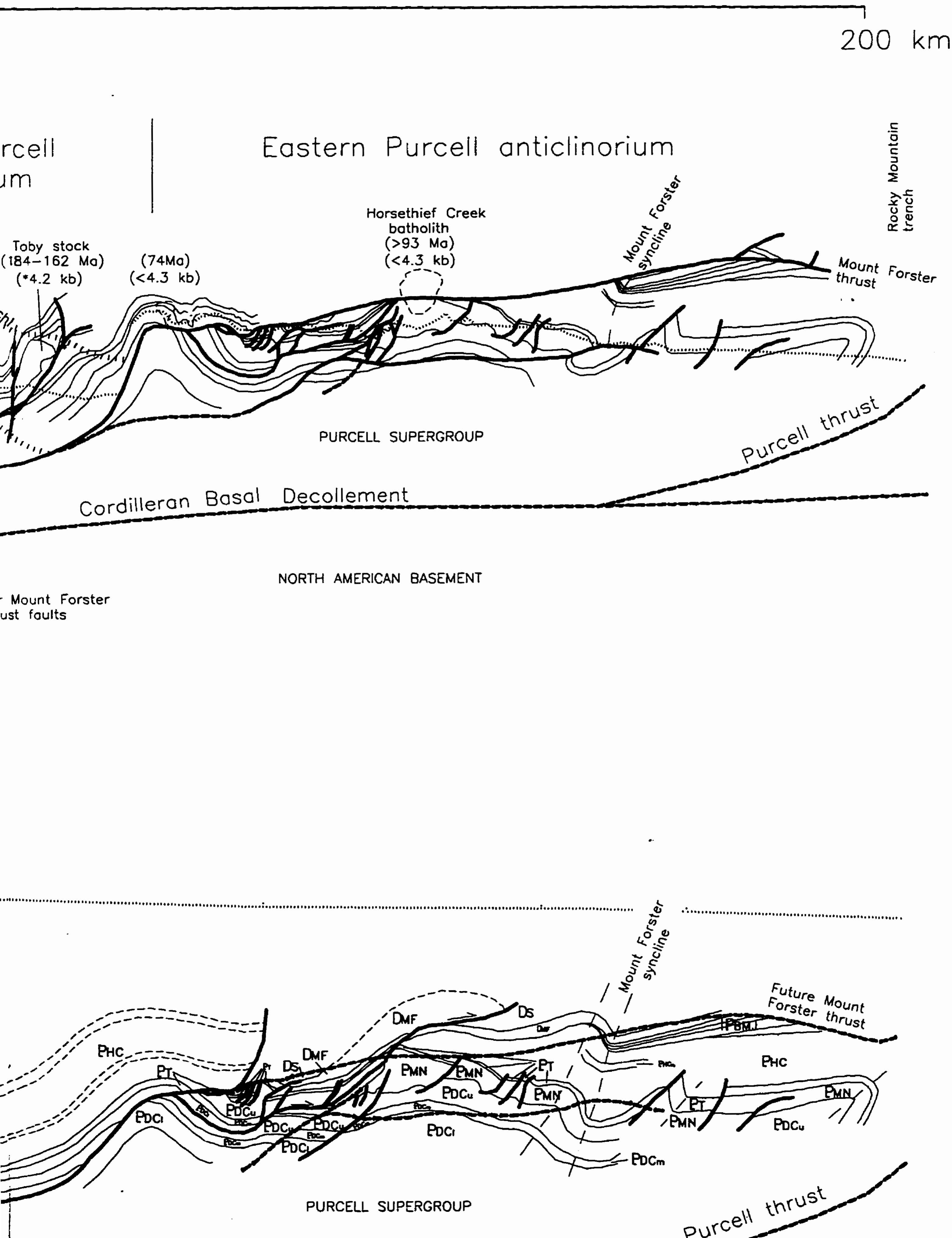
END


100
restored sections = 1:250,000
No vertical exaggeration
1) PRESENT
Kootenay Arc
Kootenay Lake
Late normal faults
Kootenay Arc boundary fault
Cooper Creek stock (180+/-7 Ma) (3-4kb)
5.3 kb
4.5-5.9 kb (Mid-Cretaceous?)
8.0-8.7 kb
Meadow Creek anticline
Duncan anticline
2) RESTORED TO ABOUT 135 Ma
Constraints: 157 Ma - 93 Ma (Purcell anticlinorium);
Post-164 Ma (Kootenay Arc)
Inferred paleotopog
IPUL
IPJ
IPI
ЄMB
ЄHu
ЄHl
ЄD
PHC
MM
Future late



150
ay Arc
Western Purcell anticlinorium
Eastern Purcell anticlin
Late normal faults
Kootenay Arc boundary fault
Glacier Creek stock (>157 Ma) (*3.8–5.4 kb)
Toby stock (184–162 Ma) (*4.2 kb)
(74Ma) (<4.3 kb)
Horsethief Creek batholith (>93 Ma) (<4.3 kb)
8.0–8.7 kb
8.8 kb
Duncan anticline
PURCELL SUPERGROUP
Cordilleran Basal Decollement
NORTH AMERICAN BASEMENT
Root zone for Mount Forster and other thrust faults
Inferred paleotopographic surface
PHC
PT
DMF
DS
PMN
PDCu
PDCl
Jgd
€D
€Hu
€Hl
PURCELL SUPERGROUP



200 km
rcell
um
Eastern Purcell anticlinorium
Rocky Mountain trench
Horsethief Creek batholith (>93 Ma) (<4.3 kb)
Mount Forster syncline
Toby stock (184–162 Ma) (*4.2 kb)
(74Ma) (<4.3 kb)
Mount Forster thrust
PURCELL SUPERGROUP
Purcell thrust
Cordilleran Basal Decollement
NORTH AMERICAN BASEMENT
Mount Forster
ust faults
Mount Forster syncline
Future Mount Forster thrust
DMF
DS
PHC
PT
PMN
PDCu
PDCl
PDCm
PURCELL SUPERGROUP
Purcell thrust

## MESOZOIC

| | |
|---|---|
| Kqm | Granite and quartz monzonite |

MIDDLE JURASSIC

| | |
|---|---|
| Jgd | Granodiorite |
| | Granitic dikes |

## PALEOZOIC

### WESTERN SUCCESSION

MISSISSIPPIAN

| | |
|---|---|
| MM | Milford Group |

LOWER PALEOZOIC

| | |
|---|---|
| lPUL | Upper Lardeau Group, undivided (Broadview, Ajax, Triune, Sharon Creek Formations) |
| lPJ | Jowett Formation |
| lPI | Index Formation (west) |
| ЄD | Upper Donald Formation (east) |

LOWER CAMBRIAN

| | |
|---|---|
| ЄMB | Badshot and Mohican Formations |
| ЄHu | Upper Hamill Group |

## PROTEROZOIC

NEOPROTEROZOIC TO LOWER CAMBRIAN

| | |
|---|---|
| ЄHl | Lower Hamill Group (impure clastic and volcanic in Kootenay and Basal Grit, Lower Quartzite and Middle Pelite in Purcell a |

NEOPROTEROZOIC (Windermere Supergroup)

| | |
|---|---|
| PHC | Horsethief Creek Group, undivided |
| PHCm | Horsethief Creek Group Marker Unit |
| PT | Toby Formation |

MESOPROTEROZOIC (Purcell Supergroup)

| | |
|---|---|
| PMN | Mount Nelson Formation |
| PDCu | Dutch Creek Formation (Upper) |
| PDCm | Dutch Creek Foramtion (Middle) |
| PDCl | Dutch Creek Formation (Lower) |

UPPER

Ds

MIDDLE

DMF

LOWER

lPBMJ

LOWER

Єc

mations)

ast)

canic in Kootenay Arc
Pelite in Purcell anticlinorium)

## EASTERN SUCCESSION

UPPER DEVONIAN

| | |
|---|---|
| Ds | Starbird Formation |

MIDDLE DEVONIAN

| | |
|---|---|
| DMF | Mount Forster Formation |

LOWER PALEOZOIC

| | |
|---|---|
| lPBMJ | INCLUDES: Beaverfoot Fm. (O–S), MacKay Fm. (U€–O) and Jubilee Fm. (M€–U€) |

LOWER CAMBRIAN

| | |
|---|---|
| €c | "Cranbrook Formation" (considered equivalent to the Upper Hamill Group) |

