THE UNIVERSITY OF CALGARY

Structural Evolution of the Isaac Lake Synclinorium, Cariboo Mountains, British Columbia

by

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ABSTRACT

The Isaac Lake Synclinorium is a structural depression in the Cariboo Mountains which preserves the Neoproterozoic Windermere Supergroup stratigraphy at a high structural level. The synclinorium is bounded by dextral-oblique strike-slip faults which include the Isaac Lake and Winder faults. The deformational history of the Isaac Lake Synclinorium can be sub-divided into three main events D_1, D_2, and D_3. D_1 includes local east-verging isoclinal folds with an associated axial planar cleavage and east-verging thrust faults. Peak metamorphism was attained during D_1 and reached lower greenschist facies conditions. D_2 formed the main regional anticlinoria and synclinoria structures that include the Isaac Lake Synclinorium, which occurred during the early to mid-Jurassic based on the relationships of D_2 structures to the 174 Ma Hobson pluton. D_3 structures include dextral-oblique strike-slip faults and associated folds and extension faults, which were initiated by the mid-Cretaceous. The dextral strike-slip faults in the Isaac Lake Synclinorium, and throughout the Cariboo Mountains form a regional dextral strike-slip fault system and are interpreted to transfer southward a portion of the displacement on the Northern Rocky Mountain Trench Fault.

ACKNOWLEDGEMENTS

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CHAPTER 1: INTRODUCTION

1.1 Introduction

The Isaac Lake Synclinorium is located in the northern Cariboo Mountains within the Omineca Crystalline Belt of the western Canadian Cordilleran Orogen. The Omineca Crystalline Belt is a morphogeological belt which contains the internal core zone to the orogen and is comprised of low to high-grade polydeformed metamorphic rocks of North American affinity, and intrusive igneous complexes of several different ages. It is bounded to the east by the Foreland Fold and Thrust Belt and to the west by accreted allochthonous terranes of the Intermontane and Insular Superterranes, an arrangement which is typical of orogenic belts around the world (Moores and Twiss, 1996). To fully understand the origin of crystalline core zones of orogenic belts such as the Omineca Crystalline Belt, there are a number of key questions which need to be addressed. What is the relationship of the high-grade metamorphic rocks of the crystalline core zone to low-grade metamorphic rocks of the crystalline core zone? How is strain distributed throughout the crystalline core zone and what is its relationship to larger tectonic processes? What is the relationship between timing of deformation in the crystalline core zone and the adjacent foreland fold and thrust belt?

Rocks of the Cariboo Mountains provide one of the best examples of the transition of structural styles from shallow to deep structural levels in the Omineca Crystalline Belt.

The Isaac Lake Synclinorium is located in the north-central Cariboo Mountains and within it are some of the shallowest structural and stratigraphic levels preserved in the Omineca Crystalline Belt. This provides the opportunity to examine the effects of deformation at shallow structural levels, and to address some of the critical questions about the Omineca Crystalline Belt as stated above.

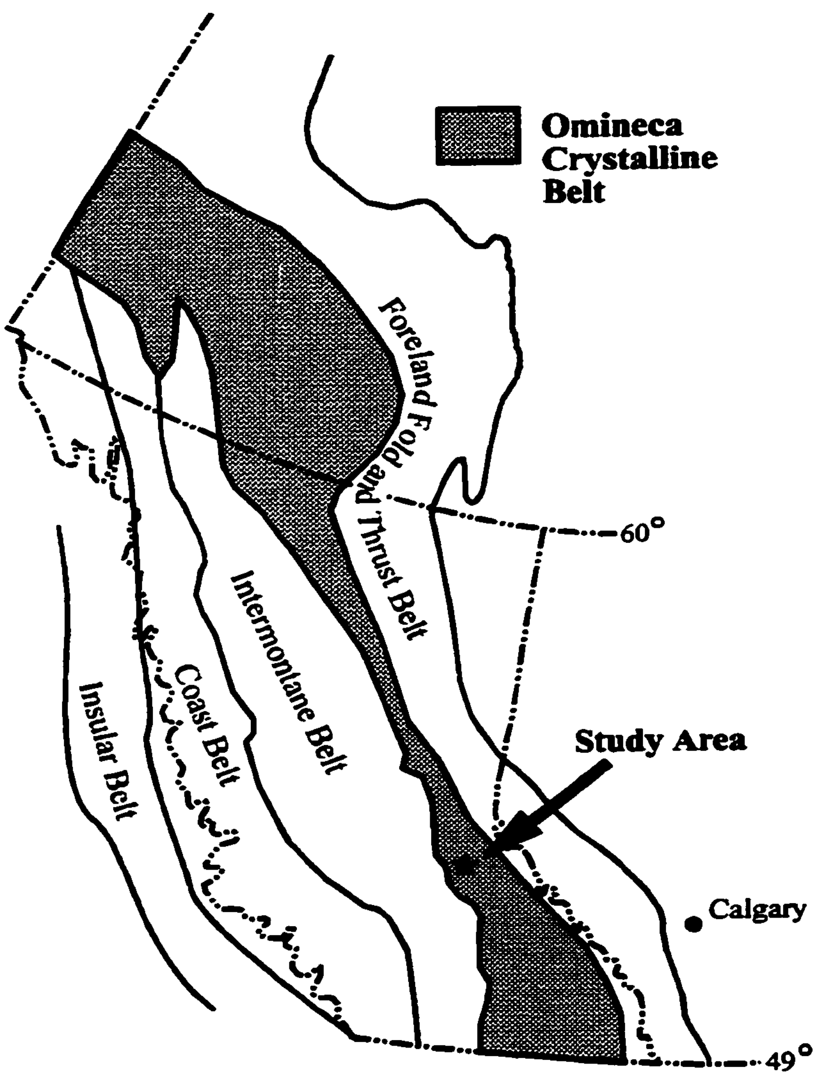

Figure 1-1: Morphogeological Belts of the Western Canadian Cordillera showing the location of the study area within the Omineca Crystalline Belt.

1.2 Regional Geology

The western Canadian Cordilleran orogen formed as a result of the northeasterly directed collision of allochthonous terranes with the rifted margin of western North America. Collision and accretion took place from the mid-Mesozoic to Tertiary. The orogen can be subdivided into five major morphogeological belts. From east to west they are, the Foreland Fold and Thrust Belt, the Omineca Crystalline Belt, the Intermontane Belt, the Coast Plutonic Belt and the Insular Belt (Monger et al., 1972; Figure 1-1). The Omineca Crystalline Belt and the Foreland Fold and Thrust Belt contain deformed rocks of North American affinity onto which the allochthonous terranes of the Intermontane and Insular Belts were accreted.

Although chiefly a morphologic distinction, the Omineca Crystalline Belt also differs from the Foreland Fold and Thrust Belt in structural style. The Omineca Crystalline Belt consists of regional Barrovian metamorphic complexes, which range from chlorite to sillimanite grade, and metamorphic "core" complexes bound by low-angle, Eocene extensional shear zones and related faults. At shallow structural levels, the rocks of the Omineca Belt are dominated by ductile to brittle-ductile structures including polydeformed folds and faults. At deep structural levels, the Omineca Crystalline Belt is dominated by ductile structures including large thrust or fold nappes that are complexly polydeformed. Campbell (1970) recognized this structural contrast and referred to it as the suprastructure to infrastructure transition.

The Omineca Crystalline Belt has a complex Jurassic to Eocene structural and metamorphic history. For the past two decades it has been studied by numerous university and government research groups which have focused on the high-grade metamorphic complexes which include the Monashee, Valhalla and Priest River-Selkirk Crest complexes in the southern Omineca Crystalline Belt (Brown, 1980; Brown et al., 1991; Carr 1991, 1992; Carr et al., 1987; Johnson, 1994; Journey, 1986; Journey & Brown, 1986; Parkinson, 1991; Parrish, 1984, 1992; 1995; Scammell, 1986, 1992, 1993; Sevigny et al., 1989, 1990; Crowley et al., 1996; Colpron et al., 1996). The core complexes provide a window to deep crustal events which occurred during formation of the western Canadian Cordillera, thus, allowing for a greater understanding of the mechanics and timing of deformation of deeper crustal levels. The relationship between high-grade metamorphic rocks and the low grade rocks of the Omineca Crystalline Belt remains poorly understood, yet offers a critical link between the evolution of the Omineca Crystalline Belt and the Foreland Fold and Thrust Belt.

The Foreland Fold and Thrust Belt is dominated by thin-skinned thrusting and detachment folding. The main thrust faults are interpreted to flatten into a common basal detachment fault that formed above the basement rocks and which is interpreted to extend into the Omineca Crystalline Belt as the Monashee Decollement (Bally et al., 1966; Price, 1981; Cook et al., 1988; van der Velden and Cook, 1996; Brown et al., 1991). The Foreland Fold and Thrust Belt and the Omineca Crystalline Belt are separated by the Rocky Mountain Trench, a large linear valley which is divided into two distinct segments each with a different kinematic history (northern and southern Rocky Mountain Trench, NRMT and SRMT respectively) (Figure 1-2). The NRMT is composed of a large-scale Cretaceous to Eocene dextral strike slip fault system with a minimum displacement of 450 km (Gabrielese, 1985). In contrast, the southern Rocky Mountain Trench contains only localized normal, transcurrent and reverse faults. Correlation of structures and stratigraphy across the SRMT suggests that displacement along the SRMT is much less than along the NRMT (Wheeler, 1963; Campbell, 1968; Price and Mountjoy, 1970; Simony et al., 1980; McDonough and Simony, 1988; Murphy 1990; Ferguson, 1994) Recent geophysical work has shown that the SRMT may be coincident with the hinge line of a pre-existing, west-dipping basement ramp. This basement structure-controlled fault location and geometry resulted in the formation of the SRMT (van der Velden and Cook, 1996).

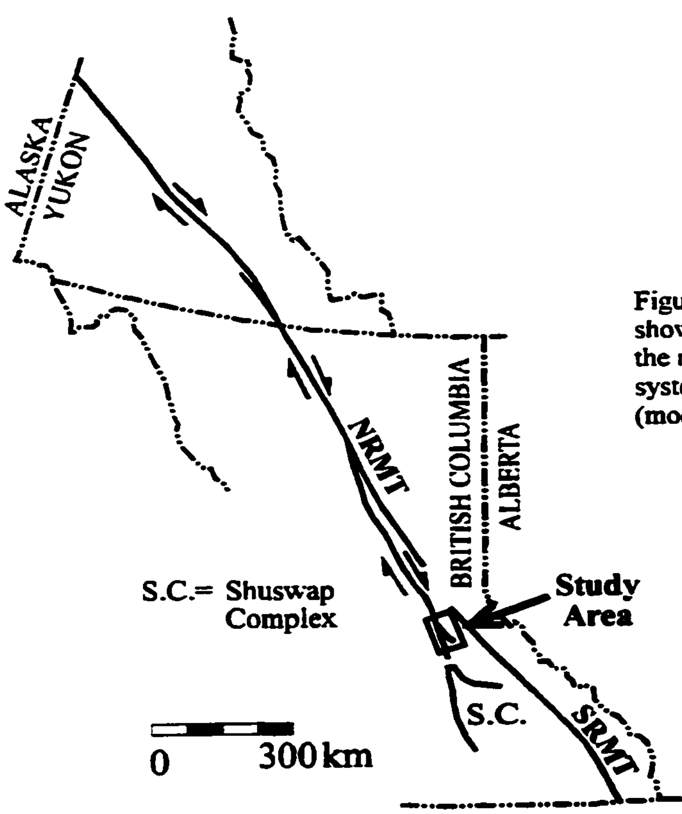

Figure 1-2. Map of the western Canadian Cordillers showing the position of the study area relative to the northern and southern Rocky Mountain Trench systems (NRMT and SRMT respectively). (modified from Gabrielese, 1985)

1.3 Geology of the Cariboo Mountains

The Cariboo Mountains, located in east-central British Columbia approximately 150 km northwest of Jasper, are the northernmost range of the Columbia Mountains (Figure 1-3). The Cariboo Mountains extend southward from the "big bend' of the Fraser River to the North Thompson River. The predominant rock units of the Cariboo Mountains are those of the Neoproterozoic Windermere Supergroup overlain by a relatively thin Paleozoic succession. All these units are of North American affinity and represent the toe of the Neoproterozoic continental margin wedge. Deformation of the strata in the Cariboo Mountains mainly took place during the Jura-Cretaceous Columbian Orogeny as a result of the collision and obduction of allochthonous terranes on the western rifted margin of North America (Monger et al., 1982).

The structures of the north-central Cariboo Mountains can be subdivided into a series of northwest-trending anticlinoria and synclinoria, which include from east to west; the Premier Anticlinorium, Isaac Lake Synclinorium, Lanezi Arch, Black Stuart Synclinorium and the Lightning Creek Anticlinorium (Sutherland-Brown, 1957; Sutherland-Brown, 1963; Campbell, 1970). These large-scale folds flatten and die out to the north into a fault-dominated region as a result of a change to a shallower structural level and more competent rock rheology.

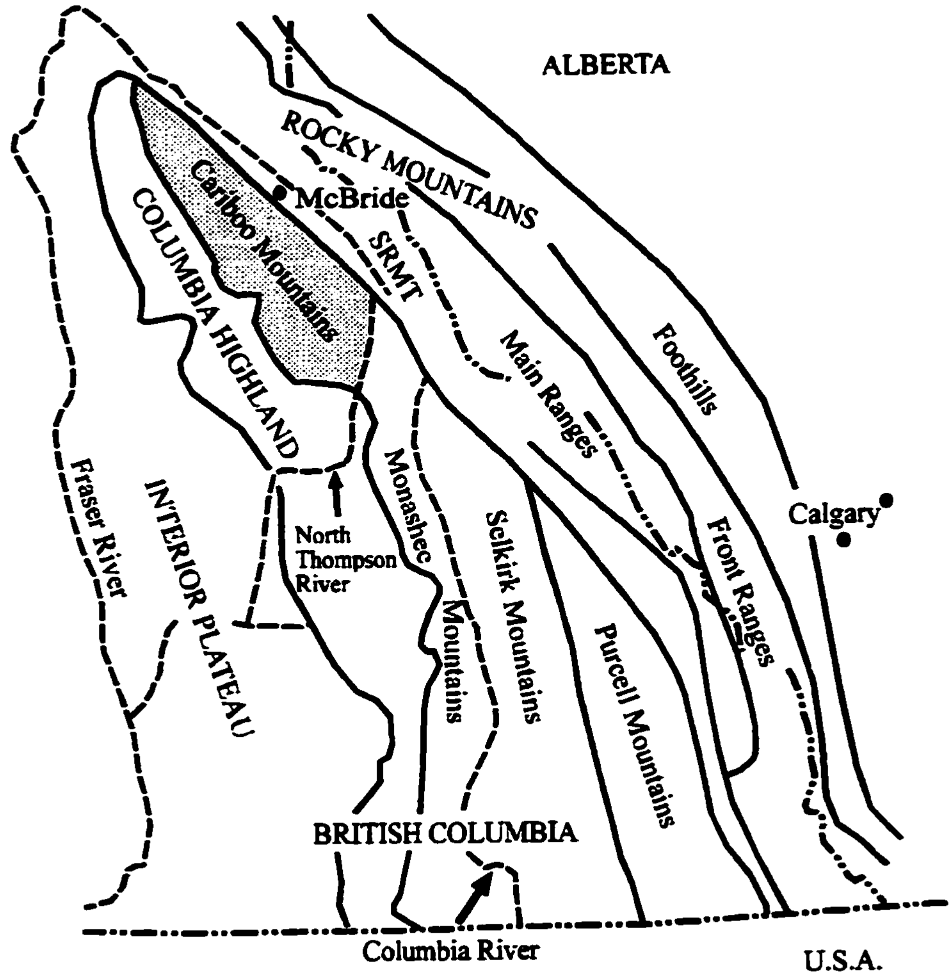

Figure 1-3. Geographic map of the southern Omineca Crystalline Belt showing the location of the Cariboo Mountains (shaded area), the northernmost range of the Columbia Mountains.

The recognition of transcurrent faults in the north-central Cariboo Mountains has been documented by recent workers (Ferguson, 1994; Gerasimoff, 1985; Murphy, 1987; Struik, 1985). Fault zones such as the Isaac Lake, Matthew, Willow River, Stony Lake, Narrow Lake and Knutson Creek Fault zones suggest that transcurrent faulting in the Cariboo Mountains was widespread, and may have significant tectonic implications. All of the transcurrent faults in the Cariboo Mountains have been found to be post-metamorphic faults (Cooley, pers.comm., 1996; Ferguson, 1994; Gerasimoff, 1985;

Murphy, 1987; Struik, 1985). The Matthew Fault is interpreted to cross-cut the mid-Jurassic (174 Ma) Hobson Pluton, implying that the age of transcurrent faulting in the Cariboo Mountains is younger than 174 Ma (Gerasimoff, 1988; Pigage, 1977). A lack of definitive age constraints has made regional correlation and interpretation of these faults difficult. A Cretaceous age could suggest that transcurrent faulting is related to collision of the second allochthonous terrane, the Insular Terrane, and the formation of the NRMT fault. An Eocene age for the transcurrent faults may imply that they are related to late motion along the NRMT fault system and may represent a transition zone (step-over zone) between the NRMT and the Fraser River-Straight Creek Fault zones to the southwest (Price and Carmichael, 1986). Recent geochronological work by Cooley (pers. comm. 1996) suggests that movement on the Matthew Fault and related faults is mid-Cretaceous. Figure 1-4 is a generalized geological map of the Cariboo Mountains showing the anticlinoria-synclinoria type structures, major thrust and transcurrent faults and the proximity to the accreted Quesnellia terrane of the Intermontane superterrane.

In the Cariboo Mountains, an overall northwest plunge of the regional anticlinorial and synclinorial structures results in the southward transition to deeper structural levels and a change in structural style. The structural style of folding changes markedly from upright to recumbent from north to south with deeper structural levels and increasing metamorphic grade (Campbell, 1970; Murphy, 1985). This transition was recognized by Campbell (1970) who referred to the two main structural zones as the suprastructure and infastructure. The suprastructure refers to the low-grade rocks at shallow structural levels, and the infastructure refers to the high-grade rocks at deeper

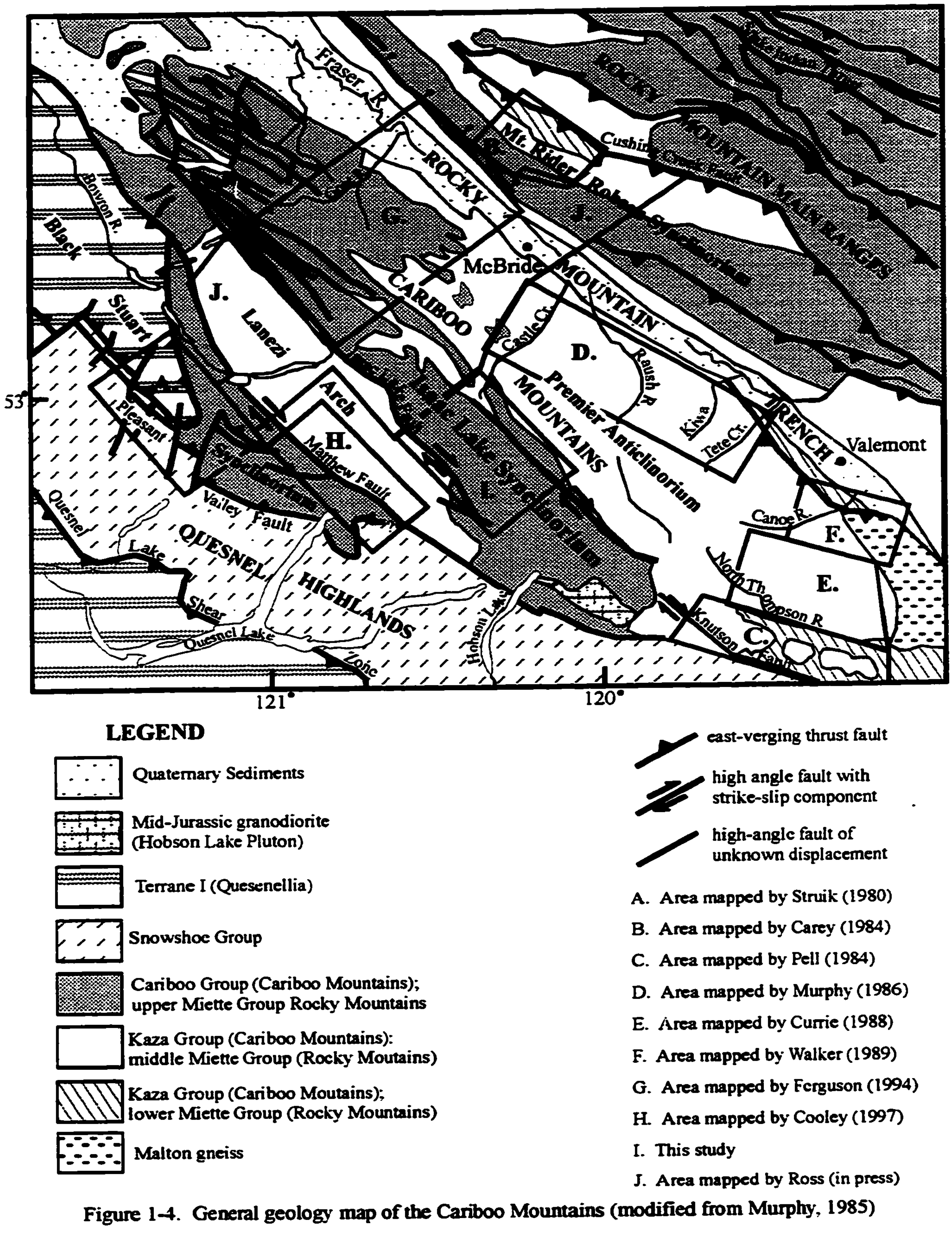

Figure 1-4. General geology map of the Cariboo Mountains (modified from Murphy, 1985)

structural levels. The northern Cariboo Mountains contain brittle to semi-ductilly polydeformed, greenschist facies Windermere Supergroup stratigraphy dominated by upright folds and faults. Continuing into the southern Cariboo Mountains, deeper structural levels expose sillimanite grade rocks that experienced ductile shear and recumbent folding (Campbell, 1970; Pell, 1984; Murphy, 1985; Currie, 1988; Walker, 1989). The north-central Cariboo Mountains contains structures transitional to those found to the north and south, thus providing an opportunity to examine a structural transition zone within the Omineca Crystalline Belt. It is in the northern to north-central Cariboo Mountains where the highest structural and stratigraphic levels of the Omineca Crystalline Belt are preserved. Understanding the deformational processes which occurred at high structural levels of the Omineca Crystalline Belt is important for two main reasons. First, it allows for the potential correlation of structures within the Cariboo Mountains to those of the western Main Ranges of the Rocky Mountains, a relationship which is poorly understood. Second, it provides the means to compare the effects of deformation between high and low structural levels of the Omineca Crystalline Belt, a relationship which also is poorly understood.

1.4 Purpose

This project focuses on the stratigraphy and structural geology within the Isaac Lake Synclinorium. The study area is located between Bowron Lake and Wells Gray Provincial Parks approximately 50 km west of the town of McBride, British Columbia

(Figure 1-7). The Isaac Lake Synclinorium is a structural depression bound to the west by the Isaac Lake Fault Zone, a northwest-trending dextral-oblique fault and to the east by the Winder Fault, a northwest-trending dextral oblique fault. Preserved within the Isaac Lake Synclinorium is a continuous succession of polydeformed Neoproterozoic Windermere Supergroup and overlying lowermost Cambrian formations at lower greenschist facies. The purpose of this study is fourfold:

1. To map the fault boundaries and internal structures of the Isaac Lake Synclinorium.
2. To document and quantify the effects of strain partitioning within the different lithologic units of the Neoproterozoic Windermere Supergroup exposed in the study area.
3. To examine the Isaac Lake Fault Zone south of Isaac Lake and its related structures, in order to document the effects of transcurrent faulting in the south-central Canadian Cordillera.
4. To determine the relative timing of metamorphism, and the metamorphic grade of the north-central Cariboo Mountains.

1.5 Previous Work/Methods

The earliest work recorded in the Cariboo Mountains is that of Amos Bowman (1889) who worked around Barkerville, in the gold placer district. Subsequent work was carried out in the Barkerville area by Johnston and Uglow (1926), Hanson (1938), Holland (1948), and Sutherland-Brown (1957, 1963). The Cariboo Mountains were first mapped at a reconnaissance scale of 1:250 000 which was published as a Geological Survey of Canada Open File map (Campbell, 1978). The Campbell et al. (1973) publication of the Geology of the McBride Map Area provides a good overview of the northern Cariboo Mountains. More recent work carried out by various university and government research groups include Fletcher, (1977); Struik, (1980); Murphy, (1985, 1987), Pell (1984); Gerasimoff, (1988); Currie (1988); Walker, (1989); Ross (1991), Ross et al., (1988, 1989), Ferguson (1994), and Cooley, (in progress). See figure 1.4 for locations of study areas.

Field work for this project was carried out during July and August of 1994 and 1995 through a series of 4-6 day helicopter supported camps. Detailed mapping of the structures and lithologies throughout the area was done at a 1:25 000 scale and compiled at a 1: 50 000 scale, and a total of three vertical cross-sections were drawn throughout the study area to illustrate the structures (Figure 1-5 and 1-6). Topographic map sheets which the study area encompasses are Mount Winder (93A16) and the Mitchell Lake (93A15) map areas. Rock samples were collected during the field seasons for the purpose of petrographic and microstructural analysis. Approximately 45 oriented thin sections were cut in order to examine the microstructural relationships between cleavages. Another 30 thin sections were cut for the purpose of examining lithologies.

Maps and cross-sections were constructed using data collected during the summer and previous data collected by Dr. Gerry Ross, Geological Survey of Canada in Calgary during the summers of 1990 and 1991.

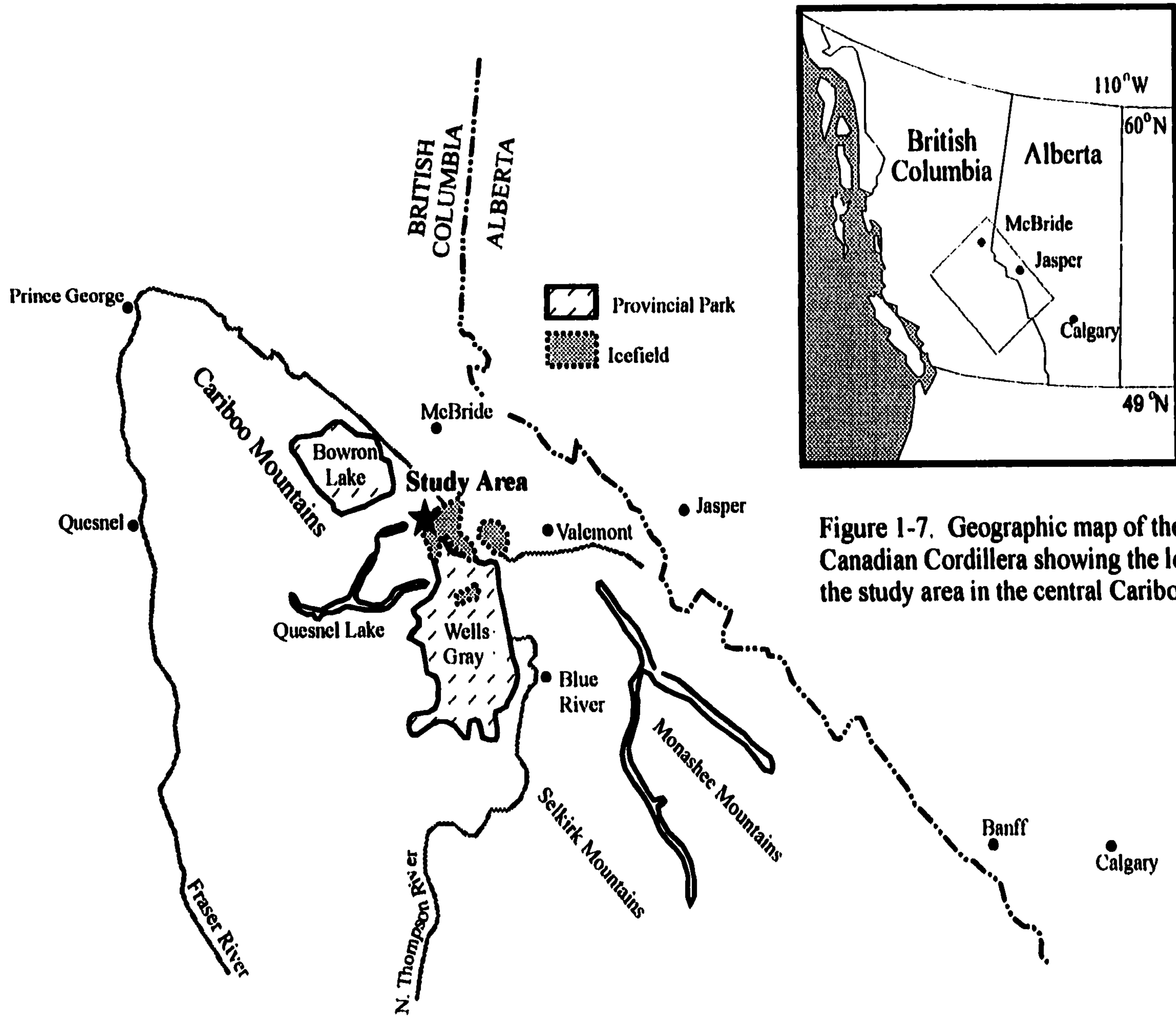

Figure 1-7. Geographic map of the southern Canadian Cordillera showing the location of the study area in the central Cariboo Mountains.

CHAPTER 2: STRATIGRAPHY OF THE ISAAC LAKE SYNCLINORIUM

2.1 Introduction

The Neoproterozoic Windermere Supergroup sedimentary rocks occur along the length of the North American Cordillera and range from 3 to 11 km in original thickness (Figure 2-1). It is a westward thickening sequence of dominantly siliciclastic strata with intervening carbonate rocks which is interpreted to have been a passive margin sequence which was deposited as a result of Proterozoic rifting and break-up of the supercontinent Rodinia (Stewart, 1972; Gabrielese, 1972; Eisbacher, 1981; Ross, 1991).

In the southern Canadian Cordillera, the Windermere Supergroup is composed of deep-water siliciclastic and carbonate rocks with shallower shelf strata locally preserved. Some of the thickest preserved portions of the deep water Windermere Supergroup are found in the southern Canadian Cordillera. A lack of biostratigraphic markers and varying metamorphic grade have made correlations between different regions difficult. As a result, the nomenclature and stratigraphic divisions vary from region to region. In the Cariboo Mountains, the Windermere Supergroup is divided into the Kaza and Cariboo Groups; the Horsethief Creek Group in the Selkirk, Purcell and Monashee Mountains and the Miette Group in the Rocky Mountains. The only regionally accepted marker unit is found in the lower Windermere Supergroup where it is known as the Old

Fort Point Formation in the Miette Group, the "marker" in the Kaza Group, the Comedy Creek unit in the Selkirk Mountains and the Baird Brook division in the Purcell Mountains (Ross and Murphy, 1988; Aitken, 1969; Grasby and Brown, 1993; Poulton and Simony, 1980; Kubli, 1990; Charlesworth et al., 1967). The marker unit of the lower Windermere Supergroup is characterized by a distinct and abrupt facies change from coarse-grained mass flow deposits below the marker unit to thinly bedded shale and limestone succession of the marker unit. Table 1 summarizes the different divisions of the Windermere Supergroup throughout the southern Canadian Cordillera and shows general correlations between them using the Old Fort Point Formation and equivalents as a datum.

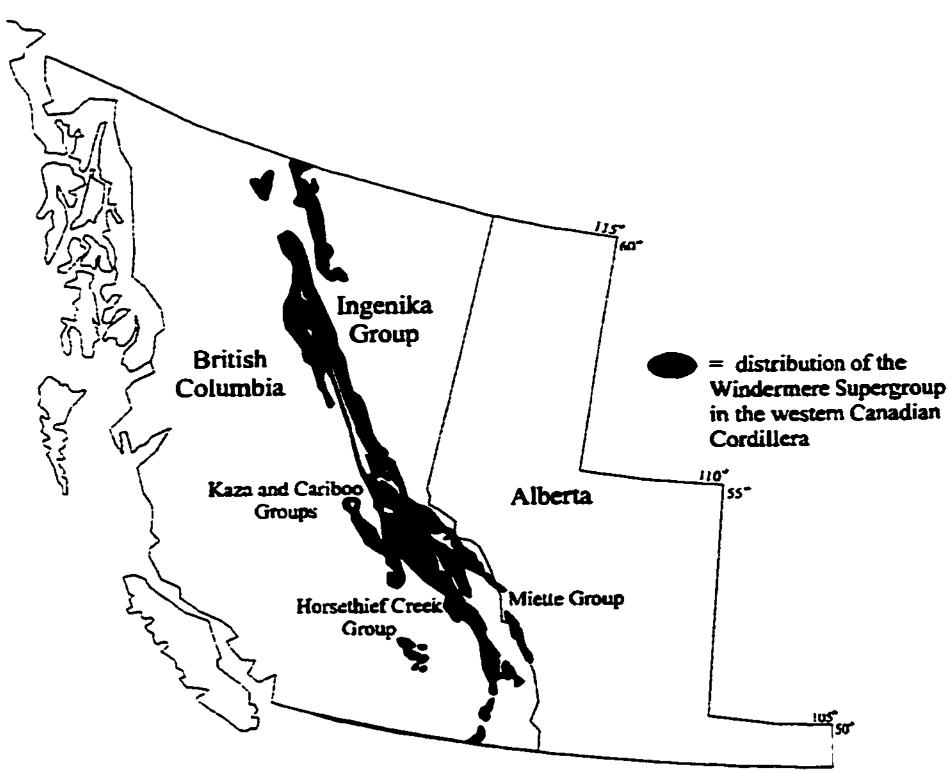

Figure 2-1. Map of western Canada showing the distribution of the Windermere Supergroup (modified from Hoffman, 1989)

The Windermere Supergroup is interpreted as a passive margin stratigraphic sequence that was deposited on the rifted western margin of North America (Stewart, 1971,1972; Gabrielese, 1972; Ross, 1991). The lower portion of the Windermere Supergroup, which includes the Kaza Group, lower Miette Group and lower Horsethief Creek Group consists predominantly of coarse-grained arkosic turbidite beds ("grits") with interbedded sandstone and shale. The grits are interpreted to be deep marine basinal sediments deposited within rift basins (Poulton and Simony, 1980; Carey and Simony, 1985; Arnott and Hein, 1986; Murphy, 1985). Detrital zircon geochronology shows a North American provenance link throughout the turbidite system, indicating that the grits were part of a single regional depositional system sourced from the east (Ross and Bowring, 1990; Ross and Parrish, 1991). This massive turbidite system, which can reach up to seven kilometres in thickness, is punctuated by the shale-dominated Old Fort Point Formation marker unit. This abrupt facies change has been attributed to a rapid basin-wide marine transgression (Ross and Murphy, 1989).

The turbiditic grit system fines upwards into a predominantly shale unit with intervening carbonate units. This unit is correlative with the shale units of the lower Cariboo Group (Isaac Formation), middle Horsethief Creek Group (Slate division) and the middle Miette Group (see Table 1). The shale beds fine upwards and have been interpreted as deep-marine slope turbidite deposits punctuated by slump and talus deposits from a shallow marine platform (Ross, 1991; Ferguson, 1994).

Stratigraphically above the turbiditic shales, where not eroded by the sub-Cambrian unconformity, is a shallow marine, carbonate platform unit (Cunningham

	Cariboo Mountains	Central Purcell Mountains	Northern Selkirk Mountains	Western Rocky Mountains	Western Rocky Mountains	Eastern Rocky Mountains
Ref.	(Campbell at. al., 1973; Ferguson, 1994)	(Poulton and Simony, 1980; Kubli, 1990)	(Grasby and Brown, 1993)	(Hein and McMechan, 1994)	(Charlesworth et al., 1967; Slind and Perkins, 1966)	(Aitken, 1969)
lower Cambrian	Yanks Peak Fm.	Hamill Group	Hamill Group	Gog Group	Gog Group	Gog Group
Neoproterozoic – Windermere Supergroup	Cariboo Group – Yankee Belle Fm. – Betty Wendle Member*	Horsethief Creek Group – Upper Clastic Division			Miette Group – Jasper Fm.	
	Cariboo Group – Yankee Belle Fm. – Amos Bowman Member*	Upper Clastic Division			Jasper Fm.	
	Cunningham Fm.	Carbonate Division			Byng Fm.	
	Isaac Fm.	Slate Division	Horsethief Creek Group – Upper Pelite Unit	Miette Group – East Twin Fm.	East Twin Fm.	Miette Group – Hector Formation
	Kaza Group – upper	Upper Grit Division	Central Clastic Unit	McKale Fm.	McKale Fm.	Hector Formation
	the "marker" (OFP)	Baird Brook Fm.(OFP)	Comedy Creek Unit	Old Fort Point Fm.	Old Fort Point Fm.	Old Fort Point equivalent
	middle	Lower Grit Division	Basal Pelite Unit	Cushing Creek Fm.	unnamed	Corral Creek Formation
	lower	Toby Fm.	Toby Fm.	base not exposed	base not exposed	base not exposed
	base not exposed	Purcell Supergroup	Purcell Supergroup			

Table 1. Correlation chart of the Windermere Supergroup stratigraphy in the Southern Canadian Cordillera. Old Fort Point Formation and equivalents are shown. Pinstriped areas represent unconformities. (* informal stratigraphic names)

Formation, Byng Dolomite and "Carbonate division"), (see Table 1). This carbonate unit consists predominantly of shallow water limestones and dolostones, which include resedimented pisolitic and oolitc packstones and wackestones (Ross et al., 1989). Where preserved, the shallow water carbonate platforms are overlain by a shallow-water, mixed carbonate-siliciclastic platform, interpreted to have been deposited in a sub-tidal environment (Yankee Belle Formation), (Lemieux, 1996; Ferguson, 1994; Young, 1979).

2.2 Windermere Supergroup in the Isaac Lake Synclinorium

The Cariboo Mountains are underlain by the Kaza and Cariboo Groups of the Neoproterozoic Windermere Supergroup (Sutherland-Brown, 1957; Campbell et al., 1973; Pell and Simony, 1987; Currie, 1988; Walker and Simony, 1989; Murphy, 1985; Ross and Murphy, 1988; Ross, 1991; Ferguson, 1994). Earliest Cambrian strata of the Cariboo Group are also exposed in the study area. The Kaza Group is subdivided into the lower, middle and upper Kaza Group and contains the only regionally correlative unit in the Windermere Supergroup, known as the "marker" unit, presumed to be equivalent to the Old Fort Point Formation (Ross and Murphy, 1988). The Cariboo Group is subdivided into the Neoproterozoic Isaac, Cunningham, and Yankee Belle formations, and the Paleozoic Yanks Peak, Midas, Mural, Dome Creek and Black Stuart formations.

The Isaac Lake Synclinorium provides an ideal area to examine the upper Neoproterozoic Windermere Supergroup stratigraphy as it preserves a thick, apparently conformable succession that extends from the deep water upper Kaza Group to the shallow water Yanks Peak Formation of the Cariboo Group. Figure 2-2 is a generalized stratigraphic section of the Windermere Supergroup exposed in the Isaac Lake Synclinorium, and Figure 1-5 shows the distribution of units throughout the study area.

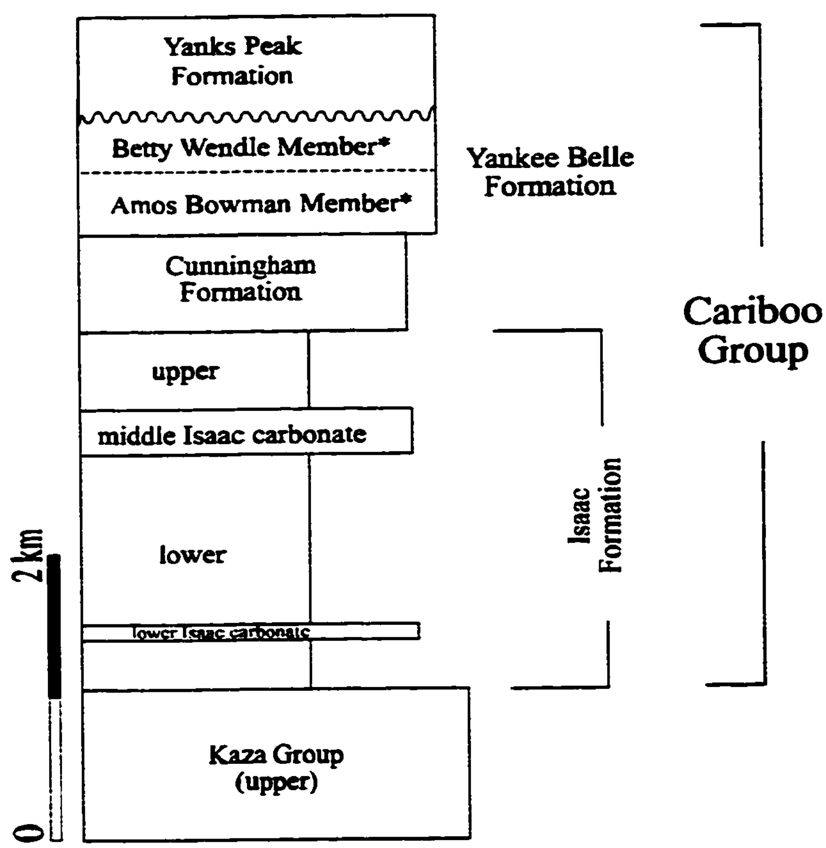

Figure 2-2. Generalized stratigraphic section of the Windermere Supergroup stratigraphy exposed in the Isaac Lake Synclinorium. The lowermost unit is the upper Kaza Group above the "marker" unit (Old Fort Point Formation).

* informal stratigraphic divisions

The rocks in the Isaac Lake Synclinorium have undergone greenschist-grade metamorphism and multiple phases of deformation. Nonetheless, original sedimentary structures are well preserved, allowing for stratigraphic and sedimentological work to be performed. The stratigraphic units exposed in the Isaac Lake Synclinorium are described in the following sections. The descriptions emphasize lithology and field characteristics which aids in structural mapping. Therefore, in this chapter, the units are named based on their mineralogical and grain size composition using the Dunham classification system for carbonate rocks, and the Folk classification system for siliciclastic rocks (Dunham, 1962; Folk et al., 1970).

2.3 Kaza Group

The uppermost strata of the Kaza Group, above the Old Fort Point Formation are exposed on the eastern and western flanks of the Isaac Lake Synclinorium. Figure 2-4 is a composite stratigraphic section of the upper Kaza Group and lower Isaac Formation in the study area with an accompanying legend (Figure 2-5). The Kaza Group in these areas is composed predominantly of thick (10-30 m) successions of coarse-grained sub-arkose sandstone granule and pebble conglomerate beds with minor amounts of interbedded siltstone and shale. The coarse-grained sub-arkose sandstone and pebble conglomerate beds are massive or normally graded and range from 0.5-1.5 m in thickness. The high percentage of sandstone and pebble conglomerate beds makes the Kaza Group a structurally competent unit within the stratigraphic sequence.

The massive sandstone beds contain no internal structures and are commonly composed of coarse to medium size sand grains of quartz and feldspar. The normally graded beds fine upwards from either coarse sand to fine sand or silt size grains; or from gravel or pebble size grains at the base to fine sand at the top. The basal contacts of the beds are commonly sharp and erosive (Plate 1). Pebbles are immature as they are poorly sorted, sub-angular and are composed of single quartz grains and aggregates of polycrystalline quartz and albite feldspar grains that range from 0.75-1.5 cm in size. The pebbles are supported within a matrix of fine-grained quartz, muscovite and chlorite.

Interbedded within the coarse-grained beds of the upper Kaza Group are fine-grained siltstones and shales. These units are composed of silt or mud size quartz grains

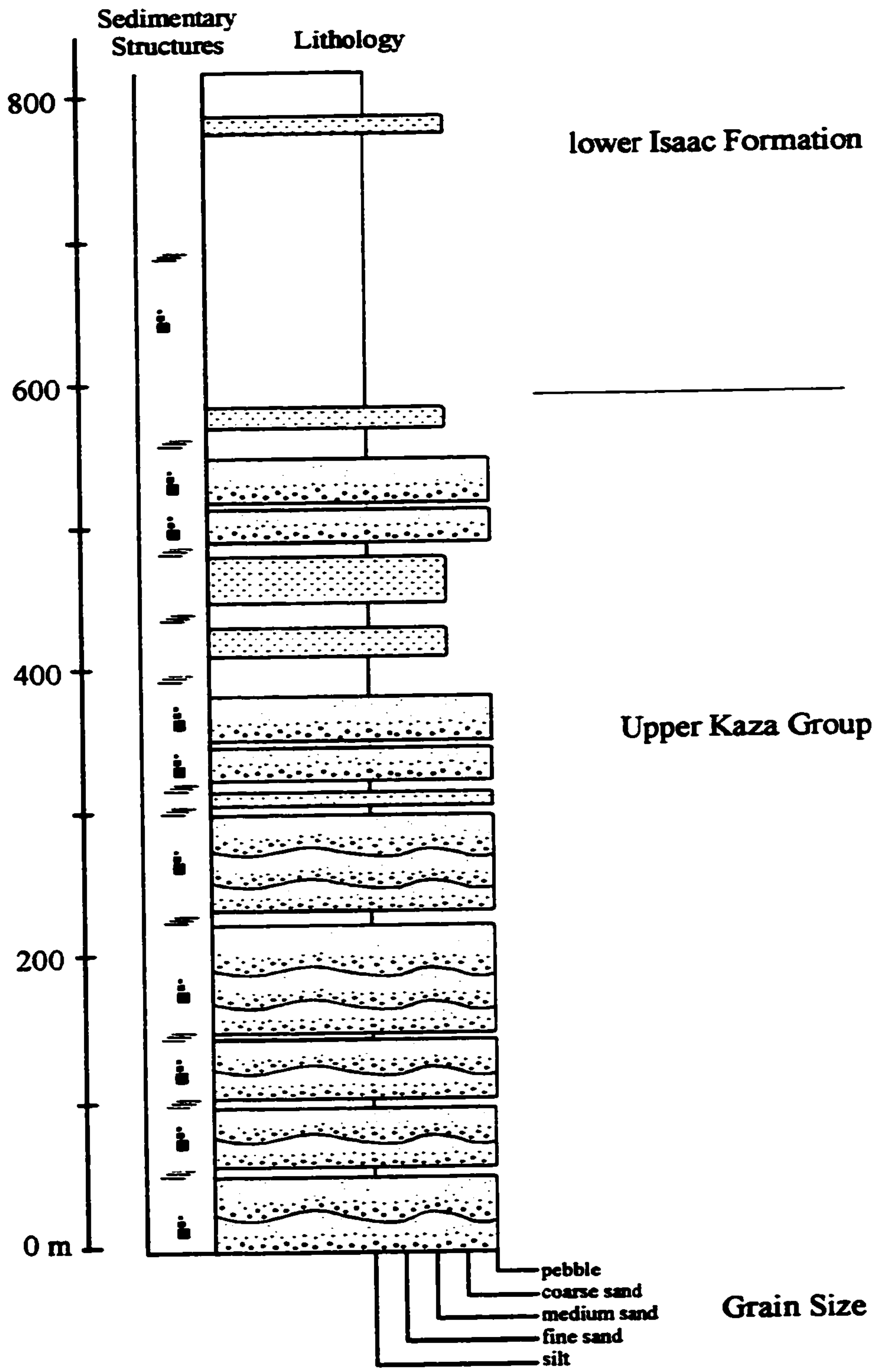

Figure 2-3. Schematic stratigraphic section of the upper Kaza Group exposed in the study area. The upper Kaza Group is composed predominantly of massive and graded coarse pebble conglomerates with interbedded shales. Towards the contact with the lower Isaac Formation, the amount of sandstone and pebble conglomerate decreases and the amount of shale increases until the lithology is composed of predominantly shale of the lower Isaac Formation.

