SEDIMENTOLOGY OF THE WAPIABI FORMATION AND EQUIVALENTS (UPPER CRETACEOUS), CENTRAL AND NORTHERN FOOTHILLS, ALBERTA

SEDIMENTOLOGY OF THE WAPIABI FORMATION AND EQUIVALENTS (UPPER CRETACEOUS), CENTRAL AND NORTHERN FOOTHILLS, ALBERTA

By

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ABSTRACT

This thesis provides detailed sedimentological descriptions and interpretations for the Upper Cretaceous Wapiabi Formation of the central and northern Alberta Foothills. The Wapiabi Formation is made up of seven members (Muskiki, Marshybank, Dowling, Thistle, Hanson, Chungo, and Nomad) which form a thick sequence (up to 700 m) of predominantly marine shales, siltstones, and sandstones.

During Muskiki time (late Turonian to Coniacian), deposition in the Alberta Basin was largely restricted to marine shales. These shales were deposited during a period of relative orogenic quiescence in the rising Cordillera to the west.

Subsequent to deposition of the Muskiki Member, siltstones of the early Santonian aged Marshybank Member (and sandstones of its equivalent to the north, the Bad Heart Formation) were deposited. Paleocurrent data suggest that this clastic pulse prograded from the northwest. A facies change from nearshore sands to deeper marine siltstones occurs in the area just north of Hinton, Alberta.

A major transgression occurred subsequent to Marshybank deposition. The resulting sequence of Dowling and Thistle shales (early to late Santonian) was deposited in a quiet marine basin during a long period of tectonic inactivity. The Hanson and Chungo Members (late Santonian aged), together, record a single progradational phase prior to the initial major progradation of the Belly River-Paskapoo Assemblage. The bioturbated shales and siltstones of the Hanson Member were

deposited well below storm wave base in a quiet marine environment. The lower Chungo Member is characterized by a regressive sequence of storm influenced sandstones which exhibit turbidites and hummocky and swaley cross-stratification. The upper Chungo Member at Mt. Yamnuska records tidal dominance within an environment protected from storm events (outer eustuary, protected bay?). At the remaining Chungo exposures the upper Chungo is interpreted in terms of a storm dominated, northeasterly prograding shoreline sequence that was influenced by a strong longshore current that flowed towards the southeast. The top of the Chungo regression is regionally capped by a thin, transgressive pebble conglomerate which marks the transition into deep marine shales of the Nomad Member.

The late Santonian to early Campanian aged Nomad Member records the last inundation of the Wapiabi Sea. The basal shale beds are transgressive in nature, being deposited above both marine and non-marine Chungo sediments, while the transitional upper beds reflect the fast rate at which the non-marine Belly River Formation prograded over the Nomad shales.

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TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

CHAPTER 1 - INTRODUCTION

Purpose of Study

The Upper Cretaceous Wapiabi Formation, exposed in the Alberta Foothills, is a 350 m to 700 m thick sequence made up of marine shales, siltstones, and sandstones. This formation is of particular sedimentological interest in that it records the final stages of marine deposition prior to the massive, coarse clastic influx of the predominantly non-marine Belly River - Paskapoo Assemblage (Eisbacher et al., 1974).

In several recent sedimentological studies, parts of the Wapiabi Formation of southwestern Alberta have been described and interpreted in some detail (Hunter, 1980; Bullock, 1981; Rosenthal, in prep.). However, despite this work, studies of the Wapiabi Formation in the central and northern Foothills of Alberta have been limited to a stratigraphic nature. This thesis, therefore, involves a sedimentologically oriented study of the Wapiabi Formation in the central and northern Foothills regions. A particular emphasis has been placed on the marine to non-marine transitional beds of the Chungo Member.