Contact

Fault

Axial Surface

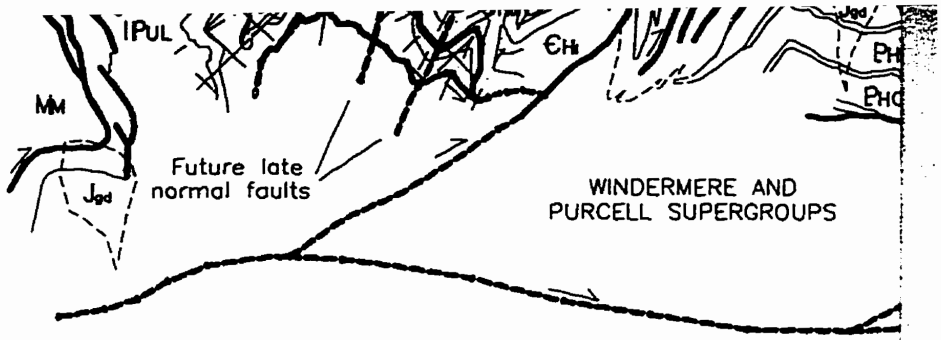
Inferred or restored fault

Inferred contact or
restored axial surface

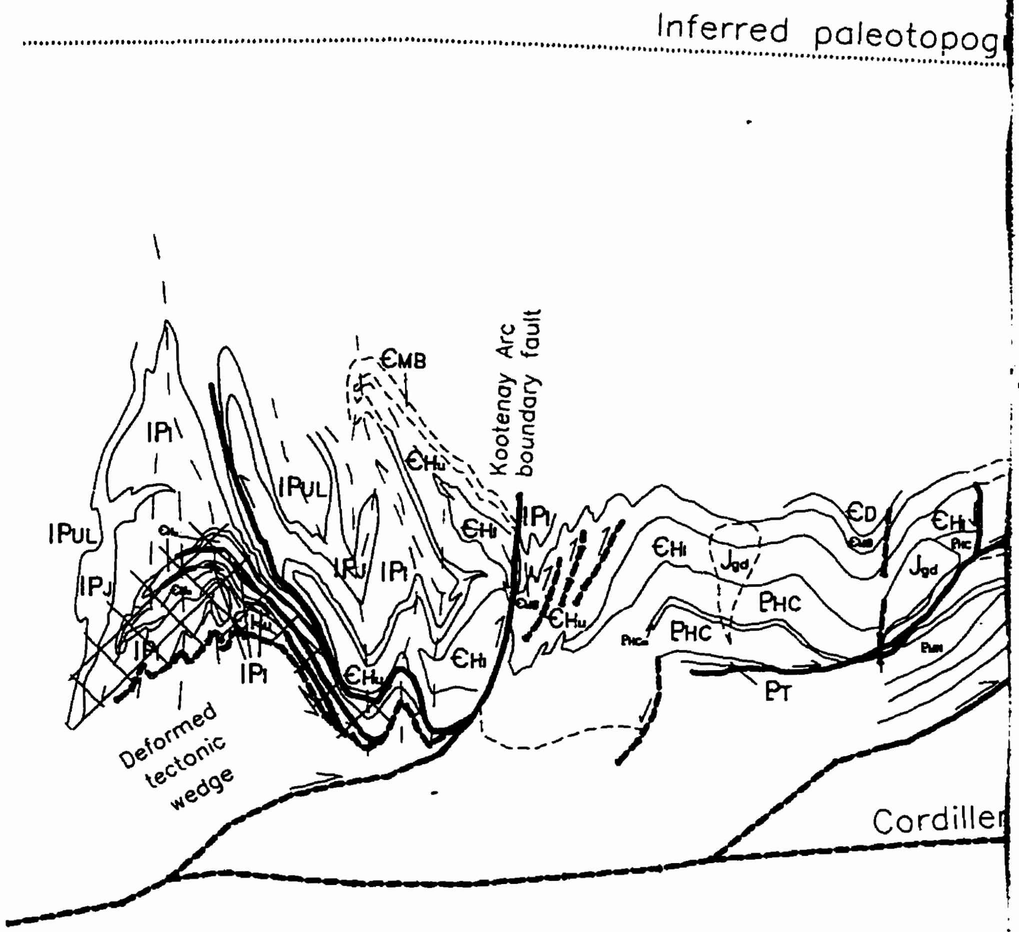
Regional metamorphic isograd

Topographic surface

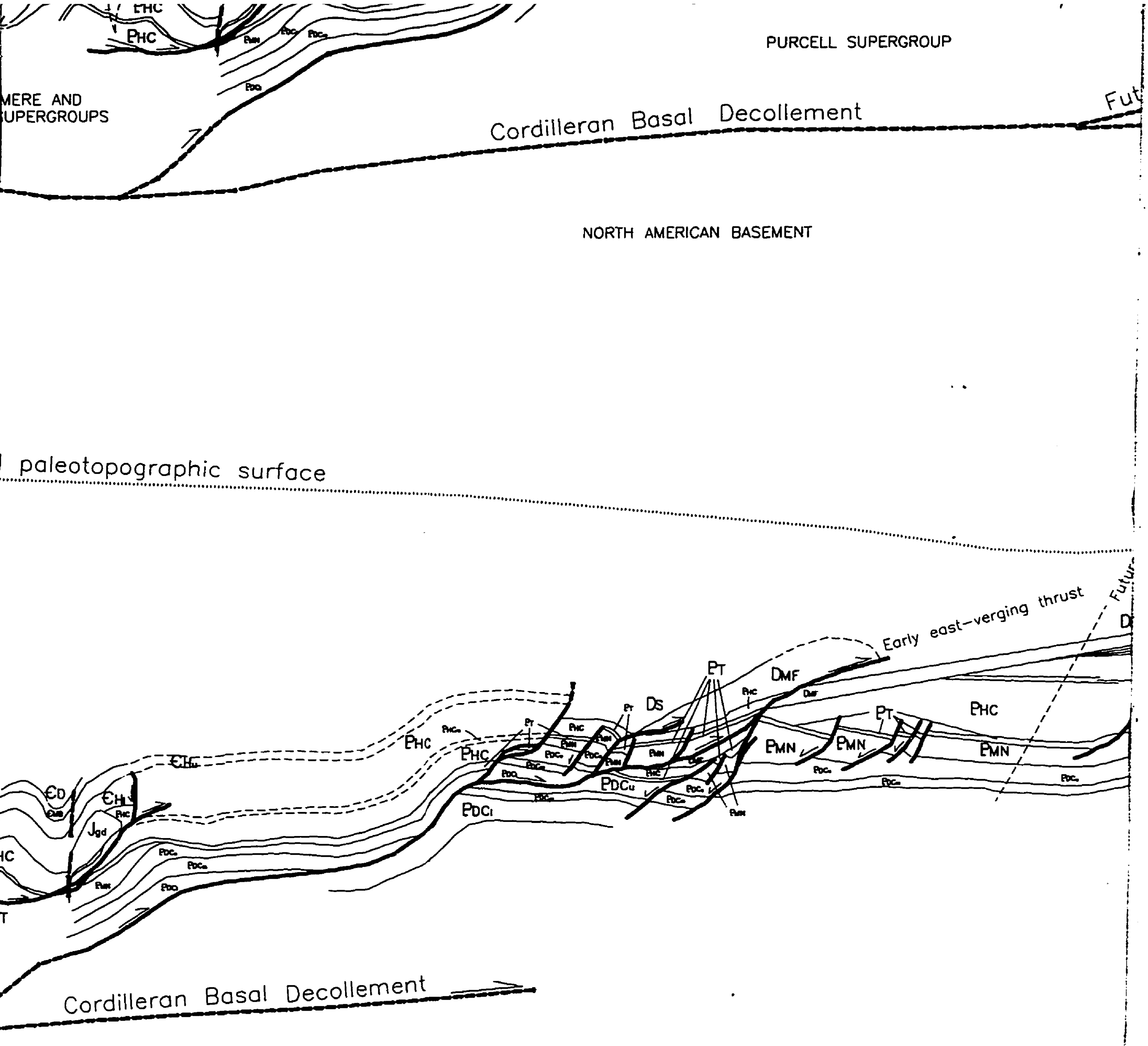
Inferr

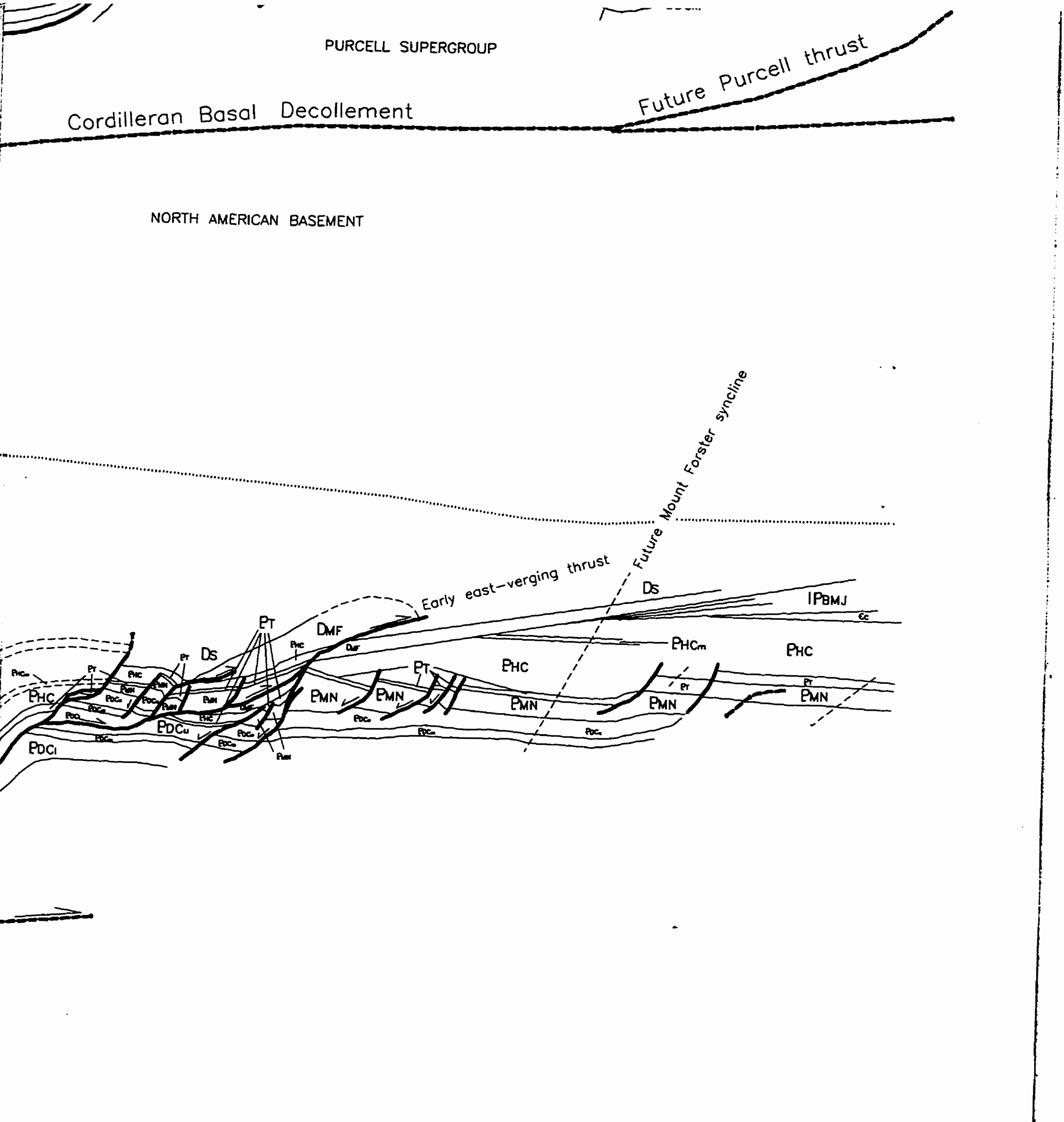


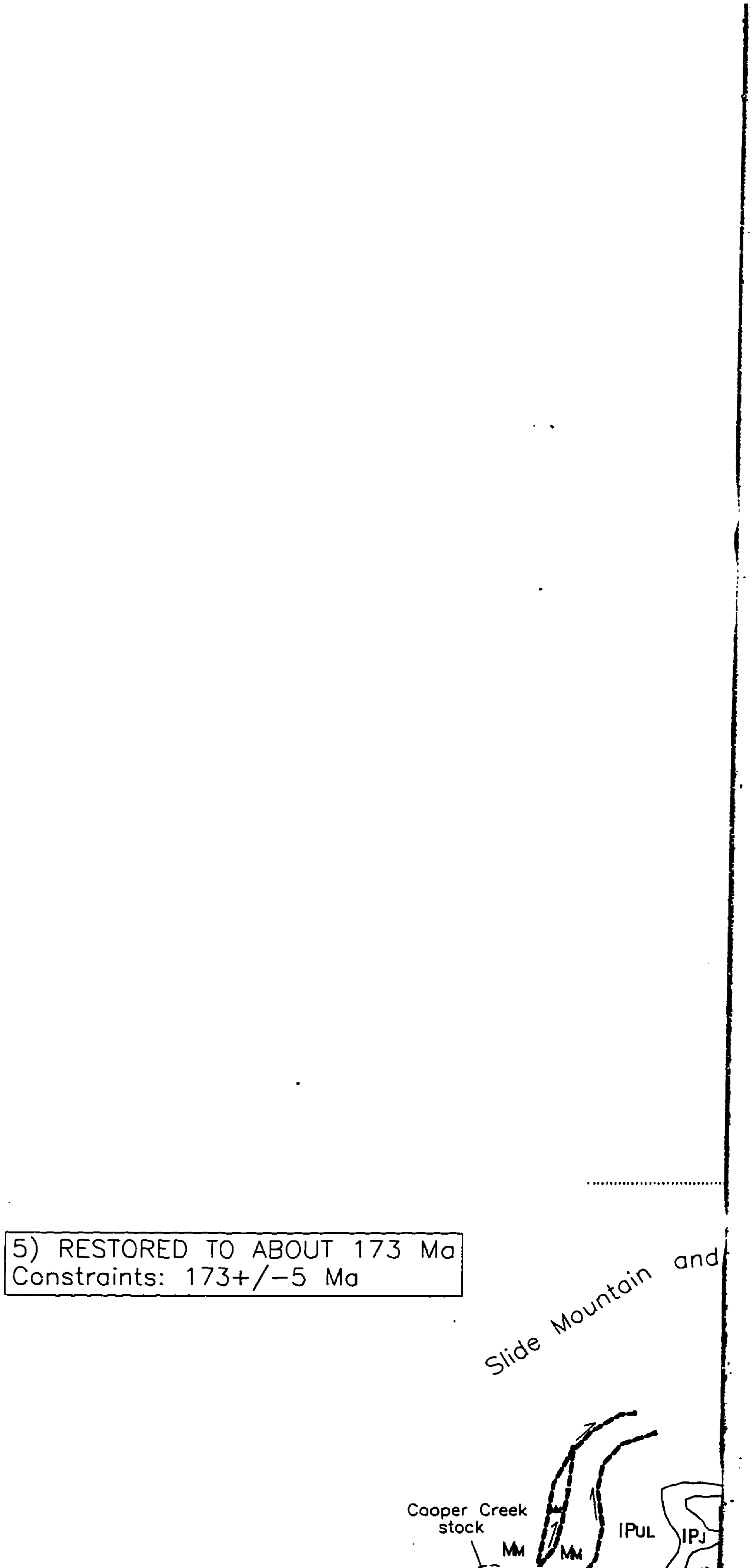

3) RESTORED TO ABOUT 155 Ma
Constraints: 157 Ma – 93 Ma (Purcell anticlinorium)
Post–164 Ma (Kootenay Arc)