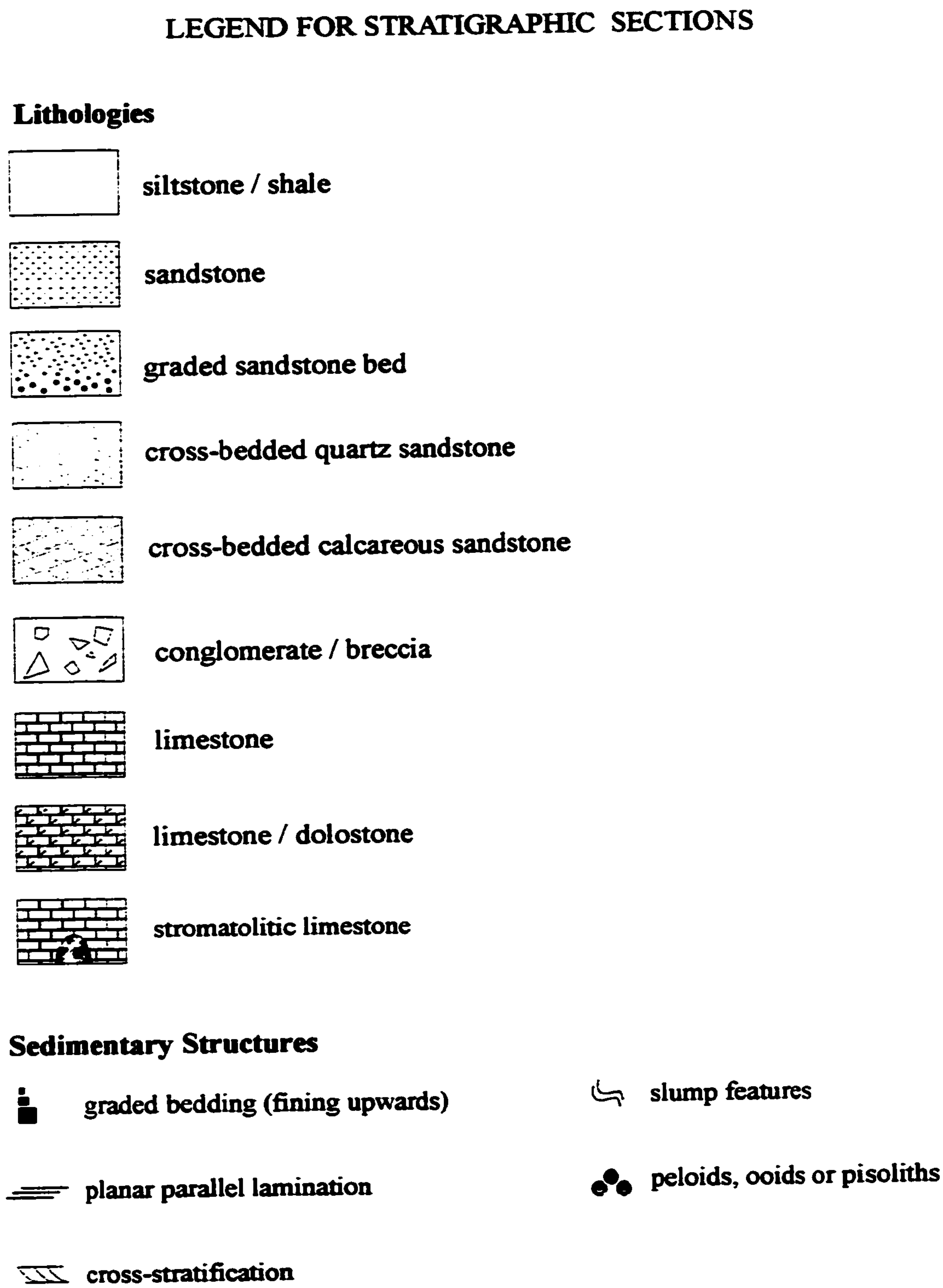

Figure 2-4. Legend for stratigraphic sections. (Figures 2-3, 2-5, 2-6, 2-7 and 2-8).

with fine-grained chlorite, muscovite and pyrite. These intervening siltstone and shale successions are less abundant than the coarse grained sandstone beds and range from 10 to 30 cm in thickness. Planar and cross laminae commonly occur in the siltstones and shales. Near the Kaza Group - Isaac Formation contact, the percentage of siltstone and shale increases relative to the amount of coarse grained sandstone beds. The composition of the shale beds also change towards the contact from gray-green coloured siliciclastic shale beds to black-blue carbonaceous shale beds and show an increase in pyrite. The contact between the Kaza Group and Isaac Formation is a conformable, gradational contact. In the study area, the contact is taken at the base of a thick, shale interval which contains within it a mappable carbonate interval, and only minor coarse grained sandstone intervals.

2.4 Isaac Formation

The Isaac Formation is the lowermost formation of the Neoproterozoic to Cambrian Cariboo Group which conformably overlies the Kaza Group. The Isaac Formation is approximately 2.5 km thick and consists of calcareous mudstones and siltstones with minor amounts of sandstone and carbonate rocks. The Isaac Formation in the study area can be subdivided into four informal units which include the lower Isaac Formation, lower Isaac carbonate, the middle Isaac carbonate and the upper Isaac

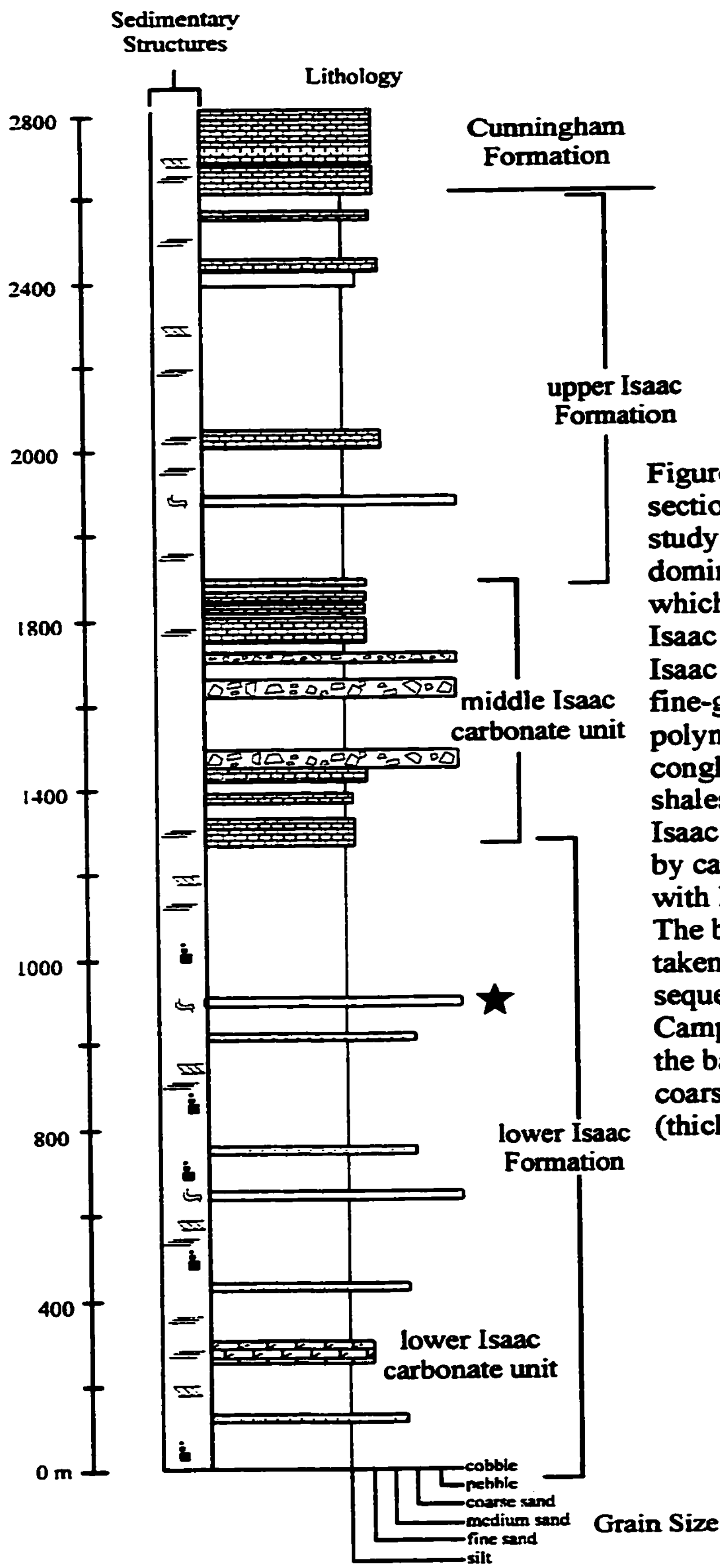

Figure 2-5. Composite stratigraphic section of the Isaac Formation in the study area. The lower Isaac is dominated by turbiditic shales which is punctuated by the lower Isaac carbonate unit. The middle Isaac carbonate consists of massive fine-grained limestone units, polymict and monomict conglomerates with intervening shales and siltstones. The upper Isaac Formation is characterized by calcareous siltstone and shale with local slumped carbonate slabs. The base of the Isaac Formation is taken at the base a thick shale sequence which contrasts with that of Campbell et. al. (1973) who take the base at the top of the last coarse-grained sandstone bed (★). (thickness based on map sections)

Formation. These map units are described in the following sections. Figure 2-5 is a composite stratigraphic section of the Isaac Formation in the study area. Overall, the Isaac Formation is a structurally incompetent unit of siltstone and shale with the exception of the lower and middle carbonate units which form rigid competent units within it.

2.4 a) <u>lower Isaac Formation</u>

The lowermost Isaac Formation, from the Kaza Group-Isaac Formation contact to the middle Isaac carbonate, consists mainly of shale and siltstone beds, rare sandstone beds and a distinct mappable carbonate interval. The shale and siltstone beds dominate the sequence and range in thickness from 2 to 10 cm and are massive or more commonly fine upwards in grain size with planar parallel laminations (Plate 2). These beds approximate classic Bouma turbidite beds on the basis of thickness and structure (Tb and Tc beds), (Bouma, 1962). Small scale crossbeds are seen rarely. The shales are made up of quartz (50-60%), muscovite (5-15%), chlorite (2-10%), feldspar (1-5%), calcite (<1-3%), and pyrite (<1%). Within the lower 500 m of the Isaac Formation, local sandstone units are encountered which include interbedded siltstone and shale. These coarser grained siliciclastic packages in the lower Isaac Formation range in thickness from 10-20 m. Rare slump beds are also found in the lower Isaac Formation which contain clasts of siltstone and shale supported in a mud matrix and show convolute folding (Plate 3).

2.4 b) lower Isaac carbonate

The lower Isaac Formation carbonate unit occurs near the base of the lower Isaac Formation punctuating the dominantly fine-grained siliciclastic sequence described above. It consists of four types of carbonate rock and is approximately 30 to 40 m thick. The carbonate unit consists of 1) an orange weathering lime-mudstone which contains fine, planar parallel laminations and a dark gray interior, 2) a tan colored lime-mudstone unit which contains fine-grained recrystallized calcite with minor quartz grains (2%), 3) a beige weathering lime-mudstone which is interbedded with a light gray weathering lime-mudstone, and 4) a massive light gray weathering lime-mudstone with a dark gray recrystallized interior. Beds range from 1-5 cm in thickness and contain planar laminations. Both beige and gray mudstones have a dark gray fresh surface. All of the units are separated by thin intervals of finely laminated blue-gray shales ranging in thickness from 20 to 50 cm.

Above the lower carbonate interval in the Isaac Formation is a thick sequence of blue-gray coloured shales with minor sandstone. Individual beds range from 2-5 cm thick and are commonly planar parallel laminated beds which fine upwards, and include rare cross laminated beds. Punctuating the overall shale dominated sequence are rare massive sandstone beds which range from 5-20 cm thick. Approximately 500 m below the contact with the middle Isaac carbonate unit, there is an increase of carbonate (up to 15%) in the shale beds, which weather a distinct orange colour.

2.4 c) middle Isaac carbonate unit

The middle Isaac carbonate unit is the second mappable carbonate-rich interval within the dominantly fine-grained siliciclastic Isaac Formation. The upper and lower contacts of the middle Isaac carbonate are difficult to distinguish as there is a gradational increase in carbonate in the lower Isaac Formation towards the lower contact with the middle Isaac carbonate unit. The contact is generally taken at the base of a massive gray-weathering limestone unit. Figure 2-6 contains a composite stratigraphic section of the middle Isaac carbonate in the study area. The lithology of the middle Isaac carbonate changes laterally and the thickness of the unit can vary from 100-400 m. The unit is characterized by the presence of massive gray-weathering limestones, and less common buff weathering dolostones within orange-weathering calcareous shale units. These intervals are thinly bedded with individual beds ranging from 2-5 cm and are massive to finely laminated. Unique to the middle Isaac carbonate unit are laterally discontinuous poorly sorted polymict and monomict conglomerates that form 1.0-2.5 m thick lenticular sheets (Plate 4). The polymict conglomerate beds contain clasts which include well preserved oolitic and pisolitic packstones and wackestones, cross-laminated calcareous sandstones, buff-weathering fine-grained dolostones, and shale (Plate 5). Clasts range in size from 5-50 cm and are sub-angular to sub-rounded. The polymict

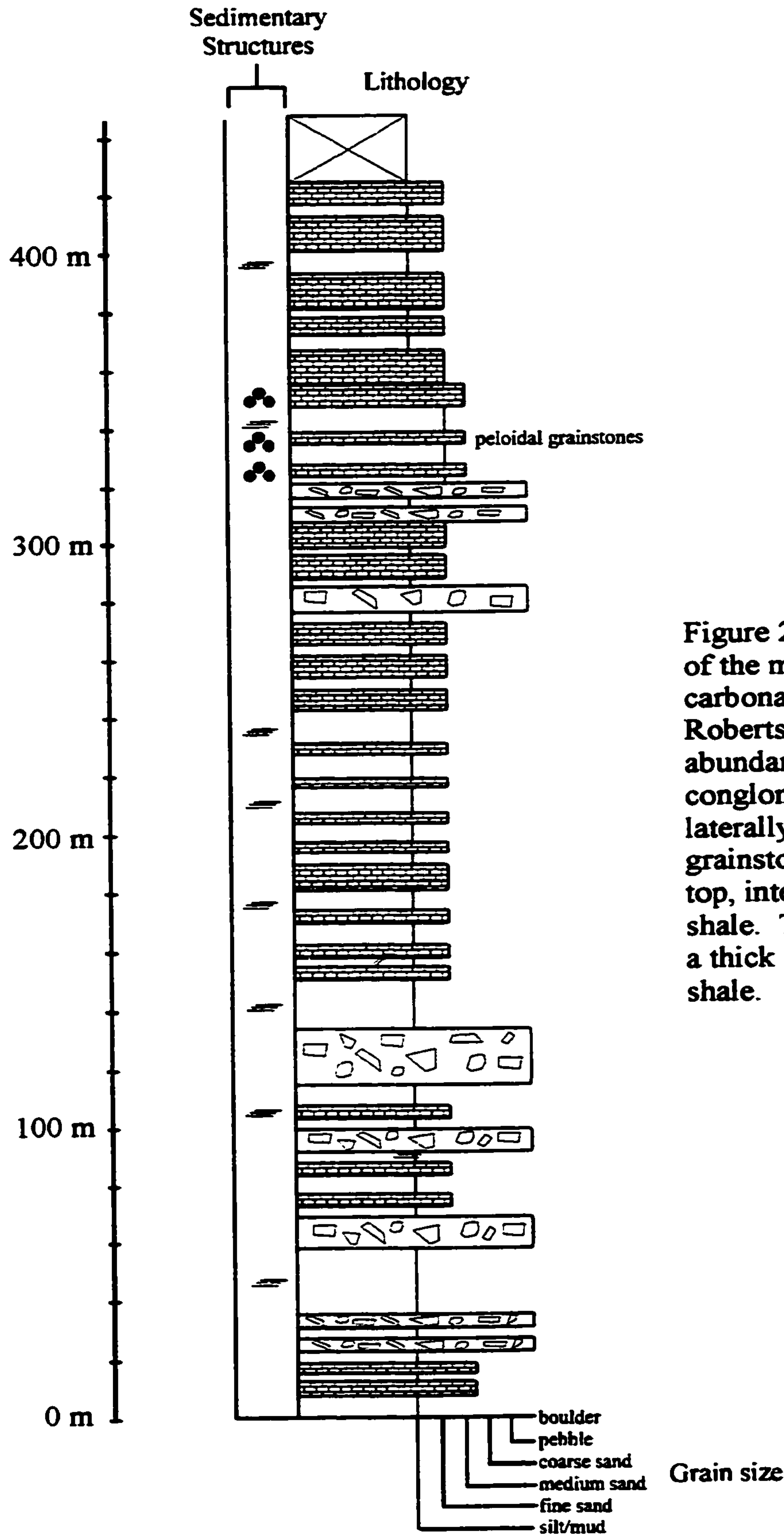

Figure 2-6. Stratigraphic section of the middle Isaac Formation carbonate unit in the Mount Roberts area. The unit contains abundant polymict and monomict conglomerate beds which are laterally discontinuous. Peloidal grainstones are common near the top, interbedded with calcareous shale. The unit is overlain by a thick sequence of black shale.

conglomerates are matrix-supported within a fine-grained gray weathering calcareous, oolitic matrix. The monomict conglomerates consist of a fine-grained carbonate matrix with buff-colored clasts of dolostone which range from 10-50 cm. The top of the middle Isaac carbonate is taken at the base of a black siliciclastic shale unit which marks the base of the upper Isaac Formation.

2.4 d) upper Isaac Formation

The upper Isaac Formation consists of predominantly calcareous, normally graded silt and shale beds which range from 2-5 cm thick (Plate 6). The shales commonly contain planar parallel laminations and rare crossbeds. Slump beds are common in the upper Isaac Formation which contain rip-up clasts of siltstone and shale and internal convolute folding similar to those found in the lower Isaac Formation. Fine to medium-grained massive sandstone beds are also found throughout the upper Isaac Formation. Above the middle Isaac carbonate the amount of carbonate in the shale beds increases (15-25% calcite) towards the limestone dominated Cunningham Formation. Within the upper 200 m of the Isaac Formation, thin fine-grained gray weathering limestone beds become interlayered with the calcareous shales. Locally within the upper Isaac Formation, large 20 m thick slump blocks of massive gray-weathering carbonate of uncertain origin occur. The contact with the overlying Cunningham Formation is a gradational conformable contact and it is taken at the point where the lithology is dominated by thick (up to 3 m), massive gray limestone beds with minor interbedded shale (Plate 7).

2.5 Cunningham Formation

The Cunningham Formation is a 500 m thick succession of limestone with minor amounts of interbedded shale and siltstones, which makes it an extremely rigid and competent unit. Figure 2-7 is a composite stratigraphic section of the Cunningham Formation in the study area. The Cunningham Formation conformably overlies the Isaac Formation and their contact is gradational, over approximately 20 m. The limestone units of the Cunningham Formation include massive to finely laminated lime-mudstones, and recrystallized peloidal packstones and grainstones, and show an overall gradation from mudstone dominated at the base to dominantly peloidal grainstone at the top of the formation. All carbonate units examined petrographically show the presence of detrital quartz ranging from <1-3%.

The mudstone units contain mud to silt size recrystallized calcite with minor amounts of quartz, muscovite and pyrite. Beds range from 0.5-2 m thick and can be massive or show fine, internal, planar laminations and occasionally cross laminations (Plate 8). The peloidal packstones and grainstones consist of individual peloids and intraclasts of peloidal grainstones in a mud matrix. The packstones and grainstones can be massive, but may also contain internal, planar, parallel laminations or rare high-angle cross-laminations. The peloids are strongly recrystallized although faint, concentric

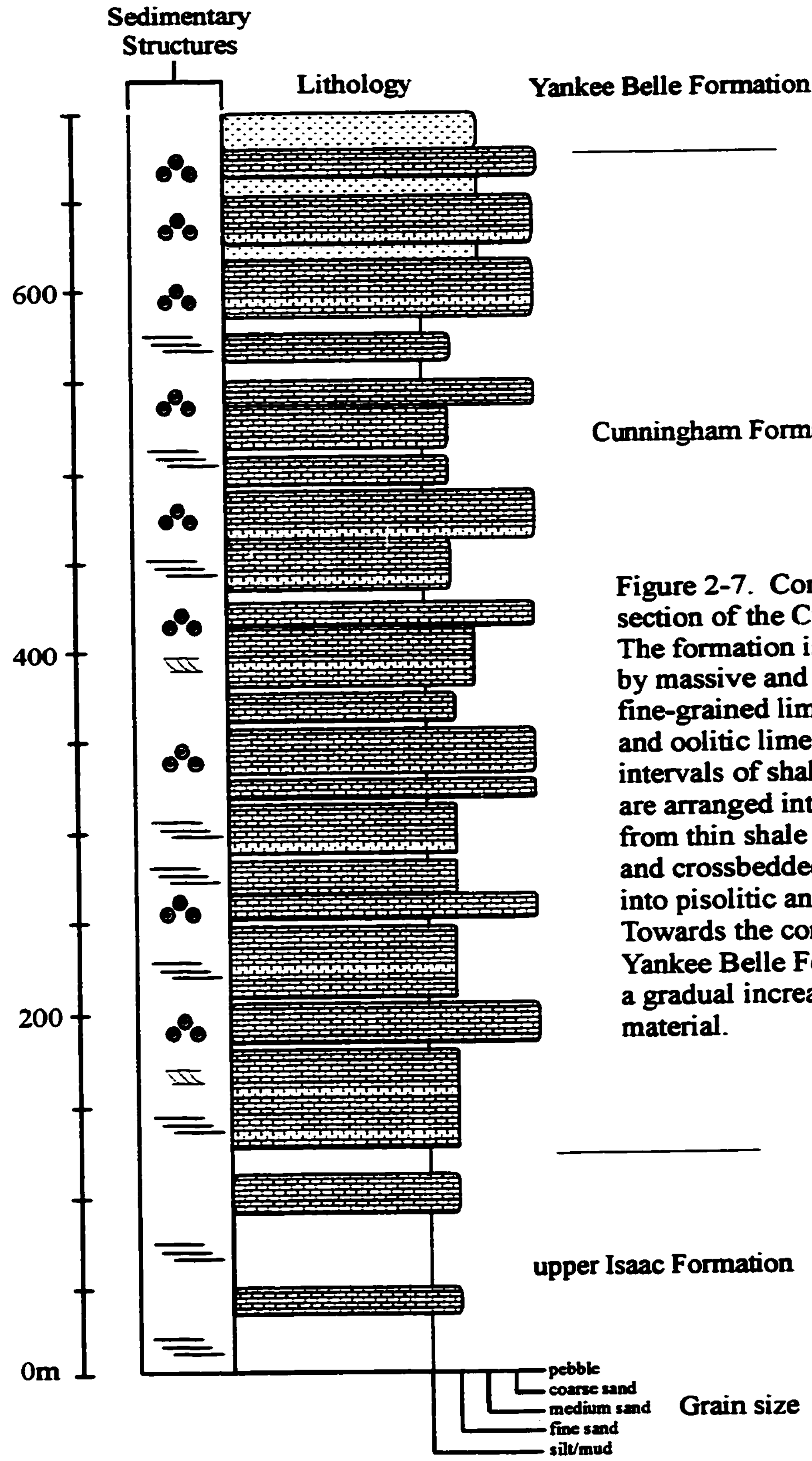

Figure 2-7. Composite stratigraphic section of the Cunningham Formation. The formation is dominated by massive and thinly laminated fine-grained limestones, pisolitic and oolitic limestones with thin intervals of shale. The lithologies are arranged into cycles which grade from thin shale intervals into massive and crossbedded gray limestone into pisolitic and oolitic limestones. Towards the contact with the Yankee Belle Formation there is a gradual increase in siliciclastic material.

laminae around a core of sparry calcite are recognizable under the petrographic microscope (Plate 9). The peloids are best defined on weathering surfaces as they weather a distinct orange color. The peloids are rounded to sub-rounded and range from 0.3-1.5 cm in diameter (Plate 10). Thin shale intervals (10-30 cm thick) are found between the mudstones.

The upper contact of the Cunningham Formation with the Yankee Belle Formation is a gradational conformable contact. In the top 100 m of the Cunningham Formation there is a increase in interbedded shales, siltstones and fine-grained sandstones. The contact between the Cunningham Formation and Yankee Belle Formation is taken at the base of a thick siliciclastic interval of interbedded shales, siltstones and sandstones with no intervening carbonate units.

2.6 Yankee Belle Formation

The Yankee Belle Formation is a mixed carbonate-siliciclastic sequence which conformably overlies the Cunningham Formation in the study area.. The competency of this formation is variable depending on the location and thickness of the carbonate units within it. The entire formation is approximately 800 m thick and can be subdivided into two distinct map units based on lithology and their inferred depositional characteristics. These two map units are informally known as the lower cyclic member (or Amos Bowman Member) and the Betty Wendle Member. Figure 2-8 is a composite

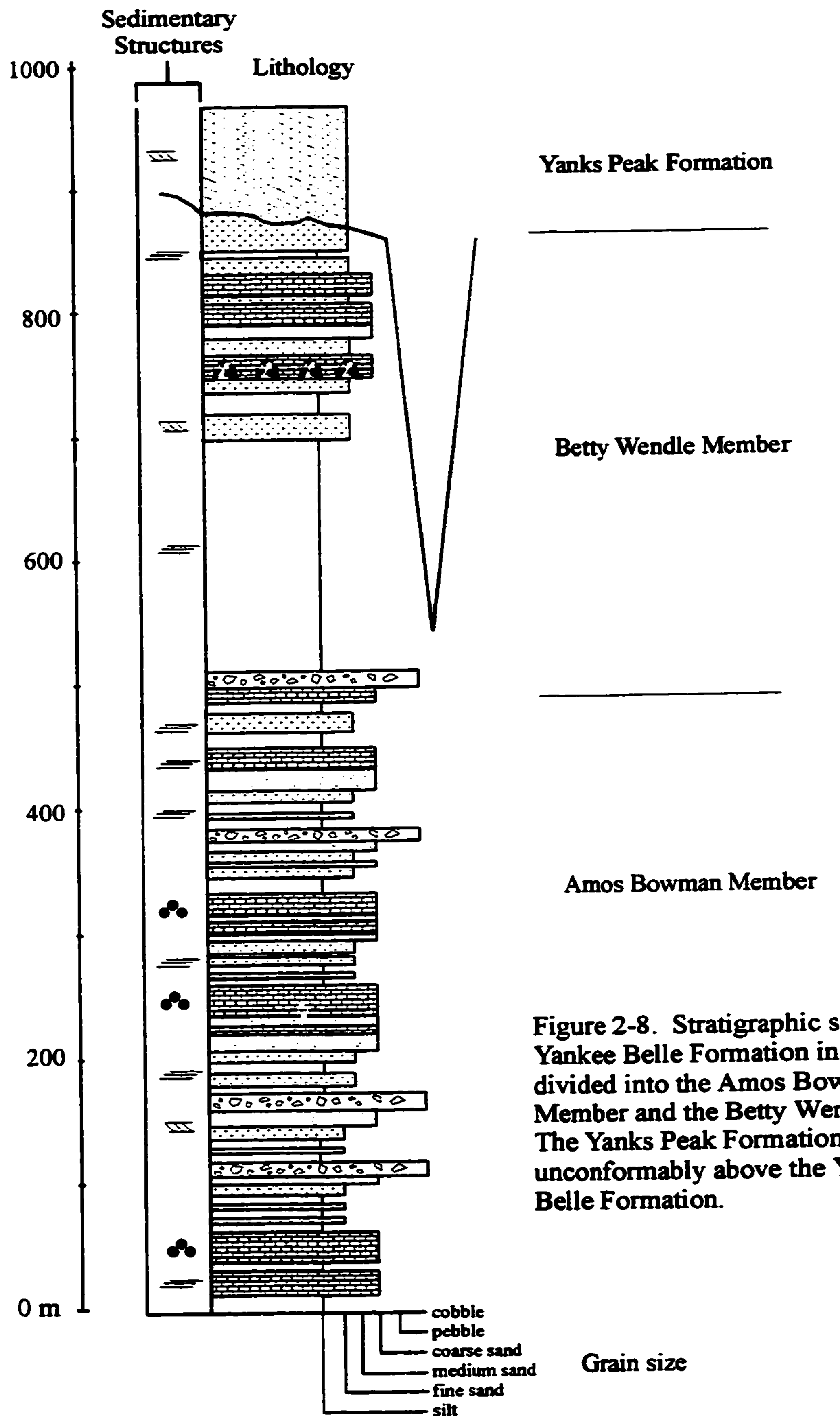

Figure 2-8. Stratigraphic section of the Yankee Belle Formation informally divided into the Amos Bowman Member and the Betty Wendle Member. The Yanks Peak Formation lies unconformably above the Yankee Belle Formation.

stratigraphic section of the Yankee Belle Formation in the study area. The Amos Bowman Member and the Betty Wendle Member will be described individually in the following sections as the Betty Wendle Member is a local stratigraphic marker with potential regional significance.

2.6 a) Amos Bowman Member

The Amos Bowman Member is the lower map unit of the Yankee Belle Formation and is approximately 580 m thick. The unit consists of a number of siliciclastic and carbonate lithologies arranged into repeated cycles. Individual cycles can range from 2 to 60 m in thickness. The cycles begin with interbedded light-medium gray siltstone and shale units and end with limestone. The siltstone beds show grading from silty to fine-sand bottoms which fine upwards into shale tops. The siltstone and shale packages are thinly bedded (1-5 cm thick), and individual beds commonly contain internal parallel laminations. The siltstone and shale are commonly pyritic with small euhedral pyrite crystals ranging from 0.1-0.5 cm in diametre. Silt-shale packages are commonly interbedded with fine to medium grained sandstones which contain low-angle, tabular cross stratification. The sandstone beds range from 1 to 40 cm and can be laterally discontinuous and lens shaped or continuous for hundreds of meters (Plate 11). The siltstone and shale packages interbedded with the sandstone beds make up the lower siliciclastic portion of a cycle. Overall, a gradation from predominantly siltstone and shale with minor, thin sandstone interbeds to predominantly sandstone beds with minor siltstone-shale interbeds occurs in the lower siliciclastic portion. The thickness of the siliciclastic portion varies dramatically between cycles as does the proportions of siltstone-shale to sandstone beds.

The sandstones that mark the top of the siliciclastic portion of a cycle commonly grade upward into calcareous sandstones that contain low-angle, planar and trough cross stratification and planar parallel laminations (Plate 12). The calcareous sandstone beds range in thickness from 1 to 50 cm in thickness, and commonly weather a brown to orange color. The calcareous sandstone beds are either sharply overlain by breccia beds or grade into carbonate mudstones or peloidal packstones. The breccia beds show sharp erosive contacts with the calcareous sands below. The breccia beds range in size from 0.3 to 1 m in thickness. They contain angular to sub-angular clasts of calcareous sand beds and quartz sandstones (Plate 13). The beds commonly extend for tens of metres laterally and eventually pinch out, giving them a lenticular shape. The breccia beds represent the top of a cycle and are subsequently overlain by siltstone-shale beds as previously described. The calcareous sandstone beds are overlain by tabular cross bedded peloidal packstones or mudstones, and the transition is gradational. The carbonate beds have a gray weathering matrix with orange weathering peloids that range from 0.1-2 cm in diameter. These beds can range from 0.3-1 m thick with thin interbedded intervals of calcareous sandstone or shale. The carbonate beds represent the top of a cycle and in turn are overlain by siltstones and shale packages and the cycle begins again.

2.6 b) Betty Wendle Member

The Betty Wendle Member conformably overlies the Amos Bowman Member, but does not have the repetitive cyclicity like the Amos Bowman Member. The basal Betty Wendle Member is characterized by pyritic limestone beds with an anastamosing ("chickenwire") weathering texture interbedded with breccia beds (Plate 14). The chickenwire limestone is everywhere restricted in its stratigraphic position and marks the base of the Betty Wendle member. The limestone and breccia beds range from 30 to 75 cm in thickness and are interbedded with green and purple coloured shales. The breccia beds contain clasts of the chickenwire limestone, green weathering shale and cross-bedded calcareous sandstone. This entire interbedded interval is 15 to 20 m thick and is abruptly overlain by a thick package of green shales and siltstones. The shales are predominantly siliciclastic and contain quartz, muscovite, chlorite, pyrite, minor calcite, albite and tourmaline (<1%). The green shales and siltstones are approximately 200 m thick. Higher in the succession, fine-grained sandstones are interlayered with the shales and siltstones.

Above the basal 100 m of interbedded fine-grained sandstones, siltstones and shales, thin limestone beds are interlayered and grade upwards into a stromatolitic boundstone unit. The stromatolites occur as 2-3 m high bioherms composed of laterally linked and vertically stacked hemispheroids. Above the stromatolitic unit, the lithology becomes predominantly siliciclastic including fine to medium size grains, cross-bedded sandstones and siltstones and minor limestone beds (Plate 15). The upper contact of the

Yankee Belle Formation with the Yanks Peak Formation is taken as the base of a thick clean quartz arenite unit.

2.7 Yanks Peak, Midas and Mural Formations

The boundary between the Yankee Belle Formation to the Yanks Peak Formation is represented by an abrupt facies change from shales, sandstones and carbonate rocks of the Yankee Belle Formation to clean quartz arenites and pebble conglomerates of the Yanks Peak Formation. The Yanks Peak Formation was observed in two localities in the study area. In both localities, only the lower 20 m or less of the Yanks Peak Formation is present. At the locality west of Mount Hogg, the lower Yanks Peak Formation consists of clean quartz arenites with high angle, large-scale cross stratification (Plate 16). These quartz arenites overlie interbedded sandstones and siltstones of the upper Betty Wendle Member, and the transition to the lower Yanks Peak is rapid. At the locality south of the Castle Creek-Niagara Creek junction, the Yanks Peak Formation consists of quartz pebble conglomerates which show planar laminations and low angle cross-stratification. The quartz pebble conglomerates of the Yanks Peak Formation lie above the green shales and fine sandstones of the lower Betty Wendle Member. The relationship between the quartz arenites and pebble conglomerates of the Yanks Peak was not observed.

The Midas Formation lies conformably above the Yanks Peak Formation and is exposed in the eastern part of the study area (Campbell, 1978), (Figure 1-5). The Midas

Formation was mapped in this locality by Campbell (1978) and consists of interbedded shale and siltstone with minor sandstone.

2.8 Summary / Discussion

The Neoproterozoic Windermere Supergroup stratigraphy exposed in the Isaac Lake Synclinorium forms a five kilometre thick, conformable succession from the upper Kaza Group (above the marker unit) to the top of the Yankee Belle Formation of the Cariboo Group, with the Yanks Peak sitting unconformably above the Yankee Belle. The stratigraphic divisions of the Cariboo Group, include the Isaac, Cunningham, Yankee Belle and Yanks Peak Formations which form distinct map units. The Isaac Formation can be further subdivided into four informal map units including the lower Isaac Formation, lower Isaac carbonate unit, the middle Isaac carbonate unit, and the upper Isaac Formation. These units may thicken and thin laterally, but each form distinct, continuous map units in the study area. The Yankee Belle may also be further subdivided into two informal map units including the Amos Bowman Member (Lemieux, 1996), and the Betty Wendle Member (Ferguson, 1994). The stratigraphic relationships of the formations to each other are similar to those to the northeast as mapped by Ferguson (1994) with one major difference. Northeast of the study area, the Cunningham-Yankee Belle Formation contact is locally unconformable and the top of the Cunningham is marked by karst surfaces, and the base of the Yankee Belle

Formation consists of a clean quartz sandstone known as the Zig Zag Member (Ferguson, 1994). This contrasts with the Cunningham Formation-Yankee Belle Formation contact in this study area, which is a conformable and gradational contact and the Zig Zag Member is not present. Sedimentary structures found within the various formations are consistent with the depositional interpretations for the Windermere Supergroup being a passive margin succession deposited on the western margin of North America as summarized in the introduction of this chapter.

All of the formation contacts within the Windermere Supergroup in the study area are gradational and conformable with the exception of the Yankee Belle Formation-Yanks Peak Formation contact which is interpreted in the study area to be an unconformity based on the abrupt facies change from the upper Yankee Belle Formation to the lower Yanks Peak Formation. Evidence for this is best seen at the locality south of the Castle Creek-Niagara Creek junction where pebble conglomerates of the Yanks Peak Formation sit on top of the green siltstones and shales of the Betty Wendle Member approximately 50 m above the chickenwire carbonate. This implies that in some localities a minimum of 200 m of upper Yankee Belle Formation has been lost to erosion below the sub-Yanks Peak unconformity. In the locality along the western margin of Wells Gray Provincial Park, the contact is more subtle, where the quartz arenites of the Yanks Peak are in contact with fine sandstones and siltstones of the upper Betty Wendle member. The age of the lower Yanks Peak Formation remains unknown as no Cambrian trace fossils were found. Cambrian-age fossils do occur in the upper half of the Yanks Peak Formation, thus the age of the upper portion is known (Ferguson, 1994).

The presence of the Betty Wendle member throughout the study area and the remarkable similarity to the stratigraphy described by Ferguson (1994) to the north, has significant regional importance for the Windermere Supergroup. The Betty Wendle Member is a regional marker unit distinguished by the rapid transition from the shallow-water "chickenwire" carbonate and carbonate breccias to the deep-water green and purple shales of the lower Betty Wendle Member. This transition has been interpreted to represent a major marine transgression (Ferguson, 1994; Lemieux, 1996). The Betty Wendle member may then be a regional marker unit within the upper Windermere Supergroup stratigraphy where preserved.

Plate 1. Base of coarse-grained arkosic turbidite bed from the upper Kaza Group. The base of this bed scours the thin-bedded turbidite beds below which is coarse-grained at the bottom and fines upwards into the laminated siltstone top.

Plate 2. Thinly bedded and internally laminated turbidite beds from the lower Isaac Formation. Beds grade from silty bottoms to shaley tops, and contain abundant pyrite porphyroblasts.

Plate 3. Slump bed from the lower Isaac Formation. Matrix consists of shale and silt size sand with clasts of laminated calcareous sandstone and siltstone.

Plate 4. Polymict conglomerate from the middle Isaac carbonate unit. Clasts consist of buff colored dolostone, oolitic and pisolitic limestone, laminated siltstone and cross-laminated calcareous sandstone, and range from 5 to 50 cm. Matrix consist of an oolitic calcareous silt.

Plate 5. Rounded oolitic limestone clast from a polymict conglomerate bed in the middle Isaac carbonate (Plate 4). Upper left corner contains a clast of laminated calcareous sandstone. (Note: coarse sandy matrix and predominence of clasts of shallow water derivation.)

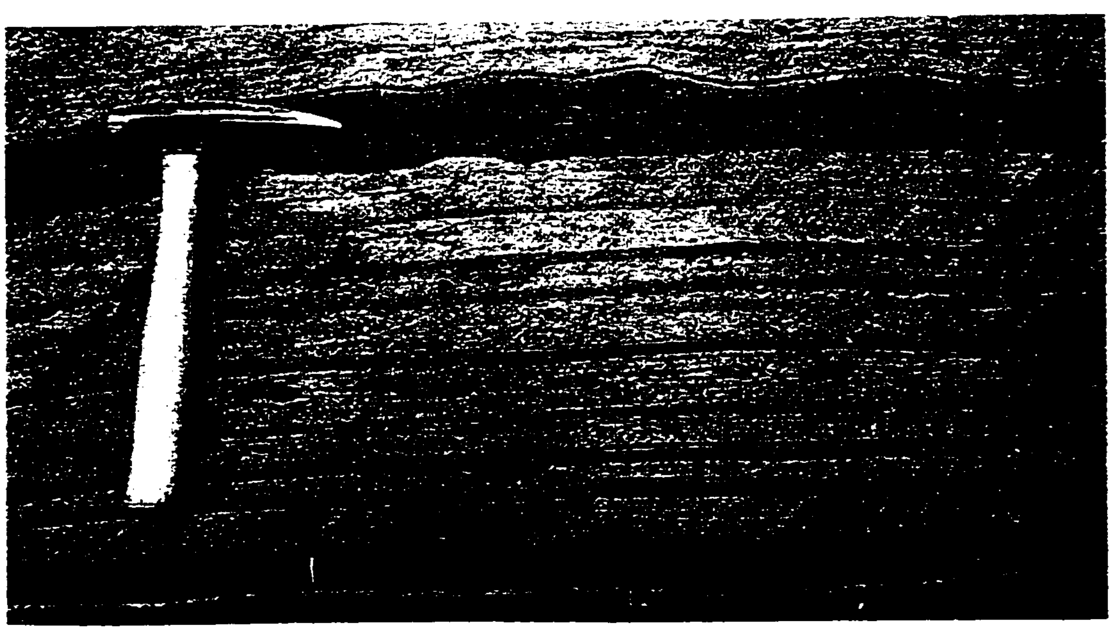

Plate 6. Pyritic, calcareous siltstone beds from the upper Isaac Formation. Beds are graded and range from silty bottoms to shale tops.

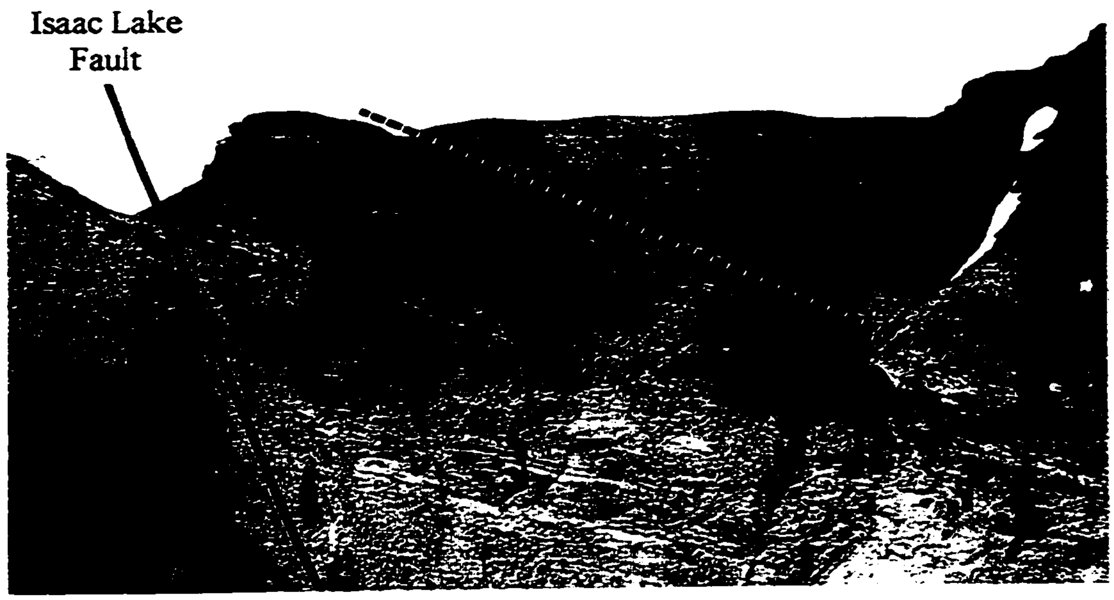

Plate 7. Slump block of massive carbonate in the upper Isaac Formation. Carbonate block is laterally discontinuous. Cliff face is approximately 1500 ft from the base of the photo to the ridge. (Note: apparent discontinuities in beds within the block and apparent onlap of Isaac shales onto bed)

Plate 8. Cross-laminated limestone bed from the lower Cunningham Formation.