In a comprehensive study of several Wapiabi Formation sections in the south, Rosenthal (in prep.) reported the domination of storm deposits in the transitional Chungo Member. Turbidites, hummocky cross-stratification (HCS), and swaley cross-stratification (SCS) were all reported. Rosenthal (in prep.) also showed how the Chungo Member sandstone thinned and shaled out in a northerly direction (towards the present study area). The sandstones were shown to thin and interfinger with silty and shaly facies of the more distal Hanson Member. However, these results are difficult to understand in light of studies by Stott (1963) who reported a thick and sandy Chungo Member in the central and northern Foothills. As well, the storm dominated shoreline interpretation of Rosenthal (in prep.) is at odds with a study by Lerand (1982) who, in one depositional summary, interpreted portions of the Chungo Member exposed at Mt. Yamnuska as being deposited in a North Sea type tidally dominated environment. This study will attempt to document the reoccurrence of the Chungo Member north of Rosenthal's (in prep.) field area with the intent of defining the regional trends of the Chungo Member. The study will also attempt to reconcile the apparent discrepancies between the environmental interpretations of Lerand (1982) and Rosenthal (in prep.).

As well as relating the Chungo Member in the central and northern Foothills with the Chungo Member in the southern Foothills, the problem of understanding the lateral relationships of the Chungo, the geometry of the Chungo sand body, and how the sand body dies out basinwards will be addressed. Understanding these factors may have economic significance as equivalent strata in the Milk River gas field are known to contain large gas reserves in excess of 5 trillion cubic feet recoverable. Moreover, the development and understanding of depositional models in the central and northern Foothills can act only as an aid to exploration in relatively unexploited areas.