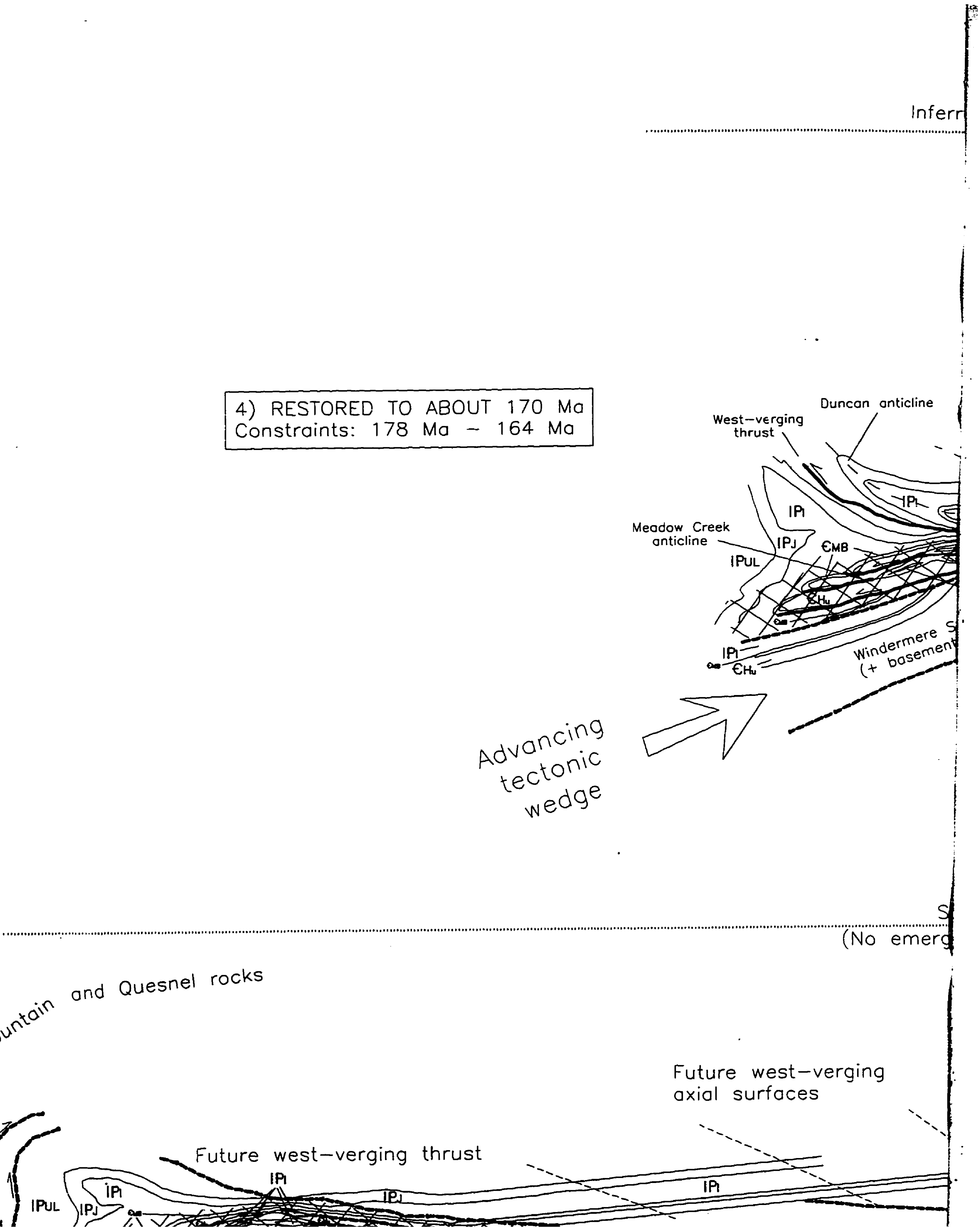



Inferred paleotopographic

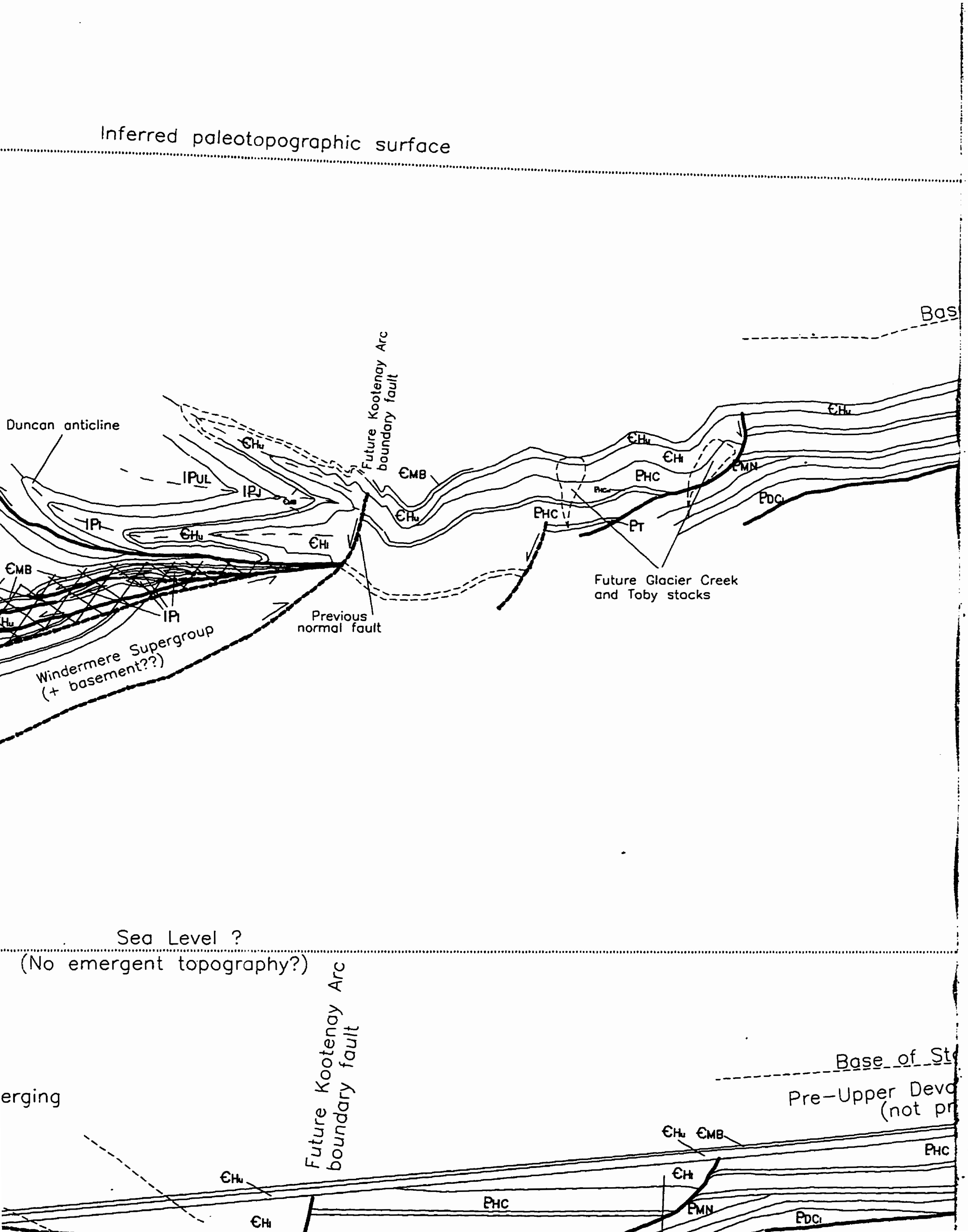

PHC
MERE AND
UPERGROUPS
PURCELL SUPERGROUP
Cordilleran Basal Decollement
NORTH AMERICAN BASEMENT
paleotopographic surface
Early east-verging thrust
PT
DMF
DS
PHC
PMN
PMN
PMN
PHC
PDCu
PDCl
Jgd
Cordilleran Basal Decollement


PURCELL SUPERGROUP
Cordilleran Basal Decollement
Future Purcell thrust
NORTH AMERICAN BASEMENT
Future Mount Forster syncline
Early east-verging thrust
Ds
IPBMJ
PHCm
PHC
DMF
PT
PMN
PDCu
PDCl

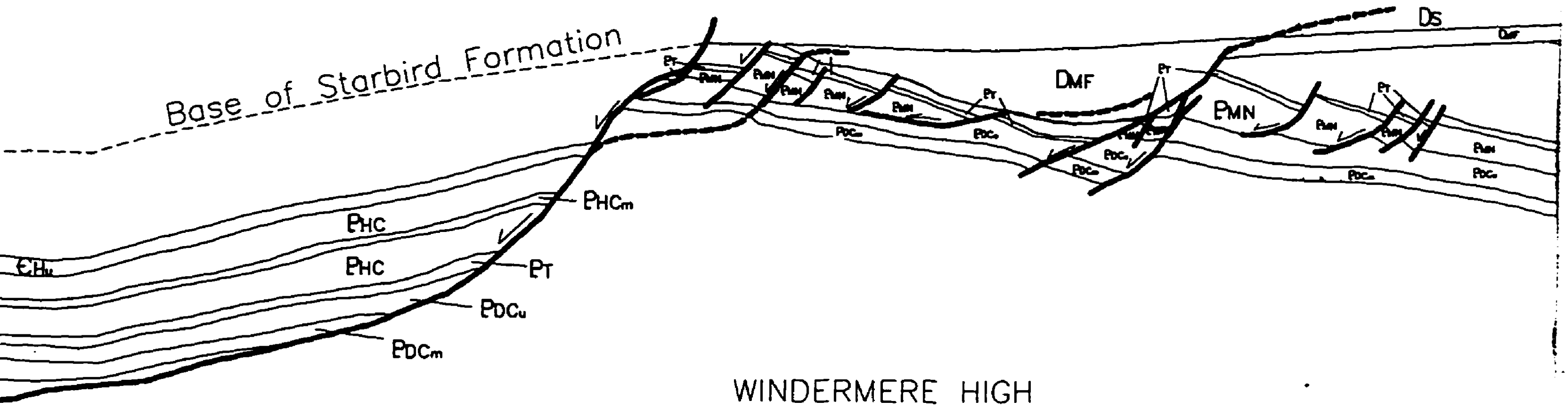

5) RESTORED TO ABOUT 173 Ma
Constraints: 173+/-5 Ma
Slide Mountain and
Cooper Creek stock
MM
MM
IPUL
IPJ