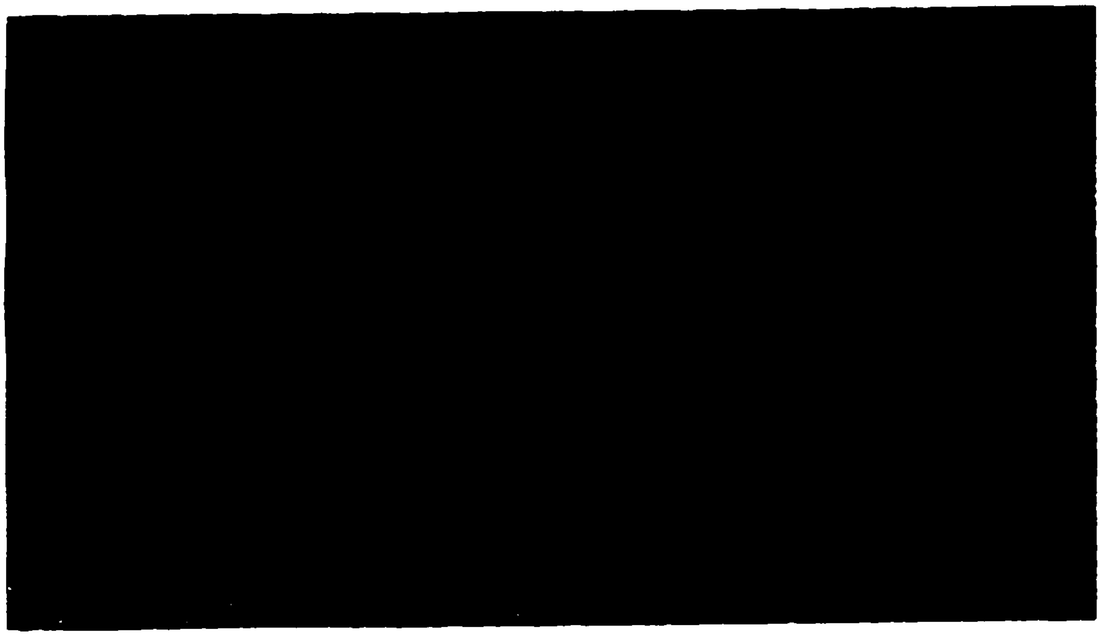

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Plate 9. Photomicrograph of recrystallized peloid from the middle Cunningham Formation. Peloids contain coarse sparry calcite cores with micritic rims (PPL)

Plate 10. Peloidal packstone with cross-cutting dolomite vein from the middle Cunningham Formation. Peloids contain coarse crystalline calcite with concentrically laminated rims.

Plate 11. Interbedded siltstone and sandstone beds from the lower Amos Bowman member of the Yankee Belle Formation. Sandstone beds are commonly lenticular and laterally discontinuous.

Plate 12. Trough-cross stratified calcareous sandstone beds from the Amos Bowman Member.

Plate 13. Breccia bed overlying the trough-cross stratified calcareous sandstone beds of the Amos Bowman Member. Clasts consist of laminated sandstone and cross-bedded calcareous sandstone, and range in size from 1 to 10 cm. Matrix consists of silt size sand.

Plate 14. Chickenwire carbonate from the lower Betty Wendle Member of the Yankee Belle Formation. Beds range from 30 to 70 cm thick, and consist predominantly of calcite. The texture of the carbonate beds is distinct to the beds of the lower Betty Wendle Member.

Plate 15. Thinly laminated sandstones and siltstones of the upper Betty Wendle Member

Plate 16. Cross-bedded quartz arenites from the lower Yanks Peak Formation.

CHAPTER 3: METAMORPHISM

3.1 Metamorphism in the Cariboo Mountains

Although the Omineca Crystalline Belt is characterized by high-grade metamorphic rocks, typical of orogenic hinterlands, it also includes large tracts of rocks deformed and metamorphosed under low-grade conditions at relatively shallow structural levels (10-15 km). One of the greatest exposures of rocks deformed and metamorphosed under such conditions occurs in the northern Cariboo Mountains. The main regional structures of the Cariboo Mountains are a series of northwest-plunging anticlinoria and synclinoria-type structures that define a regional structural depression such that rocks that are structurally shallow are exposed in the northwest and rocks at deeper structural levels are exposed in the southeast. This change in structural depth is accompanied by a change in metamorphic grade from southeast to northwest in the Cariboo Mountains, as first recognized by Campbell (1970). Overall, there is a northwest to southeast progression from older to younger peak metamorphism as well as an increase in metamorphic grade (Campbell, 1970; Pigage, 1977; Pell, 1984; Murphy, 1985; Gerasimoff, 1988; Currie, 1988). The rocks at higher structural levels in the northwestern Cariboo Mountains contain greenschist facies mineral assemblages, and rocks at deeper structural levels in the southeast contain amphibolite facies mineral assemblages. Peak metamorphism of the low-grade rocks in the northern Cariboo Mountains is interpreted to have occurred pre-174 Ma based on the cross-cutting relationship of the 174 Ma Hobson pluton to peak metamorphic structural and metamorphic fabrics (Pigage, 1977; Gerasimoff, 1988). In the southern Cariboo Mountains, peak metamorphism occurred at 135 +/-4 Ma based on a U/Pb age of metamorphic monazite in the Allan Creek area (Currie, 1988). Figure 3-1 is a metamorphic facies map showing the different metamorphic assemblages found within the Cariboo Mountains.

The stratigraphy of the Windermere Supergroup within the Isaac Lake Synclinorium is preserved at sub-biotite greenschist facies. The following section contains petrographic descriptions of the different metamorphic minerals found within the Isaac Lake Synclinorium. Each mineral is described individually based on texture and occurrence. Forty-five structurally oriented thin sections were examined in order to determine the metamorphic mineral assemblages and the relation of these minerals to the structures found in the study area. From these forty-five thin sections, eight were chosen which contained important structural relationships between metamorphic minerals and structural fabrics and a second set of oriented thin sections were made cut perpendicular to the first set. The preceding chapter described the rocks according to their original sedimentological names. The rocks in this area have been altered by low-grade metamorphism and will herein be referred to by their true metamorphic names. Shales and siltstones are metamorphosed to slates or phyllites, and sandstones metamorphosed to quartzites.

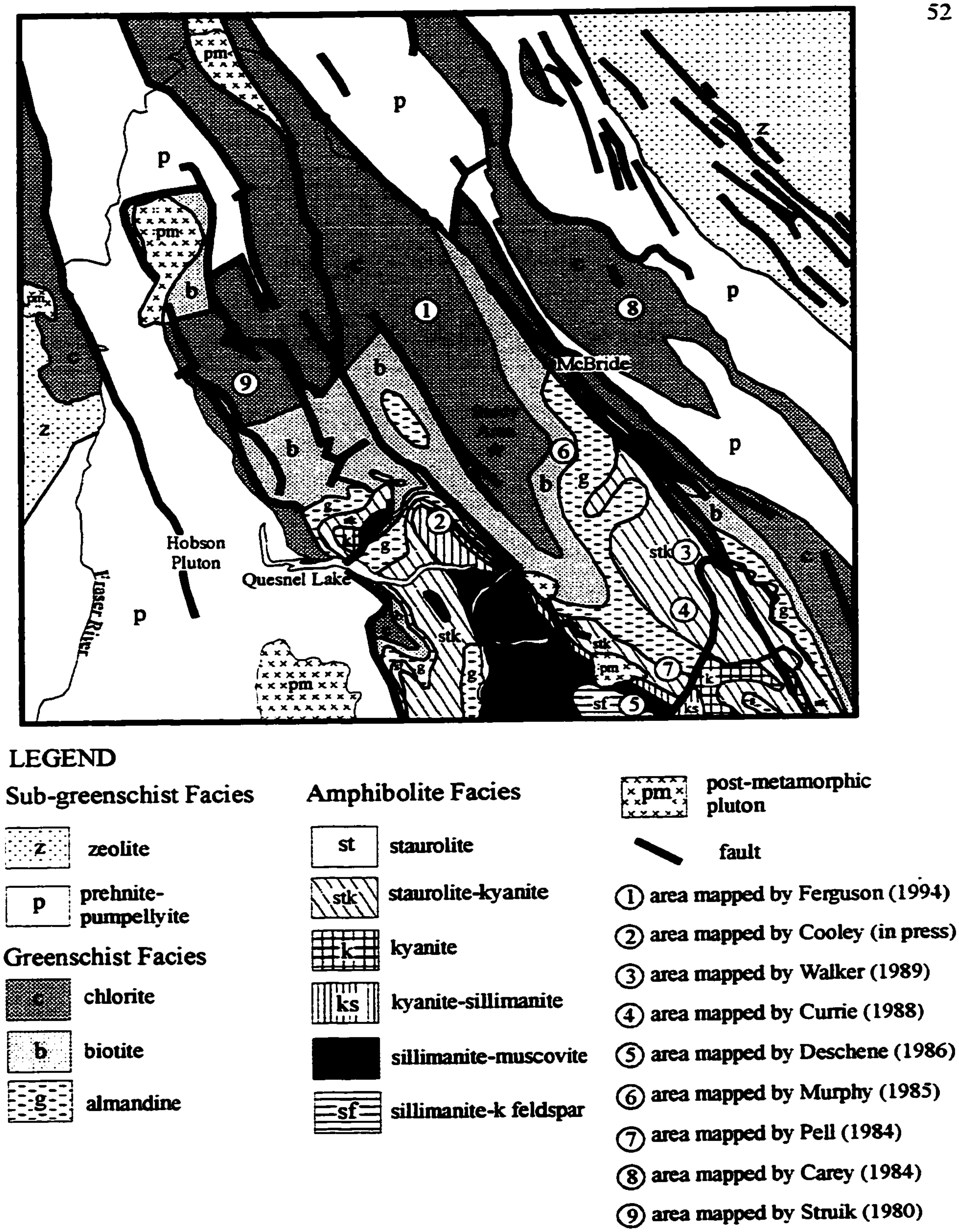


Figure 3-1. Metamorphic facies map of the Cariboo Mountains and adjacent Rocky Mountains. (modified from Read et al., 1991)

3.2 Metamorphic mineral assemblages

3.2 a) Quartz

Quartz is present in all lithologies except for very pure carbonate rocks. Quartz grains occur both as a recrystallized matrix mineral, and in pressure fringes around rigid porphyroblasts such as pyrite and chlorite-muscovite aggregates. Quartz grains show weak to moderate deformation and different degrees of recrystallization depending on the position within the structural package of deformed strata. Strongly foliated slates and phyllites in the Isaac Formation contain quartz grains with strong undulose extinction and interstitial quartz grains have a preferred shape orientation with long axes lying in the cleavage plane, parallel to the intersection lineation (Plate 17). Quartz in weakly foliated pelites of the Yankee Belle Formation shows weak undulose extinction and has sutured grain boundaries where in contact with other grains (Plate 18).

Quartz also occurs within pressure shadows and pressure fringes around rigid porphyroblasts within the matrix. According to Spry (1969), pressure shadows do not have an internal fibrous structure whereas pressure fringes do. The quartz pressure shadows are found only around chlorite-muscovite aggregates where a strong foliation is developed. The pressure shadows are composed of polycrystalline quartz, that grew parallel to the intersection lineation direction within the cleavage plane. Quartz in pressure fringes occurs around pyrite porphyroblasts and show a preferred orientation with their long axes parallel to the inferred extension direction within the cleavage plane and perpendicular to the intersection lineation. The development and intensity of the pressure fringe depends on the location relative to the main regional structures, and location within the stratigraphic succession. Pyrite porphyroblasts in the Isaac Formation show complex fringe development and are moderately to completely recrystallized (Plate 19). Fringes have a length to width ratio ranging from 12:1 to 3:1. In some cases, quartz in earlier formed pressure fringes are completely recrystallized and are composed of well annealed polygonal quartz grains (Plate 20). Units of similar lithology in the Yankee Belle Formation with pyrite porphyroblasts have only weakly developed pressure fringes (Plate 21). Where fringes are developed, they have a length to width ratio of 2:1 and the quartz does not appear to be recrystallized but does show undulose extinction.

3.2 b) Feldspar

Plagioclase feldspar occurs in minor amounts in the pebble conglomerates of the Kaza Group and in sandstones of the Yankee Belle Formation. In the Kaza Group the plagioclase feldspar occurs as small grains which show original albite twinning (0.5-2 mm) (Plate 22), and as larger (2-5 mm) chessboard albite grains (Plate 23). The grains commonly have serrated recrystallized edges and weak undulose extinction. Plagioclase feldspar in the Yankee Belle Formation occurs within fine sandstones as fine grains (0.5-2 mm) which contain original albite twins. The grains show minor recrystallization around their edges and straight to weakly undulose extinction (see Plate 21).

3.2 c) Muscovite

Muscovite is present in all lithologies except for very pure carbonate rocks. It is found commonly as preferentially aligned grains ranging from 10 to 80 μm in length (Plate 24) and within chlorite-muscovite porphyroblasts. Percentages of muscovite range from 50-60% in pelitic phyllites, 20-30% in semi-pelitic phyllites and 1-2% in carbonate rocks. Muscovite grains show weak to moderate undulose extinction and are sub-idioblastic to xenoblastic in shape (Plate 25). Larger grains (>40 μm) are kinked and broken as a result of deformation post-dating muscovite growth. In addition to metamorphic muscovite, detrital muscovites are also found in the Kaza Group and Yankee Belle Formation quartzites. These detrital flakes are distinguished from metamorphic muscovite on the basis of their large size (200-400 μm) and their occurrence as bent and broken grains around quartz grains (see Plate 21).

3.2 d) Chlorite

Chlorite occurs within slates and phyllites throughout the stratigraphic succession, but overall is less abundant than muscovite. Percentages of chlorite range from 5-15% in phyllites in the Isaac Formation, and up to 40% in phyllites of the Yankee Belle Formation. Chlorite occurs as preferentially aligned grains which show undulose extinction and range from 25-50 μm, and are sub-idioblastic to xenoblastic in shape (Plate 26). In the Yankee Belle Formation, chlorite is more abundant than in the other units and comprises the main phyllosilicate. Chlorite, which displays anomalous purple/blue interference colors is also found within the Yankee Belle Formation, particularly within phyllites of the Betty Wendle Member (Plate 30). Chlorite also occurs within pressure fringes with quartz around rigid pyrite porphyroblasts.

3.2 e) Chlorite-muscovite aggregates

Chlorite-muscovite aggregates are found within pelitic and semi-pelitic phyllites and quartzites of the Kaza Group and the Isaac and Yankee Belle Formations. The porphyroblasts are composed predominantly of chlorite with fine intergrowths of muscovite, and rare quartz, parallel or sub-parallel to the 001 basal cleavage plane in chlorite (Plate 27). The chlorite and muscovite within the aggregate usually show weak to moderate undulose extinction, however, some chlorite grains do have sharp extinction. The aggregates range from 50-500 μm in their long dimension, and are notably larger than the other grains in the rock. The aggregates have an overall ellipsoidal shape with aspect ratios of 3:1 which can be determined on the basis of their shape in mutually perpendicular thin sections from the same sample. This shape of the aggregate is dependent on the intensity of cleavage development, and on the relationship of structural fabrics to the orientation of the aggregate. Weakly foliated rocks contain aggregates with less prominent shape-preferred orientation. In most cases, the chlorite and muscovite have grown parallel to their basal (001) cleavage plane. However, some aggregates contain kinked or broken chlorite or muscovite grains which are at an angle to the basal (001) cleavage planes of other phyllosilicates within the aggregate (Plate 28).

3.2 f) Pyrite

Pyrite porphyroblasts are found most often within pelitic and semi-pelitic phyllites and slates of the Kaza Group, Isaac and Yankee Belle Formations, and less commonly within carbonate units in the Cunningham and Yankee Belle Formation. Porphyroblasts are 0.5 - 10 mm in size, idioblastic to sub-idioblastic in shape, and are larger than the surrounding matrix grains (Plate 29). The morphology of the pyrite porphyroblasts depends on the position within the stratigraphic succession. Pyrite porphyroblasts within the Kaza Group and Isaac Formation are generally large and idioblastic (up to 2 cm). Pyrite porphyroblasts in the Yankee Belle Formation are mostly xenoblastic and rarely idioblastic and are generally smaller than those in the Kaza Group and Isaac Formation (Plate 30). At all levels, the pyrite porphyroblasts show weak to intense alteration to goethite.

3.3 Relative timing of mineral growth

Three main phases of deformation have affected the rocks within the Isaac Lake Synclinorium and are referred to as D_1, D_2 and D_3 events. Associated with each of these deformation events is folding and formation of an axial planar cleavage (S_1, S_2 and S_3 respectively). The D_1 event formed local folds with an associated axial planar cleavage (S_1). The D_2 event formed the main regional structures, which include overturned southwest-verging kilometre-scale folds (F_2), and an axial planar crenulation cleavage (S_2). The third foliation (S_3) found in the study area cross-cuts and crenulates all previous fabrics and has had only a minor effect on the morphology and orientation of the metamorphic mineral assemblages. It is a predominantly post-metamorphic deformation event and will not be discussed further. The relationship of the S_1, S_2, and S_3 cleavages to each other can be seen both in the field and with the petrographic microscope. The following section describes the relationship of the metamorphic minerals with the axial planar cleavages (S_1, S_2, and S_3) in order to constrain the timing of mineral growth relative to regional deformation events. The structural significance and morphology of the cleavages and deformation events is described and discussed in detail in the following chapter.

The earliest fabric in the rocks is associated with D_1 which produced an axial planar cleavage (S_1) that is oriented sub-parallel to original bedding surfaces (S_0). S_1 is defined by preferentially aligned muscovite and chlorite with their basal (001) cleavage plane parallel to S_1 (Plate 26). Where the S_2 cleavage has strongly overprinted the earlier S_1 cleavage, S_1 is preserved within the microlithons of S_2, where muscovite and chlorite found parallel to S_1 are folded or kinked by the later S_2 cleavage (Plate 31). In some strongly foliated phyllitic rocks of the Isaac Formation all evidence of S_1 is destroyed and muscovite and chlorite matrix minerals are preferentially aligned parallel to the S_2 foliation.

The relationship of the chlorite-muscovite aggregates to the S_1 and S_2 cleavages depends on the intensity of cleavage development, and the position of aggregates relative to the F_2 folds (i.e. upright limb versus the overturned limb). The following descriptions are taken from sections cut perpendicular to the S_2 cleavage and the L_2 intersection lineation. In rocks where the S_1 is the main cleavage, the majority of chlorite-muscovite aggregates have their long axis parallel to the (001) basal cleavage plane, and to the S_1 cleavage plane (Plate 26). This relationship is seen in the Yankee Belle Formation, which has largely been unaffected by S_2 cleavage development, and in phyllites and slates of the Isaac Formation where S_2 is weakly developed and widely spaced. In the Isaac Formation, the orientation and shape of the chlorite-muscovite aggregates with respect to the S_2 cleavage depends on the position relative to the F_2 folds and the intensity of the cleavage development. On the long, upright, east-dipping limbs, the aggregates are oriented with their basal (001) cleavage planes at a moderate to high angle to the S_2 cleavage plane (Plate 32). The long axis of the aggregate also changes from parallel to the predominant sheet silicate (001) orientation in the host phyllite, as seen in weakly deformed Yankee Belle phyllites to perpendicular to (001). In the hinge regions of the folds, chlorite-muscovite aggregates are oriented with their basal (001) cleavage planes at a high-angle (sub-perpendicular) to the S_2 cleavage plane (Plate 33). On the steep, overturned limb of the F_2 folds, chlorite-muscovite aggregates are oriented with their basal (001) cleavage plane parallel to sub-parallel to S_2, and their

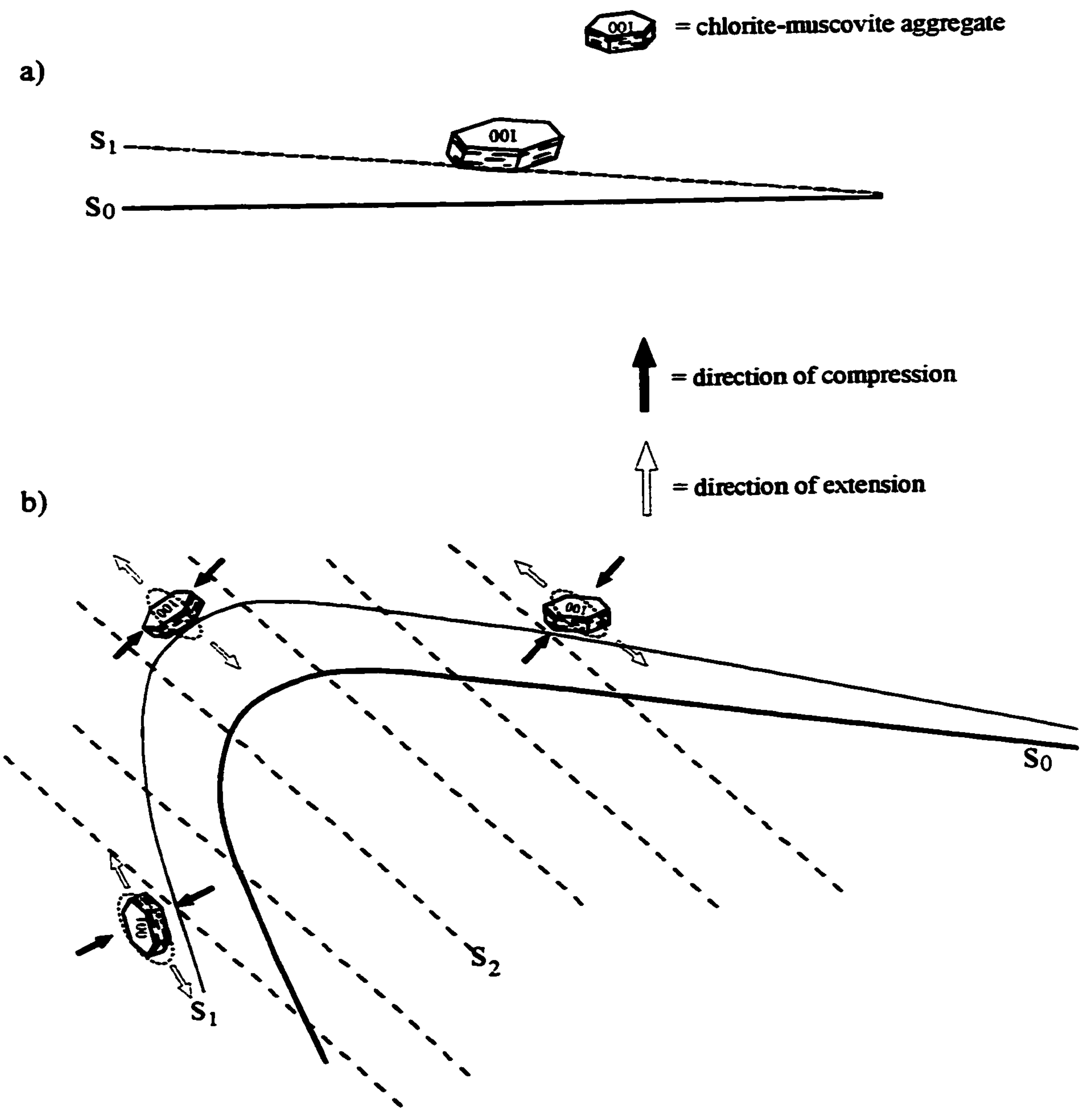


Figure 3-2. a) Diagram showing the possible relationships of the chlorite-muscovite aggregates to the S_1 cleavage planes. Aggregates are predominantly oriented parallel to the S_1 cleavage. b) Diagram showing the orientation of the chlorite-muscovite aggregates relative to the S_2 cleavage planes in the limb and hinge regions of an F_2 fold. Arrows show the compression and extension directions acting on the aggregates during the formation of the S_2 cleavage, and the dashed oval shows how the shape the aggregates is modified during D_2 deformation.

long axis parallel to S_2 also (Plate 34). The relationship of these aggregates to the F_2 fold structures are illustrated in Figure 3-2. It is possible that chlorite and muscovite grew during the development of the S_2 cleavage. This is based on the observation of thin chlorite layers that show sharp extinction within aggregates that contain chlorite and muscovite that are deformed.

The morphology of the pyrite porphyroblasts and intensity of pressure fringe development around them depends on the stratigraphic and structural position. In the Yankee Belle Formation, porphyroblasts are sub-idioblastic to xenoblastic and have only weakly developed pressure fringes around them. In the Isaac Formation, pyrite occurs as idioblastic porphyroblasts with a complex array of pressure fringes whose relationships to the S_1 and S_2 cleavages are difficult to ascertain, especially in strongly foliated rocks. The complexity of the pressure fringes depend on the position relative to the main regional F_2 folds. In all instances, the fringes are face-controlled, growing perpendicular to the pyrite crystal faces which further complicates interpreting their relationship to regional deformation. On the long, upright limbs of the F_2 folds, the pressure fringes are oriented parallel to S_1, and are also severely recrystallized with the quartz showing strong undulose extinction suggesting formation early in the structural and metamorphic history (Plate 35). Additionally, in some instances, the quartz fringes are completely recrystallized and contain annealed polygonal quartz grains. On the steep, overturned limbs of the F_2 folds, the fringes are considerably more complex. The fringes are oriented sub-parallel, oblique and perpendicular to the S_2 cleavage planes, show multiple phases of growth, and are moderately to weakly recrystallized (Plate 36).

The fringes oriented sub-parallel to parallel to the S_2 foliation are elongate with length to width ratios of 10:1 in sections cut perpendicular to the S_2 cleavage plane and L_2 intersection lineation. In sections cut perpendicular to the S_2 cleavage and parallel to the L_2 intersection lineation, the fringes have a length to width ratio of 4:1. Quartz pressure fringes which are oriented perpendicular to the S_2 cleavage plane have length to width ratios of only 2:1. Quartz grains which define the pressure fringes are weakly to moderately deformed showing bending, kinking and undulose extinction. The portion of the fringes adjacent to the matrix are the most strongly deformed or completely recrystallized. In some instances, the fringes have become detached from the pyrite grains and new fringes have begun to grow adjacent to the pyrite crystal faces (Plate 36). The fringes growing perpendicular to the S_2 cleavage are short and poorly developed. In some instances, the fringes are more developed and longer. Where this occurs, the fringes curve into the S_2 cleavage plane.

3.4 Discussion / Summary

3.4 a) Age of metamorphism and relation to regional deformation.

Previous studies show that there is an overall younging of the peak metamorphic events from northwest to southeast in the Cariboo Mountains (Pigage, 1977; Struik, 1980; Pell, 1984; Murphy, 1985; Currie, 1988; Walker, 1989; Ferguson, 1994; Reid, this study). At low metamorphic grade and shallow structural levels in the northwestern Cariboo Mountains, peak metamorphism is pre-D_2 deformation (Struik, 1980; Ferguson, 1994; Reid, this study). This implies that M_1 metamorphism took place pre-174 Ma based on the relationship of structures to the Hobson pluton south of the study area (Pigage, 1977; Gerasimoff, 1988). In the central Cariboo Mountains at intermediate structural levels, peak metamorphism occurred at the same time or later than D_2 deformation, also pre- 174 Ma based on the Hobson pluton age (Murphy, 1985). At high metamorphic grade and deep structural levels, peak metamorphism followed D_2 and is estimated at 135 +/- 4 Ma based on metamorphic monazite (Currie, 1988). At high metamorphic grades in the southern Cariboo Mountains, evidence for two metamorphic events is found (Pell, 1984; Currie, 1988). The earlier metamorphic event may be related to the early to mid-Jurassic (> 174 Ma) peak metamorphic event at shallow structural levels, but no geochronological age constraints exist. At this time, a thermal link between the shallow and deep structural levels cannot be made. The thermal history of the Cariboo Mountains differs with structural level and metamorphic grade, and is inherently complex.

A lack of geochronologic age constraints on metamorphism in the western Main Ranges of the Foreland Fold and Thrust Belt makes the correlation of the thermal and structural events in the Omineca Crystalline Belt difficult. The ages of metamorphism and deformation in the Main Ranges west of the Cariboo Mountains (Cushing Creek area, Figure 3-1) is inferred to have occurred between the late Jurassic or early Cretaceous based on the relationship of structures (i.e. Snake Indian Thrust) which cut mid-Jurassic stratigraphy, and the age of formation of the Porcupine Creek Anticlinorium to the southwest of the Snake Indian Thrust, which was formed between the late Jurassic to early Cretaceous (Carey, 1984; Price and Mountjoy, 1970; Mountjoy, 1978; Simony et al., 1980; Kubli 1990; Lickorish, 1992). This indicates that peak metamorphism occurred in late Jurassic-early Cretaceous based on the interpretation of the relationships of metamorphic minerals to main regional structures of the western Main Ranges in this area (Carey, 1984). Geochronological age constraints on peak metamorphic biotite from the Tete Jaune Cache area in the western Main Ranges yields a mid-Cretaceous age of mineral growth (100-78 Ma), (Van Den Driessche and Maluski, 1986). Similarly, rocks of the Selwyn range in the western Main Ranges are interpreted to have reached peak metamorphism in the mid-Cretaceous based on the age constraint of Van Den Driessche and Maluski (1986).

3.4 b) Metamorphism in the Isaac Lake Synclinorium

Petrographic examination of the metamorphic minerals and their relationship to structural fabrics allows for a number of conclusions to be drawn about metamorphic conditions and timing of events in the Isaac Lake Synclinorium. The common metamorphic mineral assemblage in pelitic rocks is quartz + muscovite + chlorite + albite + pyrite, indicating peak metamorphic conditions in the study area were in the chlorite zone of the greenschist facies. From Figure 3-3, this indicates temperatures between

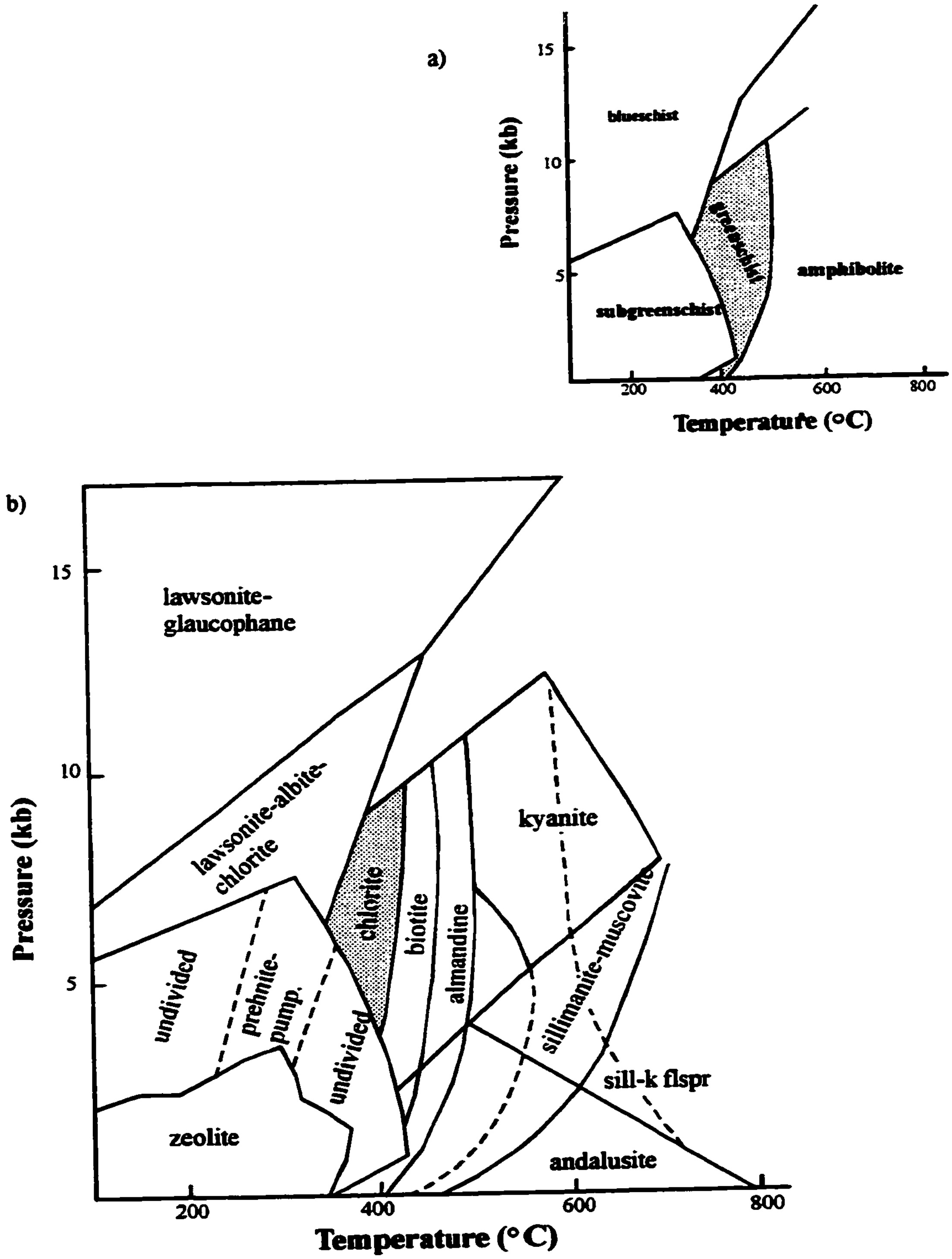


Figure 3-3. a) Pressure-temperature diagram showing the conditions under which greenschist metamorphic facies form. b) Pressure-temperature diagram showing the mineral assemblage zones of the different metamorphic facies. The rocks of the study area were metamorphosed in the chorite zone which is shown in gray. (modified from Greenwood et al., 1992)

350-450°C and pressure between 4-9 kbar (Greenwood et al., 1991). The presence of chlorite and muscovite preferentially aligned with S_1 indicates mineral growth occurred during D_1, suggesting that peak metamorphism may have occurred during this deformation episode. Whether the growth of phyllosilicate minerals continued during the formation of S_2 is not as clear-cut. In samples where phyllosilicate minerals are parallel to S_2, the grains show undulose extinction and have a sub-idioblastic to xenoblastic shape indicating that they were significantly deformed. Whether or not metamorphic mineral growth in the matrix continued into D_2 deformation remains uncertain.

The chlorite-muscovite aggregates are interpreted to have been present before the formation of S_2. This is based on the occurrence of porphyroblasts oriented with their long axis and (001) cleavage planes parallel or sub-parallel to S_1. This relationship of the chlorite-muscovite aggregates is preserved in phyllites of the Yankee Belle Formation, which have been largely unaffected by S_2 cleavage development. In phyllites of the Isaac Formation, which have been strongly affected by the development of S_2 cleavage, the shape of the aggregates has been significantly modified. The degree of modification depends on the position relative to the main regional F_2 folds, but all show a change from the long axis of the aggregate being parallel to the basal (001) cleavage plane to perpendicular to the basal (001) cleavage plane. Modified aggregates have their long axis parallel to the S_2 cleavage direction, and the angle between the (001) cleavage plane and the S_2 cleavage plane is moderate to high, depending on the position relative to the F_2 folds (see Figure 3-2). The modification of the chlorite-muscovite aggregates involves shortening perpendicular to the main regional cleavage direction (S_2), and extension parallel to the S_2 cleavage direction. Extension of the aggregates can be seen in thin sections cut both perpendicular to the L_2 intersection lineation and parallel to the L_2 intersection lineation indicating an overall flattening of the aggregates in the S_2 cleavage plane. Overall, the aggregates appear to maintain their sub-parallel to parallel relationship to the S_1 cleavage direction. This interpretation does not involve rotation of the aggregates with respect to the S_2 cleavage planes, which has been the suggested mechanism for the formation of these aggregates (Clark and Fischer, 1994).

The paragenetic sequence of metamorphic mineral growth in the Isaac Lake Synclinorium in relationship to the main deformation events is show in Figure 3-4.

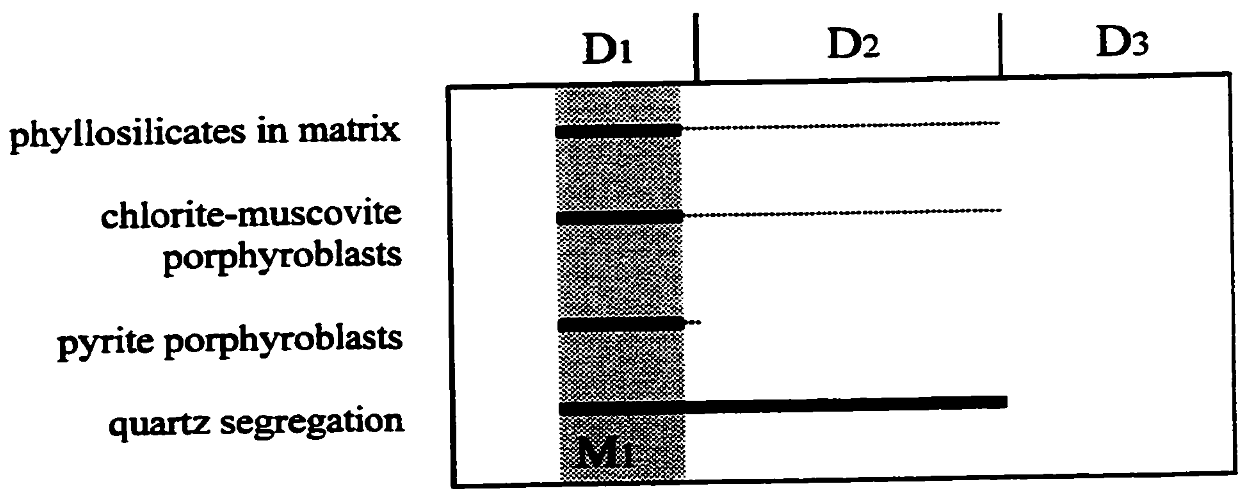


Figure 3-4. Diagram showing the paragenetic sequence of metamorphic minerals in relationship to the major deformation events in the Isaac Lake Synclinorium. Solid black line represents where there is clear evidence of mineral growth, dotted line represents possible additional growth/recrystallization during later deformation.

Similar chlorite-mica aggregates have been recognized and studied in low-grade slates belts around the world (Clark and Fisher, 1994; Li et al., 1994; Milodowski and Zalasiewicz, 1990; Dimberline, 1986; Gregg, 1986; van der Pluijm and Kaars-Sijesteijn, 1984). There is disagreement on whether the interlayering of the phyllosilicates is a primary feature (sedimentary) or secondary feature (metamorphic) (Beutner, 1978; Voll, 1960; van der Pluijm and Kaars-Sijpesteijn, 1984; Weber, 1981). The intergrowth of the two phyllosilicate phases in the aggregates has been interpreted as a result of weathering and transport processes (Beutner, 1978), metamorphic processes (Weber, 1981), or a mix of diagenetic and metamorphic processes during the formation of the aggregates (Voll, 1960; van der Pluijm and Kaars-Sijpesteijn, 1984). Although these authors disagree on the origin of the aggregates, most agree that the shape of the aggregates are modified by later deformation (Brooks and Fisher, 1994; Li et al., 1994; van der Pluijm and Kaars-Sijpesteijn, 1984).

The origin of the chlorite-muscovite aggregates in deformed phyllites and slates from the Isaac Lake Synclinorium is uncertain. Competent sandstone beds from the Kaza Group and Yankee Belle Formation contain a high percentage of detrital muscovite, some with thin chlorite intergrowths. Because the sandstones are competent relative to the interbedded pelitic rocks, the beds have been relatively unaffected by cleavage formation and deformation. This suggests that muscovite may have been a significant detrital constituent for lithologies throughout the Windermere Supergroup stratigraphy, and may have been the nucleus for the aggregates. However, the fact that the aggregates are parallel to the S_1 cleavage direction may indicate that there were either original detrital micas parallel to S_0 that rotated into the S_1 cleavage planes or that the aggregates are composed of metamorphic phyllosilicates that grew early in the formation of the S_1 cleavage.

There is a widespread occurrence of the chlorite-muscovite aggregates throughout the low-grade Windermere Supergroup strata in the southern Canadian Cordillera (Ferguson, 1994; Carey, 1984). Ferguson (1994) identified chlorite-muscovite aggregates which were also interpreted to have been present before the formation of the S_2 cleavage and used them to correlate metamorphic events from his study area to that of Carey (1984) in the western Main Ranges, east of the Cariboo Mountains. Such a correlation has significant implications as this would imply a Cretaceous age for peak metamorphism in the northern Cariboo Mountains based on Carey's interpretation of the timing of chlorite-muscovite aggregate growth to deformation in her study area. Carey (1984) interpreted the aggregates to have grown during the formation of the regional late Jurassic to early Cretaceous structures of the western Main Ranges of the Rocky Mountains. She similarly described and illustrated the aggregates as being oriented with their long axis parallel to the S_1 foliation plane and the basal (001) cleavage plane of the chlorite perpendicular to the foliation plane. This is a seemingly odd orientation for phyllosilicate minerals interpreted to have grown during deformation, as phyllosilicates are expected to grow with their (001) basal cleavage planes preferentially aligned parallel to the foliation (e.g., Spry, 1969). Based on the description of aggregates in Carey's area, an alternate interpretation would be that the aggregates were in fact present before the main regional deformation and subsequently modified to their present configuration. The orientation of these aggregates may preserve the orientation of an earlier cleavage not recognized in the western Main Ranges east of the Cariboo Mountains. If the chlorite-muscovite aggregates in the western Main Ranges grew before the Cretaceous deformation, they in fact may be correlative with the chlorite-muscovite aggregates in the Cariboo Mountains. Although these relationships are of regional significance, the regional interpretations are based on the interpretation of local fabrics which are subject to debate. The lack of geochronological age constraints of deformation and metamorphism makes correlating events between the western Main Ranges and the northern Cariboo Mountains tenuous.