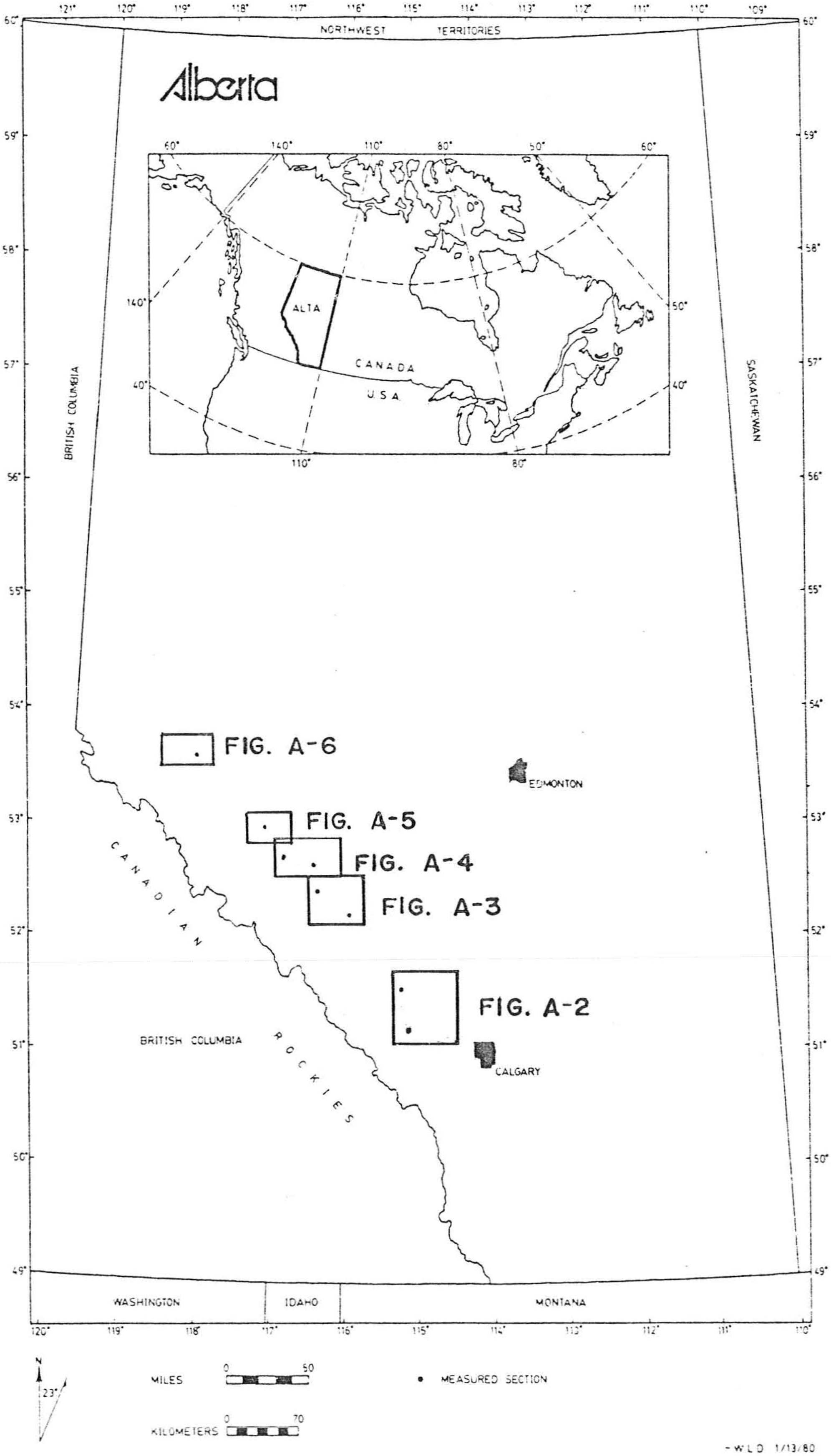
Location of Study Area

Eight sections of the Upper Cretaceous Wapiabi Formation were measured and studied in detail for the purposes of this study. These outcrops were exposed along the northwest-southeast oriented strike of the central and northern Foothills in an area bounded by Exshaw, Alberta in the southeast and Grande Cache, Alberta in the northwest. The approximate location of each section is shown in Figure 1-1.

Format of Thesis

The results of this thesis have been derived from the measurement and mapping of 8 stratigraphic sections of the Wapiabi Formation located in the central and northern Foothills (Figure 1-1). The detailed measured sections can be found in the back pocket. The first two chapters act as an introduction to the study and discuss the geologic and stratigraphic settings. The third and fourth chapters contain reviews of the stratigraphic nomenclature of the study area and previous studies, respectively. Facies descriptions and interpretations from the present study are presented in chapters 5 and 6. Chapters 7 through 10 outline section descriptions, paleoflows recorded from the section, petrology of the Chungo Member, and regional interpretations with depositional summaries. Chapter 11 lists all the major conclusions reached in the study.

Figure 1-1. Location map for measured sections. Modified from Duke (in prep.).


Alberta
NORTHWEST TERRITORIES
BRITISH COLUMBIA
SASKATCHEWAN
ALTA
CANADA
U.S.A.
FIG. A-6
FIG. A-5
FIG. A-4
FIG. A-3
FIG. A-2
EDMONTON
CALGARY
CANADIAN ROCKIES
BRITISH COLUMBIA
WASHINGTON
IDAHO
MONTANA
MILES 0 50
KILOMETERS 0 70
23°
• MEASURED SECTION
WLD 1/13/80

CHAPTER 2 - GEOLOGIC SETTING

Structural Setting

The Wapiabi Formation shales were deposited during a relatively quiescent period between two times of vast orogenic activity (approximately 76 Ma to 87 Ma, early Conacian to early Campanian). The Colombian Orogeny, Upper Jurassic to earliest Upper Cretaceous, was responsible for the tectonic and structural setting within which the Wapiabi Formation was deposited. The later Laramide Orogeny, late Upper Cretaceous to early Oligocene, was the event which led to the development of the Foothills Subprovince of the Rocky Mountain Thrust Belt, the area in which this study was centred.