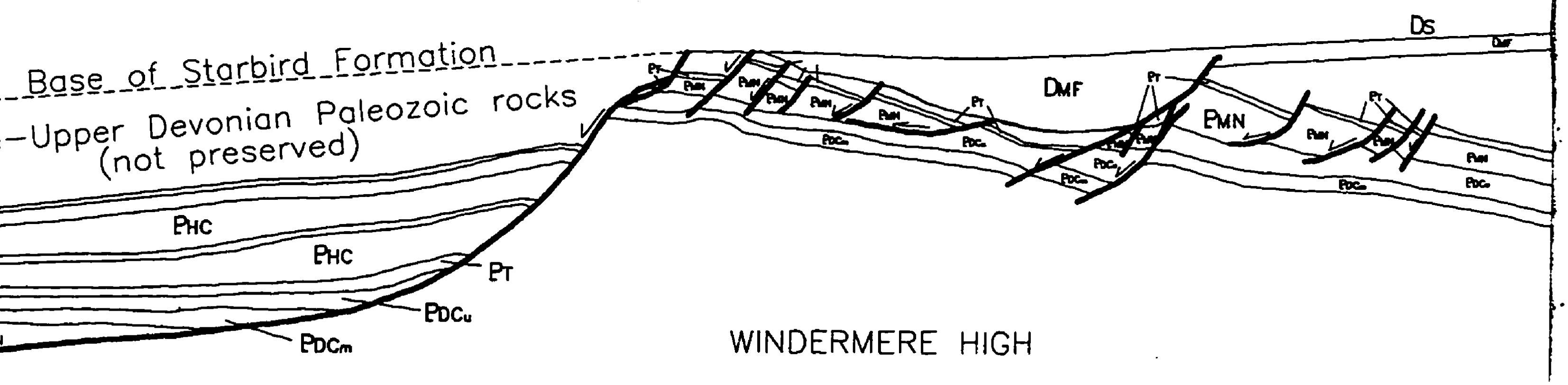
4) RESTORED TO ABOUT 170 Ma
Constraints: 178 Ma – 164 Ma
Duncan anticline
West–verging thrust
Meadow Creek anticline
Windermere
Advancing tectonic wedge
and Quesnel rocks
Future west–verging axial surfaces
Future west–verging thrust



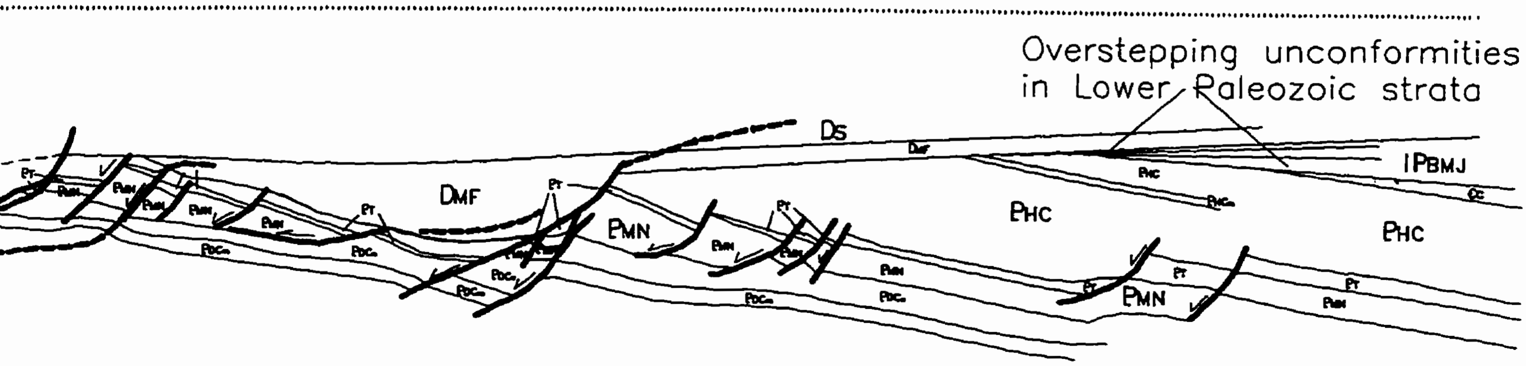
Inferred paleotopographic surface
Duncan anticline
Future Kootenay Arc boundary fault
€MB
€Hu
€Hl
IPUL
IPJ
IPl
PHC
PMN
PDCl
PT
Future Glacier Creek and Toby stocks
Previous normal fault
Windermere Supergroup (+ basement??)
Sea Level ?
(No emergent topography?)
Future Kootenay Arc boundary fault
€Hu €MB
PHC
€Hl
PMN
PDCl



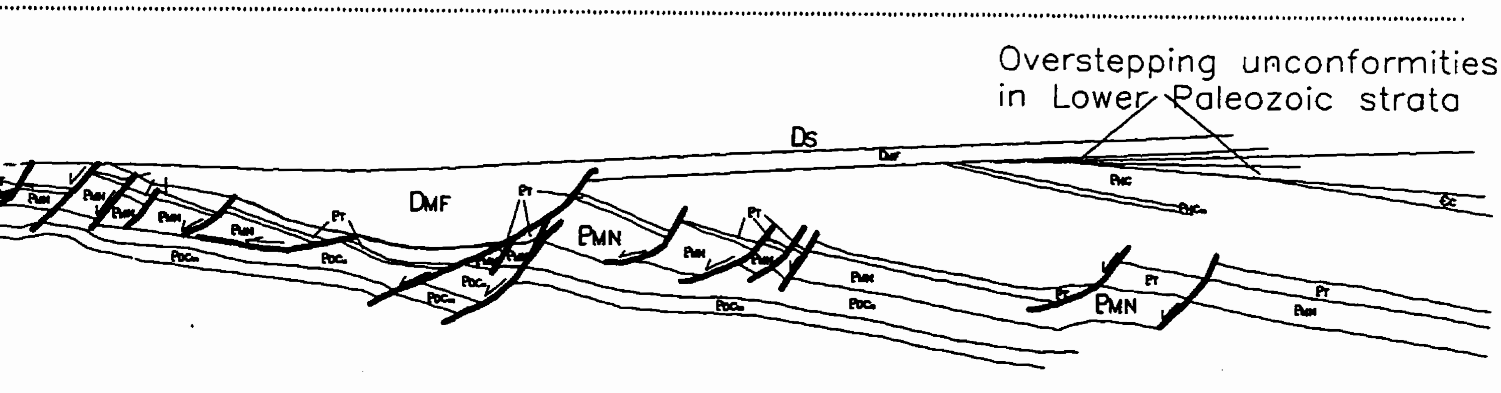
Base of Starbird Formation
Ds
DMF
PMN
PHCm
PHC
PHC
PT
PDCu
PDCm
WINDERMERE HIGH