Plate 17. Elongate quartz grains that show undulose extinction from the lower Isaac Formation. Grains are elongate parallel to the intersection lineation of the main cleavage (S_2) with bedding surfaces (S_0). (XPL)

Plate 18. Quartz grains with sutured and partially recrystallized grain boundaries from a fine sandstone of the Yankee Belle Formation. Quartz shows weak to moderate undulose extinction. Detrital muscovite is common throughout and is bent and broken around quartz grains. (XPL)

Plate 19. Elongate quartz pressure fringes around a pyrite porphyroblast from the lower Isaac Formation. Quartz shows moderate to strong undulose extinction and is recrystallized where adjacent to the matrix material. (XPL)

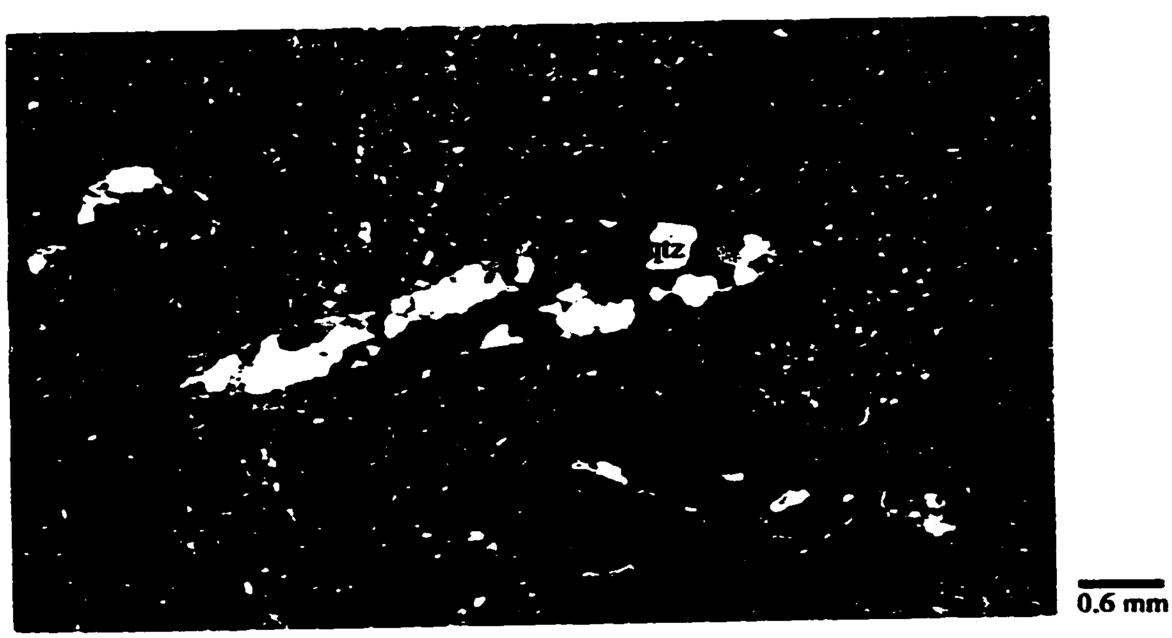


Plate 20. Recrystallized quartz pressure fringes surrounding a pyrite porphyroblast in the lower Isaac Formation. The contact between the quartz grains are sharp and polygonal. (XPL)

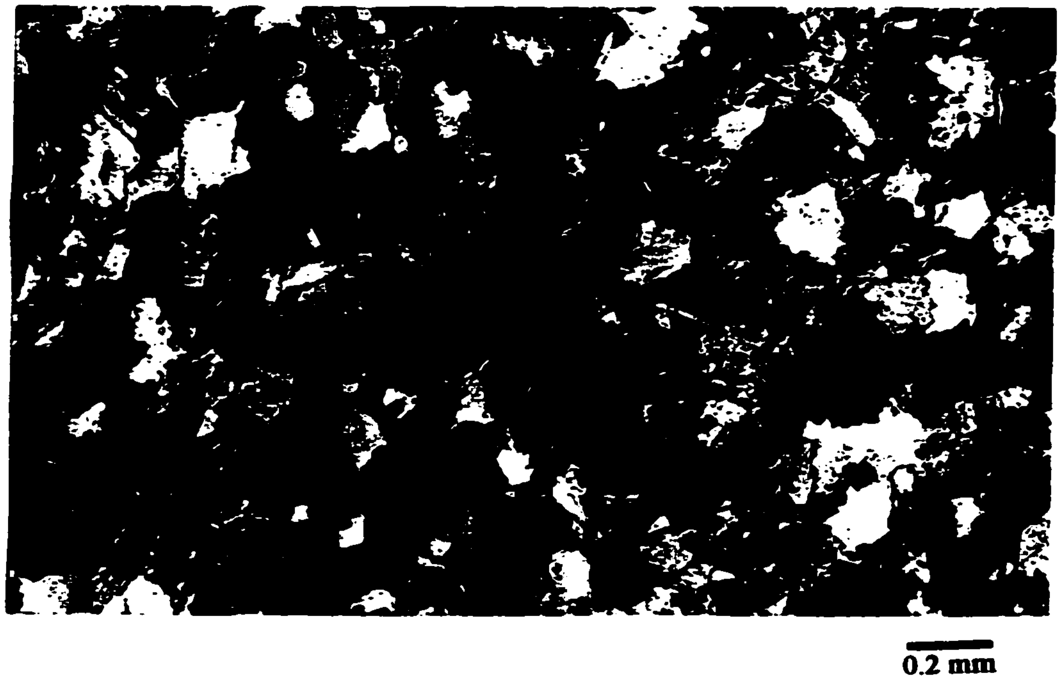
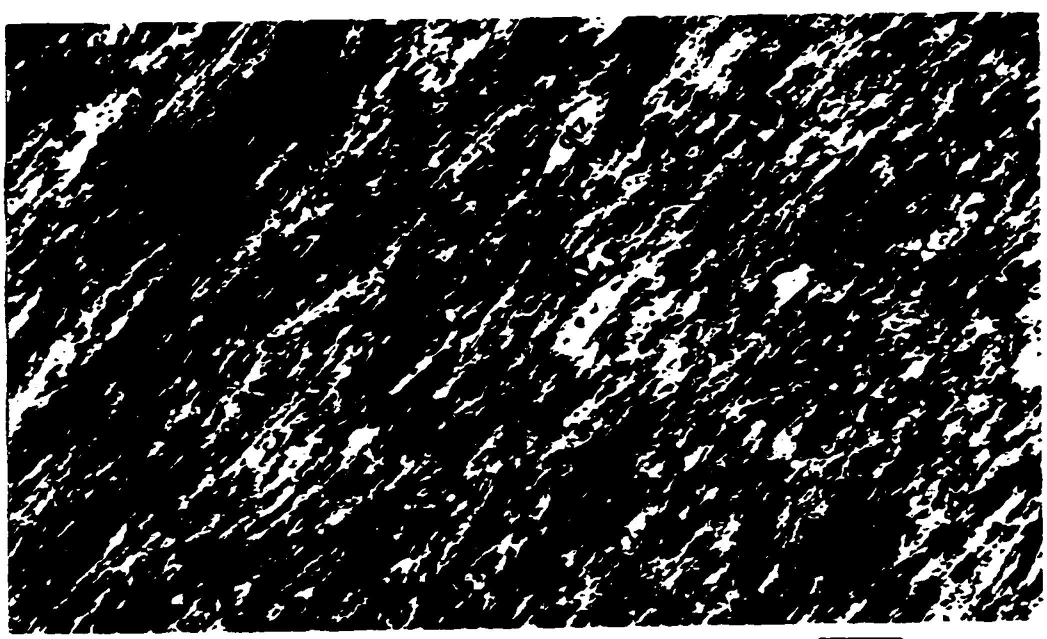


Plate 21. Weakly developed quartz pressure fringe around pyrite porphyroblast from a recrystallized sandstone in the Yankee Belle Formation. (XPL)

Plate 22. Albite grains from a coarse-grained sandstone bed in the upper Kaza Group. Albite twins within the grains show weak to moderate undulose extinction and recrystallization. (XPL)

0.6 mm

Plate 23. Chessboard albite within a coarse-grained sandstone of the upper Kaza Group. Grains contain inclusions of quartz and pyrite, and the edges are recrystallized. (XPL)



0.2 mm

Plate 24. Preferentially aligned muscovite in the Isaac Formation with muscovite oriented parallel to the regional cleavage direction (S_2). Section is cut perpendicular to the S_2 cleavage direction and parallel to the intersection lineation of S_2 with S_0 (L_2). (XPL)

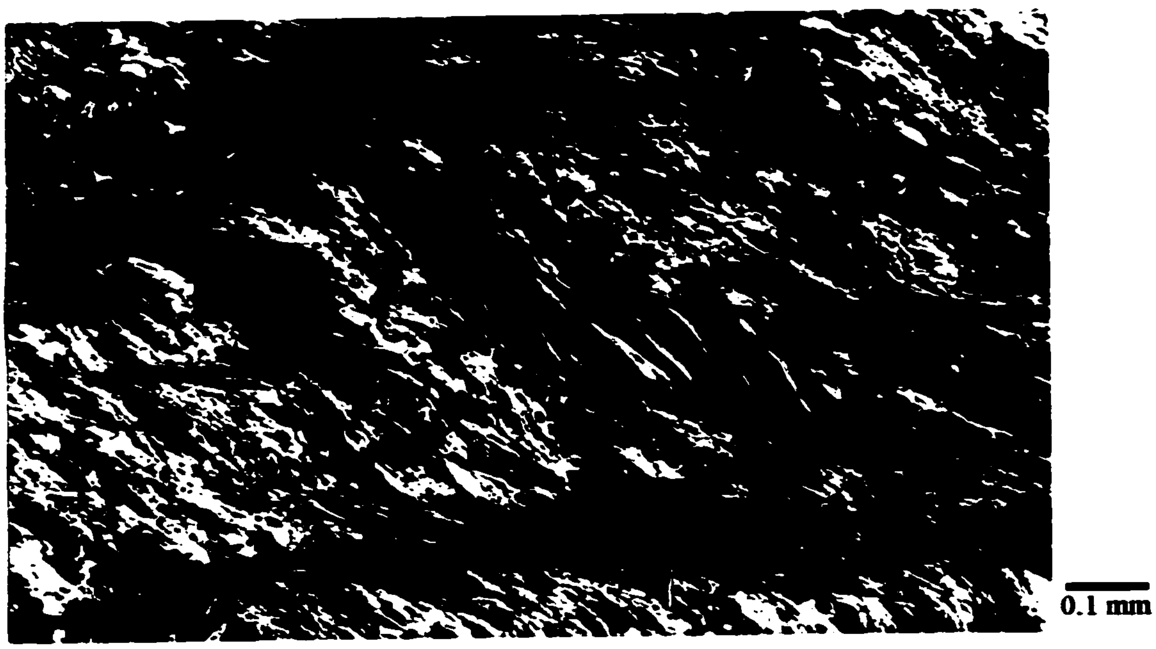


Plate 25. Folded and deformed muscovite from the Isaac Formation with muscovite oriented parallel to an earlier cleavage that is folded by the main regional cleavage (S_2). Section is cut perpendicular to the S_2 cleavage direction and perpendicular to the intersection lineation of S_2 with S_0 (bedding), (L_2). (XPL)

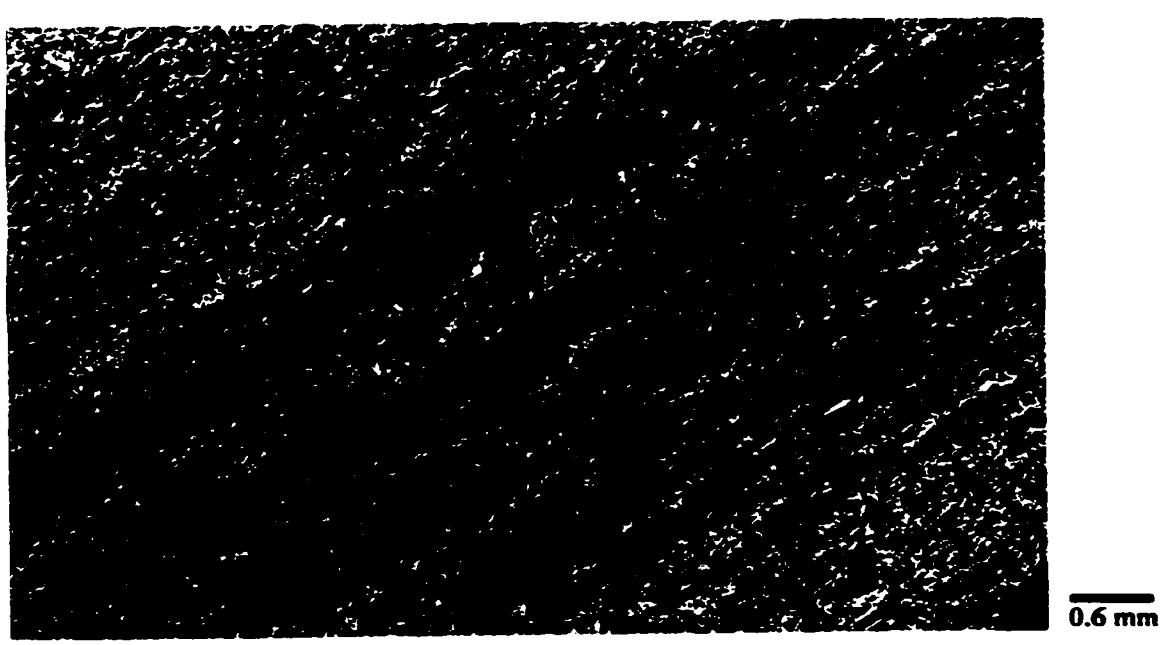
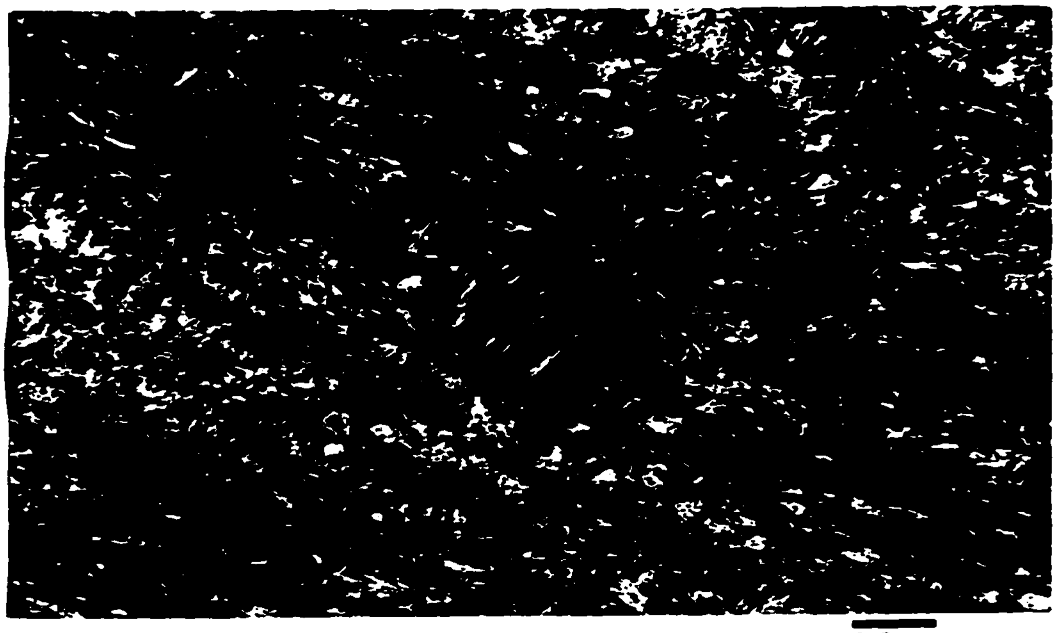


Plate 26. Fine-grained chlorite and muscovite appears to be preferentially aligned to the earliest cleavage (S_1). Individual chlorite grains range from 25-50 µm, and can only be resolved at higher magnification powers. (XPL)



0.6 mm

Plate 27. Chlorite-muscovite aggregate found in the lower Isaac Formation. The chlorite and muscovite layers are aligned parallel to sub-parallel to each other and at a high angle to the main regional cleavage (S_2). Thin quartz intergrowths are found within the aggregate as well. (XPL)

0.2 mm

Plate 28. Chlorite-muscovite aggregate with bent and broken layers, and layers oriented at a high-angle to each other. The overall orientation of the aggregates are at a high-angle to the main regional cleavage (S_2). (XPL)

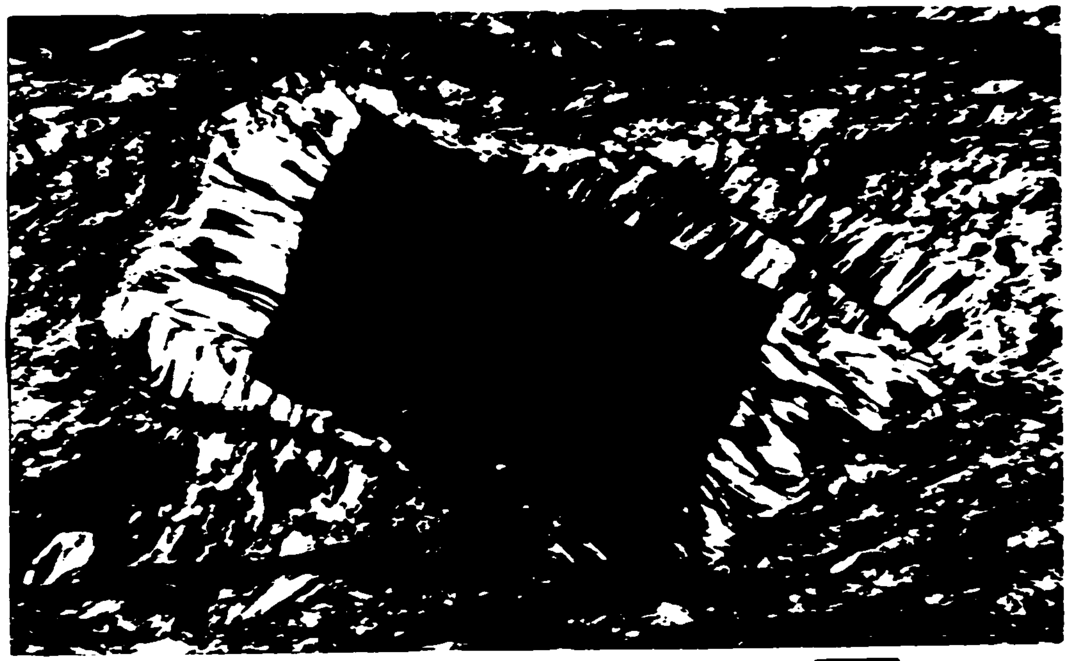


Plate 29. Idioblastic pyrite porphyroblast from the upper Isaac Formation, with quartz pressure fringes formed around the different faces shown in the thin section. (XPL)

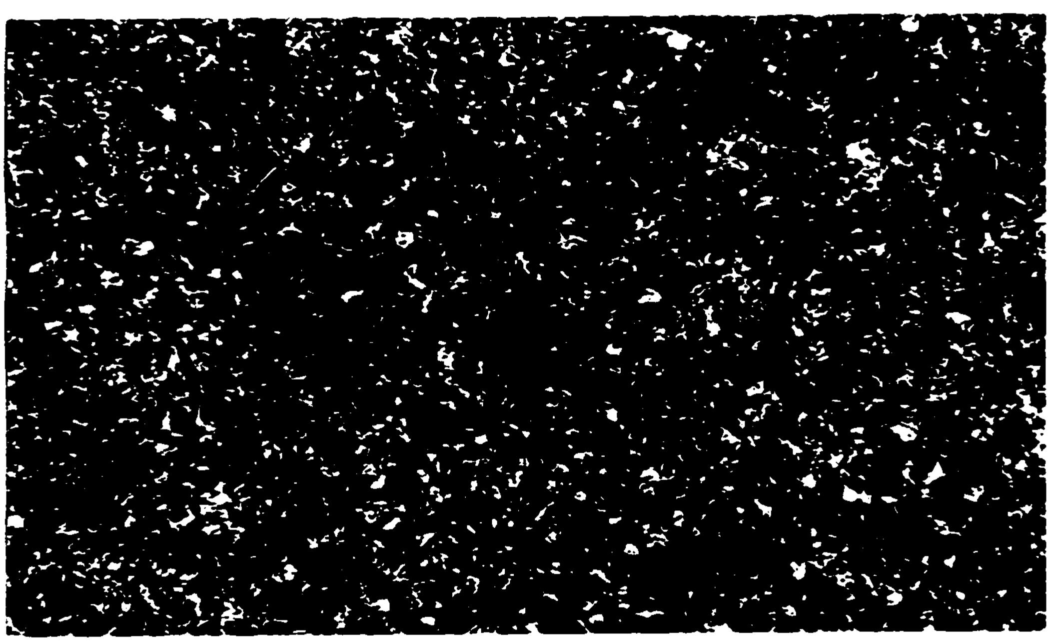
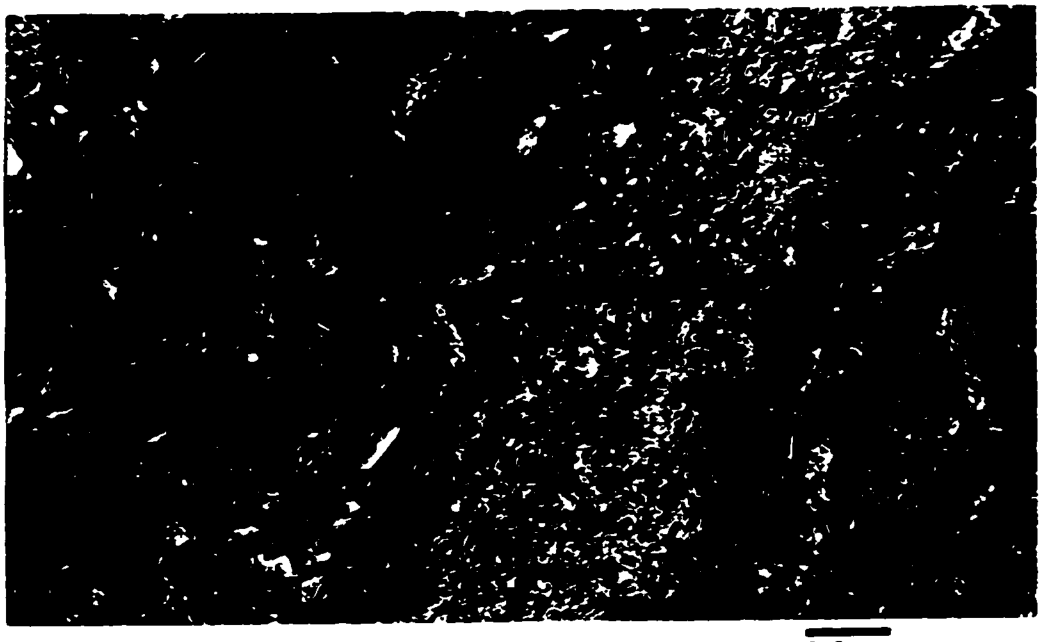


Plate 30. Sub-idioblastic and xenoblastic pyrite porphyroblasts from the Yankee Belle Formation. (XPL)

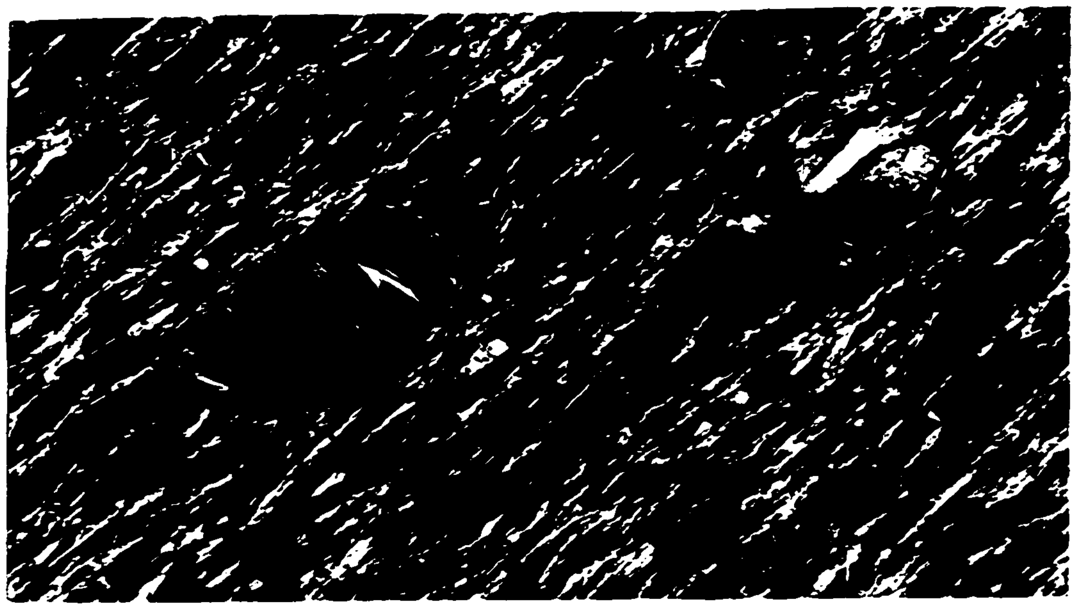


0.6 mm

Plate 31. Muscovite oriented parallel to S_1 which is subsequently folded by the main regional cleavage (S_2). (XPL)

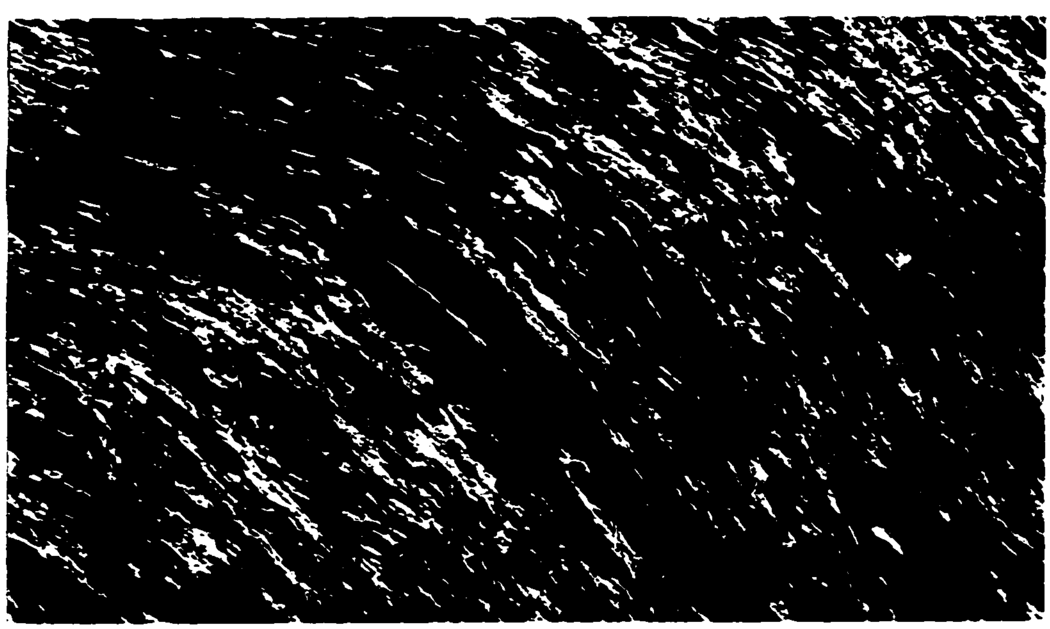
0.2 mm

Plate 32. Chlorite-muscovite aggregate from the flat, upright limb of an F_2 fold. Layers within the aggregate are oriented at a moderate to high-angle to the S_2 cleavage. Section is cut perpendicular to S_2 and the intersection lineation (L_2). (PPL)



75 μm

Plate 33. Chlorite-muscovite aggregate in the hinge region of a regional F_2 fold. Layers within the aggregate are oriented at a high-angle to the S_2 cleavage. Section is cut perpendicular to S_2 and the intersection lineation (L_2). (XPL)



0.1 mm

Plate 34. Chlorite-muscovite aggregate on the steep overturned limb of a regional F_2 fold. Layers within the aggregate are oriented at a moderate to shallow angle to the S_2 cleavage. Section is cut perpendicular to S_2 and the intersection lineation (L_2). (XPL)

Plate 35. Quartz pressure fringe around pyrite on the flat, upright limb of an F_2 fold. Quartz shows undulose extinction and is recrystallized along the boundary with the matrix. (XPL)

Plate 36. Quartz pressure fringes around pyrite on the steep, overturned limb of an F_2 fold. The quartz adjacent to the pyrite crystal faces is fibrous and shows only weak undulose extinction. The quartz fringes furthest from the pyrite grain shows strong undulose extinction and is recrystallized where adjacent to the matrix. (XPL)

CHAPTER 4: STRUCTURAL EVOLUTION OF THE ISAAC LAKE SYNCLINORIUM

4.1 Introduction

The Cariboo Mountains are part of the Omineca Crystalline Belt of the Western Canadian Cordillera which were deformed primarily during the mid-Jurassic to Cretaceous obduction and accretion of allochthonous terranes from the west (Monger, 1982). The main regional structures of the Cariboo Mountains can be divided into a series of kilometre-scale northwest trending anticlinorial and synclinorial structures (Campbell, 1970; Campbell et al., 1973). The anticlinoria and synclinoria have a regional northwest plunge, and as a result, deeper structural levels are exposed in the southeast and shallow levels are preserved in the northwest. This transition from shallow to deep structural levels in the Cariboo Mountains was first recognized by Campbell (1970), who called it the suprastructure-infrastructure transition, and described it as a gradational change in fold style and an increase in metamorphic grade with increasing structural depth (Campbell, 1970; Campbell, 1973; Murphy, 1985, 1987).

Three main phases of folding are recognized in the Cariboo Mountains. In rocks of the suprastructure, phase one folds (F_1) are rare, and recognized as recumbent, isoclinal east-verging folds with an associated axial planar cleavage (Ferguson, 1994; Campbell et al., 1973). In rocks of the infastructure in the southern Cariboo Mountains, F_1 folds are west-verging recumbent isoclinal folds (Pell, 1984; Deschene et al., 1984; Currie and Simony, 1986; Pell and Simony, 1987; Currie, 1988; Walker, 1989). The east-verging F_1 folds of the suprastructure are interpreted to be older than the west-verging F_1 folds of the infastructure based on the occurrence of the two phases of folding in the suprastructure-infrastructure transition zone (Murphy, 1985, 1987).

The second phase folds (F_2), which form the regional anticlinorial and synclinorial structures of the Cariboo Mountains, refold the east- and west-verging F_1 folds of the suprastructure and infrastructure in a series of kilometre-scale northwest trending folds, (Figure 4-1). These F_2 folds change from northeast-verging to southwest-verging from east to west, outlining several large-scale fan structures (Campbell, 1970; Murphy, 1985; Pell, 1984; Ferguson and Simony, 1991; Ferguson, 1994). The geometry of the F_2 folds also changes from the suprastructure to infastructure. In the suprastructure, F_2 folds are open to closed, upright to overturned folds with moderate to steeply dipping axial planes. In the infrastructure, F_2 folds are tight, overturned, near-recumbent isoclines with gently dipping axial planes (Campbell, 1970; Murphy, 1985, 1987). The transition of F_2 fold styles from the suprastructure to infastructure is gradational and interpreted to record increasing strain as a function of depth (Pell, 1984; Murphy, 1985, 1987).

Phase three (F_3) folds in the Cariboo Mountains are represented as open, upright east-southeast to west-northwest trending folds which refold all previous structures. The development of F_3 folds is pervasive in the suprastructure, but is more locally developed in the infrastructure. F_3 fold hinges are nearly co-axial with F_2 folds hinges in the southeast,

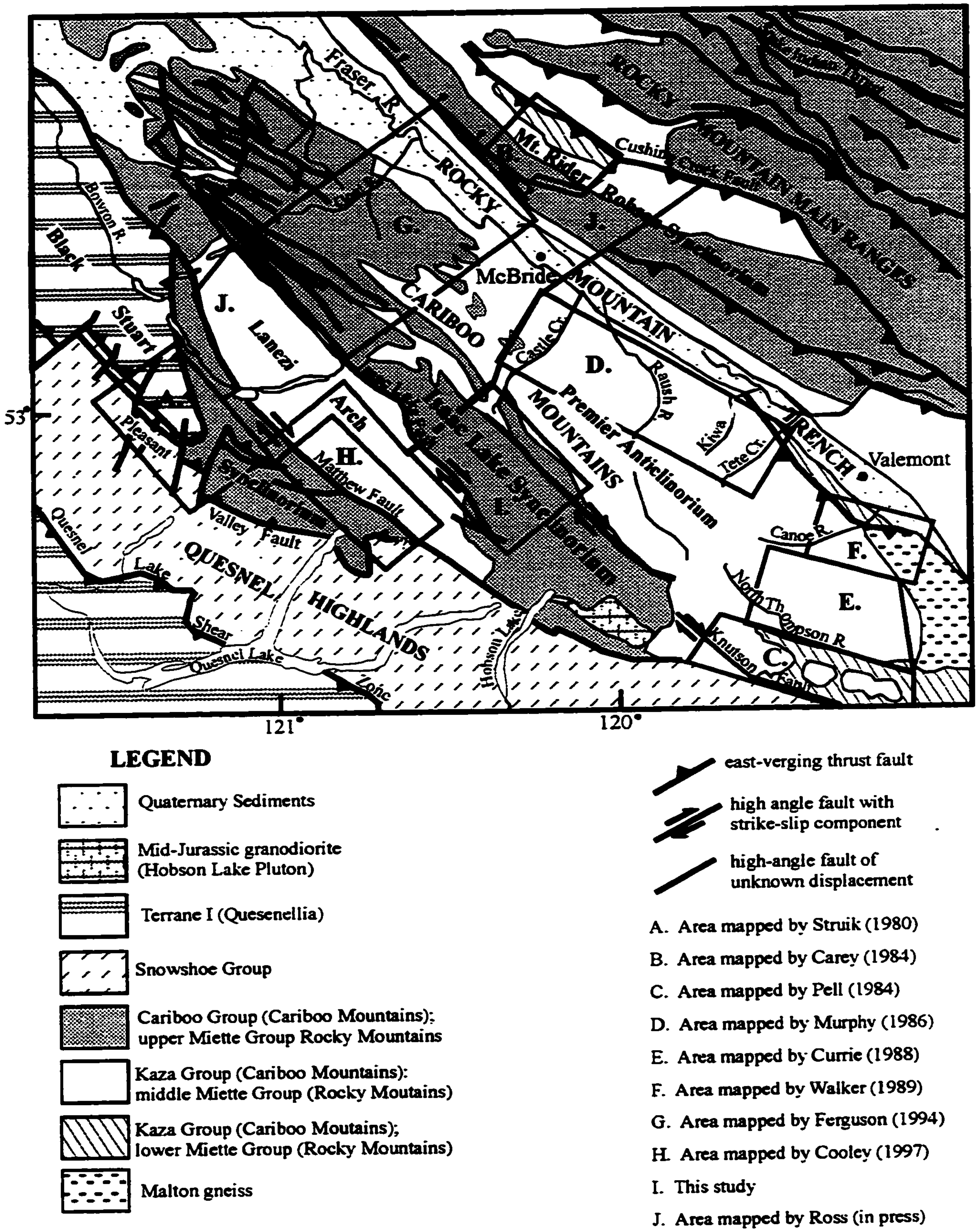


Figure 4-1. General geology map of the Cariboo Mountains (modified from Murphy, 1985)

near the Malton Gneiss, making the differentiation between F_2 and F_3 folds difficult (Pell, 1984; Murphy, 1985; Currie, 1988; Walker, 1989; Ferguson, 1994).

4.2 Structural geology of the Isaac Lake Synclinorium

The Isaac Lake Synclinorium is a regional F_2 structural depression somewhat modified in its overall geometry by fault structures along its flanks (Figure 1-6). It is fault bounded on the east side by the Winder Fault, a dextral-oblique thrust fault, and on the west by the Isaac Lake Fault zone, a dextral-oblique strike-slip fault with a minimum dextral strike-slip offset of approximately five kilometres (Figure 1-5). The Isaac Lake Synclinorium in the central Cariboo Mountains is part of the suprastructure, and provides the opportunity to study the structural evolution of greenschist facies Windermere Supergroup sedimentary rocks. The structures are dominated by open, upright to overturned F_2 folds, re-folded by open, upright F_3 folds and high-angle faults (Figures 1-5 and 1-6). Three main deformation events are recognized within the study area, and their relationships to each other are well constrained based on their field and microstructural relationships. A fourth phase of folding is also recognized mostly as a late crenulation cleavage and locally as local metre to decametre-scale folds, which post-date all earlier fabrics. Faulting is also pervasive throughout the area and four distinct types of faults are recognized. These include local east-verging thrust faults, high-angle reverse and normal faults and dextral oblique-strike-slip faults.

The development and intensity of deformational fabrics in the Isaac Lake Synclinorium has been strongly affected by the varying rheologic properties of the sedimentary succession. The Cunningham Formation, a thick-bedded limestone unit, and the Yanks Peak Formation quartzites form competent layers within the more ductile, pelite-rich Isaac and Yankee Belle Formations. This variation between relatively incompetent units to competent units in the stratigraphy creates a structural multilayer which accommodates the strain differently leading to strain partitioning. F_2 folds that are tight to overturned in incompetent Isaac Formation pelites are accommodated by broad flexures in the overlying Cunningham and Yankee Belle Formations.

The purpose of this chapter is to document and describe the geometry of these folds and faults that are exposed in the study area and to document their relationships to each other. As well, the effects of strain partitioning by the structural multilayer will be discussed. A synthesis of the major deformational events and the relation of these deformational events to other study areas will be included as an overall summary. The geometry and significance of the Isaac Lake Fault will be discussed in the following chapter.

4.3 Folds

4.3 a) <u>Phase one folds (F_1)</u>

Phase one folds (F_1) are the earliest folds found within the Isaac Lake Synclinorium. F_1 folds are recumbent folds with tight to isoclinal interlimb angles and sub-horizontal fold axes. They show hinge thickening and limb thinning and can be classified as class two similar folds (Ramsay, 1967). Mesoscopic (m-scale) evidence for F_1 folds is rare, but centimetre to metre-scale F_1 folds with an axial planar cleavage (S_1) sub-parallel to bedding along the limbs do occur locally within the Isaac Formation (Plate 37). Associated with regions of well developed F_1 folds are bedding-parallel east-verging thrust faults. These faults can have displacements from a few metres to tens of metres and are found locally in the northeast quadrant of the study area.

The S_1 cleavage is a continuous penetrative foliation defined by the parallel alignment of fine-grained muscovite and chlorite which has a maximum angle of 5° to bedding (S_0). The S_1 cleavage is seen rarely in handsample except in pelitic rocks of the Isaac and Yankee Belle Formations that have been only weakly affected by S_2 and S_3 cleavages, which tend to obscure or destroy S_1. The sub-parallel relationship between S_0 and S_1 is most evident petrographically and shows a consistent shallow westward dip of S_1 when S_0 is brought to the horizontal. S_1 is well preserved within the microlithons of S_2 crenulation cleavages in the pelitic rocks of the Kaza Group and Isaac Formation, and represents the main cleavage in pelitic rocks of the Yankee Belle Formation, which has only been locally affected by subsequent cleavages (Plate 38). Evidence for F_1 folding or S_1 cleavage development in the Cunningham Formation is inconclusive. In pelitic and semi-pelitic interbeds of the Cunningham Formation, a single weak foliation, which has not been crenulated, formed at 30-40° with bedding.

4.3 b) Phase two folds (F_2)

The second phase (F_2) folds are the main northwest-trending anticlinorial and synclinorial structures (Figure 4-1). The F_2 folds are southwest-verging asymmetric overturned similar folds with an average wavelength of two kilometres. These folds are characterized by long, upright, east dipping limbs and short, steep to overturned western limbs. They have tight to open inter-limb angles ranging from 60° to 90°. The axial planes of the F_2 folds dip to the northeast at 35 to 50° with a strike of 300 to 330° (Figure 1-6). Metre-scale folds plunge shallowly to the north-northwest at approximately 10° to 25°. Folds of this style are recognized on all scales, from microscopic (mm-scale) to mesoscopic (cm and m-scale) and macroscopic (km-scale) (Plates 39, 40, and 41). The axial planar crenulation cleavage (S_2) of the F_2 folds is seen in the Kaza Group and Isaac Formation, and locally within the Yankee Belle Formation. Lineations associated with F_2 folds (L_2) consist of the intersection lineation between the S_2 and S_0 surfaces and the hinge lines of F_2 crenulations and metre-scale F_2 folds in outcrop.

The S_2 cleavage morphology depends on the lithology of the rock. In pelitic rocks of the Kaza Group and Isaac Formation, S_2 is represented as a continuous crenulation cleavage defined by insoluble opaque material, and less commonly by the parallel alignment of muscovite and chlorite (Plate 42). In carbonate-rich intervals of the Isaac and Yankee Belle Formations S_2 cleavage is defined as spaced anastamosing solution cleavage (Plate 43). Interbedded carbonate and/or sandstone and shale commonly shows cleavage refraction, where cleavage steepens in the carbonate or sandstone layers. The

Cunningham Formation exhibits very little outcrop-scale evidence of F_2 deformation, with the exception of a weakly developed, widely spaced fracture cleavage in the lower portion of the formation. In the pelitic interlayers of the Cunningham Formation, a single, low-angle cleavage is present although no associated folds are observed. Local, bedding-parallel shear zones are also observed at the base of the Cunningham Formation.

Equal-area plots of S_2 cleavage measurements from within the Isaac Lake Synclinorium show that the poles to the S_2 cleavage planes are distributed in a girdle to which a great circle can be fitted, indicating fanning and cleavage refraction as well as re-folding by roughly cylindrical F_3 folds. Equal-area plots of L_2 intersection lineations show an overall southeast orientation with a mean trend of 112° and plunge of 3°, which contrasts with the regional northwest plunge (Figure 4-2).

4.3 c) Phase three folds (F_3)

The third phase of folding (F_3) refolds all pre-existing folds within the Isaac Lake Synclinorium. First-order F_3 folds are open and upright with a maximum wavelength of 10 km. These folds are class 2b parallel folds with open to gentle interlimb angles, upright axial surfaces and sub-horizontal fold axes (Ramsay, 1967). The strike of axial surfaces range from 090° to 110° and fold hinges trend towards 110° and 280° with sub-horizontal plunges (<10°). An axial planar crenulation cleavage (S_3) is well developed in their hinge zones and crenulates the earlier S_2 cleavage. The F_3 folds differ from the F_2 folds in

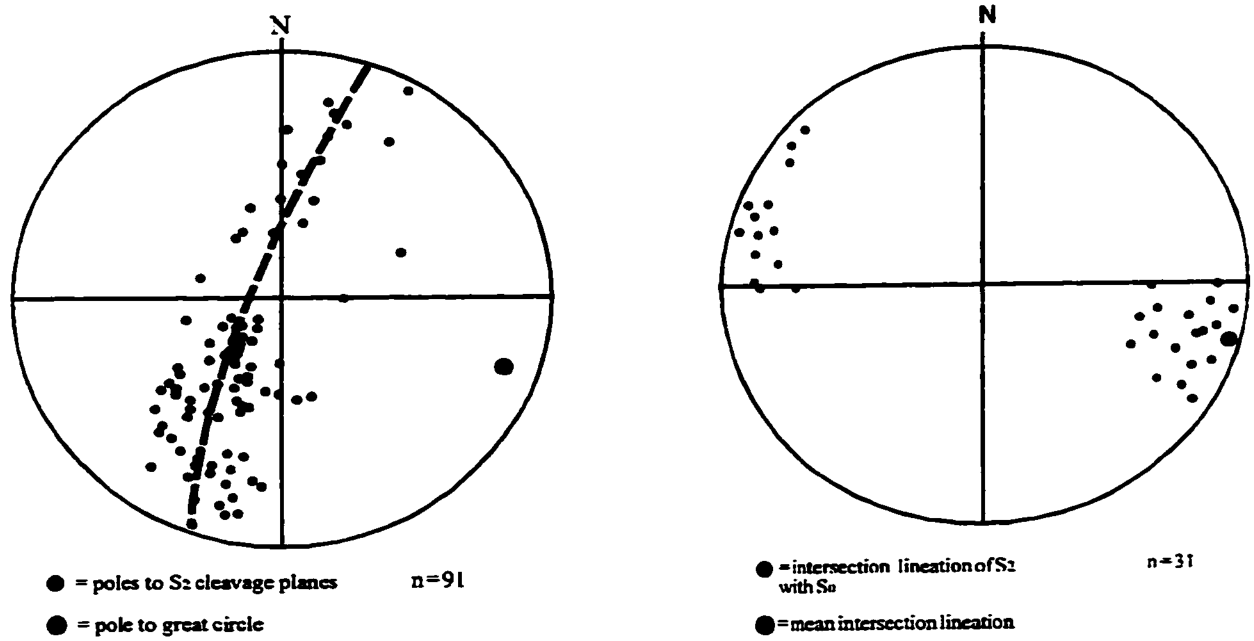


Figure 4-2. Equal-area net diagram of the poles to S_2 cleavage planes showing a distribution in a girdle as a result of F_2 fanning and F_3 refolding. Circled x represents the pole to the best fit great circle for the data. b) Equal-area net diagram of L_2 intersection lineations of S_2 and S_0. Circled x is mean intersection lineation.

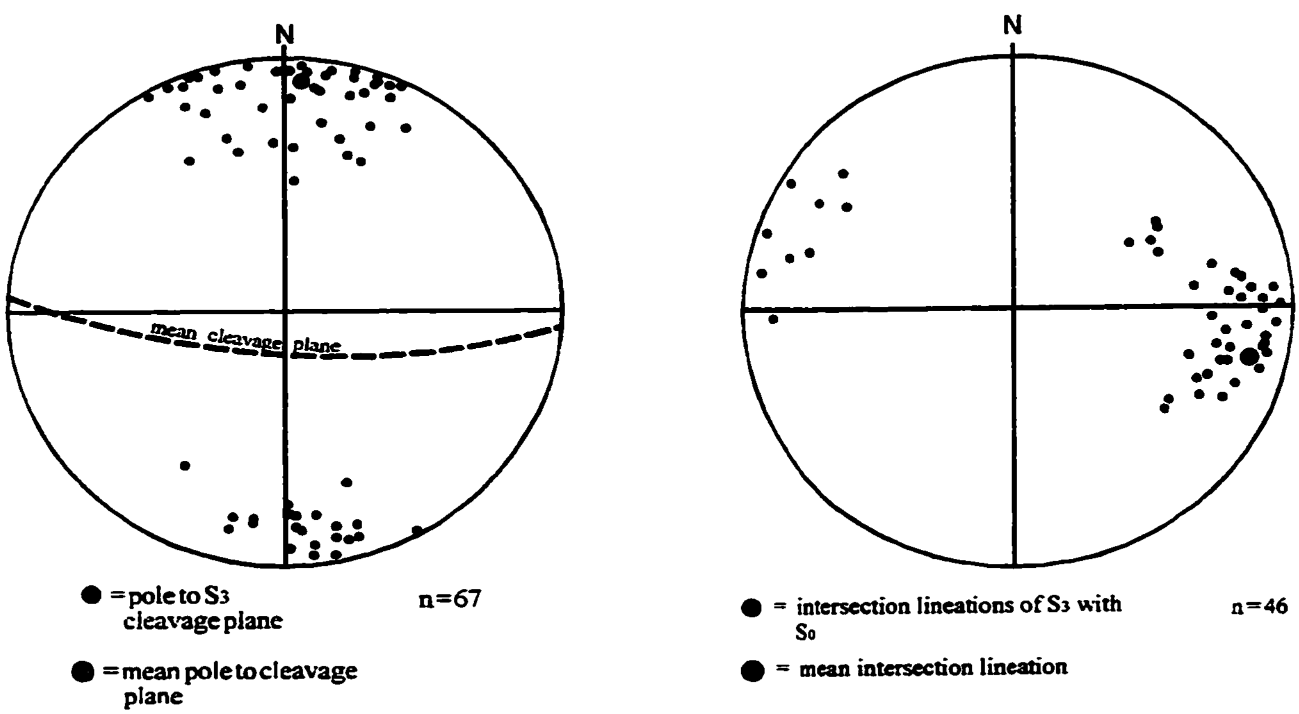


Figure 4-3. a) Equal-area net diagramof the poles to S_3 cleavage planes which give a mean pole to cleavage and shows the mean cleavage plane (dashed great circle). b) Equal-area net diagram of S_3 lineations, circled x represents the mean intersection lineation.

geometry and style of folding. Metre-scale folds associated with the large km-scale F_3 folds are well developed within the Isaac and the Yankee Belle Formations within the Isaac Lake Synclinorium, where the stratigraphy is predominantly upright (Plate 44). An axial planar cleavage related to F_3 folding can be seen in hand sample as a spaced crenulation cleavage which crenulates earlier formed S_2 fabrics (Plate 46). It is widely spaced (0.5-2cm), upright cleavage that only develops within the hinge zone of large F_3 folds. S_3 cleavage fanning in metre-scale folds is common in more competent sandstones of the Yankee Belle Formation (Plate 45). L_3 lineations associated with folding include intersection lineations of S_3 with S_0, and hinge lines of outcrop-scale F_3 folds. Petrographically, the S_3 cleavage is represented by crenulated muscovite or as a continuous spaced cleavage with gradational boundaries defined by the concentration of opaque material. S_2 is preserved within the microlithons defined by the S_3 cleavage.

All S_3 and L_3 measurements were plotted on an equal area net. Poles to the S_3 cleavage planes cluster around the north and south poles, with a mean cleavage plane that strikes 110° and dips 80° towards the south. L_3 lineations cluster in the east-southeast and west-northwest quadrants with a mean trend of 108° and plunge of 10° E (Figure 4-3).

4.3 d) Phase four folds (F_4)

A fourth phase of folding (F_4) is only present locally in the study area. These folds are on a metre- and decametre-scale and re-fold all previous structures. F_4 axial planes trend north-south to northeast-southwest, and are steep to upright. The folds have open to close interlimb angles and angular chevron-style hinge zones. A widely spaced (up to 1 m apart), kink-style crenulation is associated with this phase of folding (Plates 47 and 48).

4.4 Faults

Three types of faults occur within the study area including dextral-oblique strike-slip faults, normal faults and reverse faults; all faults have a steep dip (Figure 1-6). The dextral-oblique strike-slip faults will be discussed in the following chapter with specific reference to the main fault in the area, the Isaac Lake Fault. This section will focus on the description of the high-angle reverse and normal faults found in the study area.

Northwest-striking, northeast-dipping high angle reverse faults are found throughout the study area, although they have been observed only in the Cunningham and Yankee Belle Formations (Plate 49). Displacement along these faults ranges from tens of metres to hundreds of metres based on apparent stratigraphic throw. Fault zones are narrow, usually spanning five to ten metres, and kinematic indicators are poorly exposed. Drag folds, with amplitudes of a few metres, are seen commonly adjacent to the faults also indicating a reverse sense-of-motion. The relationship of the high-angle reverse faults to the F_1, F_2 and F_3 folding is difficult to determine. The faults are found in stratigraphic units that show little evidence of F_1 and F_2 folding, although, they could represent a manifestation of the folding seen in the ductile layers, especially those found within the Isaac Formation.