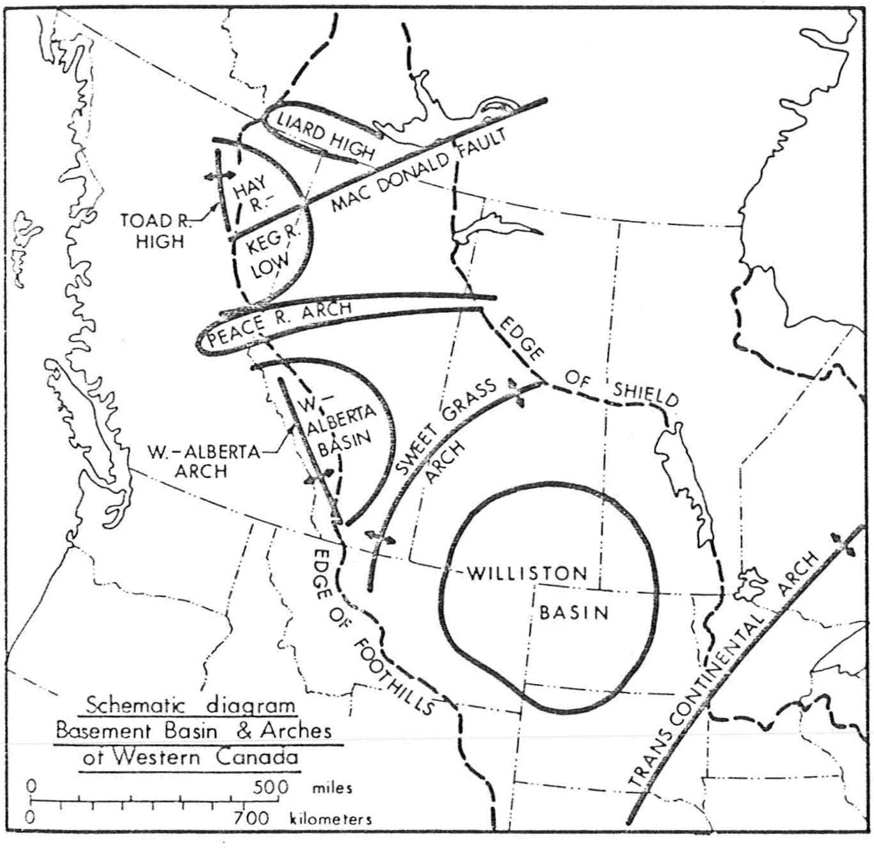
The Columbian Orogeny was largely characterized by the emplacement of granitic plutons, metamorphism in the Omineca Crystalline Belt, and significant uplifting in the Rocky Mountain region. Orogenic uplift was particularly intense in the southern Rocky Mountain area (Douglas et al., 1970). The resultant sedimentary response was the shedding of clastic material into a developing eastern foreland basin between the rising Cordillera to the west and the North American craton to the east. Isostatic adjustment due to the clastic input led to deepening of this Alberta Trough (Price et al., 1972). The Wapiabi Formation was deposited within this foreland trough during a time of relative orogenic quiescence after the major tectonic events of the Columbian Orogeny had taken place (Douglas et al., 1970). However, sedimentation was not controlled only by the geometry of the Alberta Trough. Two Precambrian basement structural features, the northeast trending Peace River Arch to the north, and the Sweetgrass Arch to the south, influenced sedimentary dispersal patterns within the basin (Figure 2-1).

The present structural style exhibited in the Foothills Subprovince (Price et al., 1972) of the Rocky Mtn. Thrust Belt was largely created during tectonic activity of the Laramide Orogeny (Price and Mountjoy, 1970). Figure 2-2 shows a structural cross-section from the present main ranges into the Alberta Interior Platform. Thrusting was initiated while sedimentation of the Late Cretaceous and early Tertiary was in progress, but continued into the early Oligocene, well after sedimentation had ceased (Douglas et al., 1970).

The allochthonous thrust belt, developed over the passive and westerly dipping undeformed Hudsonian crystalline basement at the edge of the North American craton (Bally et al., 1966, Wheeler et al., 1972), was suggested to have been the result of buoyant upwelling of a metamorphic core zone well to the west of the study area (Wheeler et al., 1972). Initial thrusting of "thin-skinned type" took place just east of this central metamorphic core zone. All subsequent thrusting then proceeded in order towards the northeast, resulting in northeasterly foreshortening (Price et al., 1972).