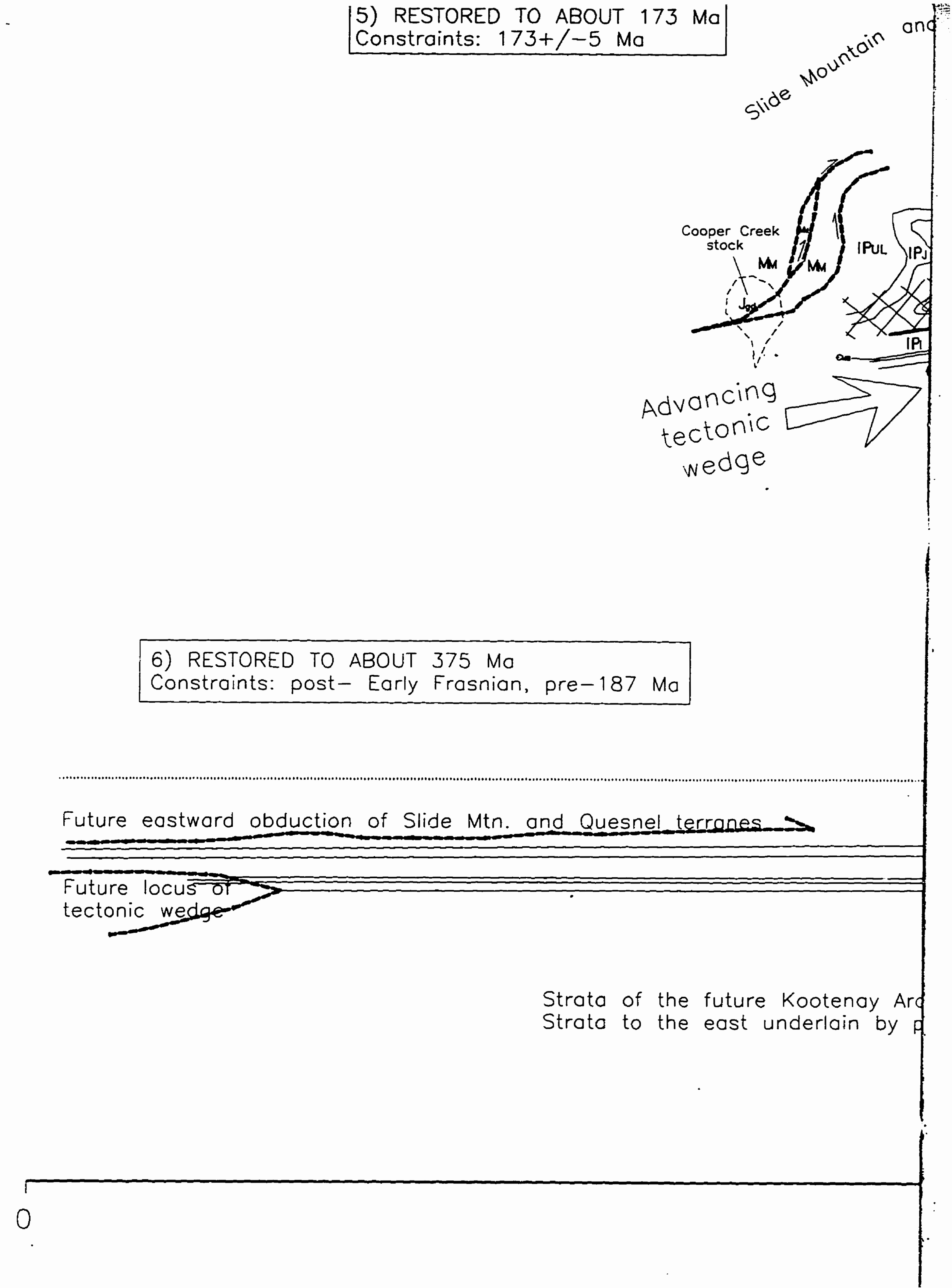



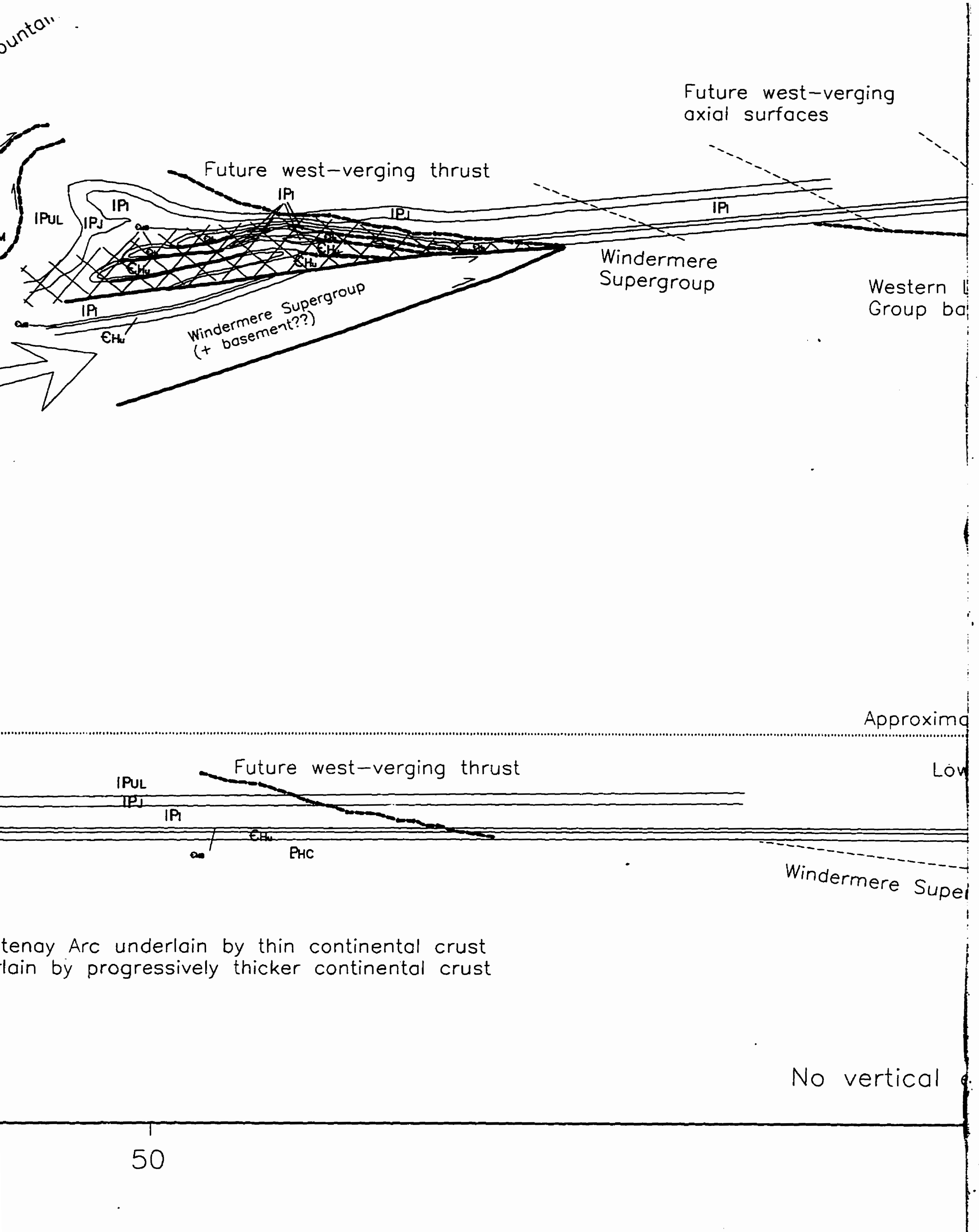
Base of Starbird Formation
—Upper Devonian Paleozoic rocks
(not preserved)
Ds
DMF
PMN
PHC
PHC
PT
PDCu
PDCm
WINDERMERE HIGH



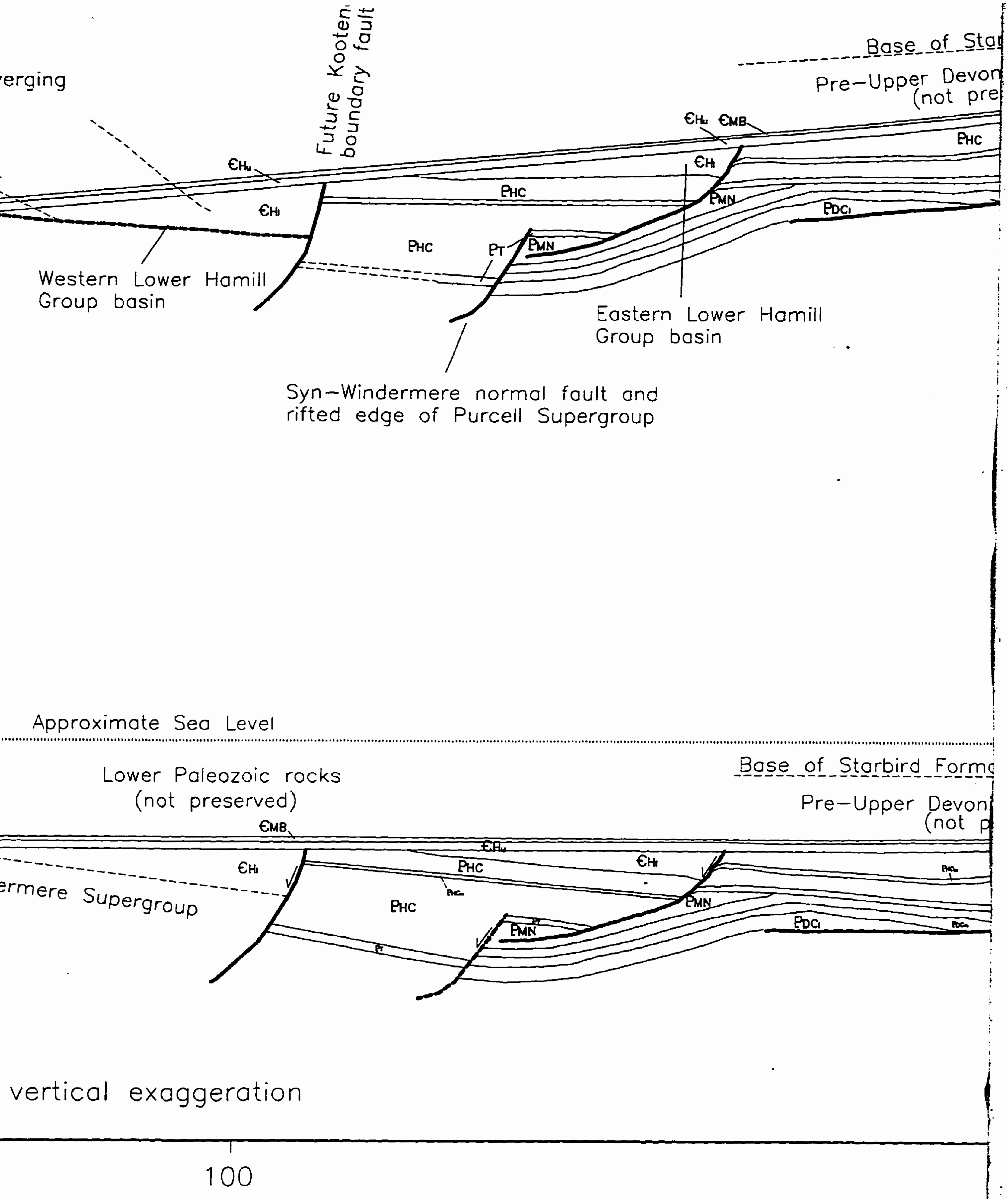

WINDERMERE HIGH



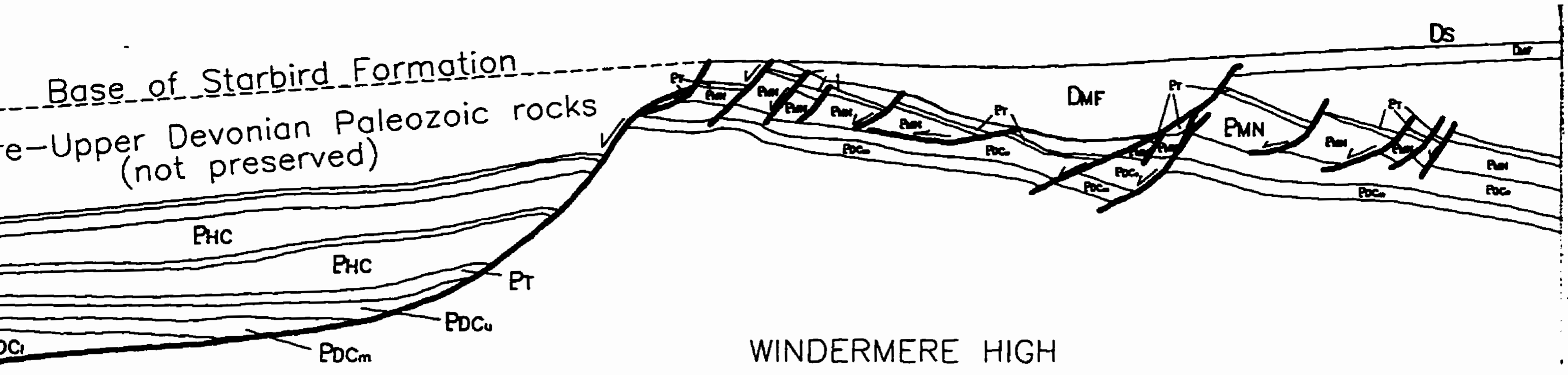

5) RESTORED TO ABOUT 173 Ma
Constraints: 173+/−5 Ma
Slide Mountain and
Cooper Creek stock
MM
MM
IPUL
IPJ
IPI
Advancing tectonic wedge
6) RESTORED TO ABOUT 375 Ma
Constraints: post− Early Frasnian, pre−187 Ma
Future eastward obduction of Slide Mtn. and Quesnel terranes
Future locus of tectonic wedge
Strata of the future Kootenay Arc
Strata to the east underlain by p
0