Northeast-striking high-angle normal faults are also found throughout the study area. These faults are seen to cross cut F_3 folds, but their relationship to S_4 crenulations and folds could not be determined. These faults are seen at different scales. Small-scale faults can be traced for tens of metres and have a few metres of stratigraphic throw. Large-scale normal faults can be traced for tens of kilometres and show stratigraphic offsets of hundreds of metres. Fault zones associated with the large-scale faults are a few metres wide, and sub-vertical slickenlines on subsidiary fault planes indicate dip-slip movement.

4.5 The Structural Multilayer

The variation in structural style of the main regional deformation (D_2) structures in the Isaac Lake Synclinorium is controlled by the stratigraphic succession. Folded rocks show a variation in fold geometry that implies strain variations. Strain variations depends on the rheological properties of the layers involved, the pressure and temperature at which the multilayer was deformed, the thickness of the layers and the nature of the contacts between adjacent layers (Ramberg, 1964; Ramsay, 1967, 1982; Cobbold et al., 1971; Ramsay and Huber, 1987; Treagus, 1988; Treagus and Sokoutis, 1992; Mazzoli and Carnemolla, 1993; Erickson, 1996) The Windermere Supergroup stratigraphy exposed in the study area consists of thick sequences of different lithologies which causes strain partitioning to occur within the sequence. The major transition in structural style occurs at the Isaac Formation-Cunningham Formation contact where there is a pronounced lithological change from predominantly thin-bedded fine-grained calcareous shale to thick-bedded limestones, respectively. This divides the stratigraphy into two main structural layers; the lower ductile layer which includes the Kaza Group and Isaac Formation, and the upper brittle layer which contains the Cunningham, Yankee Belle and Yanks Peak formations. Figure 4-4 is a generalized stratigraphic section of the Windermere Supergroup exposed in the study area, and the main rock type of each formation. Tight to overturned, metre- to kilometre-scale F_2 folds in the Isaac Formation and Kaza Group rocks do not occur in the structurally and stratigraphically higher Cunningham and Yankee Belle formations. Rather, broad kilometre-scale open folds and high-angle reverse faults are the contractional features in the Cunningham and Yankee Belle formations. Thus, evidence from field observations suggests that the amount of shortening in the lower ductile layer is greater than the amount of shortening in the upper brittle layer (Cunningham, Yankee Belle, Yanks Peak formations) suggesting that the lower ductile layer has undergone more strain during the deformational history of the Isaac Lake Synclinorium.

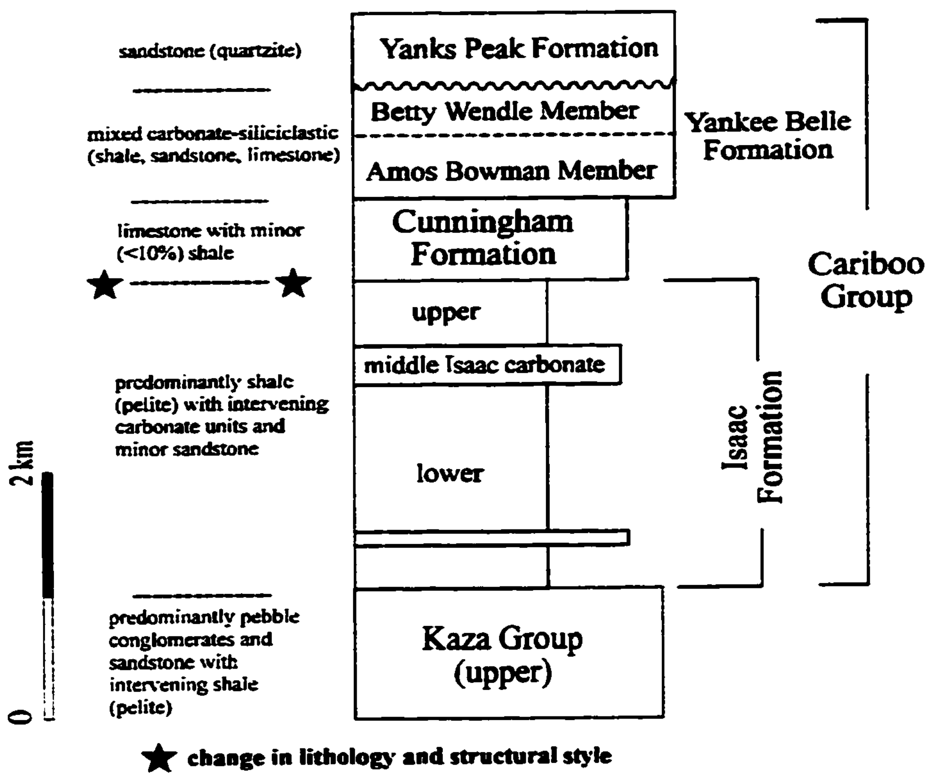


Figure 4-4. Generalized stratigraphic section of the Windermere Supergroup exposed in the Isaac Lake Synclinorium showing the main lithology of each formation and the transition in structural style between the Isaac and Cunningham Formations.

Quantitative finite strain evaluations allow an understanding of how rocks behave during deformation. The amount of strain a rock has undergone can be calculated using a number of different techniques including line-length restoration (Ramsay, 1967). This technique compares the original (undeformed) length of a line to the deformed length to give a percentage of how much the multilayers have been extended or shortened. This is done using the equation:

$$e = \frac{L_f - L_o}{L_o} \times 100$$

where,

e = percent shortening (-ve e value)
percent lengthenign (+ve e value)

L_f = final length of line

L_o = original length of line

(Ramsay, 1967)

Outcrop-scale examples of the change from tight F_2 folds in the Isaac Formation to broad open folds in the Cunningham Formation are found throughout the study area (Plate 50 and 51). These examples show that the amount of shortening in the Isaac Formation is more than the amount of shortening that occurred in the Cunningham Formation. Plate 50 is a photo showing tight folds in calcareous pelitic beds of the upper Isaac Formation and a gentle fold in the first thick carbonate unit of the Cunningham Formation. Line length calculations show that the tightly folded calcareous shale beds of the Isaac Formation have shortened by a minimum 38%, and the thick massive carbonate unit of the Cunningham Formation above by 10%. Plate 51 is a larger scale example of the tightly folded Isaac Formation below the broadly folded Cunningham Formation. Shortening in the carbonate rich beds in the Isaac Formation is approximately 30%, whereas in the broadly folded Cunningham above, the shortening is approximately 8%.

Pressure fringes around rigid markers also provide a means to determine the amount of finite strain during deformation (Elliott, 1972; Wickham, 1973; Durney and

Ramsay, 1973; Ramsay and Huber, 1983; Selkman, 1983; Ellis, 1984, 1986; Beutner and Diegel, 1985; Etchecopar and Malavielle; 1987; Hanmer and Passchier, 1991; Spencer, 1991; Aerden, 1996). Pressure fringes are comprised of elongate minerals (commonly quartz or calcite) which form adjacent to rigid grains in a deforming matrix. These fringes form as a result of strain partitioning in the matrix during deformation which is caused by the presence of the rigid grains (Durney and Ramsay, 1973). This heterogeneity within the rock sets up areas of high and low pressure around the rigid grains. The areas of low pressure, which are oriented perpendicular to the maximum principle compressive stress axis, allow for the precipitation of the fringes from solution as the matrix pulls away from the rigid grain (Hanmer and Passchier, 1991).

Pressure fringes around rigid pyrite porphyroblasts are common indicators of strain. In order to quantitatively measure strain, a number of assumptions need to be made about the pressure fringes: 1) the fibres are syntectonic; 2) fibres have not undergone subsequent deformation; and 3) the fibres are not face-controlled. Pyrite porphyroblasts are present throughout all of the stratigraphy of the Windermere Supergroup and their relationship to the structural fabrics are described in detail in Chapter 3. Pyrite porphyroblasts within the different formations of the Windermere Supergroup throughout the area provide the opportunity to compare qualitatively the difference in strain between the lower and upper layers. A quantitative estimate of strain using the pyrite pressure fringes cannot be done because the timing of formation of the fringes is uncertain (during D_1, D_2 or both), the fringes have undergone deformation, and the fringes are face-controlled. Despite these uncertainties, the fringes around pyrite porphyroblasts in the Isaac Formation are considerably more developed and complex than pressure fringes around pyrite grains in the Yankee Belle Formation, allowing for the overall cumulative strain of the different structural layers to be qualitatively examined and compared. All fringes were examined in thin sections cut in the XZ plane which is parallel to the intersection lineation of S_2 and S_0 (L_2) and perpendicular to the S_2 cleavage plane.

In the Isaac Formation, pyrite fringes are complex and show at least two phases of growth (Plates 35 and 36). The complexity of the fringes can vary depending on the position relative to the regional F_2 folds. For instance, fringes on the steep overturned limb of F_2 folds are more complex and elongate than those on the shallow upright limb (Plates 35 and 36), due to the expected higher strain in rocks on the overturned limb (Ramsay and Huber, 1987). Folding in the upper layer (Cunningham and Yankee Belle formations) does not form overturned limbs and therefore, only pyrite fringes from the flat limbs of F_2 folds in the Isaac Formation will be used for comparison with fringes in the Yankee Belle Formation. Due to the idioblastic shape of the pyrite porphyroblasts in the Isaac Formation, the fringes formed are face-controlled and growing at right angles to the pyrite faces (Plate 29). Therefore, the direction of fringe growth cannot be used to determine the maximum stretching direction, as face-controlled fringes grow irrespective of this direction. Pyrite porphyroblasts in similar lithologies in the Yankee Belle Formation show little to no pressure fringe development (Plate 21). Weak fringes can be seen around some larger (> 5mm) pyrite porphyroblasts, but overall, the pyrite porphyroblasts do not show complex fringe development as they do in the Isaac Formation.

The Cunningham Formation, which separates the relatively incompetent Isaac Formation from the Yankee Belle Formation is predominantly a thick massive carbonate unit. In the field, the Cunningham Formation exhibits few signs of small-scale strain or mesoscopic deformation. Outcrop-scale folds are not present and cleavage development in the carbonate units is rare; restricted to the lowermost portions. The intervening pelite layers show a weak cleavage development at angles of 10° to 30° to bedding, but no crenulation cleavage is present. Deformation of the Cunningham Formation is manifested in broad, kilometre-scale upright, open folds, high-angle reverse faults, and localized bedding-parallel shear zones (Plate 52). Petrographically, the Cunningham shows no internal signs of deformation. Originally spherical peloids in the limestone units show no departure from sphericity due to strain. Calcite grains also show no internal evidence of deformation such as twinning, dislocation creep or slip.

The difference in the percentage of shortening between the lower and upper layers in the Isaac Lake Synclinorium is apparent at a macroscopic, mesoscopic and microscopic scale. The mechanisms that allow for this will be discussed in the following summary / discussion section.

4.6 Summary / Discussion

4.6 a) Summary of deformation in the Isaac Lake Synclinorium

The structural evolution of the Isaac Lake Synclinorium can be divided into three main deformation events (D_1, D_2 and D_3). The first main event (D_1) formed metre-scale, east-verging, recumbent, isoclinal folds, small east-verging thrust faults, and an associated axial planar cleavage (S_1) which is sub-parallel to bedding (S_0). F_1 folds and faults are only found locally within the Isaac and Yankee Belle formations in the study area, however S_1 is pervasive throughout. Overall, the stratigraphy within the Isaac Lake Synclinorium is upright and large kilometre-scale F_1 fold closures are not found. This leads to the inference that the D_1 deformation event was widespread, affecting all of the rocks in the study area, yet it did not produce large-scale nappe structures that can be identified in the field.

The D_2 deformation event was the main, regional deformation event that produced the northwest-trending anticlinoria and synclinoria structures of the Cariboo Mountains. The Isaac Lake Synclinorium itself is a first-order F_2 fold within which second order, kilometre-scale F_2 folds occur. The F_2 folds in this area are northwest-trending, southwest-verging, asymmetric overturned folds. Meso- and macroscopic F_2 folding is restricted to the Kaza Group and Isaac Formation. Above the Isaac Formation, D_2 shortening is accommodated by broad, kilometre-scale, upright open folds, high-angle reverse faults, and local bedding-parallel shear zones. The axial planar cleavage associated with F_2 folding (S_2) crenulates S_1, and is the main cleavage found throughout the study area in the Kaza Group and Isaac Formation. In hand sample and thin section this cleavage is seen as a continuous crenulation cleavage with the cleavage plane defined by insoluble and opaque material, and locally by the parallel alignment of phyllosilicate minerals. S_2 development could not be found in the Cunningham Formation, but is found locally as a crenulation cleavage within the Yankee Belle Formation.

D_2 structures are cross-cut and deformed by D_3 structures, which include oblique-dextral strike-slip faults (Isaac Lake Fault), east-west trending upright folds, and northeast-trending extensional structures. The D_3 structures are in turn deformed by north-south trending D_4 crenulation cleavages and folds. The relationship of the D_3 structures to the dextral strike-slip faults such as the Isaac Lake Fault will be discussed separately in the following chapter on the dextral strike-slip faulting in the Isaac Lake Synclinorium.

The stratigraphy of the Isaac Lake Synclinorium forms a structural multilayer of contrasting lithologies which can be divided into a lower and upper layer. The lower layer consists of the upper Kaza Group and Isaac Formation, and the upper layer consists of the Cunningham, Yankee Belle and Yanks Peak formations. A distinct change in structural style occurs between the two layers as shown by field and microstructural observations. Pressure fringes around pyrite porphyroblasts on the flat F_2 limbs in the Isaac Formation are considerably more complex than pyrite pressure fringes in the Yankee Belle Formation indicating that the lower layer has undergone more strain than the upper layer. Line length restoration calculations on carbonate beds in the lower layer and upper layer show that the lower layer accommodated approximately 30 to 40% shortening whereas carbonate beds in the upper layer show only 8 to 10% shortening. This shortening discrepancy is most likely accommodated by localized bedding parallel detachments which are found in the uppermost Isaac Formation, and lower Cunningham Formation.

The D_4 event is minor and produced local chevron-style folds and a widely spaced crenulation cleavage that is oblique to the D_2 and D_3 fabrics. F_4 folds are found only locally within the Kaza Group and are local metre to decametre sized folds. The transverse crenulation cleavage is found in the stratigraphy throughout the study area and appears to be regionally pervasive. Overall, the effects of the D_4 event on the other structures is insignificant. Figure 4-5 pictorially summarizes the main deformation events (D_1, D_2 and D_3) in the Isaac Lake Synclinorium.

4.6 b) Age constraints

The age of deformation in the Isaac Lake Synclinorium is based on the relationship of equivalent structures with the Hobson Pluton south of the study area (Figure 4-1). The Hobson Pluton is interpreted to cut regional D_2 structures (Pigage, 1977; Gerasimoff, 1988). This brackets the age of the D_1 and D_2 events in this study area to have occurred before 174 Ma (mid-Jurassic) (Pigage, 1977; Gerasimoff, 1988), and after Upper Triassic based on the involvement of Upper Triassic sediments with D_1 and D_2 events (Brown, 1968; Campbell, 1971; Struik, 1980). The relationship of the Hobson Pluton to the regional structures has remained uncontested since the work of Gerasimoff (1988).

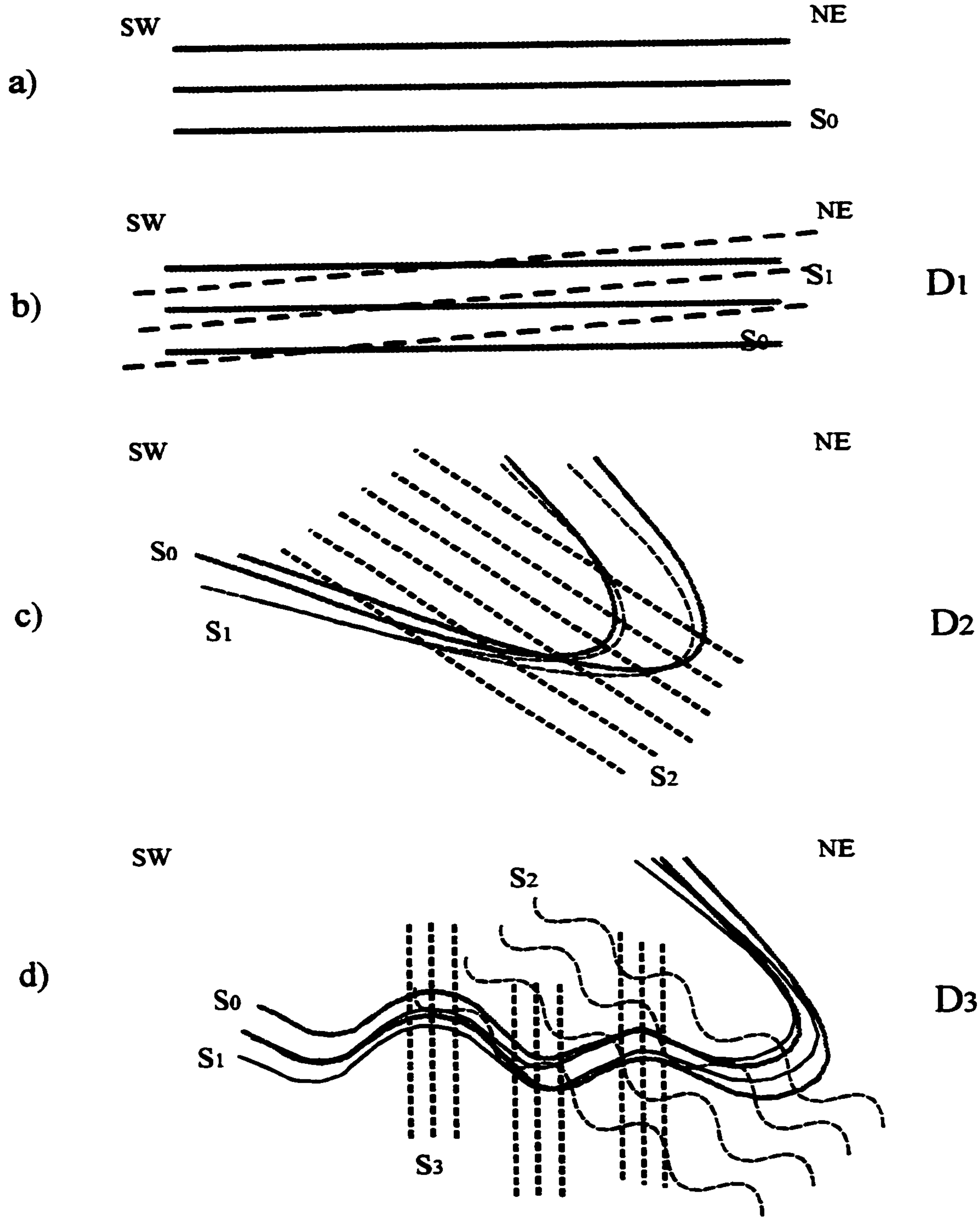


Figure 4-5. Evolution of cleavage development in the Isaac Lake Synclinorium. a) shows original stratigraphic layering (S_0), b) S_1 cleavage development at a low angle to S_0; c) regional F_2 folding and development of a penetrative axial planar cleavage S_2 which crenulates S_1; d) formation of F_3 folds and axial planar S_3 cleavage found in hinge regions of F_3 folds which refold F_2 and crenulates S_2 cleavage.

However, both Pigage (1977), and Gerasimoff (1985, 1988) describe the Hobson Pluton as deformed. The relationship of the deformational fabrics seen in the Hobson Pluton to the deformational fabrics in the country rock is not adequately discussed in either Pigage (1977) or Gerasimoff, (1988). This leaves open the possibility that the Hobson pluton is syn-kinematic with D_2 deformation. Due to the ambiguity of the field relationships between the Hobson pluton and the country rocks, and the implications it has on the interpreted ages of deformation, a re-examination of the pluton is warranted. Regardless of the uncertainty of the relationship between D_2 structures and the Hobson pluton, D_1, D_2 and peak metamorphism (M_1) in the Isaac Lake Synclinorium occurred either prior to or around 174 Ma and are therefore mid-Jurassic or older.

4.6 c) Suprastructure-Infrastructure correlation

The Cariboo Mountains have a deformational history spanning the period from the Jurassic to the Eocene. Within the suprastructure and infrastructure, structural style, metamorphic grade and timing of peak metamorphism differs making regional correlations difficult. A tentitive correlation of deformation events between the Isaac Lake Synclinorium and other study areas from both the suprastructure and infrastructure of the Cariboo Mountains is provided in Table 2 a and b. Common to all of the different study areas within the Cariboo Mountains is the regional northwest-trending structures which form the large-scale F_2 anticlinoria- and synclinoria-type structures of the Cariboo

REFERENCE		D_1	D_2	D_3	D_4
Reid (this study)	SUPRASTRUCTURE	- pre-174 Ma east-verging, isoclinal m-scale folds - peak metamorphism during D_1.	- regional, NW-striking folds upright to overturned (km-scale) - regional S_2 crenulation cleavage - post peak metamorphism	- east-west striking upright folds (km-scale) - S_3 co-axial to S_2 - dextral-oblique strike-slip faults (Isaac Lake fault)	- minor folding & cm-scale N-S striking transverse crenulation.
Ferguson, 1994	SUPRASTRUCTURE	- pre-174 Ma east-verging, isoclinal m-scale folds - peak metamorphism during D_1.	- regional, NW-striking folds upright to overturned (km-scale) - regional S_2 crenulation cleavage - post peak metamorphism	- east-west striking upright folds (km-scale) - F_3 co-axial to F_2 - dextral-oblique strike-slip faulting	- normal faulting & cm-scale N-S striking transverse crenulation.
Struik, 1980	SUPRASTRUCTURE	- pre-174 Ma east-verging, isoclinal m-scale folds - peak metamorphism during D_1.	- regional, NW-striking folds upright to overturned (km-scale) - regional S_2 crenulation cleavage - post peak metamorphism	- northwest-southeast striking folds (m to dm scale), F_2 and F_3 co-axial - northwest-striking dextral strike-slip faulting	- local m-scale folds and N-S striking kink crenulation

Murphy, 1985	TRANSITION	- east-verging, isoclinal m-scale folds	- west-verging nappes (km-scale) including Raush anticline.	- regional NW-striking folds upright to overturned (km-scale) - peak metamorphism syn-kinematic with D_2.	- local folds and crenulations	- local m-scale folds and N-S striking kink crenulation

Table 2 a) Summary of the main deformation events in the suprastructure and transition zone of the Cariboo Mountains.

REFERENCE		D_1	D_2	D_3	D_4
Pell, 1984	INFRASTRUCTURE	- **west-verging** nappes (km-scale) only developed at deeper structural levels.	- **regional NW-striking folds** upright to overturned (km-scale) - peak metamorphism syn-kinematic with D_2.	- minor randomly oriented m-scale folds, S_3 co-axial to S_2.	- normal faulting (North Thompson River Fault) - 51-45 Ma (Sevigny and Simony, 1989).
Currie, 1988		- **west-verging** nappes (km-scale, onset of M_1 metamorphism.	- **regional NW-striking folds** upright to overturned - age of D_2: 160-131 Ma constrained by U/Pb geochron. - peak metamorphism, post F_2, 134 +/- 4 Ma.	- minor randomly oriented m-scale folds, S_3 co-axial to S_2, crenulates S_2 fabrics.	- normal faulting (Allan Creek Fault) - 51-45 Ma (Sevigny and Simony, 1989).
Walker, 1989		- **west-verging** nappes (km-scale, onset of M_1 metamorphism.	- **regional NW-striking folds** overturned to isoclinal (km-scale) - peak metamorphism post-F_2 (134 +/- 4 Ma, Currie, 1988)	- minor randomly oriented m-scale folds, S_3 co-axial to S_2, crenulates S_2 fabrics.	- normal faulting (Albreda Normal Fault) - 51-45 Ma (Sevigny and Simony, 1989).
Cooley, 1997		- small-scale isoclinal folds.	- **regional NW-striking, SW-verging folds (km-scale)** - peak metamorphism post F_2.	- dextral strike-slip faulting (Matthew Fault), initiated by mid-Cretaceous.	- normal faulting - N-S striking transverse crenulation.

Table 2 b) Summary of the main deformation events in the infrastructure of the Cariboo Mountains.

Mountains. From east to west in the Cariboo Mountains, the sense of vergence of these folds changes from northeast-verging to southwest-verging respectively. This change in vergence is attributed to fanning of the fold belt (Campbell, 1970; Ferguson, 1994; Ferguson and Simony, 1992). The Isaac Lake Synclinorium is located on the western side of the fan structure and all of the F_2 structures are southwest-verging.

The geometry of the F_2 folds changes from the suprastructure to the infrastructure levels. In the suprastructure, the F_2 folds are upright to overturned with steep to moderately dipping axial planes, and re-fold earlier isoclinal F_1 folds (Campbell, 1970; Campbell et al., 1973; Struik, 1980; Ferguson, 1994; Reid, this study). F_2 folds described in the infrastructure are also upright to overturned northeast to southwest-verging folds which re-fold an earlier phase of isoclinal folds (Pell, 1984; Pell and Simony, 1987; Currie, 1988; Walker, 1989). The F_2 folds in the infrastructure have more shallowly dipping axial planes than those described in the suprastructure, but these structures do not form large-scale recumbent nappes as does the first phase of folding in the infrastructure.

The problem in correlating structures between the suprastructure and infrastructure in the Cariboo Mountains begins with the D_1 event. In the suprastructure, D_1 is related to east-verging, recumbent, isoclinal folds with an axial planar cleavage sub-parallel to S_0 (Struik, 1980; Ferguson, 1994; Reid, this study). The folds and axial planar cleavage are associated with the earliest phase of deformation evident at the suprastructure level and perhaps represent deformation in the footwall of the obducted Slide Mountain and Quesnel Terranes (Murphy, 1987). At the infrastructure level, D_1 is represented by west-verging, multi-kilometre-scale isoclinal folds and an associated axial planar cleavage (Pell, 1984; Deschene et al., 1986; Currie, 1988; Walker, 1989).

It is in the transition zone between the suprastructure and infrastructure levels where the relationship between the different D_1 structures are found. The east-verging D_1, isoclinal folds of the suprastructure are re-folded by isoclinal west-verging folds typical of the infrastructure (Murphy, 1985, 1987). Thus, D_1 in the suprastructure is not equivalent to D_1 in the infrastructure. Evidence for east-verging isoclinal structures deformed by west-verging structures is also present in the northwestern Cariboo Mountains. Struik (1980) describes the earliest deformational structures in his study area as east-verging isoclinal folds which are cut and deformed by west-verging thrust faults, all of which are deformed by the main regional northwest-trending folds.

How dramatically the D_2 structures change from the suprastructure to the infrastructure depends on how the large-scale structures are correlated. For instance, D_2 structures in the transition zone are described as isoclinal, recumbent, southwest-verging kilometre-scale folds as are D_1 structures in the infrastructure. Murphy (1985) correlates the upright to overturned D_2 structures of the suprastructure (Ferguson, 1994; Reid, this study) to the overturned, isoclinal southwest-verging folds of the transition zone, thus implying that the geometry of the folds changes dramatically from the suprastructure to the transition zone. However, the D_2 structures in the transition zone have a similar geometry to the D_1 structures of the infrastructure. Also, the D_2 structures in the infrastructure are described as overturned southwest-verging folds but not isoclinal recumbent folds (Pell, 1984; Currie, 1988; Walker, 1989). One alternate interpretation is that the overturned, isoclinal, recumbent D_2 folds of Murphy (1985) actually correlate with the D_1 structures of the infrastructure, and that the regional D_2 structures change their geometry along strike with structural level, from upright to overturned, but not as dramatic as previously interpreted by Murphy (1985). The correlation remains confusing and enigmatic, as none of the structures have been traced along strike, and significant, unmapped gaps remain between study areas of the suprastructure and infrastructure. Without geochronological age constraints and detailed structural mapping, structural fabrics in the infrastructure and suprastructure cannot be correlated without ambiguity.

As the geometry of the structures change with structural depth, so does the metamorphic grade and timing of peak metamorphism. This has made correlating structures by their association with peak metamorphism impossible. In the suprastructure, peak metamorphism is attained during D_1, before the main regional anticlinoria and synclinoria are formed (Struik, 1980; Ferguson, 1994; Reid, this study). Although the exact age of peak metamorphism is not known, it is known to be pre-174 Ma based on the relationship of the Hobson Pluton to D_1 structures (Pigage, 1977; Gerasimoff, 1988). Towards the southeast, in the transition zone, peak metamorphism is attained syn-kinematically with the D_2 structures (Murphy, 1985). In the southern Cariboo Mountains, peak metamorphism post-dates the formation of D_2 regional structures and metamorphic monazite gives an age of 135 +/- 4 Ma (Currie, 1988). This 135 +/- 4 Ma metamorphic event overprints an earlier metamorphic assemblage which may correlate with the peak metamorphic event in the suprastructure (Currie, 1988). Geochronological age constraints on the D_2 structures themselves also show a west to east younging. Although the relationship of the Hobson Pluton to the D_2 structures is debatable, D_2 deformation is either 174 Ma or older. In the southeast (infrastructure), the D_2 event is distinctly younger, than the D_2 event in the northwest (suprastructure) . Zircon fractions from a pre-D_2 pegmatite and post-D_2 pegmatite constrain the age of D_2 between 160 Ma and 131 Ma (Currie, 1988). Therefore, there is an overall younging of peak metamorphism and deformation from shallow to deep structural levels (suprastructure to infrastructure) in the Cariboo Mountains.

4.6 d) **Regional correlation of the Cariboo Mountains Suprastructure**

Although significant problems remain with correlating deformation events of different structural levels within the southern Omineca Crystalline Belt, there is an overall similarity between the upper structural levels (suprastructure) of the Cariboo Mountains and the upper structural levels of the Selkirk and Purcell Mountains. Much of the southern Omineca Crystalline Belt is comprised of low-grade slate belts similar to the rocks preserved in the Isaac Lake Synclinorium (Figure 4-6). Table 3 summarizes the deformation histories of the greenschist facies upper structural levels of the Cariboo, Selkirk and Purcell Mountains. As shown in Table 3, the Cariboo Mountains and Selkirk Mountains share a similar deformational history; the D_2 folds are the main regional structures which formed during the early to mid-Jurassic. The main structures of the Purcell Mountains however are thrust-dominated, which is similar to the structural style of

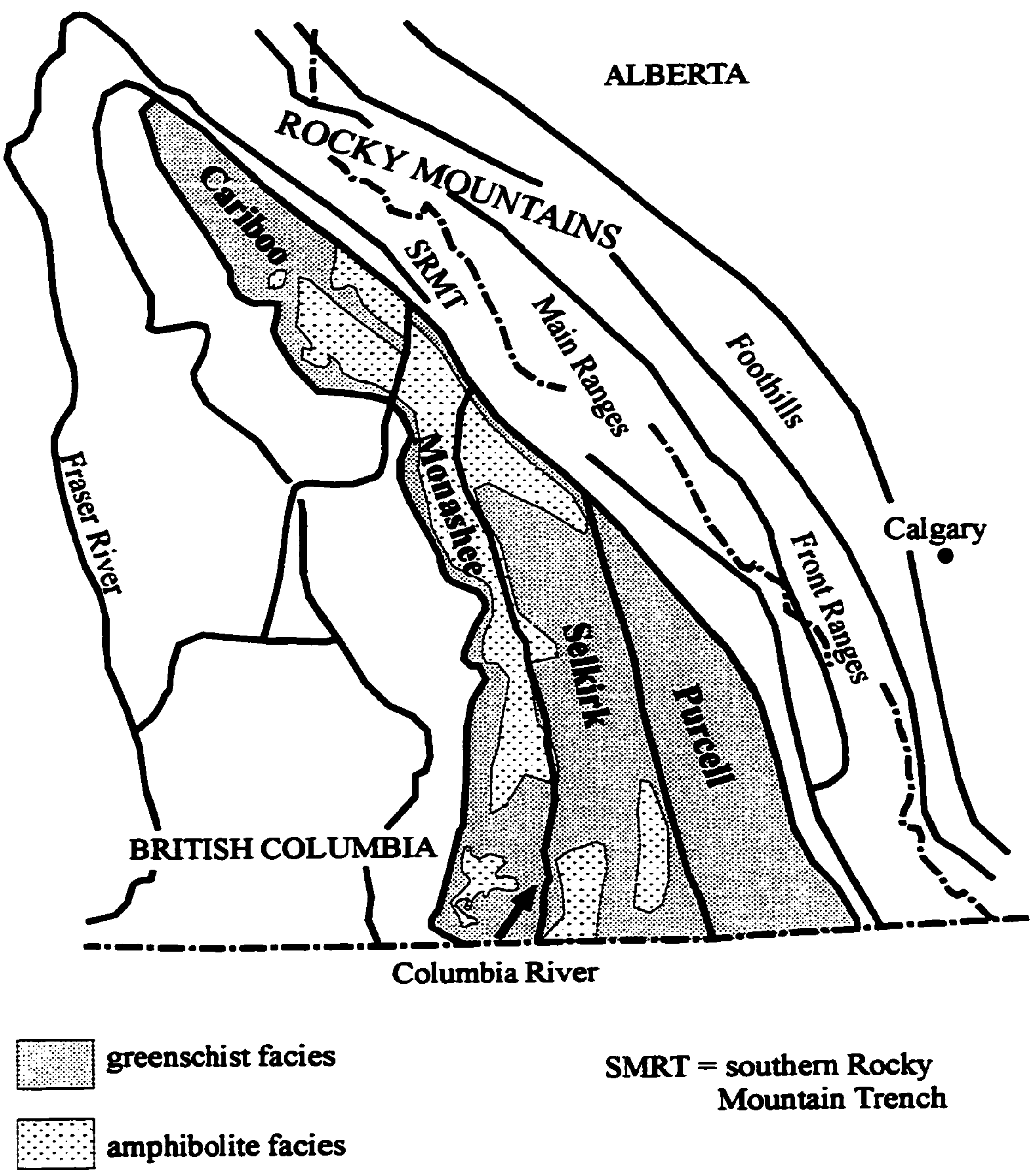


Figure 4-6. Map of the southern Canadian Cordillera showing the distribution of greenschist and amphibolite facies rocks in the southern Omineca Crystalline Belt.

REFERENCE	D_1	D_2	D_3	D_4
CARIBOO MOUNTAINS Reid (this study), Ferguson, 1994; Campbell et al., 1973	- pre-174 Ma **east-verging**, isoclinal m-scale folds - peak metamorphism during D_1.	- **regional, NW-striking SW-verging folds** - regional S_2 crenulation cleavage - post peak metamorphism - pre- 174 +/- 1 Ma.	- east-west striking upright folds (km-scale) - S_1 co-axial to S_2 - dextral oblique strike-slip faults (Isaac Lake Fault)	- minor folding & cm-scale N-S striking transverse crenulation.
SELKIRK MOUNTAINS Carr, 1991; Brown, 1991; Brown and Tippet, 1978	- **west-verging** thrust faults and nappe structures (Carnes nappe)	- **regional, NW-striking, SW-verging folds** - regional S_2 crenulation cleavage - peak metamorphism syn-kinematic - pre 168 +/- 3 Ma.	- local folding	- N-S striking extension faults (Eocene)
		D_1	D_2	D_3
PURCELL MOUNTAINS Root, 1987; Kubli, 1990		- east-verging thrust faults - regional-scale folds - peak metamorphism syn-kinematic - D_2 deformation ~ 166 Ma.	- penetrative east-west crenulation cleavage - east-verging thrust faults (pre- 95 Ma) - strike-slip component to faults	- N-S striking extension faults (Eocene)
WESTERN MAIN RANGES (Cushing Creek) Carey, 1984		- east-verging thrust faults and initiation of the Mt. Rider Synclinorium and Cushing Creek Anticlinorium (post mid-Jurassic)	- east-verging thrust faults and folds which deform previously formed structures - peak metamorphism	

Table 3. Summary of the main regional deformation events at high structural levels in the Cariboo, Selkirk, Purcell Mountains and Western Main Ranges at Cushing Creek east of McBride.

One of the major problems which remains unresolved is the correlation of rocks deformed in the suprastructure in the Cariboo Mountains to similarly deformed and metamorphosed rocks of the western Main Ranges at the latitude of McBride. The major structure in the western Main Ranges at the latitude of McBride is the Mount Rider / Robson Synclinorium (Carey, 1984; Campbell et al., 1973) (Figure 4-1). Previous workers have interpreted the Mount Rider / Robson Synclinorium to be the eastern continuation of the fold train that dominates the structures of the Cariboo Mountains (Campbell et al., 1973; Ferguson, 1994). In the Cariboo Mountains, it is known that the anticlinoria and synclinoria structures (D_2 structures) in the suprastructure are mid-Jurassic or earlier in age. Work done by Carey (1984) demonstrated that the Mount Rider / Robson Synclinorium and the Cushing Creek Anticlinorium formed as a result of northeast-verging thrusting along the Snake Indian Thrust fault. The Snake Indian thrust fault cuts the upper part of the mid to late Jurassic Fernie Group, indicating that thrusting, and folding took place no earlier than the latest Jurassic (Mountjoy, 1978; Carey, 1984). This difference in the age of formation of the fold structures in the Cariboo Mountains and those of the western Main Ranges indicates that the two sets of structures did not form synchronously.

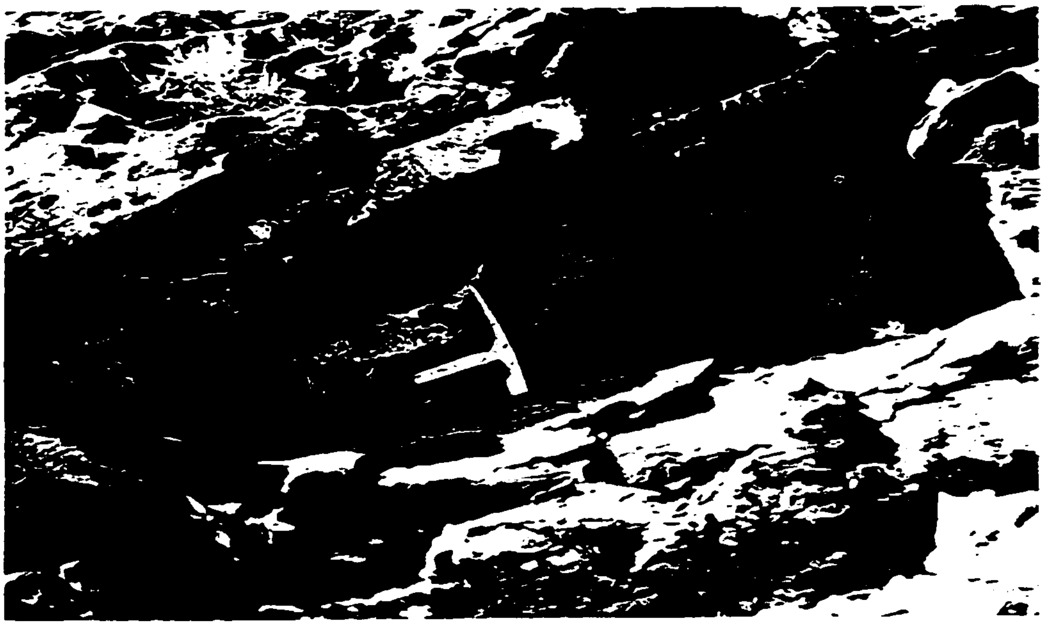


Plate 37. Northeast-verging F_1 fold in the upper Isaac Formation. Limbs of the folds are thinned, and the hinge is thickened. The folds have an axial planar cleavage (S_1) associated with them, which is crenulated by later cleavages.

Plate 38. Photomicrograph of the S_1 cleavage associated with the F_1 folds. The cleavage is defined by the parallel alignment of chlorite, muscovite and chlorite-muscovite aggregates, and is oriented sub-parallel to S_0 (bedding).

Plate 39. Photomicrograph of cm-scale southwest-verging F_2 folds.

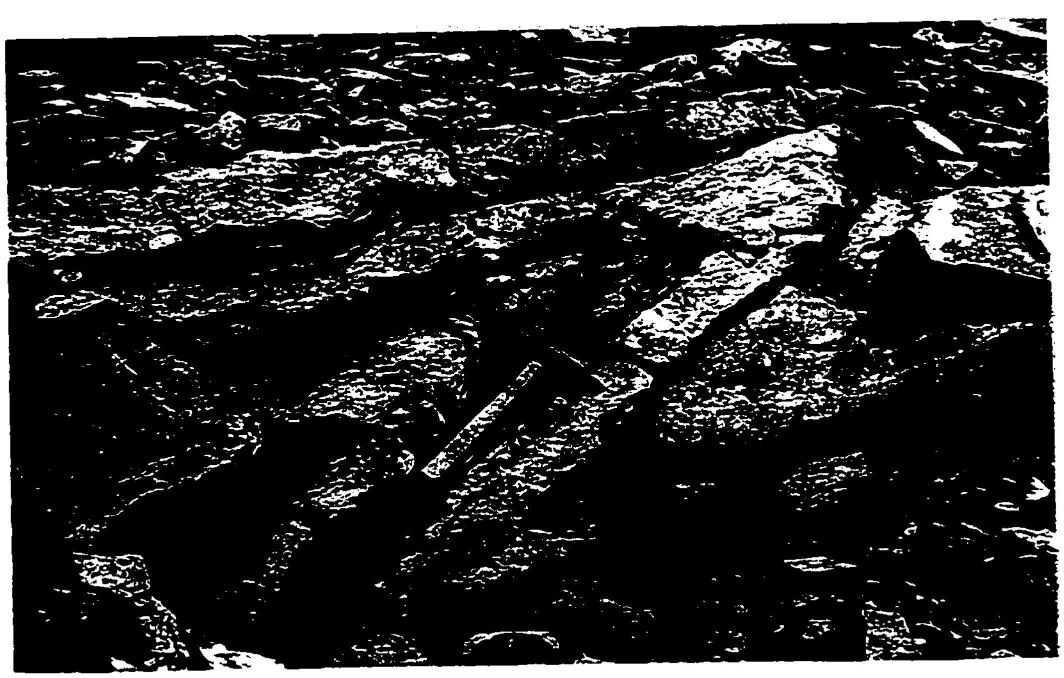


Plate 40. Metre-scale northwest-trending, southwest-verging F_2 fold in the upper Isaac Formation. The axial planar cleavage (S_2) associated with these folds crenulates the earlier formed S_1 cleavage.

Plate 41. Looking southward towards a kilometre-scale, southwest-verging overturned F_2 fold hinge preserved in the peak of Mt. Quanstrom.

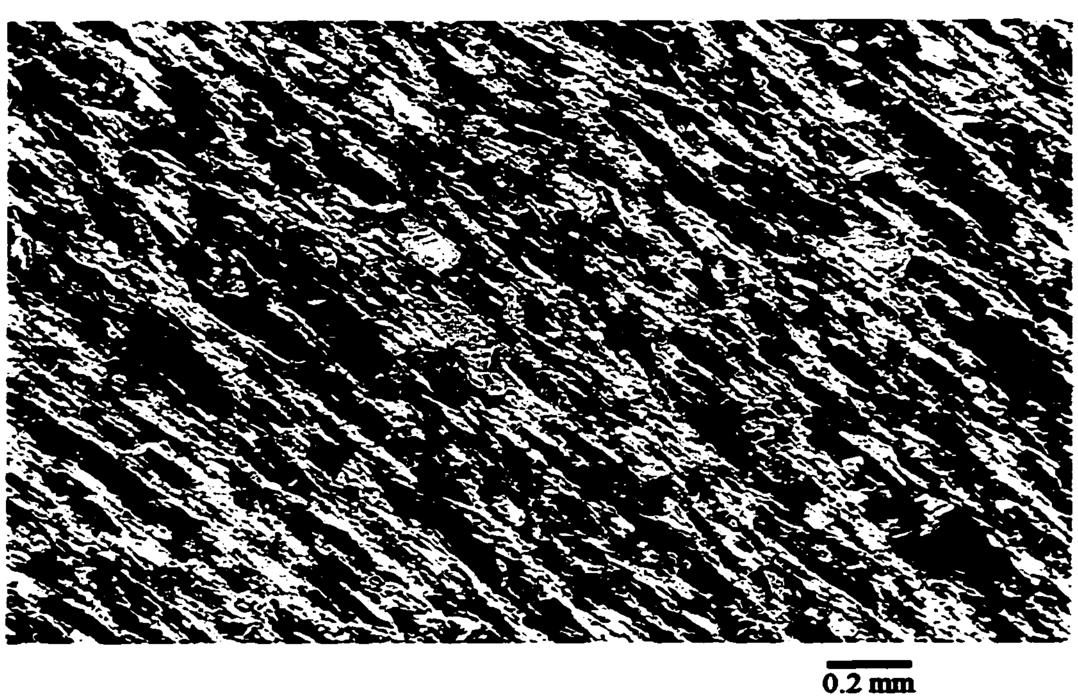
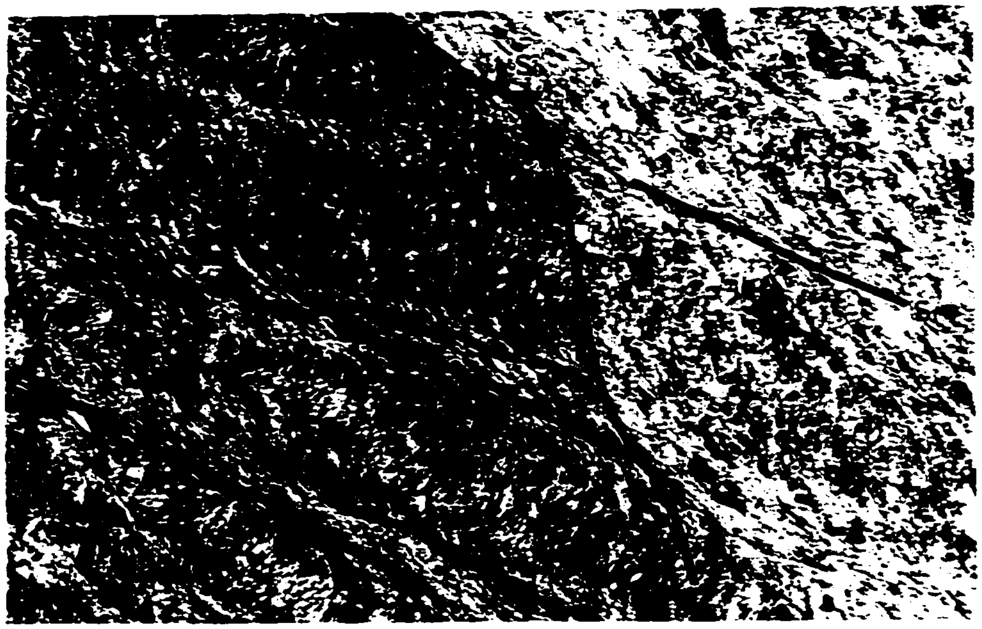


Plate 42. Continuous S_2 cleavage defined by the parallel alignment of muscovite from a pelite in the lower Isaac Formation. Quartz grains are also oriented with their long axis parallel to the S_2 cleavage direction. Section is cut parallel to the intersection lineation of S_2 with S_0 (L_2), and perpendicular to the S_2 cleavage plane.