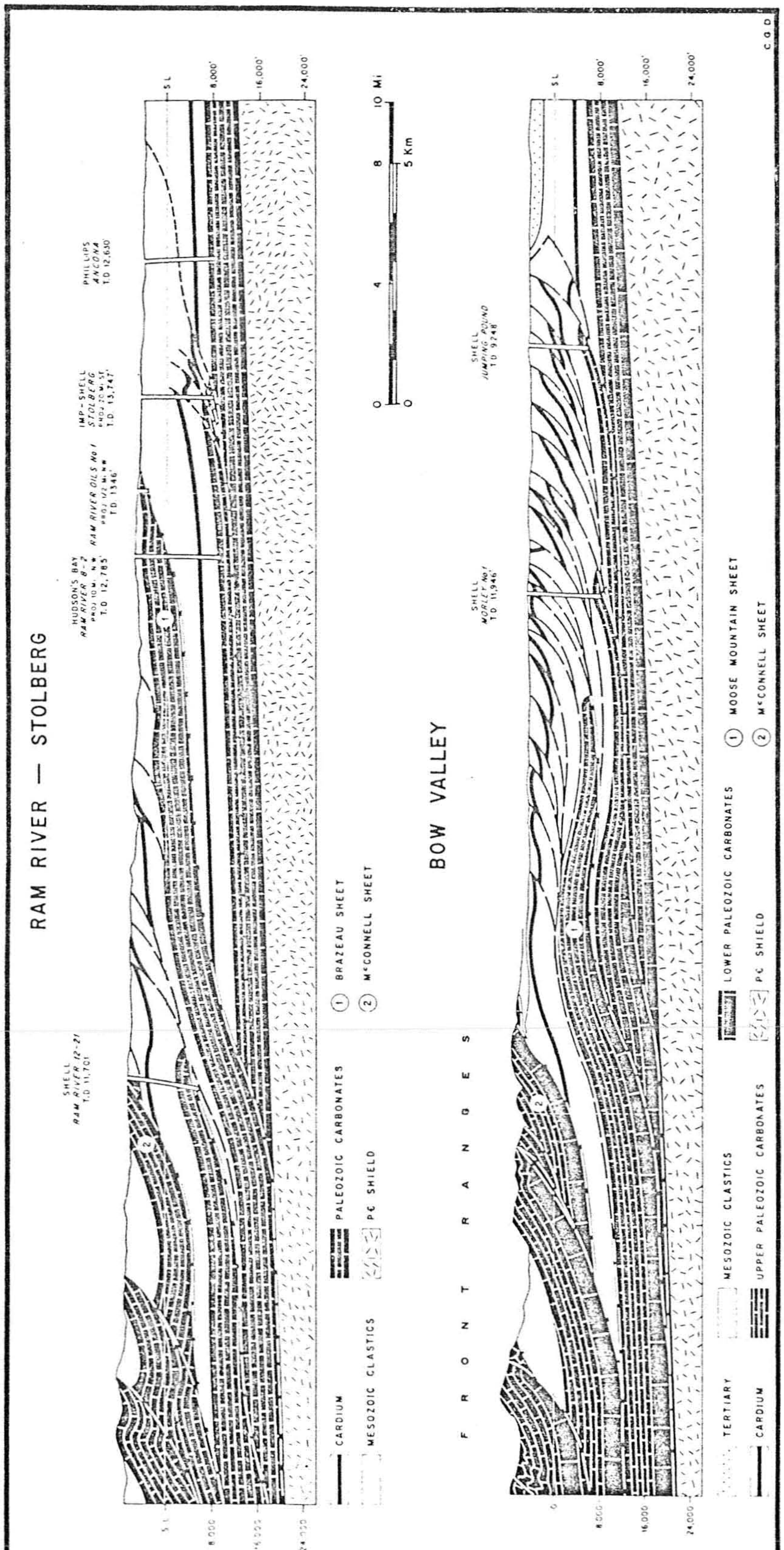
Thrusting in the Foothills area originated low in Proterozoic and Paleozoic strata. It then followed bedding zones of coal and shale (and other easily sheared sedimentary zones) up through Mesozoic and Tertiary strata to the surface. Faulting continued laterally along

Figure 2-1. Schematic diagram of Precambrian basement features of western Canada. From Stelck, 1975.