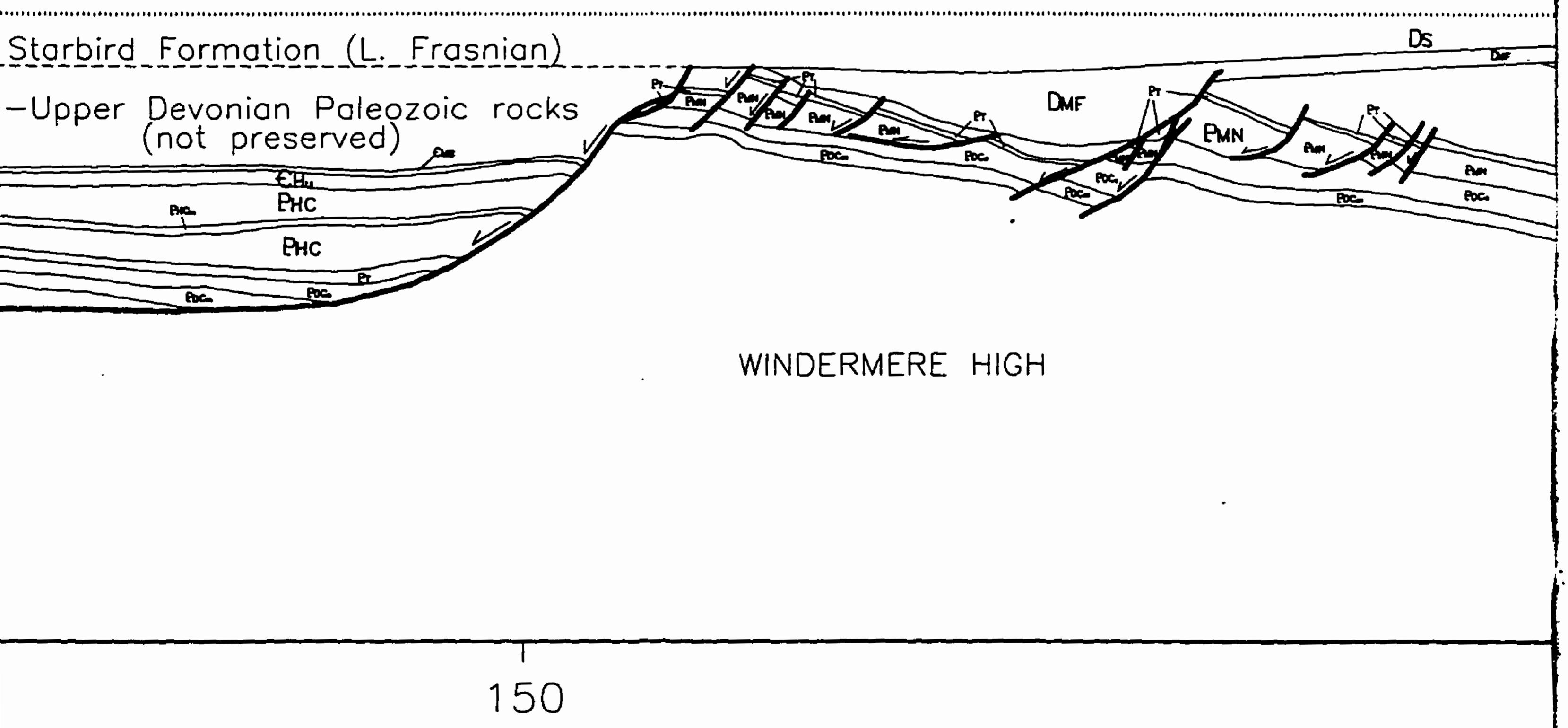
ountai
Future west-verging axial surfaces
Future west-verging thrust
IPUL
IPJ
IPI
€Hu
Windermere Supergroup (+ basement??)
Windermere Supergroup
Western L
Group ba
Approxima
Future west-verging thrust
Low
PHC
Windermere Supe
tenay Arc underlain by thin continental crust
lain by progressively thicker continental crust
No vertical
50



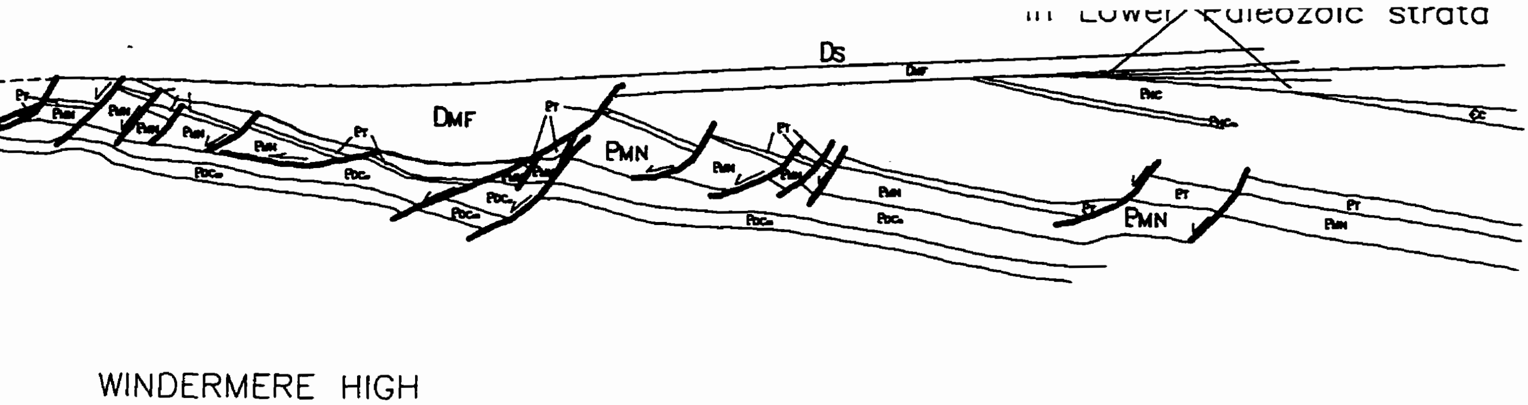
erging
Future Kooten
boundary fault
Base of Sta
Pre–Upper Devon
(not pre
ЄHu ЄMB
PHC
ЄHu
ЄHl
PHC
PDCl
ЄHl
PHC
PT
PMN
PMN
Western Lower Hamill
Group basin
Eastern Lower Hamill
Group basin
Syn–Windermere normal fault and
rifted edge of Purcell Supergroup
Approximate Sea Level
Base of Starbird Forma
Lower Paleozoic rocks
(not preserved)
Pre–Upper Devon
(not p
ЄMB
ЄHu
ЄHl
PHC
ЄHl
ermere Supergroup
PHC
PMN
PDCl
PMN
PT
vertical exaggeration
100

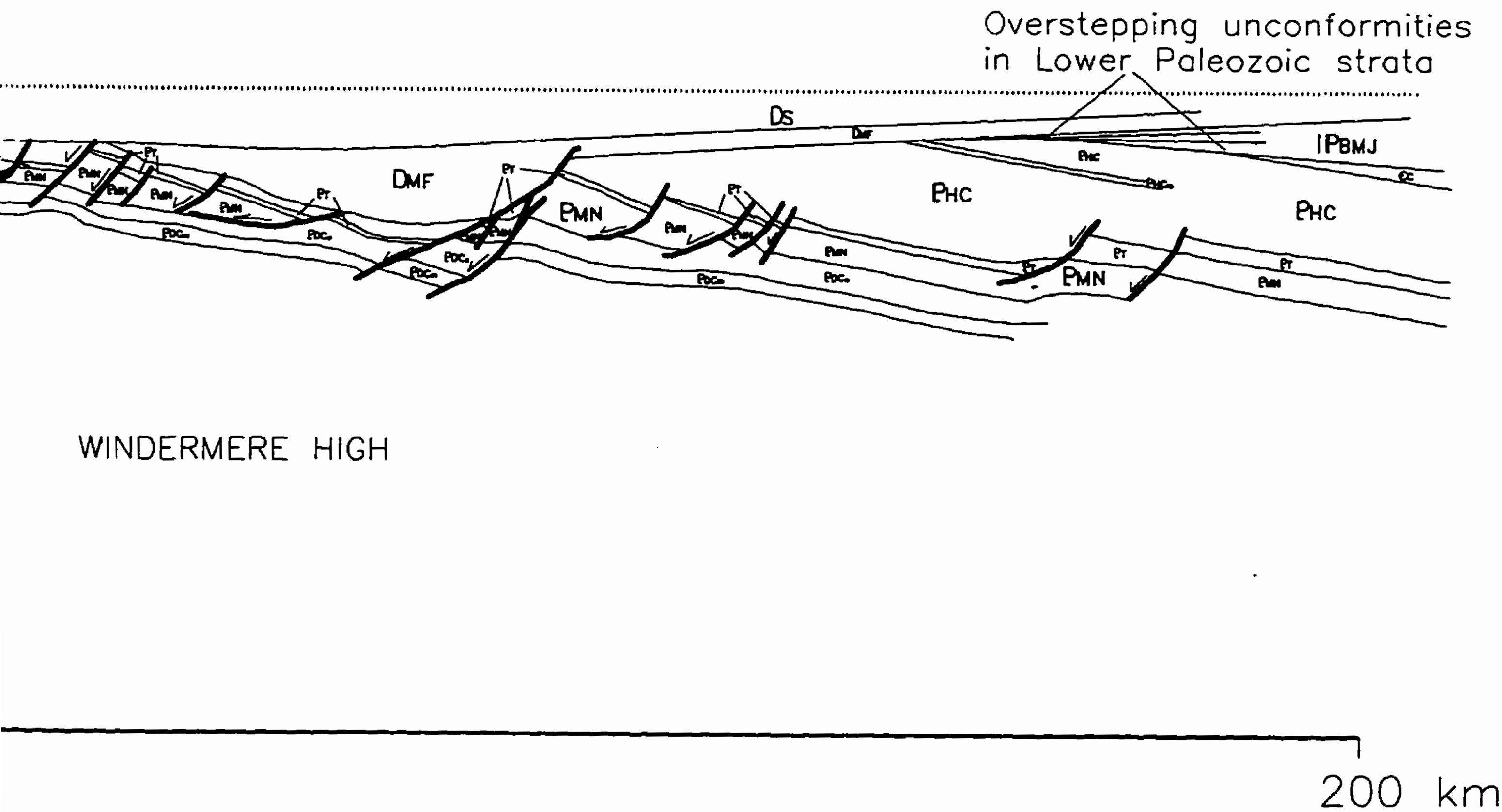


Base of Starbird Formation
e–Upper Devonian Paleozoic rocks
(not preserved)
PHC
PHC
PT
PDCu
PDCm
DMF
PMN
Ds
WINDERMERE HIGH

Starbird Formation (L. Frasnian)
–Upper Devonian Paleozoic rocks
(not preserved)
ЄHu
PHC
PHC
DMF
PMN
Ds
WINDERMERE HIGH
150


Ds
DMF
PMN
PMN
WINDERMERE HIGH


Overstepping unconformities
in Lower Paleozoic strata
Ds
IPBMJ
DMF
PMN
PHC
PHC
PMN
WINDERMERE HIGH
200 km