0.6 mm

Plate 43. Discontinuous S_2 cleavage in a calcareous siltstone from the middle Isaac carbonate unit. S_1 is crenulated around the S_2 cleavage planes.

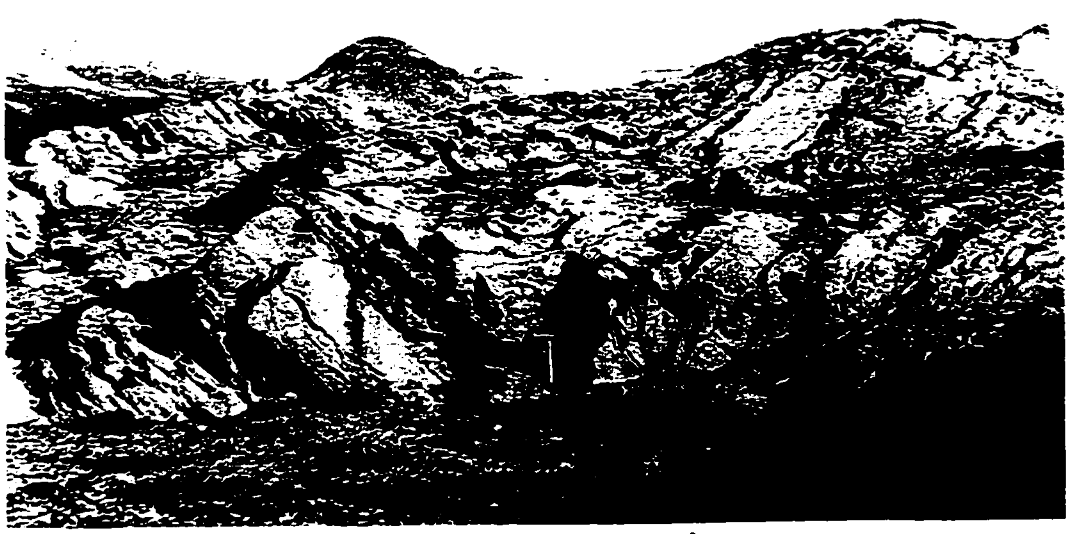


Plate 44. East-southeast trending, open, upright F_3 fold in the Yankee Belle Formation. Fold plunges gently (5°) towards the southeast, and has a spaced axial planar cleavage (S_3) developed in the hinge region.

Plate 45. F_3 fold in the Yankee Belle Formation with a spaced axial planar cleavage (S_3) developed in the hinge region. Cleavage fans outwards away from the hinge of the fold.

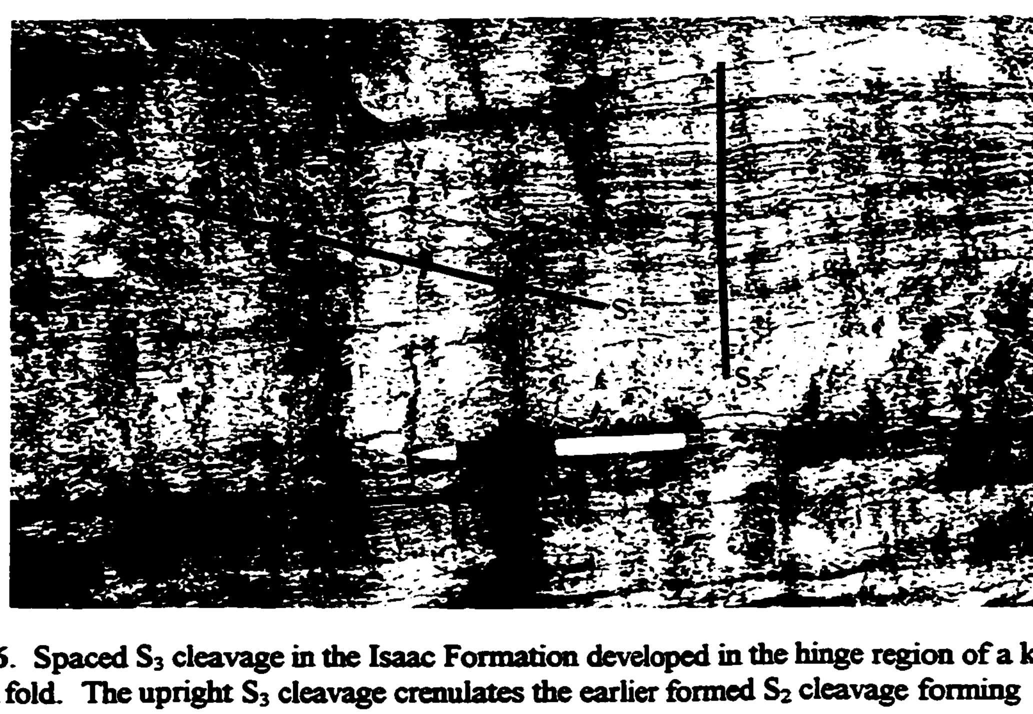


Plate 46. Spaced S_3 cleavage in the Isaac Formation developed in the hinge region of a kilometre-scale F_3 fold. The upright S_3 cleavage crenulates the earlier formed S_2 cleavage forming intersection lineations which plunge shallowly towards the southeast (3-5°).

Plate 47. North-south striking S_4 kink crenulation which cuts across S_2 and S_3 fabrics.

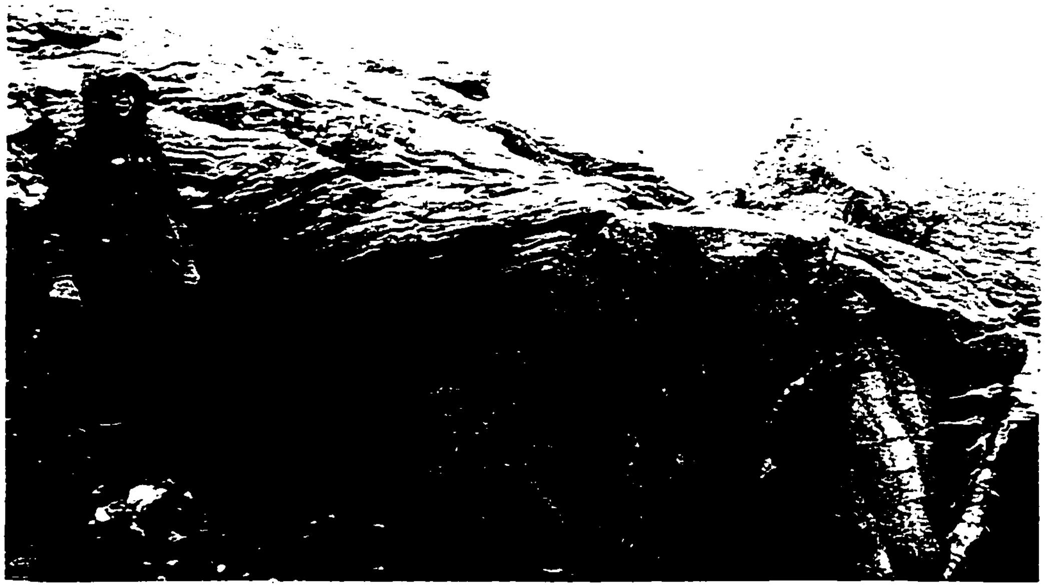


Plate 48. Local northeast-southwest trending F_4 fold which is seen refolding the intersection lineation of S_2 with S_0 (L_2) on the base of a vertical Kaza Group sandstone bed.

Plate 49. High-angle reverse fault in the Yankee Belle Formation. Elevation from the top of the scree slope to the ridge line is approximately 1000 m.

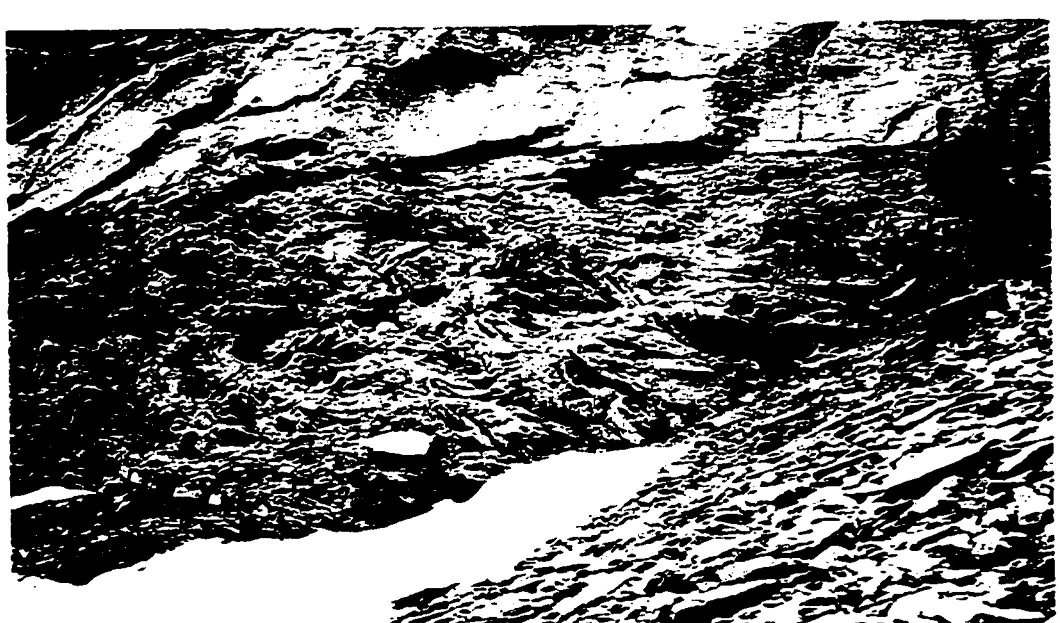


Plate 50. Thick basal Cunningham Formation carbonate bed which is broadly folded overlying the tightly folded and thinly bedded shale and carbonate beds of the upper Isaac Formation.

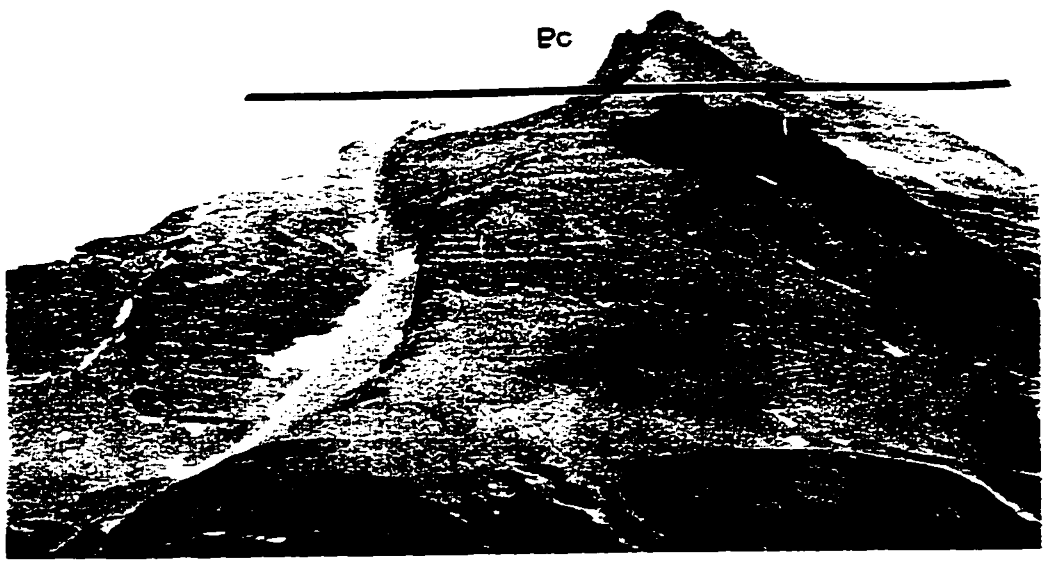


Plate 51. Tight, southwest-verging folds in the upper Isaac Formation which are not expressed in the overlying Cunningham Formation.

Plate 52. Sheared carbonate bed from the lowermost Cunningham Formation. Shear zone is 5 cm wide and runs parallel to the base of the bed.

CHAPTER 5: DEXTRAL STRIKE-SLIP FAULTS IN THE ISAAC LAKE SYNCLINORIUM

5.1 Introduction

Dextral strike-slip fault systems have been documented throughout the western Canadian Cordillera. Although the magnitude of dextral strike-slip faulting in the western Canadian Cordillera is a subject of debate among workers, these fault systems have significant tectonic implications for the evolution of the western Canadian Cordilleran orogen (Gabrielese, 1985; Price and Carmichael, 1986; Irving at al., 1985; van der Heyden, 1992; Monger et al., 1994; Basset and Kleinspehn, 1996). The formation of the orogen involved the accretion of the Intermontane and the Insular Superterranes to the western margin of North America which occurred from the Mesozoic to early Cenozoic (Monger, 1982). Two main tectonic models have been developed for the accretion of these superterranes. The first model uses paleomagnetic data and paleontological and palynological distribution data to estimate ~1100 +/- 600 km of post mid-Cretaceous northward translation of the Intermontane Superterrane relative to western North America (Irving et al., 1985). Paleomagnetic data also placed the Insular Superterrane ~1900 +/- 500 km south of the Intermontane Superterrane, therefore requiring some 1900 km of dextral displacement to bring the Insular Superterrane into its present position (Beck and Noson, 1972; Irving et al., 1985, 1995, 1996; Wynne et al., 1995). This cumulative ~3000 km of post mid-Cretaceous northward translation of the terranes relative to the

North American craton cannot all be accommodated by known dextral strike-slip faults in the Cordillera (Irving et al., 1994, 1995, 1996). Cumulative dextral displacement on faults in the western Canadian Cordillera can accommodate the ~1100 km of northward translation of the Intermontane Superterrane (Irving et al., 1996), however the additional 1900 km of dextral displacement for the Insular Superterrane has not been accounted for. The second major tectonic model does not involve significant northward translation of the superterranes relative to North America or each other (van der Heyden, 1992; Monger et al., 1994). The model of van der Heyden (1992) proposes that the Intermontane and Insular Superterranes were adjacent by Jurassic time and at the approximate longitudinal position as seen today. This is based on sedimentological and geochemical similarities of rocks within each of the superterranes. However, abundant plutonism during the mid-Cretaceous to form the Coast Plutonic Belt which straddles the Intermontane and Insular superterrane boundary obscures the critical structural relationships, and definitive links remain elusive. This model attributes dextral strike-slip faults in the western Canadian Cordillera to oblique subduction and convergence during the mid-Cretaceous.

Regardless of the conflicting models, the presence and magnitude of the dextral strike-slip faults found in the western Canadian Cordillera is significant. Some of the most prominent fault systems include from east to west: the Tintina-Northern Rocky Mountain Trench and McLeod Fault systems; the Teslin, Thibert, Kutcho and Pinchi Fault system; and the Fraser River-Straight Creek, Finlay, Takla and Ingenika, Yalakom, Pasayten, Harrison and Ross Lake Fault systems. These fault systems were active mainly from the mid-Cretaceous to Eocene (Gabrielese, 1985; Coleman and Parrish, 1991; Basset and

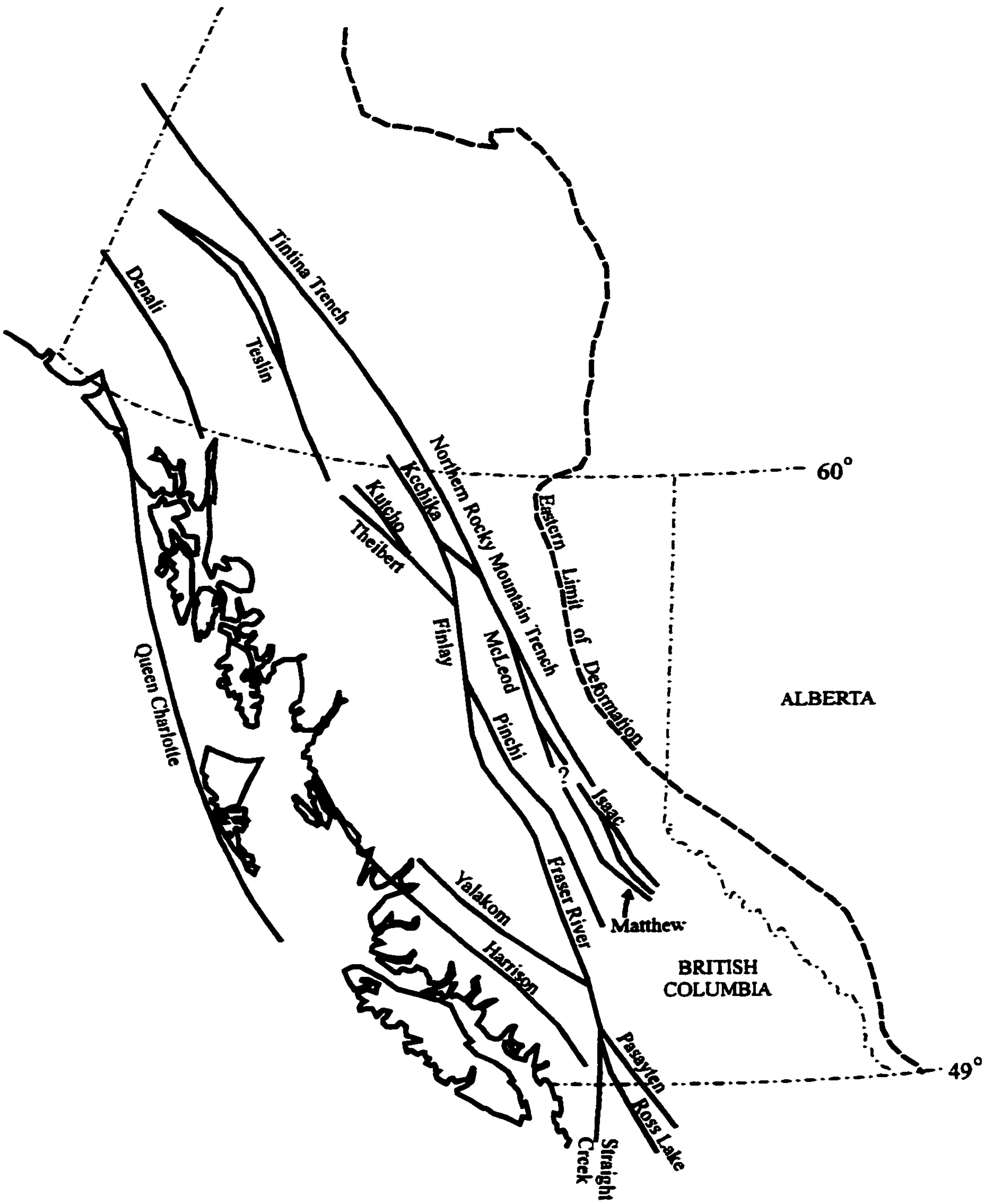


Figure 5-1. Map of the western Canadian Cordillera showing major dextral strike-slip fault systems.

Kleinsphehn, 1996) and accommodated approximately 1100 km of cumulative dextral strike-slip motion. Figure 5-1 is a map of the western Canadian Cordillera showing the distribution of major dextral strike-slip faults. One of the major debates about strike-slip faulting in the southern Canadian Cordillera is the nature of the southern extension of the northern Rocky Mountain Trench fault. The northern Rocky Mountain Trench fault is a large dextral strike-slip fault which can be traced from the Yukon to east-central British Columbia. This fault lies along strike with the southern Rocky Mountain Trench valley which does not host a significant dextral strike-slip fault (Price and Carmichael, 1986; Murphy, 1990; McDonough, 1989; van der Velden and Cook, 1996). It was suggested that motion along the northern Rocky Mountain Trench fault was transferred to the Fraser River-Straight Creek fault in a large-scale step-over zone in the Eocene (Gabrielese, 1985; Price and Carmichael, 1986). However, the dextral strike-slip faults in the Cariboo Mountains indicate that some of the motion may have been transferred southward through the Cariboo Mountains west of the southern Rocky Mountain Trench.

Transcurrent faults have been documented in the Cariboo Mountains by various workers over the last 25 years (Campbell, 1973; Murphy, 1987; Struik, 1980, 1985a, 1985b; Gerasimoff, 1985, 1988; Ferguson, 1994). These faults are predominantly northwest to north-northwest striking, dextral strike-slip faults and include the Isaac Lake, Matthew, Knutson Creek, Narrow Lake, Stony Lake, Willow River, and Pinchi faults. Strike-slip faulting in the Cariboo Mountains is interpreted to be post-mid Jurassic in age (174 +/- 4 Ma), as the Hobson pluton south of the study area is sheared by the Matthew Fault. Strike-slip faults also cross-cut the regional D_2 structures which form the large-scale anticlinoria- and synclinoria-type structures (Gerasimoff, 1988; Murphy, 1985,1987; Ferguson, 1994; Reid, this study). The regional significance of dextral strike-slip faults in the Cariboo Mountains has been overlooked in the literature due to the lack of geochronological age constraints. However, recent geochronological work by Cooley (pers. comm, 1996) on lineated hornblende in the Matthew Fault zone has shown that strike-slip faulting was initiated by the mid-Cretaceous.

Figure 5-2 is a geological map showing the location of major strike-slip faults in the Cariboo Mountains. Dextral oblique strike-slip faults define the eastern and western boundaries of the Isaac Lake Synclinorium, creating an east-dipping fault block which contains the earlier formed regional (F_2) synclinorium structure (Figure 1-5 and 1-6). These faults include the Isaac Lake Fault at the western boundary and the Winder Fault at the eastern boundary. The Isaac Lake Fault is a northwest striking, northeast dipping, dextral-oblique strike-slip fault with an east-side down component. The fault is contained within a series of large northwest-trending valleys that can be traced for approximately 100 kilometres from north of Isaac Lake southward towards Hobson Lake (Plate 53). The fault and its related kinematic indicators are best exposed north of the study area along the west shore of Isaac Lake, where chlorite-grade rocks on the east side of the fault are juxtaposed against garnet-grade rocks on the west side of the fault (Plate 54), (Ferguson, 1994, Ross and Ferguson, 1996). At Isaac Lake, the fault has a calculated dip of 60° towards the northeast (Ferguson, 1994; Ross and Ferguson, 1996). Slickenlines and stretched quartz fibres along fault surfaces at Isaac Lake plunge to the southeast at an angle of ~35- 45°, indicating dextral-oblique movement along the fault with an east side

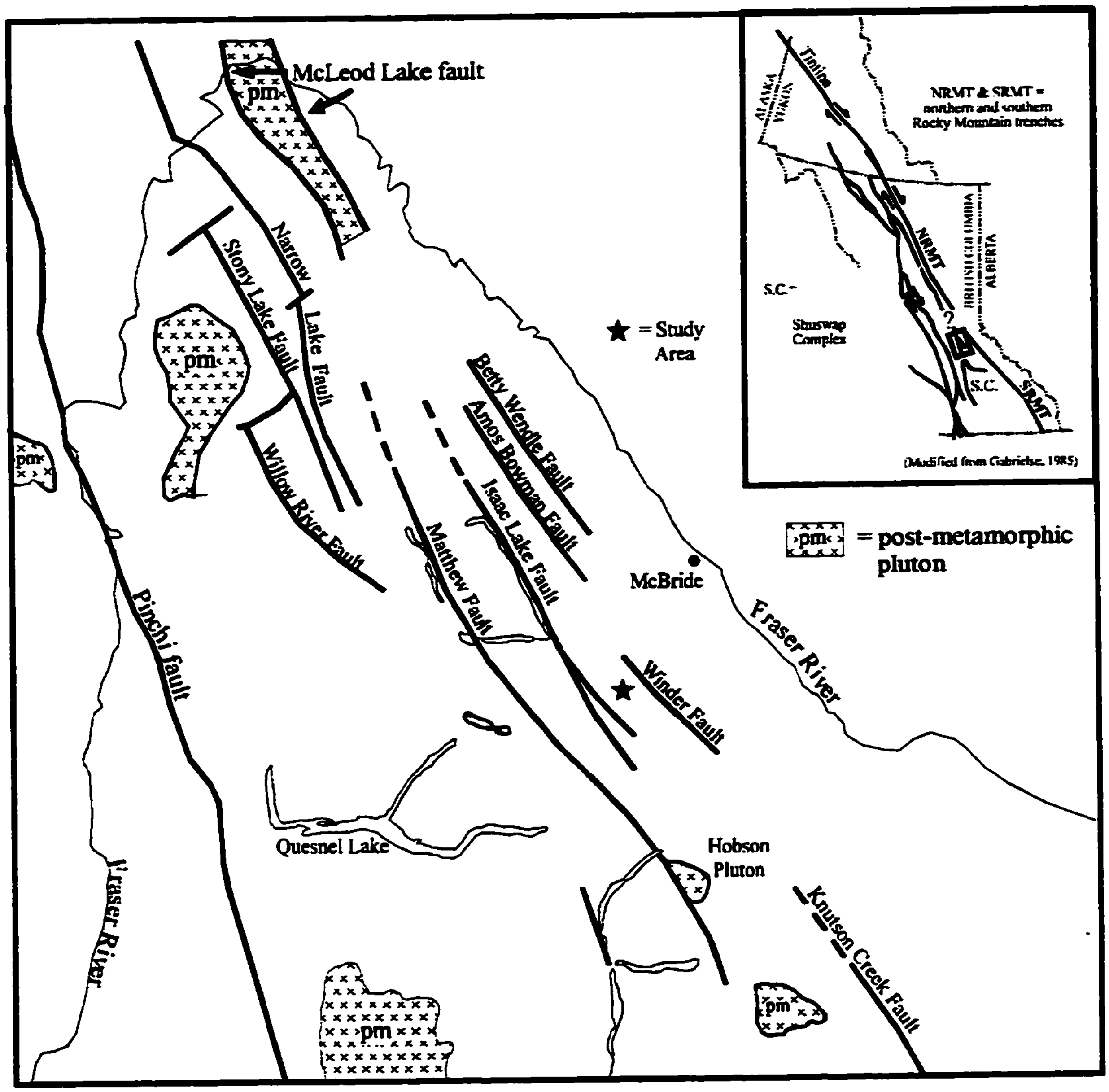


Figure 5-2. Map of the Cariboo Mountains showing the location of the major dextral strike-slip faults. Compiled from Murphy (1987), Struik (1985a, 1985b), and Ferguson (1994).

down component (Ferguson, 1994; Ross, pers. comm., 1996). Dip separation along the Isaac Lake Fault at Isaac Lake has been estimated at four to six kilometres (Ferguson, 1994). Given this information, the net-slip along the Isaac Lake Fault at Isaac Lake is calculated to be approximately 6 km with a resolved dextral strike-slip offset of 5 km.

At the southern tip of Isaac Lake, the Isaac Lake Fault bifurcates into two sub-parallel splays which can be traced through the study area (Ross and Ferguson, 1996). The following section describes the effects of dextral strike-slip faulting in the Isaac Lake Synclinorium.

5.2 Dextral strike-slip faults in the Isaac Lake Synclinorium

5.2 a) Isaac Lake Fault

In the study area, the two strands of the Isaac Lake Fault are contained within northwest-striking valleys with little to no outcrop and dip steeply to the northeast (Plate 55). There are three main locations in the study area where the faults and/or the rocks adjacent to the faults were examined. The location of these three areas are shown in Figure 1-5 and are numbered with Roman numerals i, ii, and iii.

Area i crosses the western strand of the Isaac Lake Fault in the study area. The fault juxtaposes Kaza Group stratigraphy in the footwall against upper Kaza Group / lower Isaac Formation stratigraphy in the hangingwall. Quartz veining and fracturing in the rocks adjacent to the fault is extensive. Small-scale compressional faults that displace bedding features by tens of centimetres are also found in the adjacent rock. The stratigraphy in the footwall of the fault has no marker units that can be identified in the hangingwall to provide a constraint on the throw or net-slip along the fault.

Area ii crosses the eastern splay of the Isaac Lake fault in an area with poor outcrop below tree line. On the eastern hangingwall side of the fault, scattered outcrops of the lower Isaac Formation and middle Isaac carbonate are juxtaposed against Kaza Group rocks in the footwall of the fault. The fault zone itself is obscured by vegetation, however, Figure 1-5 shows that the relationship between the Isaac Formation in the hangingwall and the Kaza Group in the footwall is not a stratigraphic contact, and therefore, the rocks must have been juxtaposed by the faulting.

Area iii contains the best exposure of the eastern splay of the Isaac Lake Fault, where it juxtaposes the middle Isaac carbonate unit in the footwall against the upper Isaac Formation / Cunningham Formation in the hangingwall (Plate 56). Both sides of the fault have good outcrop exposure. In the rocks adjacent to the fault are north-northeast oriented extension fractures filled with quartz and calcite and north-south oriented centimetre-scale compressional faults. No kinematic indicators such as slickenlines, or stretching lineations were found. Cross-section B-B' in Figure 1-6 shows approximately 1000 m of dip separation along the fault.

Given the data from the Isaac Lake faults in the study area, the amount of net slip along the western strand cannot be constrained. However, along the eastern strand the dip separation is approximately 1000 m and the net-slip direction is interpreted to be parallel to the rake of the stretching lineations and slickenlines found along the Isaac Lake

Fault at Isaac Lake (Ferguson, 1994; Ross and Ferguson, 1996). With this information, the magnitude of net-slip and dextral strike-slip displacement can be calculated using a fault plane diagram (Figure 5-3). The fault plane diagram shows that if the stratigraphic separation along the eastern strand of the Isaac Lake Fault is 1000 m, then the resolved net-slip would be 1750 m and the magnitude of dextral strike-slip offset would be 1350 m. The orientation of extension fractures, and small-scale compressional faults in the rocks adjacent to the Isaac Lake faults is also consistent with the interpretation that the Isaac Lake faults are dextral-oblique strike-slip faults.

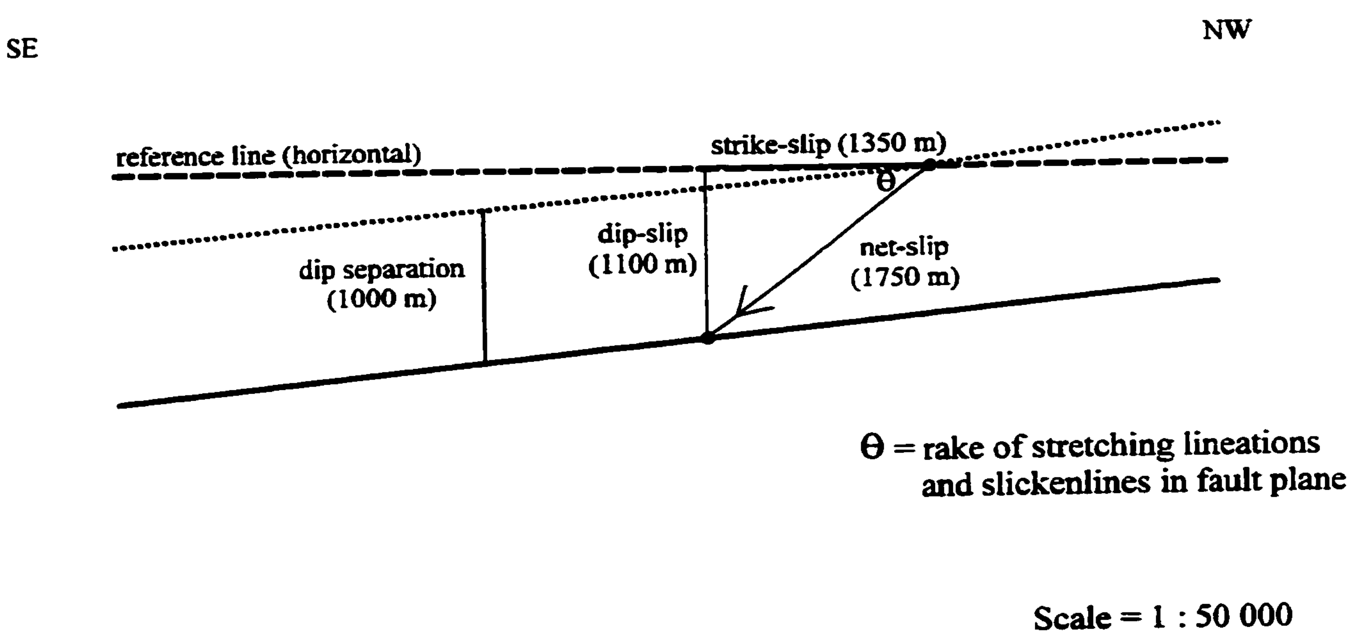


Figure 5-3. Fault plane diagram of the eastern splay of the Isaac Lake Fault. Dip separation of a hypothetical marker in the stratigraphy is illustrated by the dotted line (southwest side of the fault), and solid line (northeast side of the fault). The net-slip direction is taken to be parallel to the rake of the stretching lineation and slickenlines along the fault at Isaac Lake. Net-slip is calculated to be 1750 m and the strike-slip is 1350 m.

5.2 b) Winder Fault

The eastern boundary of the Isaac Lake Synclinorium is also a dextral-oblique strike-slip fault, with an east-side-up (thrust) component of displacement, which is well exposed across a series of north-trending ridges (Figure 1-5). The fault is northwest-striking and dips steeply to the northeast as illustrated on cross-sections A-A' and B-B' in Figure 1-6. This fault cuts obliquely across an F_2 fold (Figure 1-5), and where the fault crosses steeply dipping stratigraphy of the upper Isaac Formation in the overturned limb, original bedding appears transposed. The surfaces between layers of different composition are sharp and planar and sedimentary structures cannot be recognized. Carbonate units within and adjacent to the fault zone are highly fractured with randomly oriented quartz and dolomite veins. In the southeastern portion of the study area, fault slickenlines and stretched quartz fibres plunge moderately to the northwest at 35-45° indicating a dextral-oblique east-side-up (reverse) sense of motion (Ross, pers. comm., 1996). In rocks adjacent to the fault zone, northeast striking, steeply dipping extension fractures, and north-south striking small-scale strike-slip faults have an orientation that is consistent with the interpretation that the Winder fault is a dextral-oblique strike-slip fault (Plates 57 and 58). Along the fault ~800 m of dip separation is measured (Figure 1-6). Using a net-slip direction that is parallel to the slickenlines and stretching lineations, and the dip separation of 800 m, the magnitude of net-slip and strike-slip displacement is ~1400 m and 1100 m respectively.

5.2 c) Mitchell River Fault

The Mitchell River Fault is the third major dextral-oblique strike-slip fault in the study area. The fault is located approximately eight kilometres east of and parallel to the trend of the Isaac Lake faults, and can be traced from the headwaters of the Mitchell River to the south for 15 km where it is truncated by a northeast-striking normal fault (Figure 1-5). The fault dips steeply to the northeast with an east-side-down (normal) component of displacement, and it juxtaposes the upper Isaac Formation in the footwall against the Yankee Belle Formation in the hangingwall (Figures 1-5 and 1-6). The dextral component of displacement along the fault also displaces the axial plane of a large, upright, F_3 fold and indicates a maximum dextral strike-slip displacement on the fault of ~1000 m. This combined with a dip separation of ~750 m give an overall net-slip separation of 1250 m.

5.2 d) Related structures

The dextral strike-slip faults in the study area, including the Isaac Lake, Winder and Mitchell River faults, cross-cut D_2 fold structures, truncate D_3 structures, and are in turn deformed by D_4 structures. The northeast-striking normal faults cross-cut the F_2 folds, dextral strike-slip faults, and F_3 folds, but do not displace D_4 structures. The field descriptions and geometry of the D_2, D_3 and D_4 structures is given in Chapter 4. The angular and age relationships between the dextral-oblique faults, F_3 folds, northeast-trending normal faults, minor extensional fractures and compressional faults are shown in Figure 5-4 a. Figure 5-4 compares the structural relationships seen in the study area to that predicted by a right-lateral shear couple (Moody and Hill, 1956; Wilcox et al., 1973; Christie-Blick and Biddle, 1985). In figure 5-4 a, the angular relationship between compressional F_3 folds and the dextral strike-slip faults is similar that shown in Figure 5-4 b. Also, the orientation of extensional features such as faults and tension fractures relative to the main principle fault is similar to that shown in Figure 5-4 b. The dextral strike-slip faults, folds and extension structures all have similar age relationships to the other structures in the study area as they deform D_2 structures, but are in turn deformed by D_4 structures. Based on the angular and age similarity, the dextral-strike-slip faults, F_3 folds and northeast-trending extensional faults are interpreted to have formed during the same progressive deformation event (D_3), as a result of dextral strike-slip faulting.

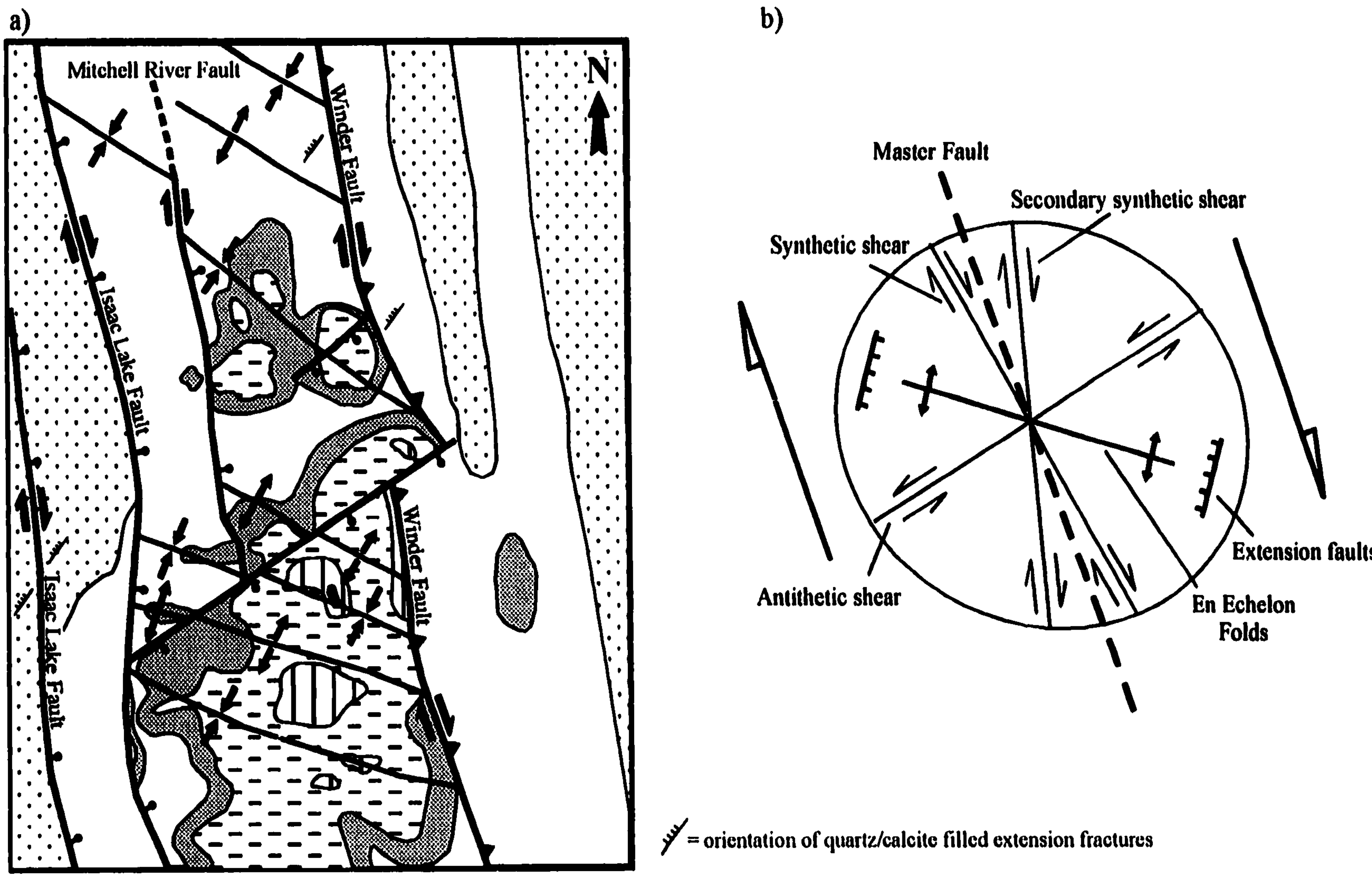


Figure 5-4. a) Diagram showing the Isaac Lake, Winder and Mitchell River faults, and D_3 structures illustrating the angular relationship between the faults, F_3 folds, normal faults and tension fractures. This is compared to a diagram (right) showing the angular relationship between structures expected to form in a dextral strike-slip fault system (Modified from Moody and Hill (1956), and Christie-Blick and Biddle (1985)).

5.3 Summary / Discussion

5.3 a) Summary of dextral-oblique strike-slip faulting

The Isaac Lake Synclinorium is bound and cut by northwest-striking dextral-oblique faults that include the Isaac Lake, Winder, and Mitchell River faults, creating an east-dipping fault block (Figure 1-6). These faults are recognized as dextral-oblique faults based on the orientation of slickenlines and stretching lineations on the fault planes, both within the study area and north of the study area at Isaac Lake, and by the orientation of small-scale extensional fractures and compressional faults adjacent to the fault zones. Dip separations across the Isaac and Winder faults are estimated at 1000 m and 800 m respectively. Using the orientation of the slickenlines and stretching lineations associated with the faults as the net-slip direction, the magnitude of net-slip and strike-slip displacement can be calculated. The net-slip and strike-slip displacement along the eastern strand of the Isaac Lake Fault is 1750 m and 1350 m respectively. On the Winder Fault, net-slip and strike-slip displacement is 1400 m and 1100 m, respectively. The net-slip and strike-slip displacement along the Mitchell River fault is calculated to be 1250 m and 1000 m respectively with a net-slip direction of 40° towards the southeast. Thus, the cumulative net-slip displacement in the study area is ~4400 m, and the cumulative dextral strike-slip displacement is ~3450 m.

The structural and angular relationship between the dextral-oblique faults, F_3 folds and northeast-trending extensional faults in the study area are similar to what is expected in a right lateral shear couple, suggesting that these structures are related to a single deformation event (D_3). The F_3 folds and extensional faults are found throughout the Isaac Lake Synclinorium, implying that the effects of dextral strike-slip faulting produced penetrative strain. This accommodation of dextral strike-slip motion through penetrative strain (i.e. folds, faults, and cleavage development) means that the amount of strike-slip deformation is more than can be calculated from regional maps.

The F_3 folds, which are related to the dextral strike-slip faulting, are themselves cut by the Mitchell River fault, which indicates that the history of dextral strike-slip faulting in the Isaac Lake Synclinorium is most likely a continuum of events which began with the formation of the Isaac Lake and Winder faults and ended with the formation of the Mitchell River fault and northeast-striking extensional faults. The Isaac Lake and Winder faults formed relatively early during D_3, along with the F_3 folds which formed preferentially between these two faults due to the orientation of the stratigraphy (see section 4.1 c). Dextral strike-slip faulting then continued which resulted in the formation of the Mitchell River fault and large-scale northeast-striking normal faults, which cross-cut the previously formed F_3 folds. Given that the timing on many of the major dextral strike-slip faults in the western Canadian Cordillera is mid-Cretaceous to Eocene, this age relationship seems reasonable.

5.3 b) Age constraints and regional correlations

The Isaac Lake Fault and other dextral strike-slip faults, including the Matthew Fault to the west, are interpreted to have formed during the same deformation event, based on the relationships of all of these faults to the main regional structures. That is, the faults cross-cut D_2 structures but not D_4 structures. Recently acquired geochronological age constraints on the Matthew Fault to the west (Cooley, pers. comm., 1996) indicate that this fault was initiated by the mid-Cretaceous. This implies that the Cariboo Mountain dextral strike-slip system would have been initiated by the mid-Cretaceous. This is coeval with the timing of dextral strike-slip motion along the Northern Rocky Mountain Trench Fault, which was active from the mid-Cretaceous to Eocene (Gabrielese, 1985).

The Isaac Lake and Matthew faults can be traced from the Quesnel Lake map area (NTS 93A) northward into the McBride map are (NTS 93H) but they have not been traced north of Isaac Lake. However, northward along strike of the Matthew and Isaac Lake faults lies the western and eastern splays of the McLeod Lake fault. The McLeod Lake fault is a north-northwest striking subsidiary fault of the Northern Rocky Mountain Trench fault which can be traced for approximately 350 kilometres from the northern tip of Williston Lake to the southeastern portion of the McLeod Lake map (NTS 93J), where it bifurcates at the southern tip of McLeod Lake (Muller and Tipper, 1969; Wheeler and McFreeley, 1991). Age constraints on the McLeod Lake fault are uncertain. The fault deforms the western margin of the Eaglet Pluton which has been dated as 43 Ma (Struik, 1993; Wanless et al., 1970), which suggests the fault was active in the Tertiary. Timing constraints on earlier movement along the fault are not available. The McLeod Lake fault also truncates northwest-striking dextral strike-slip faults, however, timing constraints on these faults are not well defined (Struik, 1990). The geometric relationship between the McLeod Lake Fault, and the Isaac Lake and Matthew faults suggests that the Cariboo Mountain strike-slip fault system could be a portion of the southern extension of the Northern Rocky Mountain Trench fault system. A quantitative estimate of the magnitude of dextral strike-slip movement in the Cariboo Mountains is difficult to ascertain for a number of reasons. Firstly, many of these faults were originally identified as normal faults, and therefore, the magnitude of dextral strike-slip motion was not calculated. Secondly, it is difficult to find geological markers that can be traced from one side of the fault to the other to provide constraints on the amount of dextral strike-slip motion that has occurred. Thirdly, dextral strike-slip movement was accommodated by both the faults and their associated structures (folds and extensional faults).