LIARD HIGH
MAC DONALD FAULT
HAY R.-
KEG R. LOW
TOAD R. HIGH
PEACE R. ARCH
EDGE OF SHIELD
W.-ALBERTA BASIN
W.-ALBERTA ARCH
SWEET GRASS ARCH
WILLISTON BASIN
EDGE OF FOOTHILLS
TRANS CONTINENTAL ARCH
Schematic diagram
Basement Basin & Arches
of Western Canada
0
500 miles
0
700 kilometers

Figure 2-2. Structural cross-sections from the Main Ranges to the Alberta Interior Platform taken along Ram River and Bow Valley. From Balley et al., 1966.


RAM RIVER — STOLBERG
SHELL
RAM RIVER 12-21
T.D 11,701
HUDSON'S BAY
RAM RIVER 8-2
PROJ 10 MI. NW
T.D 12,785'
RAM RIVER OILS No 1
PROJ 1/2 MI. NW
T.D 1346'
IMP-SHELL
STOLBERG
PROJ 20 MI. SE
T.D 13,747'
PHILLIPS
ANCONA
T.D 12,630'
S.L
8,000'
16,000'
24,000'
CARDIUM
MESOZOIC CLASTICS
PALEOZOIC CARBONATES
P€ SHIELD
1 BRAZEAU SHEET
2 McCONNELL SHEET
0
4
8
10 Mi
0
5 Km
BOW VALLEY
FRONT RANGES
SHELL
MORLEY No1
T.D 11,946'
SHELL
JUMPING POUND
T.D 9,248'
TERTIARY
CARDIUM
MESOZOIC CLASTICS
UPPER PALEOZOIC CARBONATES
LOWER PALEOZOIC CARBONATES
P€ SHIELD
1 MOOSE MOUNTAIN SHEET
2 McCONNELL SHEET
C.G.D

strike at the surface (Douglas, 1960). Upwards in the section, new faults formed and various thrusts interlocked, merged, or became connected by smaller faults (Douglas et al., 1970).

The present Foothills Subprovince is characterized by closely spaced, translational, imbricate thrust slices dipping in a westerly to southerly direction (Price et al., 1972). Local variations of dip, especially in the central Foothills, are due to pre-thrusting of long narrow synclines. Several thrust slices are terminated as concentric, flexural-slip folds of varying size.

The thrust slices upon which the measured sections in this study are located and the amount of foreshortening along these thrust slices have been determined. They are as follows:

Mt. Yamnuska - Dyson Mtn. Thrust, ≅ 25 km shortening

Burnt Timber Creek - Waiparous Thrust, ≅ 16 km shortening

Cripple Creek - Bighorn Thrust, ≅ 31 km shortening

Bighorn River - Bighorn Thrust, ≅ 33 km shortening

Blackstone River - Brazeau Thrust, ≅ 25 km shortening

Thistle Creek - Bighorn Thrust, ≅ 31 km shortening

McLeod River - Folding Mtn. Thrust, ≅ 35 km shortening

Little Berland River - Masan Thrust, ≅ 15 km shortening

Once these thrust slices were determined, palinspastic reconstructions were undertaken to find the original depositional position of each section. The method used for palinspastic reconstruction is outlined in Appendix V. The results of the reconstructions are shown in Figure 2-3. This figure may be handy to

Figure 2-3. Palinspastically reconstructed positions of measured sections from the present study area. The arrows show the direction of foreshortening.