# PLATE 4A

REGIONAL CROSS-SECTION
THROUGH THE KOOTENAY ARC
AND PURCELL ANTICLINORIUM

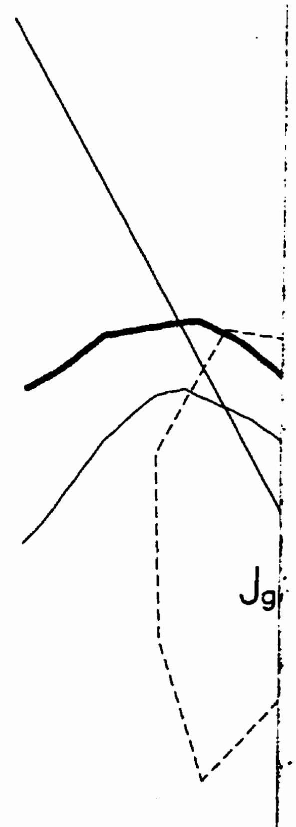
Compiled by M. J. Warren, 1996

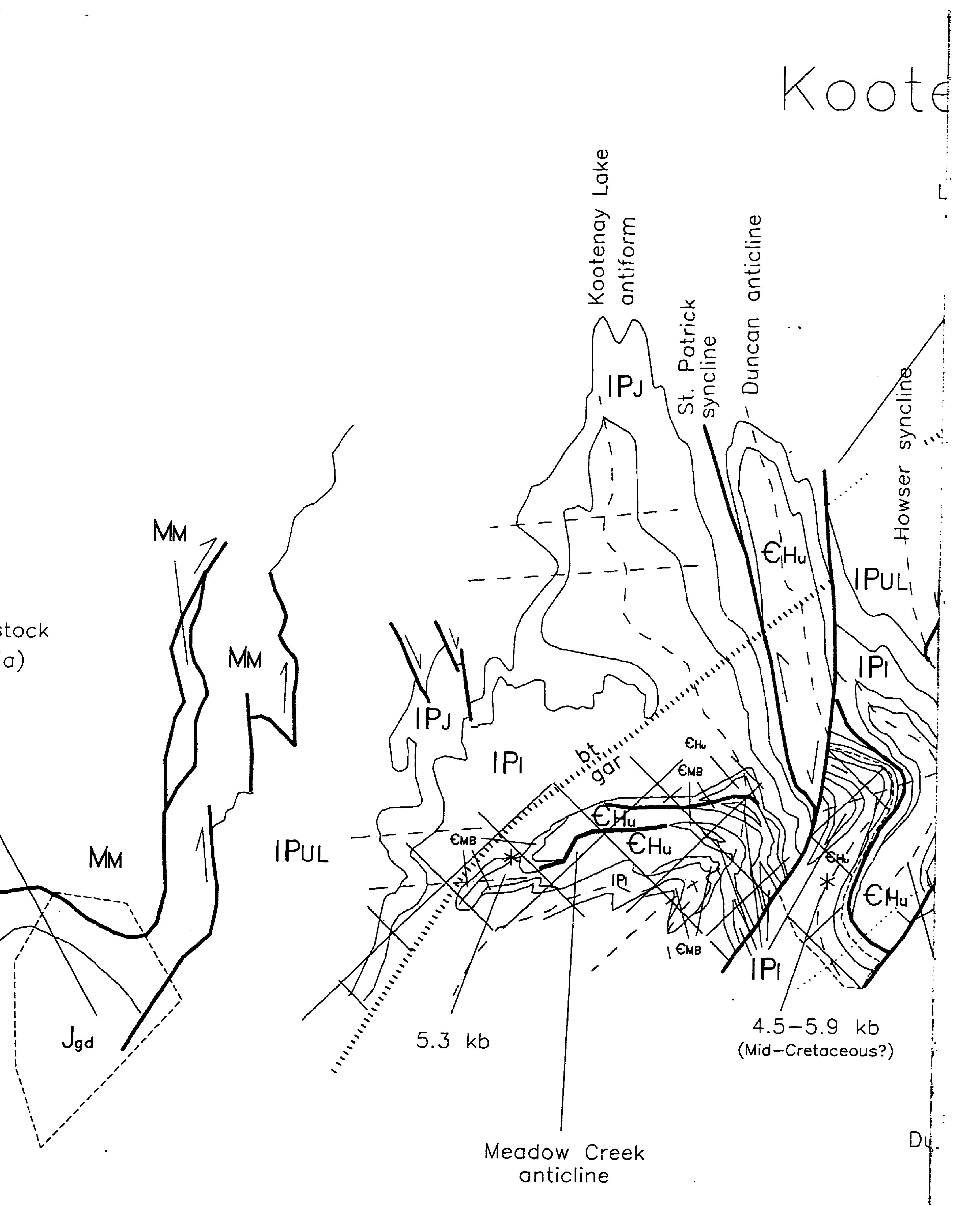
1:75,000

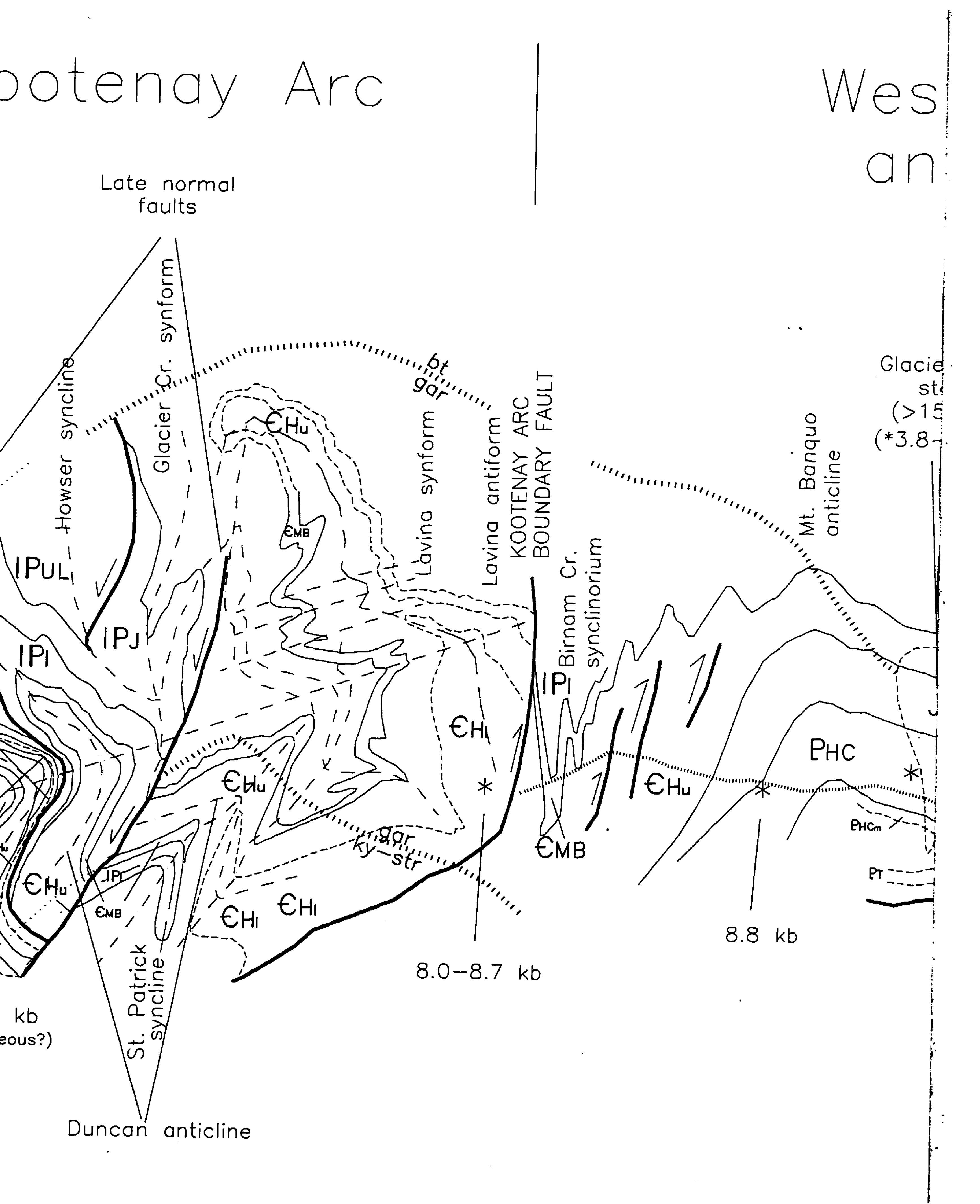
See Plate 4 for legend

Cooper Creek stock
(180+/-7 Ma)
(3-4kb)