More detailed studies on the fault zones themselves need to be undertaken, which will allow for better age constraints of the timing of fault motion, and determination of the magnitude of displacement along the faults. At present, the amount of net-slip displacement on the dextral strike-slip faults in the Cariboo Mountains has not been calculated and therefore the amount of cumulative dextral motion accommodated by faults in the Cariboo Mountains cannot yet be determined. This will be necessary in order to further pursue the regional significance of dextral strike-slip faulting in the Cariboo Mountains and southern Omineca Crystalline Belt.

Plate 53. Photo looking northwest along the length of Isaac Lake and the strike of the Isaac Lake Fault. The fault dips at approximately 60° towards the northeast and kinematic indicators found along the shore of Isaac Lake plunge at 35-35° towards the southeast. The fault juxtaposes Kaza Group in the footwall against the Cunningham Formation in the hangingwall giving a dip separation of four kilometres. (Photo from G. Ross)

Plate 54. Slickenlines found in outcrop along the shore of Isaac Lake which plunge at an angle of 40° towards the southeast indicating a dextral-oblique sense of motion to the fault.

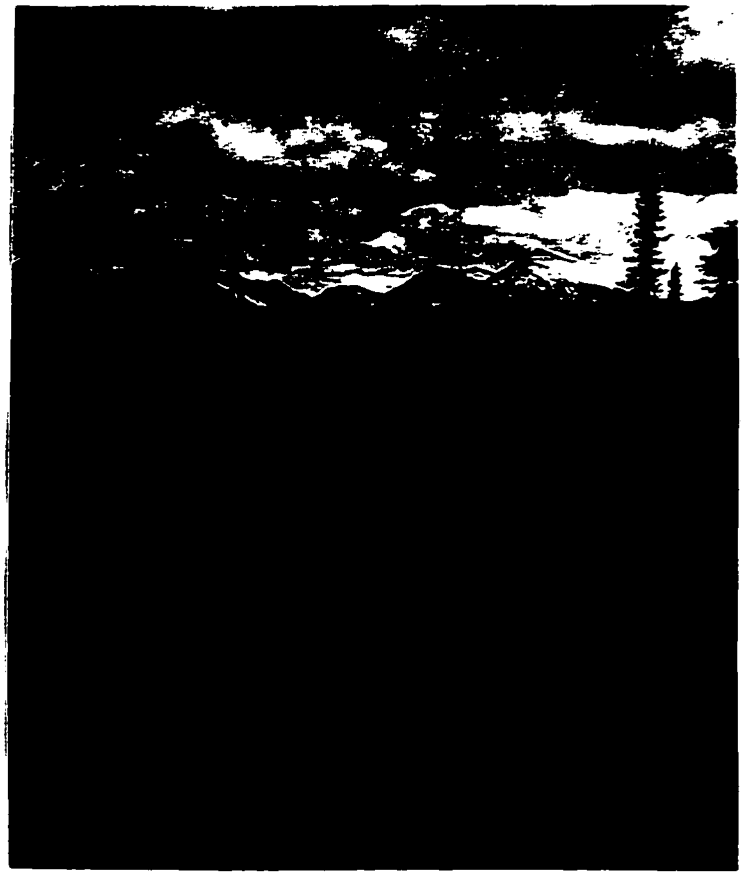


Plate 55. Photo looking northwest along the valley which contains Christian Lake (UTM 266500 E, 5860500 N) and the eastern splay of the Isaac Lake fault in the study area. The fault juxtaposes Kaza Group in the footwall against Isaac Formation in the hangingwall giving a dip separation of approximately 1000 m.

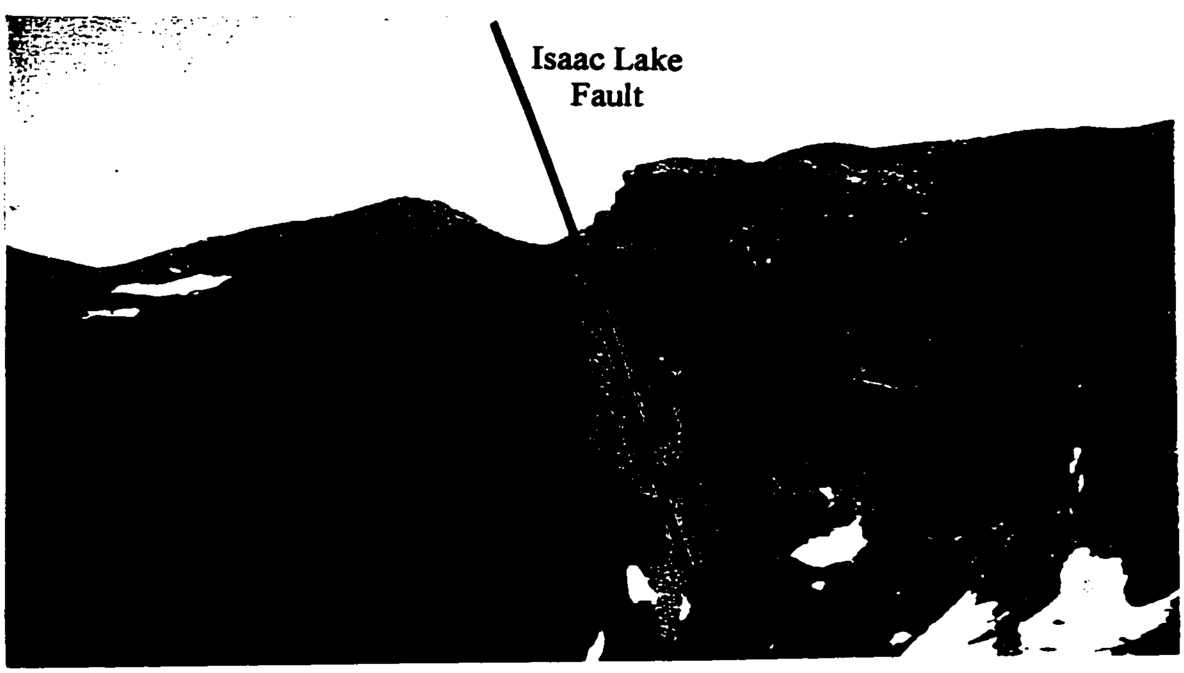


Plate 56. Photo looking northwest towards a notch that contains the eastern splay of the Isaac Lake fault (area iii). The fault juxtaposes middle Isaac carbonate in the footwall against the lower Cunningham Formation in the hangingwall giving a dip separation of approximately 1000 m.

Plate 57. Quartz filled tension fractures in Isaac Formation in the footwall of the Winder fault. Fractures strike northeast-southwest, and are vertical.

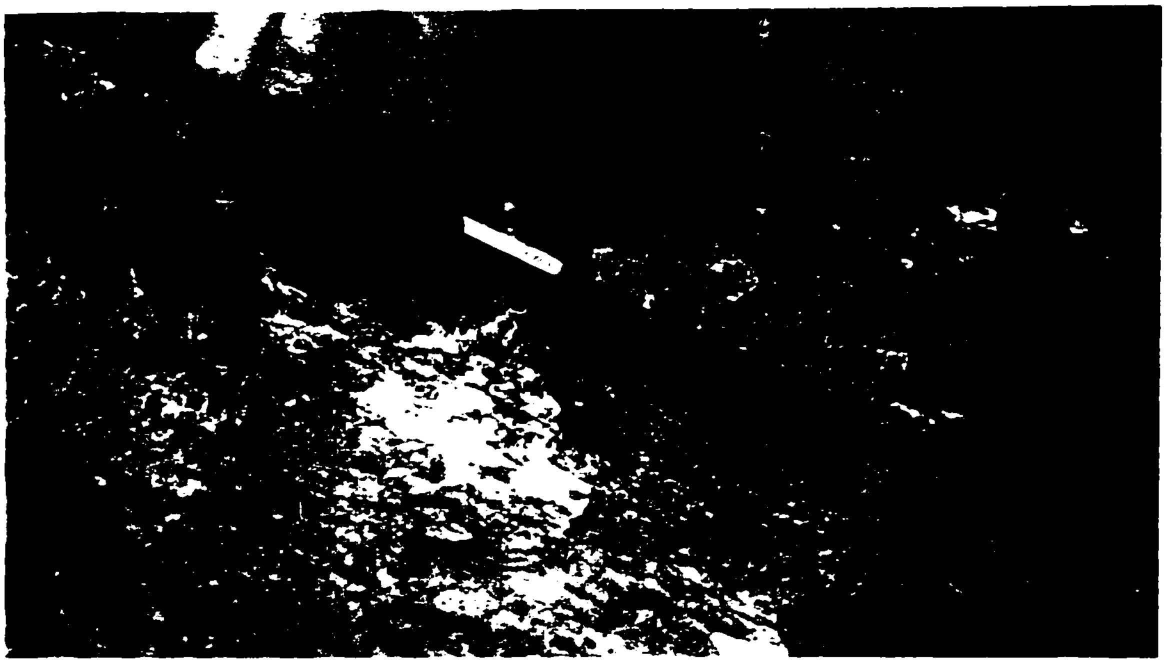


Plate 58. Photo of a small-scale dextral strike-slip fault with slickenlines that plunge at 15° towards the south. The fault strikes north-south, and displaces bedding by a few tens of centimetres.

CHAPTER 6: SUMMARY / CONCLUSIONS

The purpose of this chapter is to summarize the main conclusions of this study regarding the stratigraphic, metamorphic and structural evolution of the Isaac Lake Synclinorium. The regional significance of the Isaac Lake Synclinorium and its relationship to other study areas will then be discussed and the possibilities for future work presented.

6.1 Stratigraphy of the Isaac Lake Synclinorium

A five kilometre thick stratigraphic succession of the Neoproterozoic Windermere Supergroup is preserved within the Isaac Lake Synclinorium, which includes the upper Kaza Group and the formations of the Cariboo Group (Isaac, Cunningham, Yankee Belle, and Yanks Peak Formations). The Windermere Supergroup stratigraphy has been previously interpreted as a passive margin succession with stratigraphic units that range from deep marine basin and slope deposits to shallow marine platform deposits (Stewart, 1972; Ross, 1991). This interpretation is consistent with the Windermere Supergroup stratigraphy found in the Isaac Lake Synclinorium.

The Isaac Formation can be subdivided into four informal map units which include the lower Isaac Formation, lower Isaac carbonate unit, middle Isaac carbonate unit and the upper Isaac Formation. The Yankee Belle Formation can be divided into two map units which include the Amos Bowman and Betty Wendle members. The Betty Wendle Member represents a marine transgression, marked by a change from cyclic shallow water carbonates and sandstones of the Amos Bowman Member into the deep-water green and purple shales of the Betty Wendle Member (Lemieux, 1996). This distinct lithologic transition from the Amos Bowman Member to the Betty Wendle Member is also observed to the northeast, and may represent a potential regional marker unit for the upper Windermere Supergroup.

All of the contacts between the formations exposed in the study area are conformable with the exception of the Yankee Belle-Yanks Peak contact which is interpreted to be an unconformity. This interpretation is based on the abrupt change in facies which occurs between these two formations and evidence for stratigraphic cut-out. The lower Yanks Peak Formation lies in contact with different portions of the Betty Wendle Member of the Yankee Belle Formation in different parts of the field area. At the Castle Creek - Niagara Creek junction locality, the Yanks Peak is in contact with the lower portion of the Betty Wendle Member, where an abrupt change from the green and purple shales of the Yankee Belle Formation to quartz pebble conglomerates of the Yanks Peak Formation occurs. South of this area, at the western boundary of Wells Gray Provincial Park, clean quartz arenites of the Yanks Peak Formation are in contact with interbedded siltstones and sandstones of the upper Betty Wendle Member. Whether the Yankee Belle - Yanks Peak contact marks the Precambrian-Cambrian boundary remains enigmatic as no trace fossils were found in the lower portion of the Yanks Peak to indicate it is Cambrian in age.

The conformable relationship between the Kaza Group and Isaac Formation and the Isaac Formation and Cunningham Formation is also seen in the stratigraphy to the northeast (Ferguson, 1994). However, in the northeast the nature of the Cunningham Formation - Yankee Belle Formation contact is locally unconformable, and marked by the clean quartz sandstone of the Zig-zag Member (Ferguson, 1994). This relationship between the Cunningham Formation and Yankee Belle Formation was not observed in the study area.

The contrasting lithologies of the different formations in the study area form a structural multilayer which partitioned strain during deformation. The multilayer consists of a thick, lower ductile layer that includes the Kaza Group and Isaac Formation, and an upper brittle-ductile layer that includes the Cunningham, Yankee Belle and Yanks Peak Formations. The Cunningham Formation, a competent 500 m thick limestone unit, forms the base of the upper brittle-ductile layer.

6.2 Structural evolution of the Isaac Lake Synclinorium

The Isaac Lake Synclinorium is part of the suprastructure of the Cariboo Mountains as defined by Campbell (1970). The structural evolution of the Isaac Lake Synclinorium can be divided into three main deformation events (D_1, D_2 and D_3). The first event (D_1) formed metre to decametre-scale isoclinal folds, local east-verging thrust faults and a penetrative axial planar cleavage. The age of D_1 deformation has an upper age limit constrained by the 174 +/- 4 Ma Hobson pluton which cuts D_1 structures

(Pigage, 1977; Gerasimoff, 1988). Peak metamorphism (M_1) was attained during D_1 deformation which reached lower greenschist facies (chlorite zone) conditions, as chlorite, muscovite and chlorite-muscovite aggregates are found preferentially aligned parallel to the S_1 cleavage. The mineral assemblage in pelitic rocks consists of quartz + muscovite + chlorite + pyrite + albite. Chlorite-muscovite aggregates are also found throughout the pelitic and semi-pelitic units in the study area. The aggregates are composed predominantly of chlorite with thin muscovite and quartz interlayers. The aggregates are interpreted to have grown during M_1, as their basal cleavage plane (001) is oriented parallel to S_1. The aggregates were modified during D_2 deformation as they extended parallel to the S_2 cleavage planes and shortened perpendicular to the S_2 cleavage planes. It is possible that chlorite grew in these aggregates as they were extended during D_2 deformation, as some chlorite grains in the aggregates show sharp extinction and are not bent or broken. This implies that the growth of some of the chlorite and muscovite may have also occurred during D_2 deformation.

The D_2 event formed the main regional structures which includes the kilometre-scale anticlinoria and synclinoria found throughout the Cariboo Mountains. The F_2 folds are northwest-trending and southwest-verging with an associated axial planar cleavage that crenulates the earlier S_1 cleavage. The geometry of the F_2 folds depends on their position in the structural multilayer. F_2 folds in the lower layer are tight, overturned, southwest-verging folds, whereas F_2 folds in the upper layer are broad, open folds that have high-angle reverse faults associated with them. The S_2 crenulation cleavage associated with F_2 folds is found throughout the lower layer and locally in the upper layer where it is restricted to pelitic rocks of the Yankee Belle Formation.

Line length restorations on carbonate units in the lower and upper layers show that the lower layer accommodated a minimum shortening of 35-40%, whereas the upper layer accommodated a minimum shortening of only 8-10%. Pyrite pressure fringes also record similar disparities in strain between the lower and upper layers. Pyrite porphyroblasts in the lower layer have complex, deformed pressure fringes, whereas pyrite porphyroblasts in similar lithologies of the Yankee Belle Formation show only weak pressure fringe development. This indicates that the lower layer has undergone a more intense strain history than the upper layer. In order to account for this difference in shortening between the upper and lower layers, discrete detachment surfaces at the base of the upper layer (i.e., at the Isaac-Cunningham contact) are proposed as they have been locally observed. Such detachment zones in the upper Isaac Formation and at the base of the Cunningham Formation could have partitioned much of the strain in the form of shear while the overlying units were deformed by faulting with small displacements and broad, open folding giving the appearance of less intense deformation.

The D_2 structures were subsequently deformed by D_3 structures which include dextral-oblique strike-slip faults, southeast-northwest-striking folds (F_3) and northeast-striking normal faults. Dextral-oblique strike-slip faults bound the Isaac Lake Synclinorium and include the Isaac Lake, Winder, and Mitchell River faults. The Isaac Lake faults form the western boundary of the study area and consist of two parallel splays that are both dextral-oblique slip faults with an east-side-down (normal) component. The net-slip and strike-slip displacement along the Isaac Lake faults can only be estimated for the eastern fault, which is approximately 1750 m and 1350 m respectively. The eastern boundary is formed by the Winder Fault, also a dextral-oblique strike-slip fault. It dips steeply to the east with an east-side-up (thrust) component, and has a net-slip displacement of 1400 m and strike-slip displacement of 1100 m. The angular relationship between the dextral strike-slip faults, F_3 folds and the northeast-striking normal faults is that predicted for a right lateral shear couple. Therefore, these structures are interpreted to be co-genetic D_3 structures formed as a result of regional, dextral strike-slip faulting. The presence of dextral-strike-slip fault-related structures throughout the synclinorium indicates that strain was not only taken up by displacement along the faults, but distributed throughout the fault block as a set of structures. Fold structures that formed as a result of dextral strike-slip faulting within the fault block were easily accommodated in the Isaac Lake Synclinorium due to the sub-horizontal orientation of the stratigraphy prior to faulting.

The D_3 structures are deformed by local D_4 structures. These structures include kink folds and a widely spaced centimetre-scale crenulation cleavage. D_4 structures are the youngest structures found in the Isaac Lake Synclinorium.

6.3 Regional significance of the Isaac Lake Synclinorium

The Cariboo Mountains have a deformational history spanning 120 Ma from the Jurassic to the Eocene. The main regional structures can be subdivided into a series of northwest-trending anticlinoria and synclinoria that include from east to west: the Premier Anticlinorium, Isaac Lake Synclinorium, Lanezi Arch, Black Stuart

Synclinorium and the Lightening Creek Anticlinorium. An overall northwest plunge of these structures results in the southward transition to deeper structural levels and a change in structural style. This transition is referred to as the suprastructure-infrastructure transition, and provides a unique opportunity to examine the changes that occur from shallow to deep structural levels in the Omineca Crystalline Belt.

At shallow structural levels (suprastructure), the earliest structures (D_1) include local east-verging thrust faults and isoclinal folds. In the infrastructure these structures are not found, instead the earliest structures (D_1) include kilometre-scale west-verging isoclinal folds. In the intermediate zone (transition zone) between the suprastructure and infrastructure, the east verging structures (locally known as D_1) of the suprastructure are deformed by the west-verging structures of the infrastructure, also locally known as D_1. The localizing of the D_1 structures in the suprastructure is most likely the result of the east-verging obduction of the Intermontane Superterrane onto the western margin of North America which led to east-verging shear in the very shallow levels of the suprastructure.

Following D_1, D_2 affected both the suprastructure and infrastructure to form the main regional D_2 structures and large-scale D_2 fan structures across the Cariboo Mountains. Folds are northeast-verging in the eastern part of the Cariboo mountains and become southwest-verging in the western part of the Cariboo Mountains (Ferguson, 1994; Currie, 1988; Campbell, 1970). The geometry of the D_2 structures changes with increasing depth and metamorphic grade (i.e. from the suprastructure to the infrastructure), but the extent to which the geometry changes requires further investigation.

The timing of the formation of D_2 structures shows an overall younging from west to east in the Cariboo Mountains. D_2 structures of the suprastructure formed pre-174 Ma based on the cross-cutting relationship of the Hobson pluton. The age of formation of D_2 structures in the infrastructure occurred between 160 Ma and 131 Ma based on the ages of deformed and undeformed pegmatites respectively. (Currie, 1988). The timing of peak metamorphism with relation to the D_2 structures changes with structural depth and metamorphic grade. In the suprastructure, peak metamorphism is pre-D_2 (pre 174 Ma), in the transition zone, peak metamorphism is syn- to post-D_2, and in the infrastructure, peak metamorphism is post D_2 (135 Ma). One of the major questions about metamorphism in the Cariboo Mountains is whether or not there are two separate metamorphic events, one at pre- 174 Ma and one at 135 Ma, or one prolonged metamorphic event which ceased before D_2 at shallow structural levels and outlasted D_2 at deep structural levels. At deep structural levels in the Cariboo Mountains there is evidence for a greenschist facies metamorphic event that occurred during D_1 deformation, as S_1 is defined by the parallel alignment of chlorite and muscovite (Cooley, pers. comm., 1996, Walker, 1989; Currie, 1988; Pell, 1984). Until there are better geochronological age constraints, it will remain unclear whether this metamorphism is separate or marks the beginning of M_1.

There are several problems when making regional correlations of D_2 structures. First, the main folds (F_1, F_2 and F_3) are near co-axial in certain parts of the Cariboo Mountains making them difficult to distinguish one from the other. Second, depending on depth and lithology, the folds change geometry. Third, the timing of peak metamorphism and the formation of the F2 folds changes from the suprastructure to the infrastructure. Finally, there are significant gaps between study areas, thus making correlations between study areas difficult.

Although there are inherent differences between the structural evolution of the suprastructure and infrastructure of the Cariboo Mountains, there are remarkable similarities of the suprastructure of the Cariboo Mountains to other areas in the southern Omineca Crystalline Belt that contain low-grade metamorphic rocks deformed at shallow structural levels. For example, the Selkirk Mountains south of the Cariboo Mountains contain large tracts of lower-greenschist facies rocks. These rocks reached peak metamorphism in the early to mid-Jurassic (during D_1) which formed a pervasive cleavage sub-parallel to S_0. The D_2 structures include upright to overturned northwest-striking, northeast-verging folds in the east, and southwest-verging folds in the west, which is similar in geometry to the D_2 structures of the Cariboo Mountains. The rocks deformed at shallow structural levels in the southern Omineca Crystalline Belt share a very similar structural and metamorphic history which occurred in Jurassic time.

The relationship of low-grade rocks of the Omineca Crystalline Belt to low-grade rock of the western Main Ranges of the Rocky Mountains is not as clear. Rocks of the western Main Ranges have a similar lower-greenschist facies mineral assemblage, however, the structures are dominated by east-verging thrust faults and folds. The age of formation of the regional structures and peak metamorphism of the western Main Ranges is inferred to have occurred between the late-Jurassic and early-Cretaceous (Carey, 1984). This is clearly younger than the age of formation of the main regional structures in the suprastructure of the Cariboo Mountains. The link between structures of the Cariboo Mountains to structures of the western Main Ranges remains unresolved.

In the Cariboo Mountains, the main regional (D_2) structures are clearly cross-cut by the post-metamorphic northwest-striking dextral strike-slip faults and refolded by the associated F_3 folds. Major dextral strike-slip faults in the Cariboo Mountains include the Isaac Lake, Matthew, Amos Bowman, Betty Wendle, Knutson, Willow River, Stony River, Narrow Lake and Pinchi faults. The entire system is located southwest of the southern termination of the Northern Rocky Mountain Trench fault. Recent geochronological work on the Matthew fault has shown that dextral strike-slip faulting was initiated by the mid-Cretaceous, although a minimum age constraint on the fault was not obtained. The age of strike-slip faulting (D_3) in the Cariboo Mountains is therefore inferred to have been initiated by the mid-Cretaceous based on the age of the Matthew fault. The Isaac Lake and Matthew faults can be traced for over 100 kilometres from the northern tip of Isaac Lake, southward to Hobson Lake. These faults also lie along strike with the eastern and western splays of the McLeod Lake fault, a subsidiary fault off the Northern Rocky Mountain Trench fault. The Isaac Lake and Matthew faults are interpreted to be continuations of the McLeod Lake fault and accommodate some of the dextral motion of the Northern Rocky Mountain Trench fault. This would make the Cariboo Mountain dextral strike-slip fault system part of the southern extension of the Northern Rocky Mountain Trench Fault which was active during the mid-Cretaceous.

6.4 Future work

Over the last fifteen years in the Cariboo Mountains, a number of studies have been undertaken that have increased the understanding of the geometry of the structures, timing of metamorphic events, and the relationship of the shallow structural levels to the deeper structural levels. Two of the major problems with linking these different structural levels is the change in geometry of the folds and the lack of geochronological age constraints. Although most of the Cariboo Mountains have been mapped, gaps occur between most of the study areas and they cannot be linked together without uncertainty. The Isaac Lake Synclinorium demonstrates how structures change their geometry along strike depending on lithology and structural depth, and therefore, in order to make accurate correlations of structures at increasing structural depth, they must be carefully traced along strike so as not to be confused with other structures. This could be achieved by following major structures along strike mapped from the suprastructure level to the infrastructure level. Geochronological age constraints on individual structures is also required in order to correlate from one area to another accurately, as well as re-examining pluton and country rock relationships that remain debatable (i.e. Hobson pluton).

The relationship of metamorphic events in the suprastructure and infrastructure also remains an outstanding problem, that needs to be better constrained with radiometric ages of metamorphic minerals. This is particularly important in the transition zone which only has relative age constraints on the timing of metamorphism and is the critical link between the suprastructure and infrastructure.

The relationship between the suprastructure of the Cariboo Mountains and the western Main Ranges remains enigmatic. The presence of chlorite-muscovite aggregates in both of these areas with similar relationships to the main regional cleavages may be the common link between these two areas. At present, there are no geochronological age constraints on the timing of formation of the aggregates in either of these areas, which will be necessary in order to interpret this relationship further.

Regionally, the recognition of dextral strike-slip deformation in the Cariboo Mountains is becoming increasingly significant, but it is still in the preliminary stages of study. Tracing the Isaac Lake and Matthew faults northward to determine their relationship with the McLeod Lake fault as well as obtaining better age constraints on the McLeod Lake fault will be important in determining the significance of strike-slip faulting in the Cariboo Mountains. A re-examination of strike-slip faults in the Cariboo Mountains is also necessary in order to determine the magnitude of the cumulative dextral displacement accommodated by this fault system.

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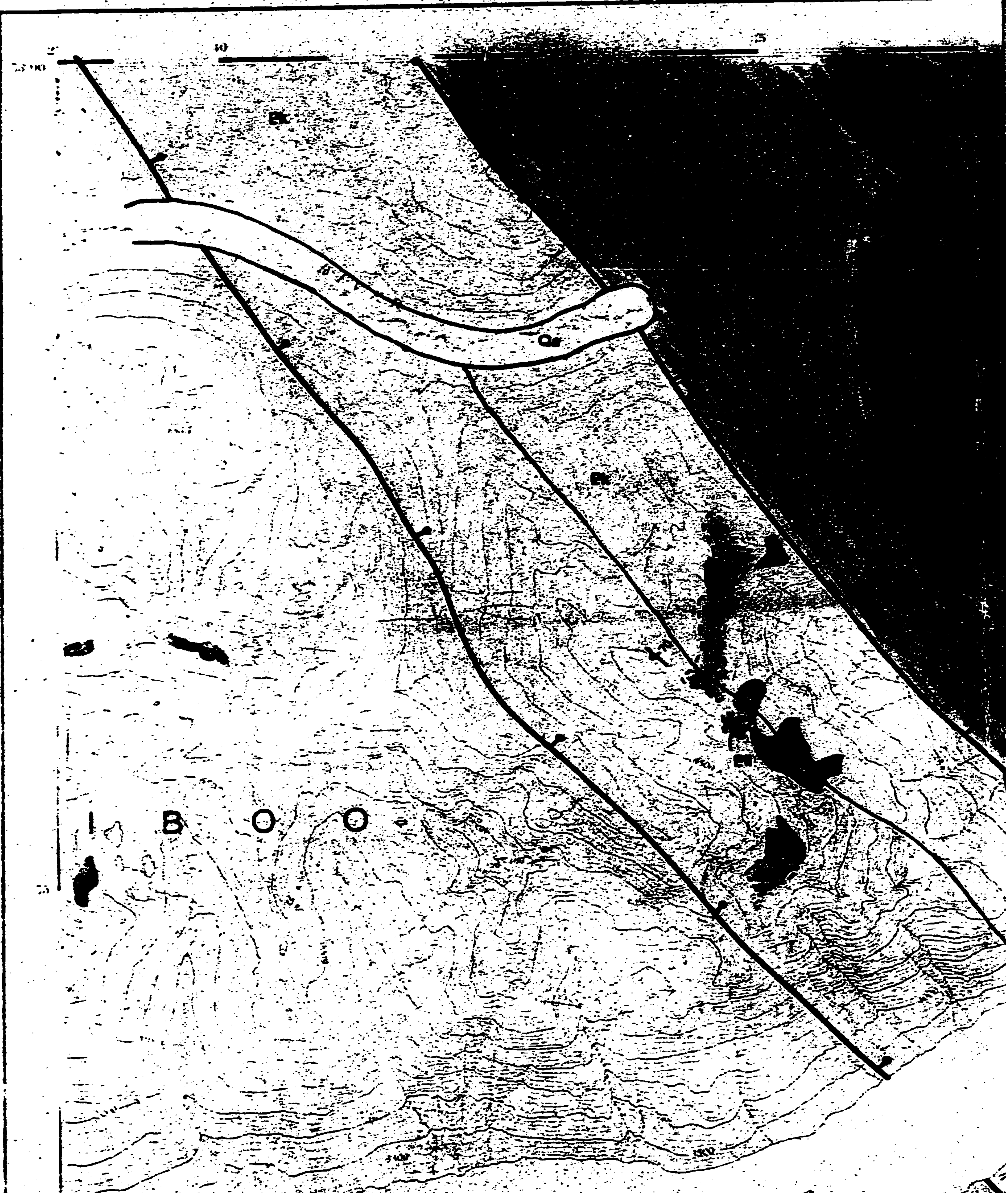
Young, F.G., 1979, **The lowermost Paleozoic McNaughton Formation and equivalent Cariboo Group of eastern British Columbia: piedmont and tidal complex.** Geological Survey of Canada Bulletin 288, 59 pp.

EDITION 2



190 05 297000m. E. 120°00'

LEGEND

QUATERNARY

Qa — Till, alluvium

LOWER CAMBRIAN
WINDERMERE SUPERGROUP
CARIBOO GROUP

Midas Formation
shale, siltstone, minor sandstone

Єyp — **Yanks Peak Formation**
quartzite, quartzose pebble conglomerate

NEOPROTEROZOIC
WINDERMERE SUPERGROUP
CARIBOO GROUP

Betty Wendle member
shale, siltstone, stromatolitic boundstone, and sandstone

Amos Bowman member
shale, siltstone, sandstone and limestone

Єyb — **Yankee Belle Formation**
undifferentiated

Cunningham Formation
massive limestone, peloidal grainstone, minor shale

upper Isaac formation
shale, minor limestone, siltstone

middle Isaac carbonate
limestone, conglomerate, siltstone and shale

lower Isaac formation
shale, siltstone, minor sandstone and grit, lower isaac carbonate unit (PIc)

Isaac Formation
undifferentiated

Kaza Group (uppermost)

70

55

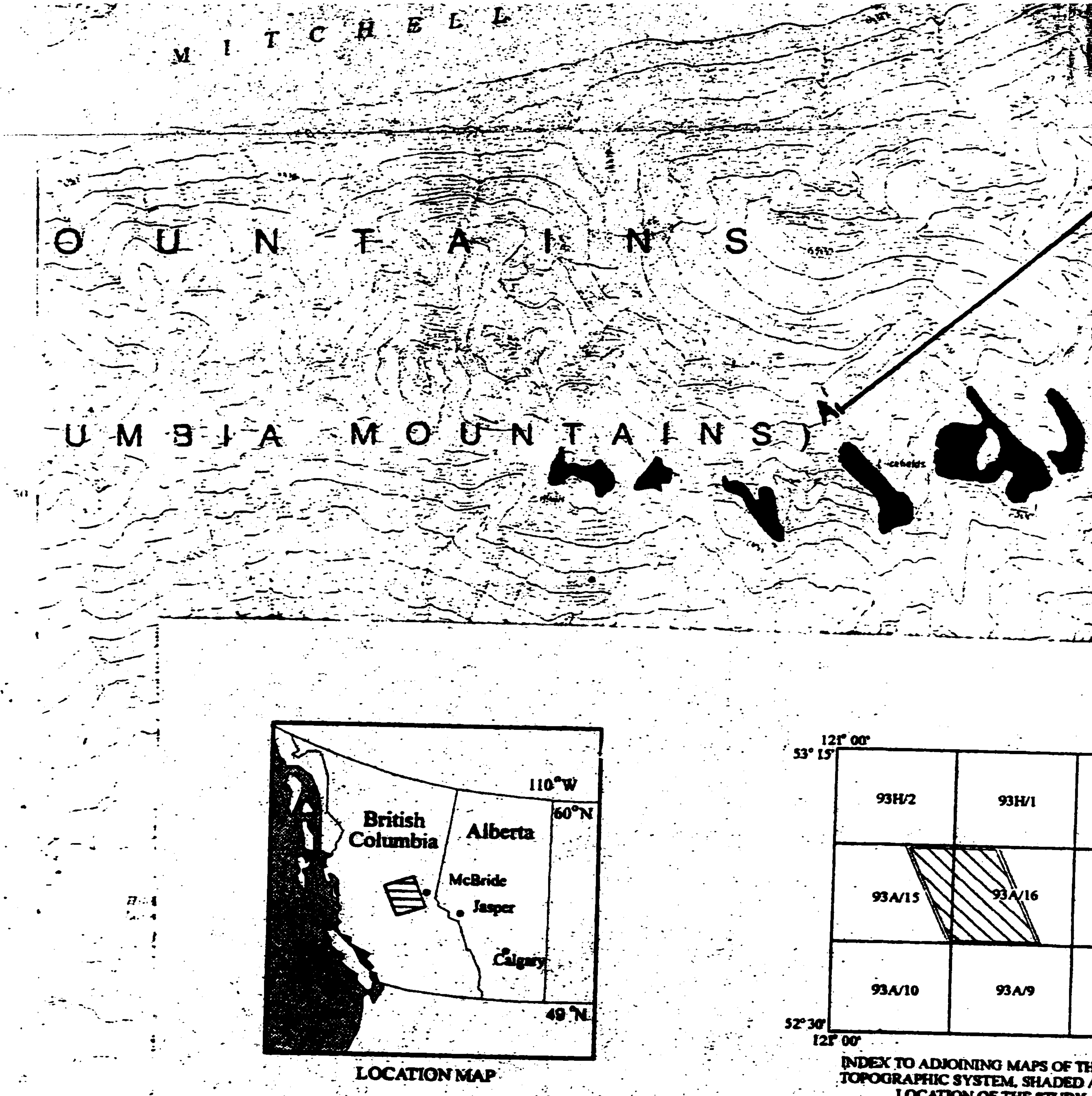
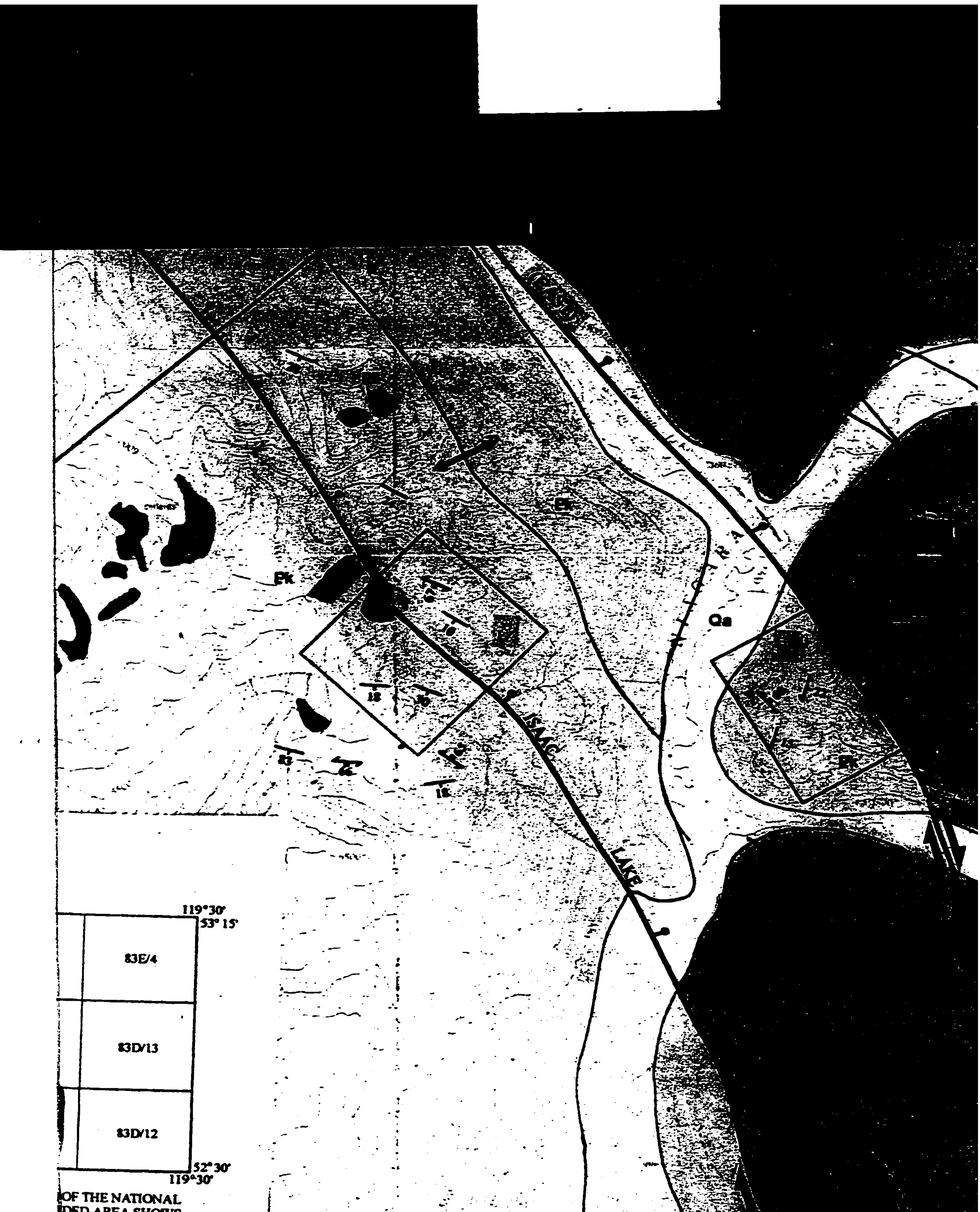
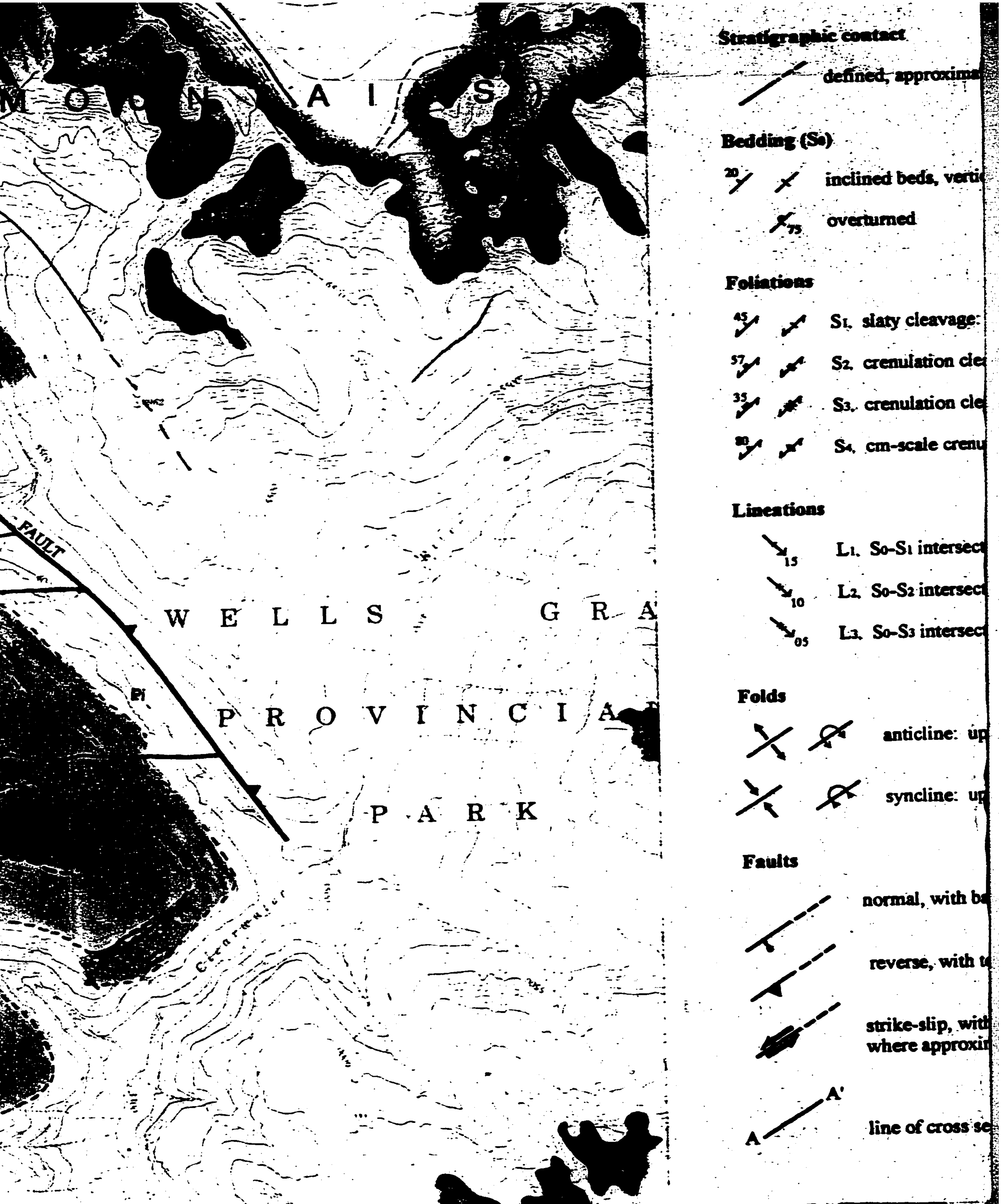
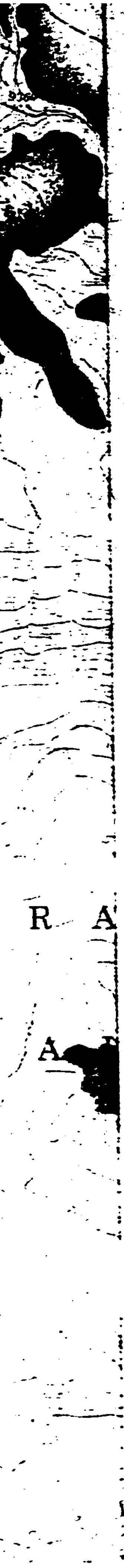


Figure 1-5. Geological map of the Isaac Lake Synclinorium


Ek
Ek
Qa
ISAAC
LAKE
18
30
83
18
119°30'
53° 15'
83E/4
83D/13
83D/12
52°30'
119°30'
OF THE NATIONAL


M O N A I S
FAULT
W E L L S G R A
P R O V I N C I A
P A R K
Stratigraphic contact
defined, approxima
Bedding (So)
inclined beds, verti
overturned
Foliations
S1. slaty cleavage:
S2. crenulation cle
S3. crenulation cle
S4. cm-scale crenu
Lineations
L1. So-S1 intersect
L2. So-S2 intersect
L3. So-S3 intersect
Folds
anticline: up
syncline: up
Faults
normal, with ba
reverse, with te
strike-slip, with
where approxim
A
A'
line of cross se



MAP SYMBOLS

Stratigraphic contact

defined, approximate

Bedding (S_0)

20 inclined beds, vertical

75 overturned

Foliations

45 S_1. slaty cleavage: inclined, vertical

57 S_2. crenulation cleavage or slaty cleavage: inclined, vertical

35 S_3. crenulation cleavage, inclined, vertical

80 S_4. cm-scale crenulation cleavage, inclined, vertical

Lineations

15 L_1. S_0-S_1 intersections

10 L_2. S_0-S_2 intersections, crenulations, F_2 fold axes

05 L_3. S_0-S_3 intersections, crenulations, F_3 fold axes

Folds

anticline: upright, overturned

syncline: upright, overturned

Faults

normal, with ball in hangingwall, dashed where approximate

reverse, with teeth in hangingwall, dashed where approximate

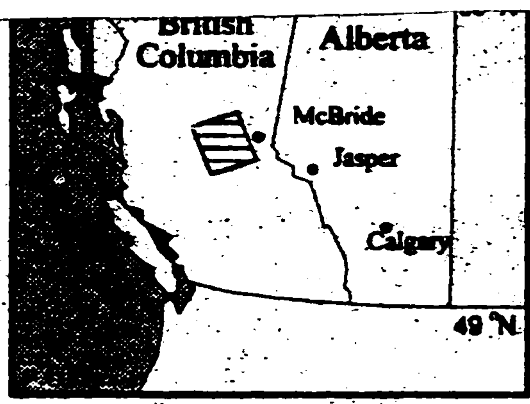
strike-slip, with arrows indicating direction of motion, dashed where approximate

A'

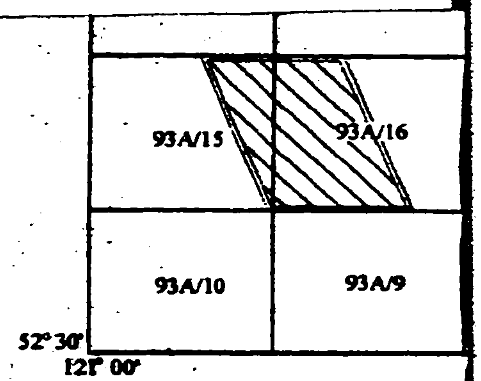
60

50

5850000m.N.