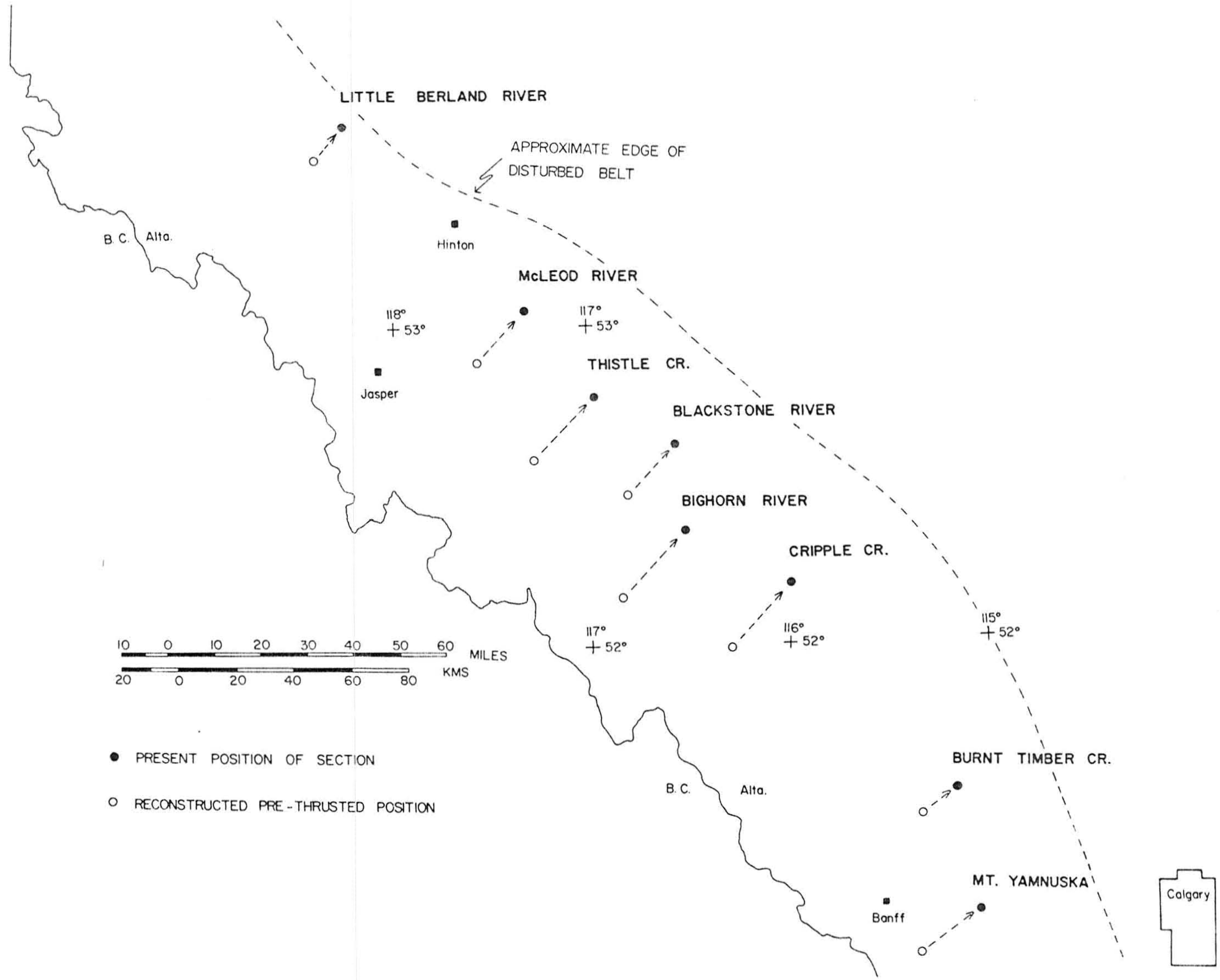

LITTLE BERLAND RIVER
APPROXIMATE EDGE OF DISTURBED BELT
B. C.
Alta.
Hinton
McLEOD RIVER
118°
53°
117°
53°
Jasper
THISTLE CR.
BLACKSTONE RIVER
BIGHORN RIVER
CRIPPLE CR.
117°
52°
116°
52°
115°
52°
10 0 10 20 30 40 50 60 MILES
20 0 20 40 60 80 KMS
PRESENT POSITION OF SECTION
RECONSTRUCTED PRE-THRUSTED POSITION
B. C.
Alta.
BURNT TIMBER CR.
MT. YAMNUSKA
Banff
Calgary

Figure 2-4. Stratigraphic column for the Foothills and Front Range west of Calgary. From Sprang et al., 1981.