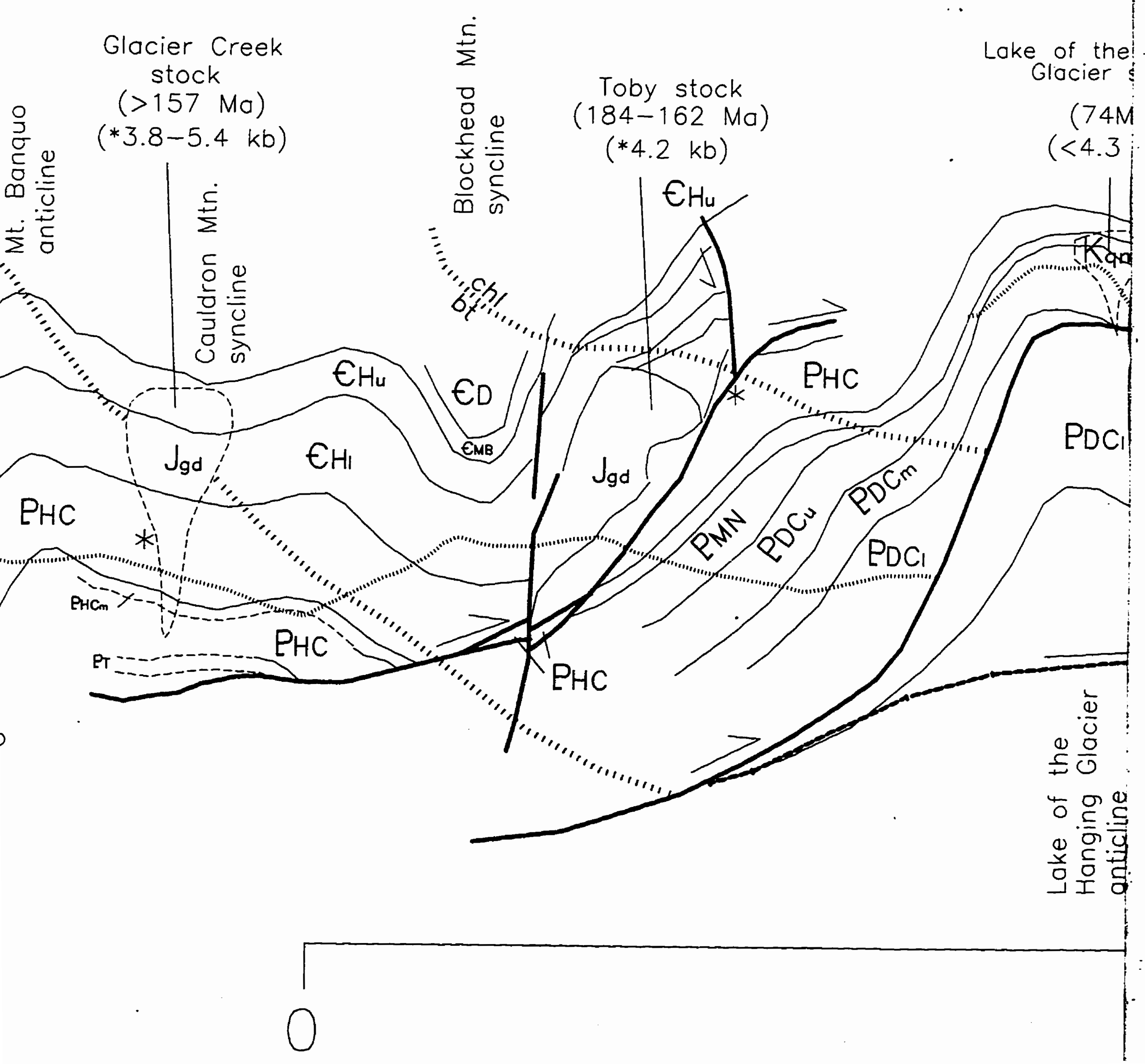
WEST



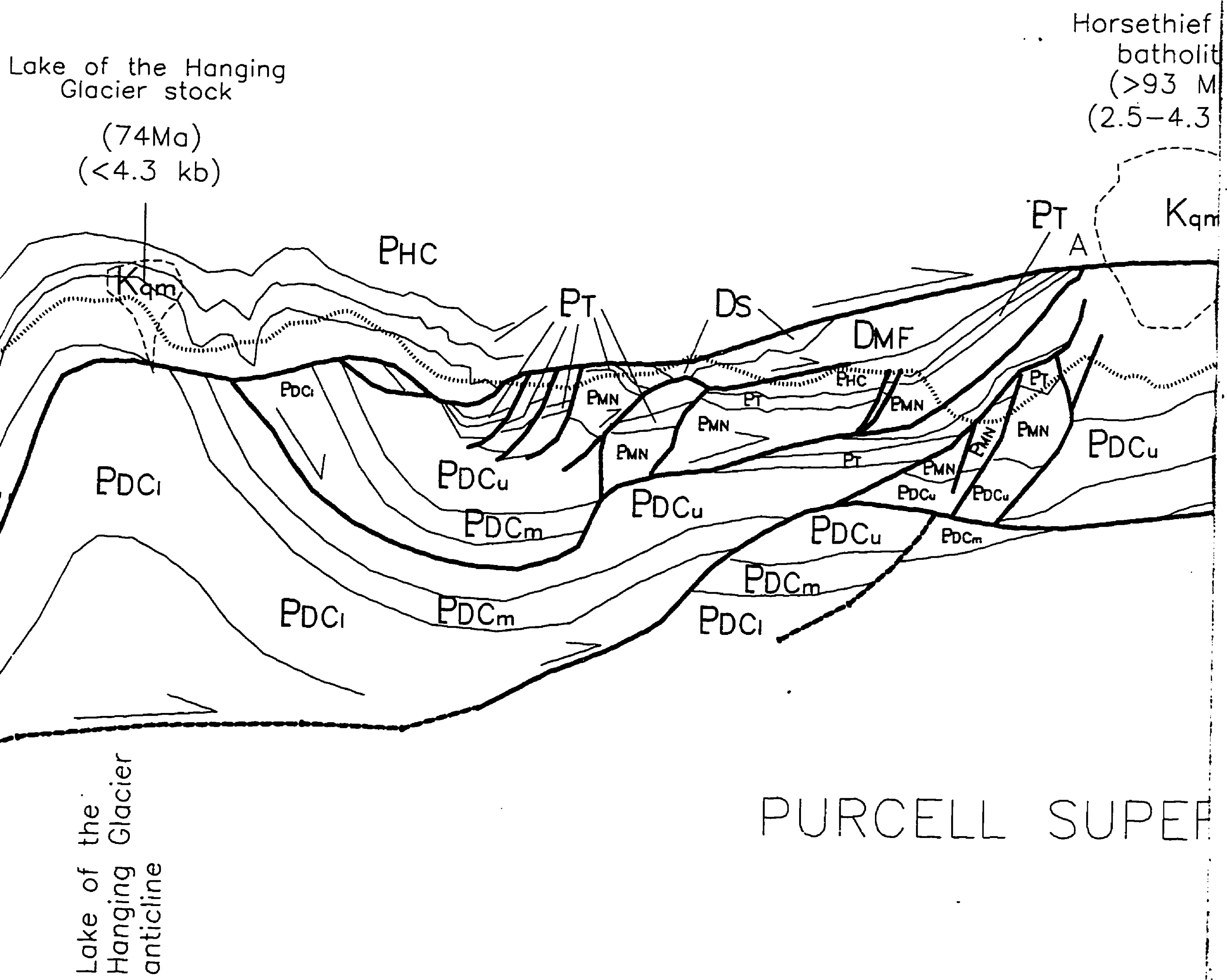

Koote
Kootenay Lake antiform
St. Patrick syncline
Duncan anticline
Howser syncline
IPJ
ЄHu
IPUL
IPl
MM
stock
Jgd
bt
gar
ЄMB
5.3 kb
4.5–5.9 kb
(Mid–Cretaceous?)
Meadow Creek anticline
Du


ootenay Arc
Wes
an
Late normal
faults
Glacier Cr. synform
Howser syncline
bt
gar
Lavina synform
Lavina antiform
KOOTENAY ARC
BOUNDARY FAULT
Birnam Cr.
synclinorium
Mt. Banquo
anticline
Glacie
st
(>15
(*3.8-
ЄHu
ЄMB
IPUL
IPJ
IPl
ЄHl
IPl
ЄHu
PHC
ЄMB
PHCm
PT
gar
ky-str
ЄHu
IPl
ЄMB
ЄHl
ЄHl
8.0–8.7 kb
8.8 kb
kb
eous?)
St. Patrick
syncline
Duncan anticline

# Western Purcell anticlinorium

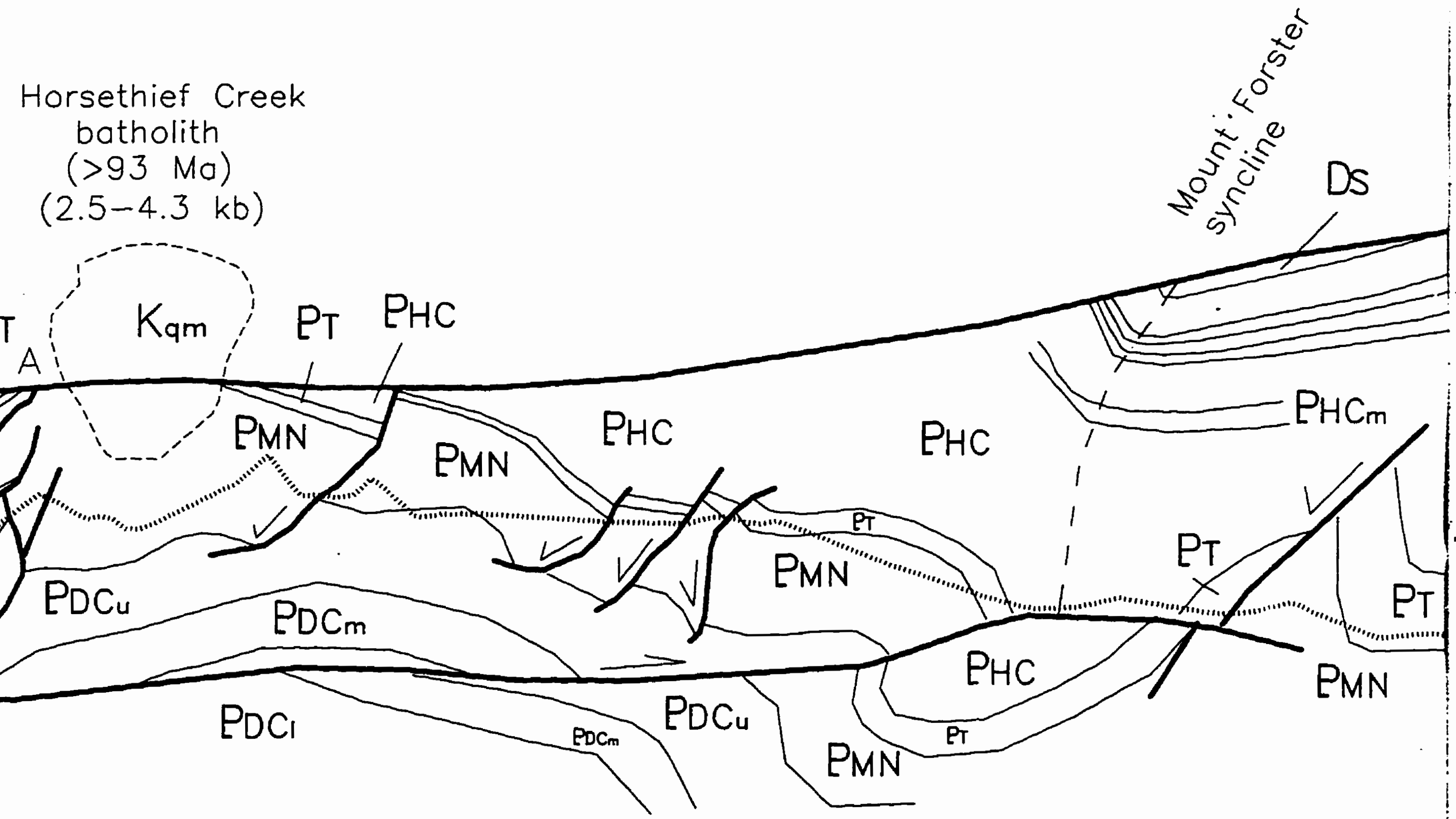


# Eastern Purcell

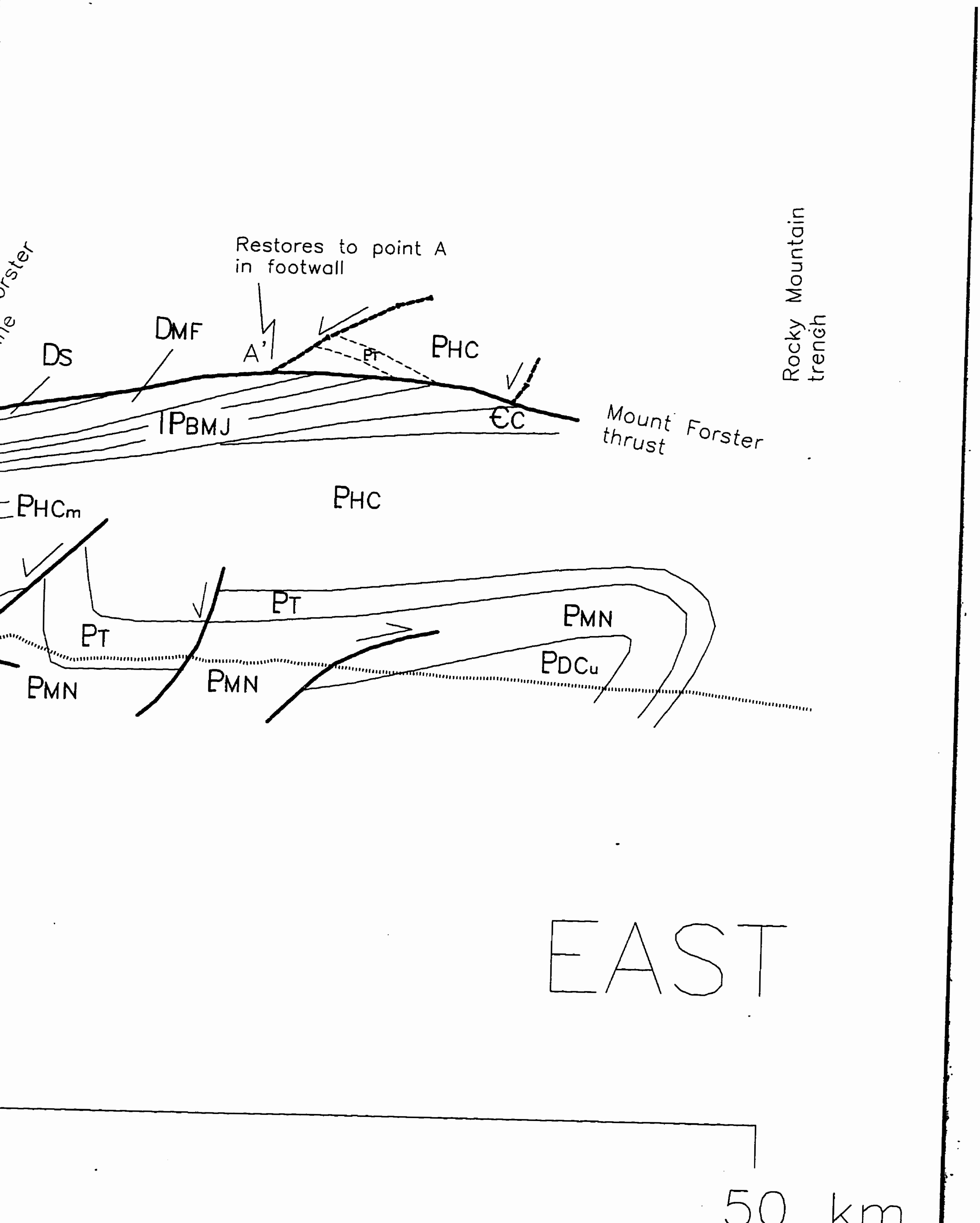



PURCELL SUPER

## cell anticlinorium




SUPERGROUP



orster
ne
Restores to point A
in footwall
DMF
Ds
A'
PHC
PT
IPBMJ
€C
Mount Forster
thrust
Rocky Mountain
trench
PHCm
PHC
PT
PMN
PT
PDCu
PMN
PMN
EAST
50 km