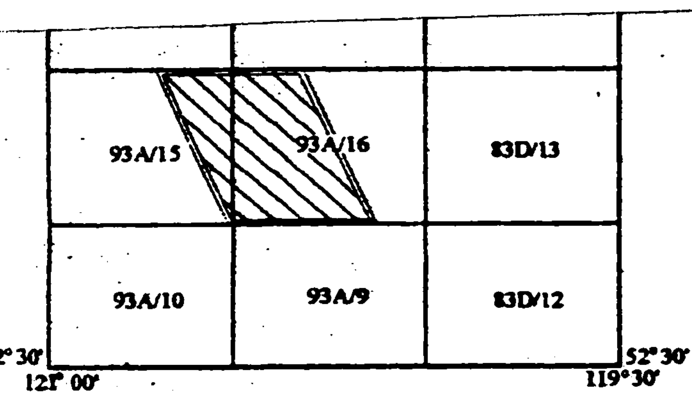
LOCATION MAP



INDEX TO ADJOINING MAPS OF
TOPOGRAPHIC SYSTEM, SHADE
LOCATION OF THE STU

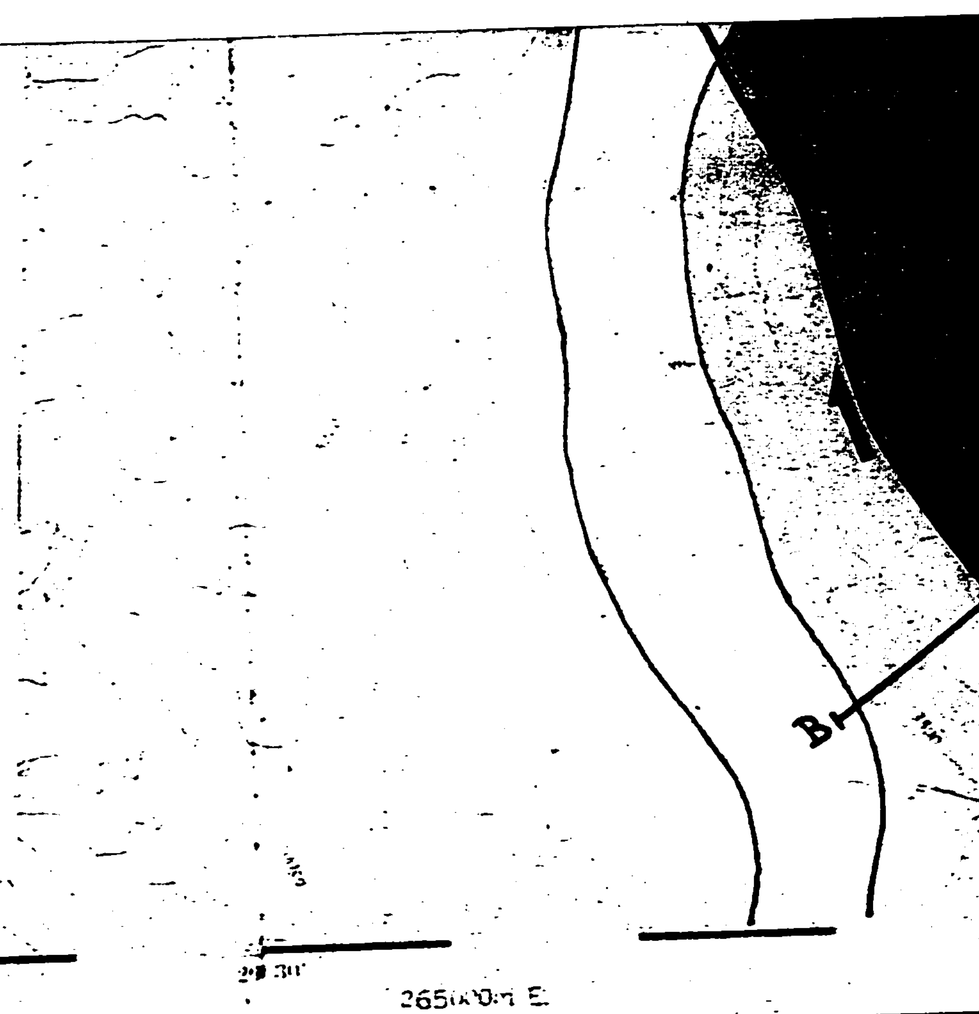
Figure 1-5. Geological map of the Isaac Lake Synclinori

MITCHELL LAKE
CARIBOO LAND DISTRICT
BRITISH COLUMBIA



INDEX TO ADJOINING MAPS OF THE NATIONAL TOPOGRAPHIC SYSTEM, SHADED AREA SHOWS LOCATION OF THE STUDY AREA

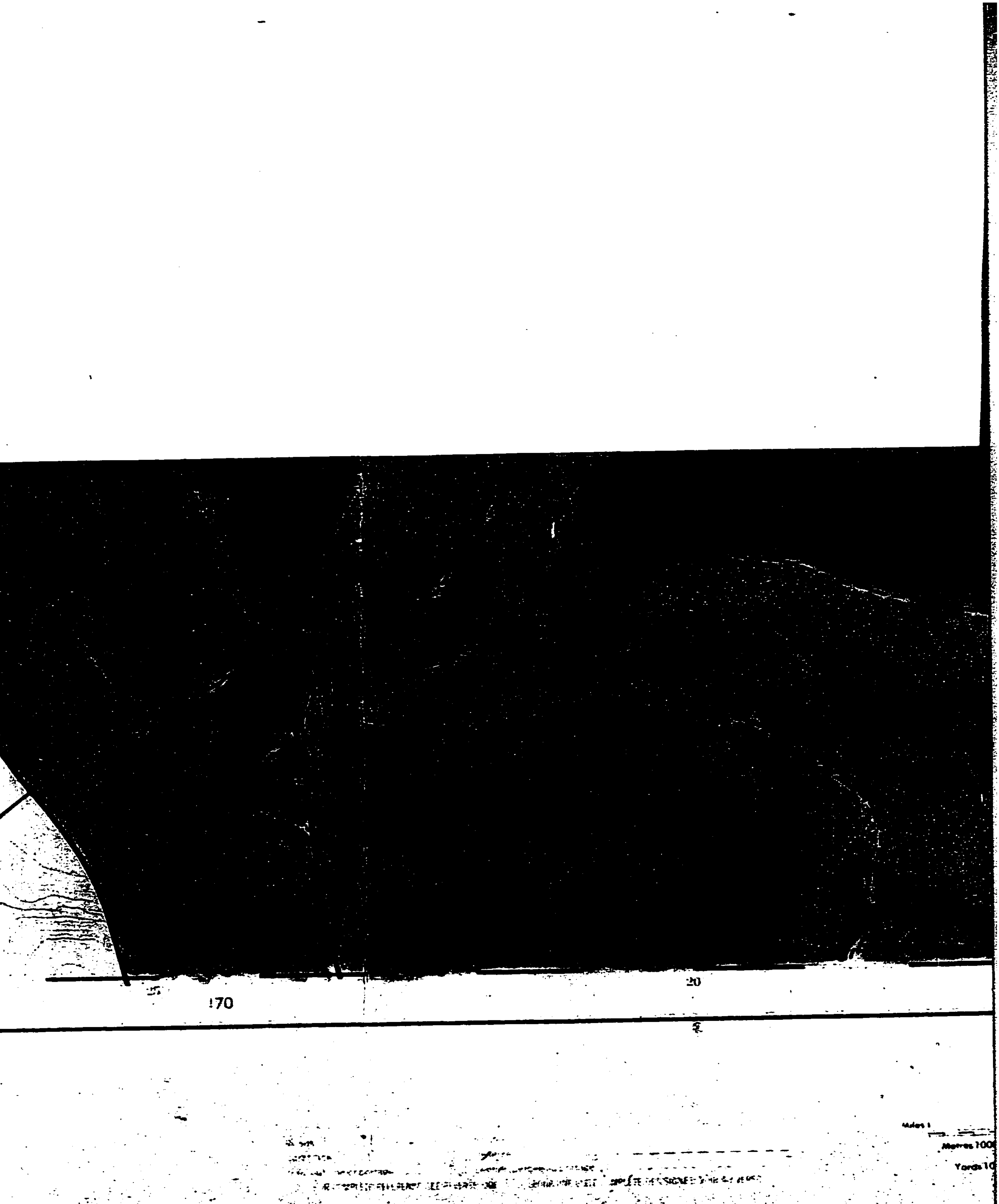
e Isaac Lake Synclinorium



LAKE

STRICT

MBIA


170
20
Miles
Metres 100
Yards 10

MOUNT WINDER

BRITISH COLUMBIA

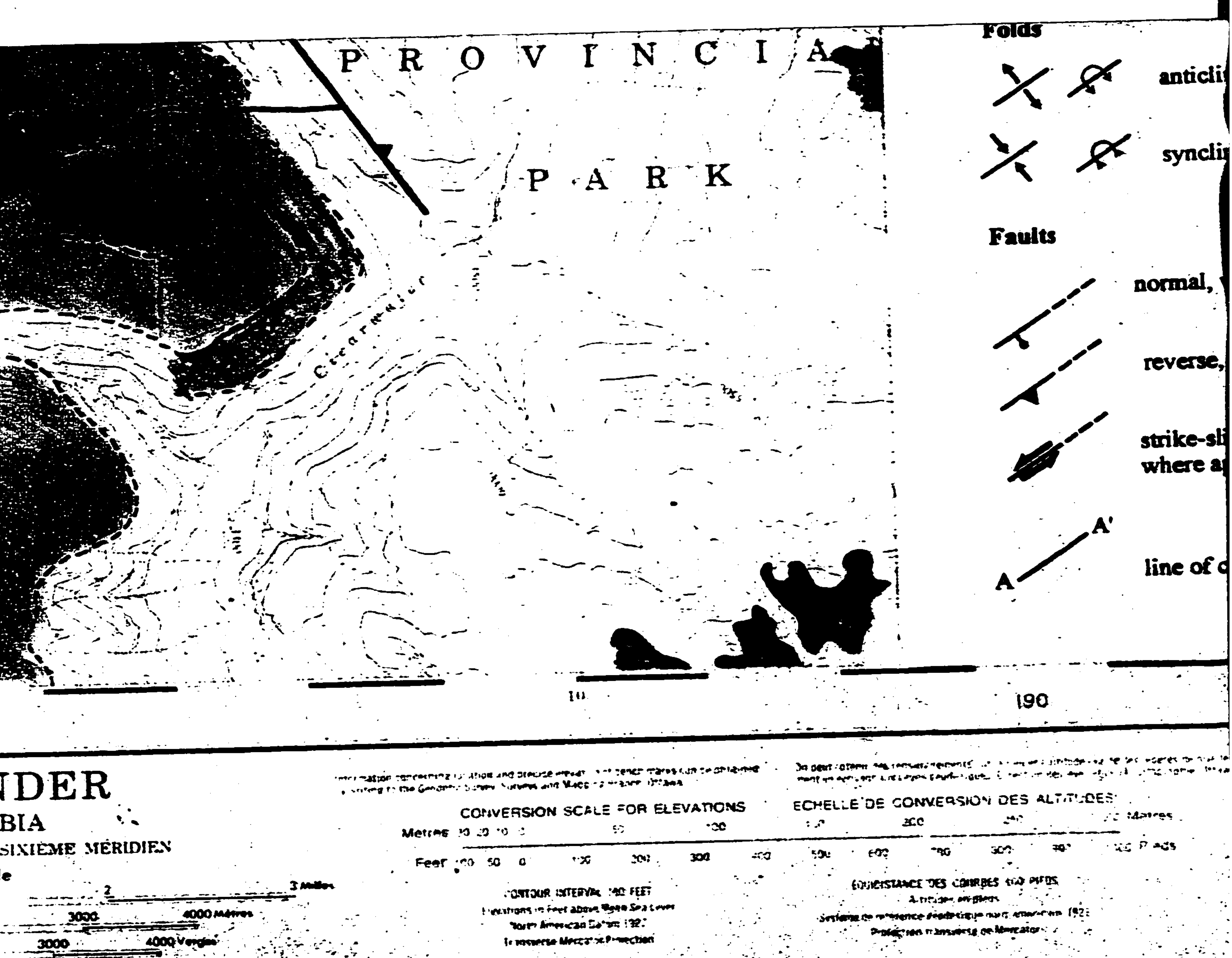
WEST OF SIXTH MERIDIAN-OUEST DU SIXIEME MERIDIEN

Scale 1:50 000 Échelle

Miles 1 0 2 3 Miles

Mètres 1000 0 1000 2000 3000 4000 Mètres

Yards 1000 0 1000 2000 3000 4000 Verges



P R O V I N C I A
P A R K
Folds
anticli
syncli
Faults
normal,
reverse,
strike-sli
where a
A
A'
line of c
190
NDER
BIA
SIXIÈME MÉRIDIEN
3 Milles
4000 Mètres
4000 Verges
CONVERSION SCALE FOR ELEVATIONS
ECHELLE DE CONVERSION DES ALTITUDES
Metres
Feet
Pieds
North American Datum 1927

Folds

anticline: upright, overturned

syncline: upright, overturned

Faults

normal, with ball in hangingwall, dashed where approximate

reverse, with teeth in hangingwall, dashed where approximate

strike-slip, with arrows indicating direction of motion, dashed where approximate

A A'

line of cross section

190

296000m. E

5850000m. N

ECHELLE DE CONVERSION DES ALTITUDES

Mètres

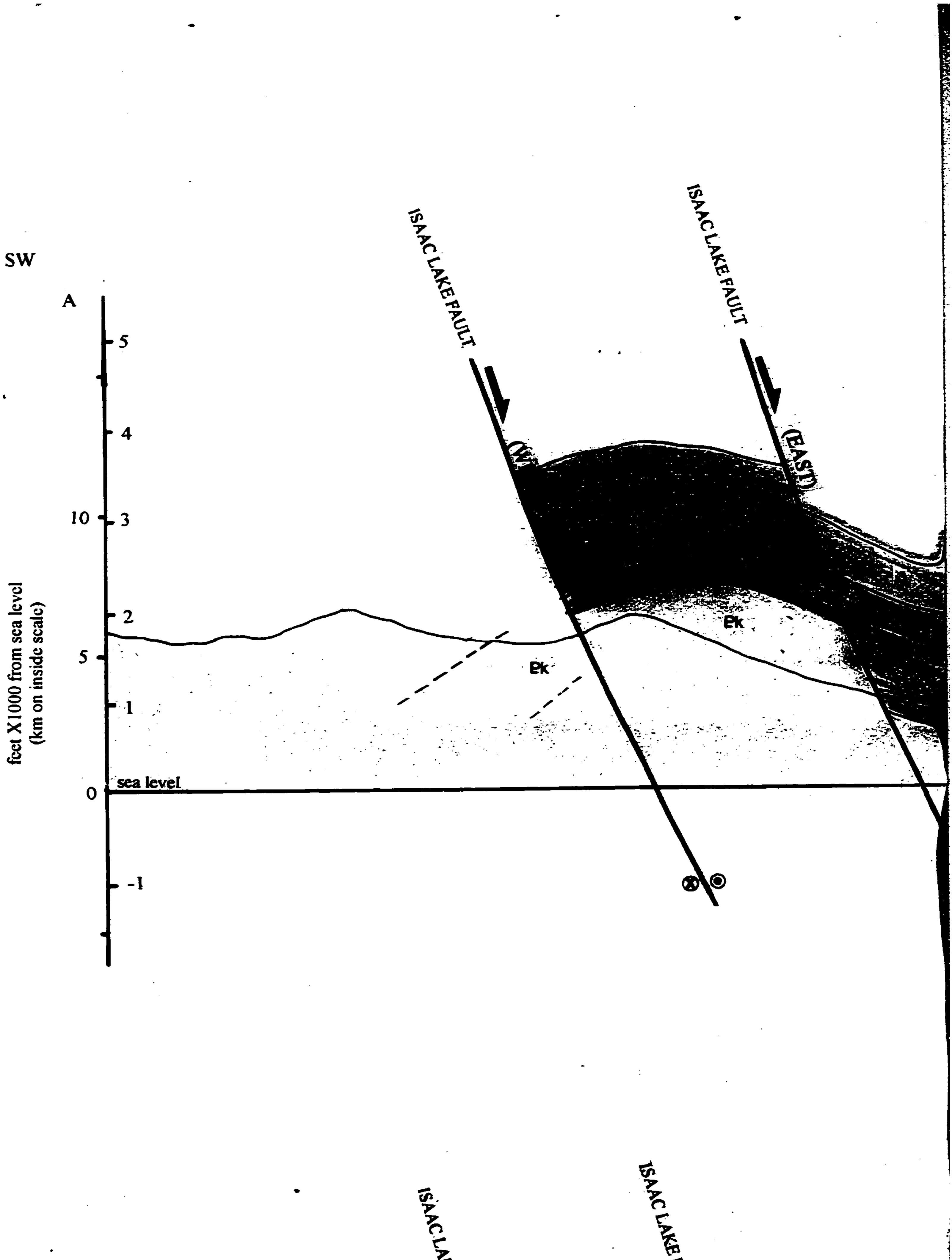
Pieds

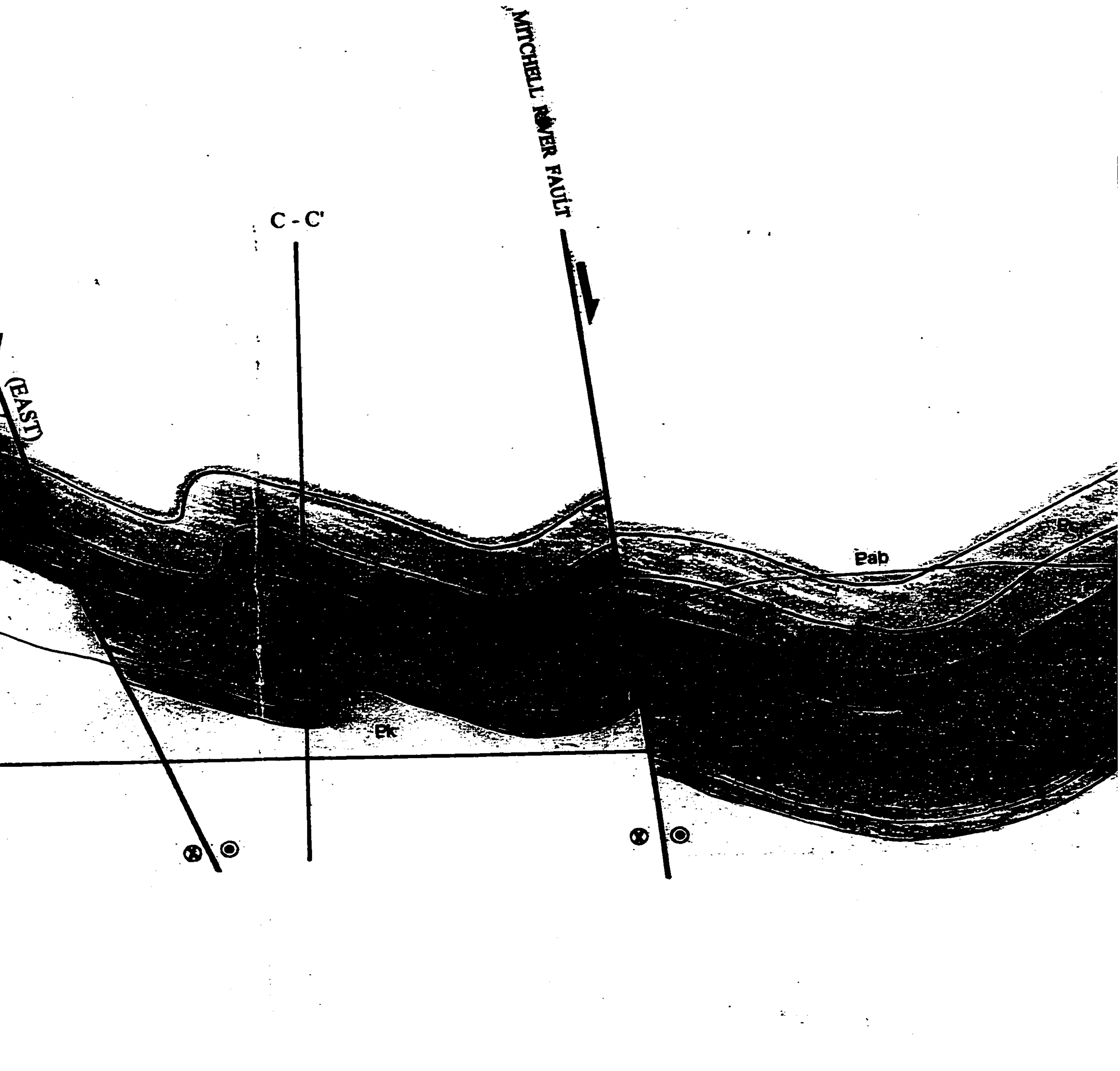
ÉQUIDISTANCE DES COURBES 100 PIEDS

Altitudes en pieds

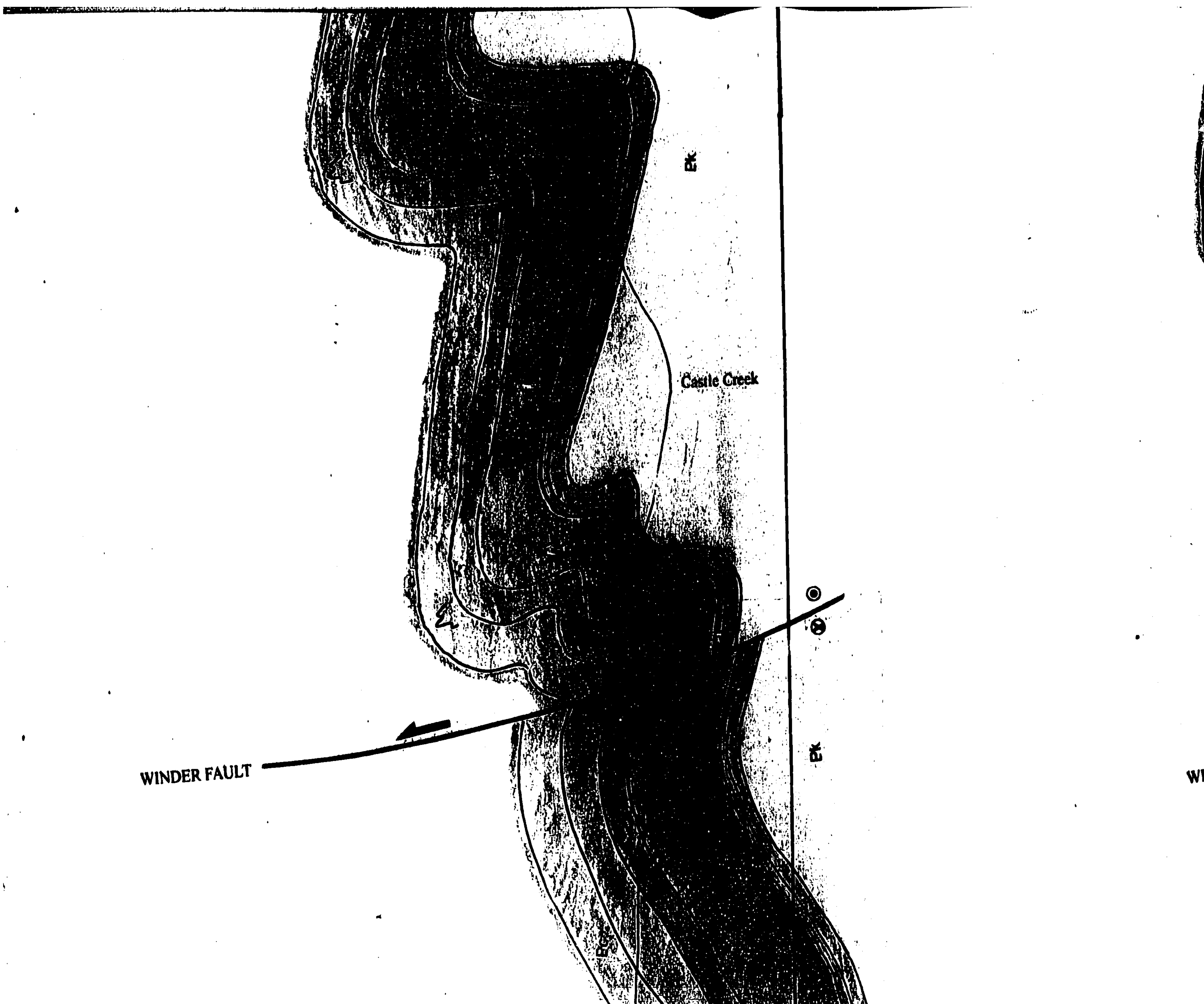
Projection transverse de Mercator

Dressé par la DIRECTION DES LEVÉS ET DE LA CARTOGRAPHIE
MINISTÈRE DE L'ÉNERGIE, DES MINES ET DES RESSOURCES


SW
A
feet X1000 from sea level
(km on inside scale)
sea level
ISAAC LAKE FAULT
ISAAC LAKE FAULT
(EAST)
Ek
Ek



MITCHELL RIVER FAULT
C - C'
(EAST)
Pab
Pk


WINDER FAULT
Castle Creek
Pk
Pk

NE

A'

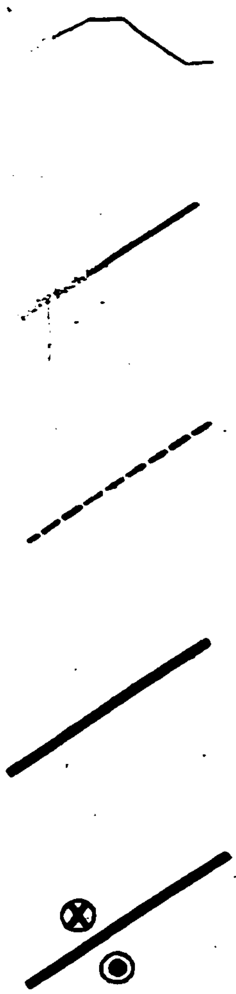
10

5

0

-5

LEGEND



topographic profile

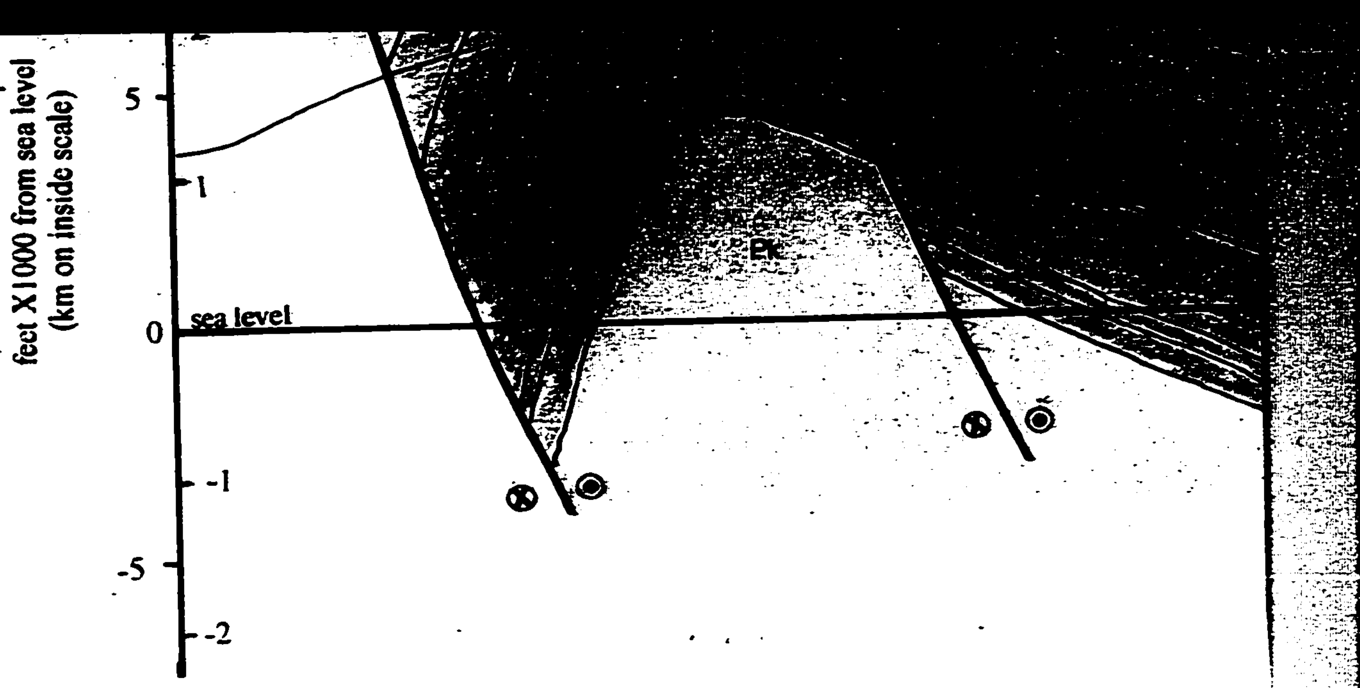
stratigraphic contact

trace of stratigraphic bedding

fault

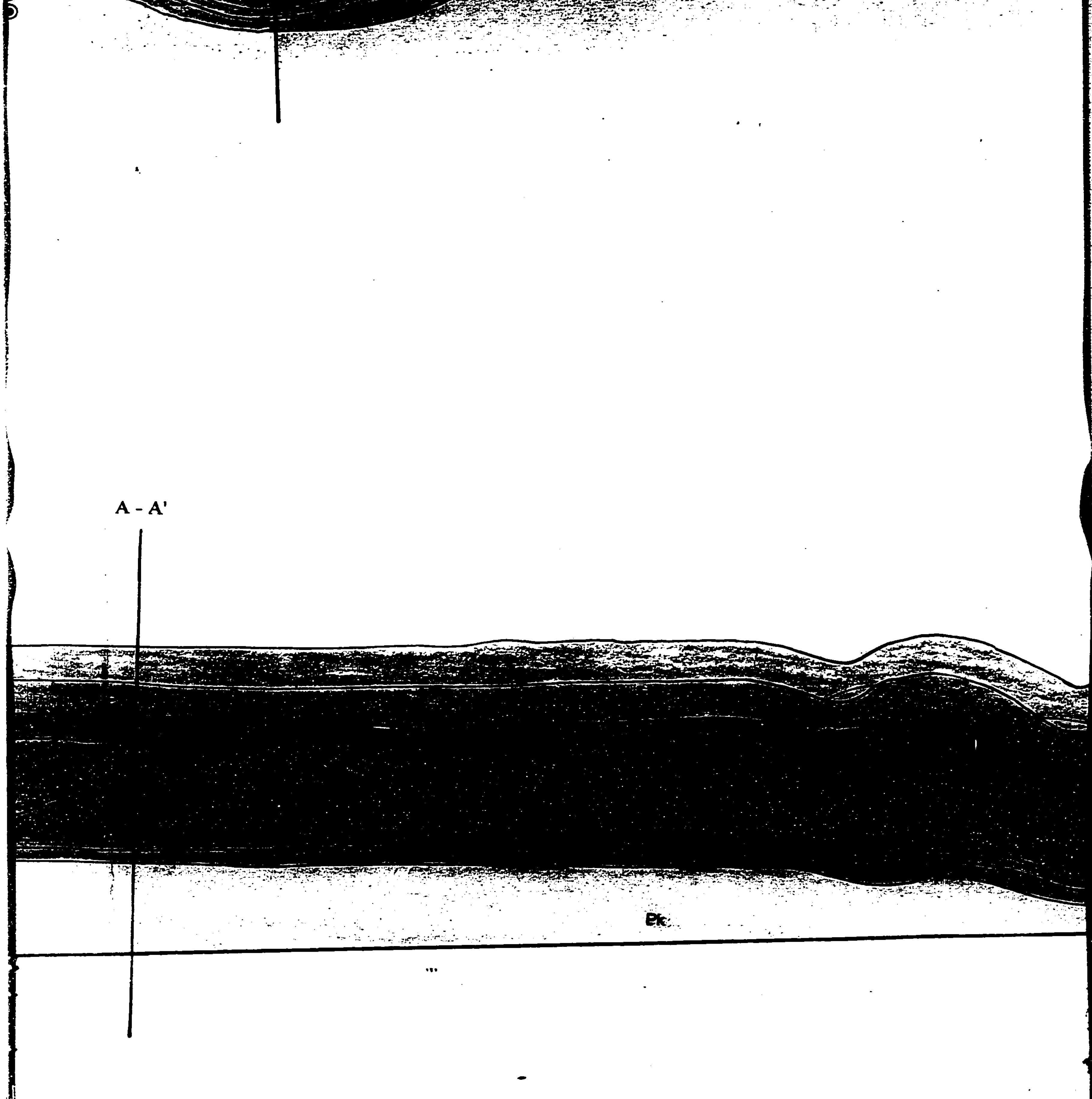
fault with strike-slip component
⊗ = motion away from reader,
◉ = motion towards reader

* see legend with Figure 1-5 for stratigraphic descriptions

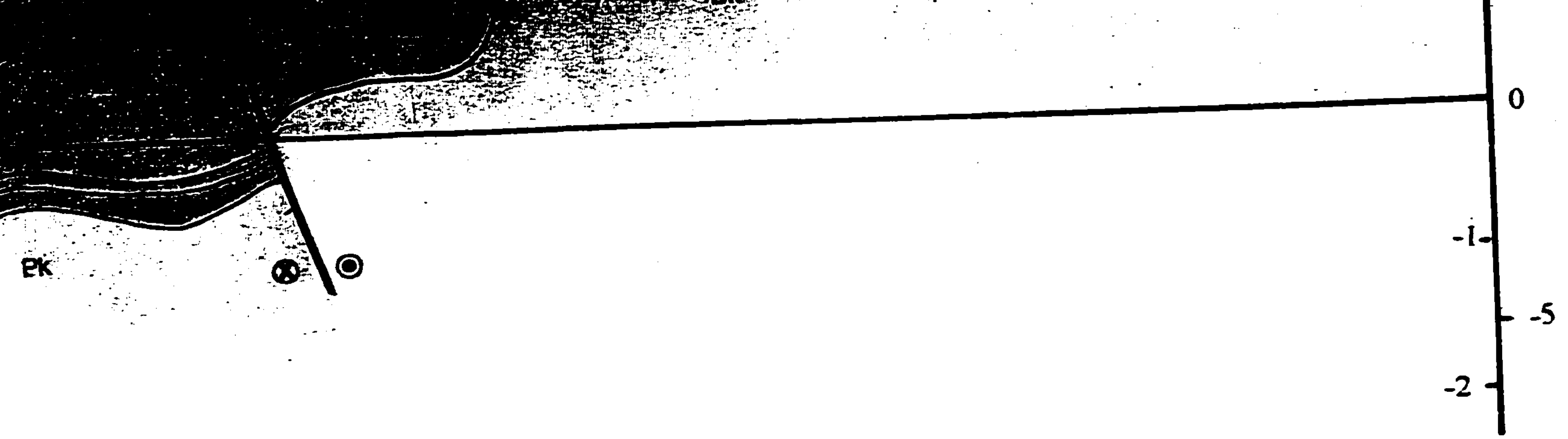


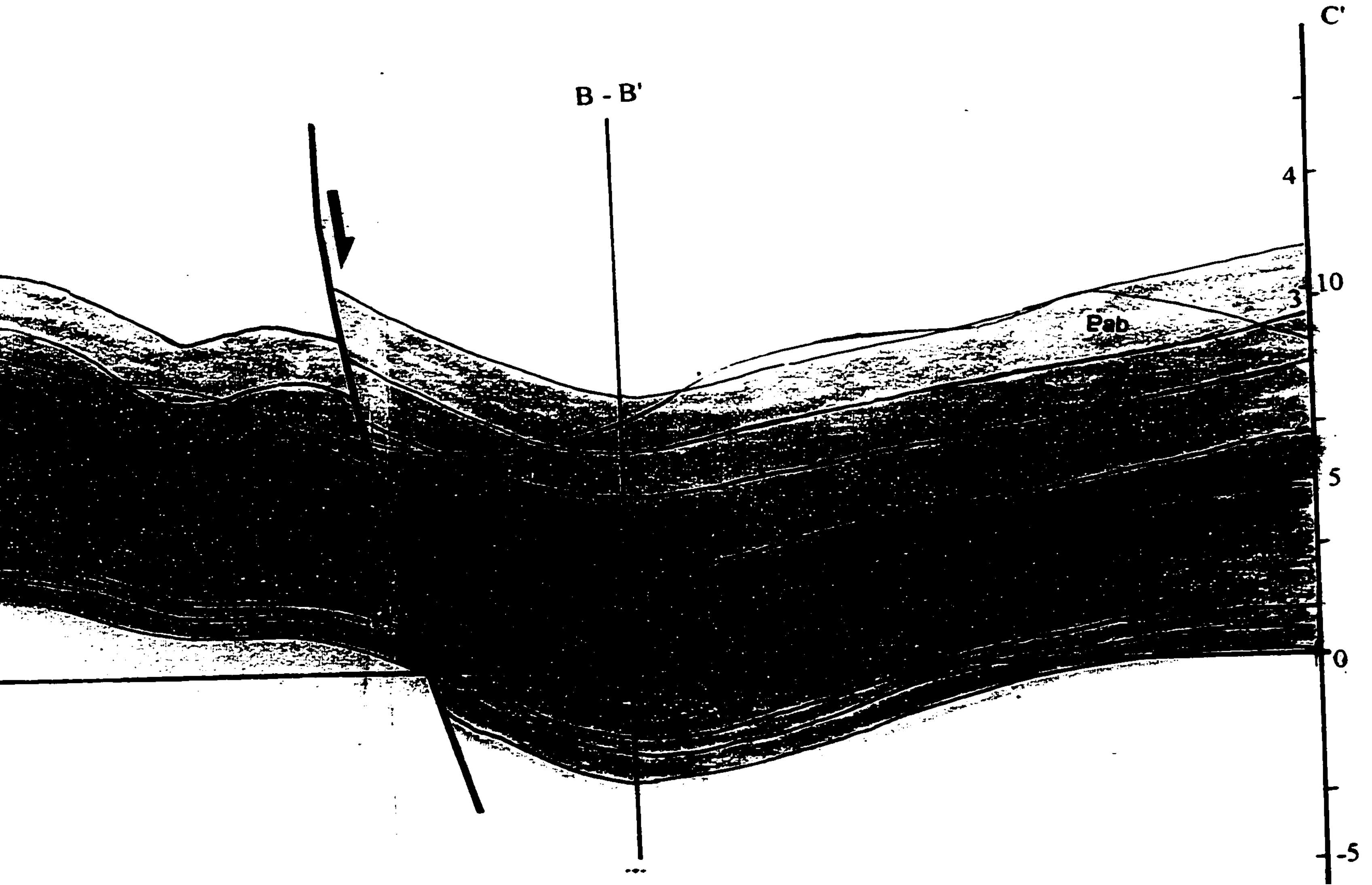
feet X1000 from sea level
(km on inside scale)
5
1
0
sea level
-1
-5
-2

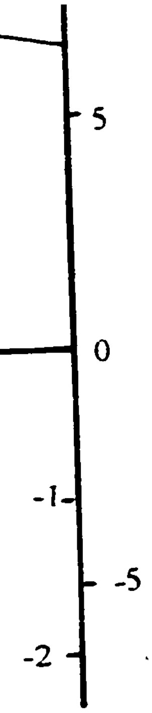
N
C
A
4
10



A - A'
Pk


Pk
0
-1
-5
-2


B - B'
C'
4
10
3
Pab
5
0
-5



SCALE 1: 50 000

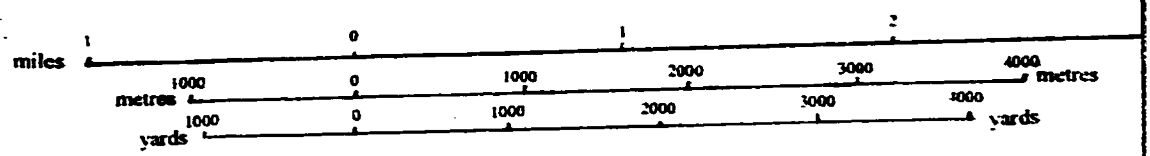
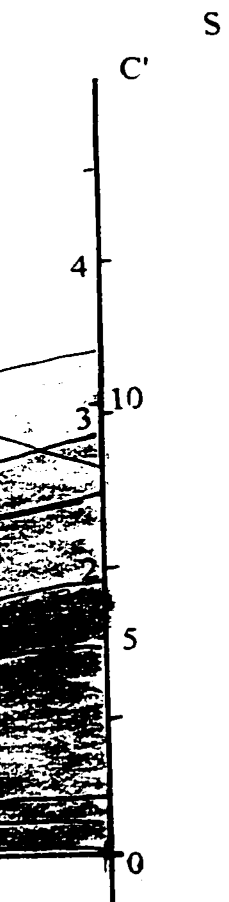
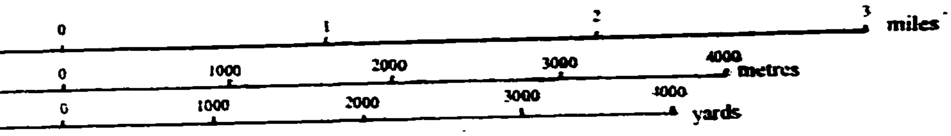


Figure 1-6. Vertical cross-sections through the Isaac Lake Synclinorium. See Figure



SCALE 1: 50 000



s through the Isaac Lake Synclinorium. See Figure 1-5 for location of cross-sections.

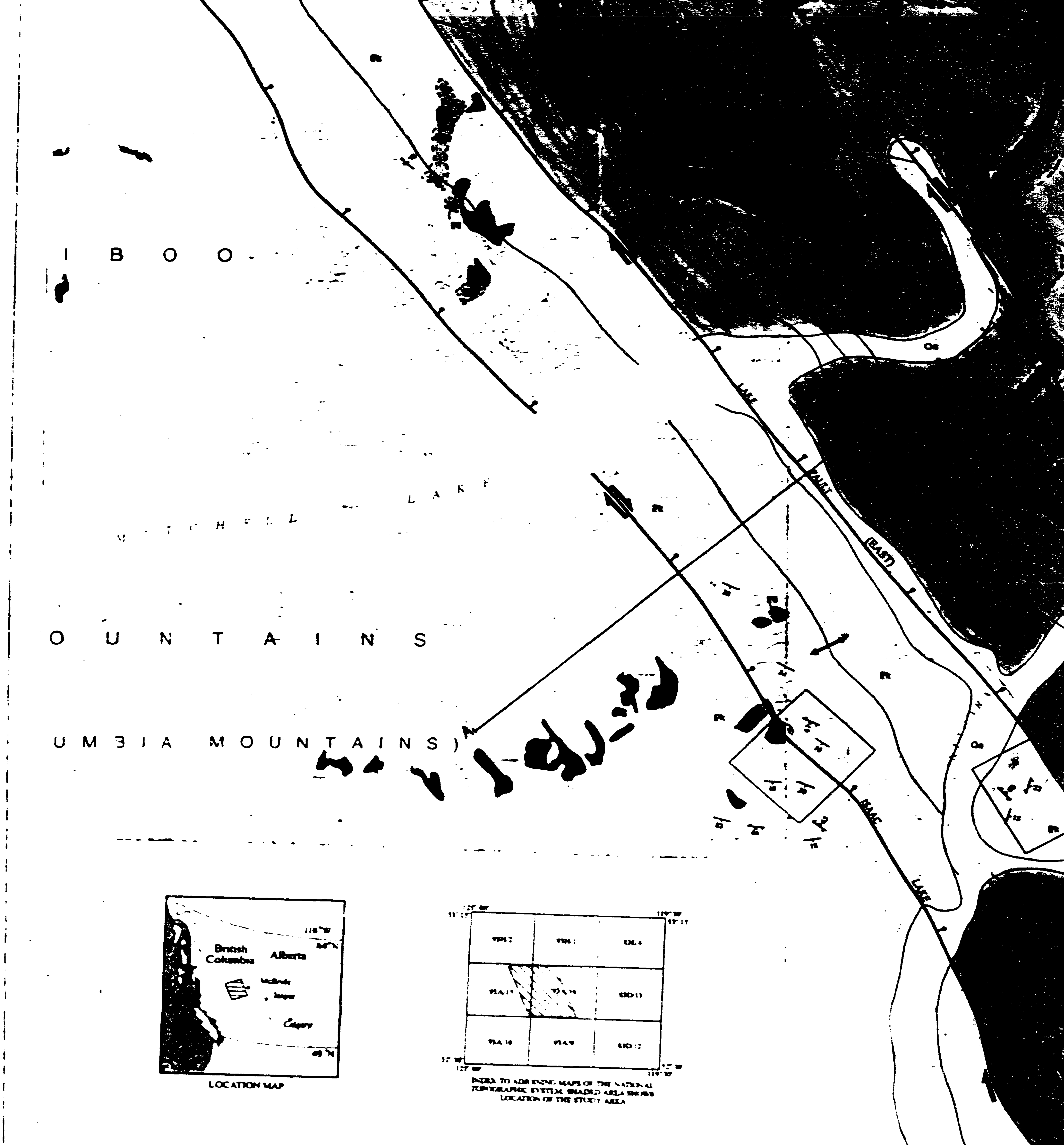


Figure 1-5. Geological map of the Isaac Lake Synclinorium

Cyp — Yanks Peak Formation
[illegible]

NEOPROTEROZOIC

WINDERMERE SUPERGROUP

CARIBOO GROUP

Betty Wendle member
[illegible]

Amos Bowman member
[illegible]

PYb — Yankee Belle Formation
[illegible]

Cunningham Formation
[illegible]

upper Isaac formation
[illegible]

middle Isaac carbonate
[illegible]

lower Isaac formation
[illegible]

Isaac Formation
[illegible]

PK — Kaza Group (uppermost)
[illegible]

MAP SYMBOLS

Stratigraphic contact

defined, approximate

Bedding (S0)

inclined beds, vertical

overturned

Foliations

S1 slaty cleavage inclined, vertical

S2 crenulation cleavage or slaty cleavage inclined, vertical

S3 crenulation cleavage, inclined, vertical

S4 cm-scale crenulation cleavage, inclined, vertical

Lineations

L1 S0-S1 intersections

L2 S0-S2 intersections, crenulations, F2 fold axes

L3 S0-S3 intersections, crenulations, F3 fold axes

Folds

anticline upright, overturned

syncline upright, overturned

Faults

normal, with ball in hangingwall, dashed where approximate

reverse, with teeth in hangingwall, dashed where approximate

strike-slip, with arrows indicating direction of motion, dashed where approximate