TABLE OF FORMATIONS

FOOTHILLS & FRONT RANGE STRATIGRAPHY WEST OF CALGARY

AGE		FORMATION OR GROUP		LITHOLOGY	THICKNESS	FACIES	AV. VELOCITIES IN FEET/SEC.
TERTIARY		PASKAPOO			2500 4000	CONTINENTAL DELTAIC	11,000
CRETACEOUS	UPPER	EDMONTON	UPPER BRAZEAU		1500 - 2000	CONTINENTAL DELTAIC	11,000 12,000
		BEARPAW			0 -150	MARINE	
		BELLY RIVER	LOWER BRAZEAU		1200 - 3500	CONTINENTAL	12,000 13,000
		WAPIABI	ALBERTA GROUP		1250 - 1500	MARINE	12,500 13,000
		CARDIUM			300 - 400	DELTAIC-LITTORAL	13,000-14,000
		BLACKSTONE			800 - 1000	MARINE	13,000 14,000
	LOWER	BLAIRMORE			1100 - 1400	CONTINENTAL MARINE	14,000 16,000
		KOOTENAY			0 - 1200	CONTINENTAL	14,000-16,000
JURASSIC		FERNIE			50 - 500	MARINE	13,500-14,000
TRIASSIC		SPRAY RIVER			0 - 400	PARTLY MARINE	14,000-16,000
PERMO-PENN		ROCKY MTN.			0 - 150	PARTLY MARINE	18,000
MISSISSIPPIAN		MOUNT HEAD	RUNDLE		200 - 300	RESTR. MARINE	19,000
		TURNER VALLEY			250 - 350	MARINE	
		SHUNDA			250 - 300	RESTR. MARINE	
		PEKISKO			250 - 300	MARINE	21,000
		BANFF			500 - 600		
DEVONIAN	UPPER	PALLISER			800 - 1000	RESTR. MARINE	19,000
		FAIRHOLME			1000 - 1300	MARINE	21,000
CAMBRAIN	? U ?	? ARCTOMYS ?			0 - 100 ?	MARINE	18,000 20,000
	MIDDLE	PIKA			150 - 400		
		ELDON			500 - 900		
		STEPHEN			200 - 250		
		CATHEDRAL			200 - 800	LITTORAL-MARINE	
PЄ		HUDSONIAN				IGNEOUS & METAMORPHIC ROCKS	

refer to when reading the "Paleocurrent Data" section of the text.

Stratigraphic Setting

The southeastern Canadian Cordillera of southwestern Alberta and southeastern British Columbia is composed of two major sedimentary rock sequences: a miogeosynclinal shelf sequence of late Precambrian to late Jurassic age and a clastic wedge of molasse type sedimentation of late Jurassic to early Tertiary age (Price and Mountjoy, 1970; figure 2-4). The Paleozoic sediments are largely made up of shallow marine carbonates with some appreciable amounts of terrigenous clastics derived from the North American craton to the northeast. However, McIlreath (1977) and Hopkins (1977) have defined some allocthonous carbonates deposited within slope and deep marine environments. The Mesozoic clastics, largely deposited within shallow marine and fluvial environments, were emplaced in response to tectonic events induced by the Columbian and Laramide Orogenies (Eisbacher et al., 1974).

The clastic miogeosynclinal sediments of the Purcell and Windermere Groups (Middle to Late Proterozoic) form the oldest exposed sediments in the southeastern Cordillera (Wheeler et al., 1972). While these groups unconformably overlie the Hudsonian crystalline base to the west, the transgressive sandstones of the lower Cambrian Gog Group overlie the Precambrian base towards the interior platform.

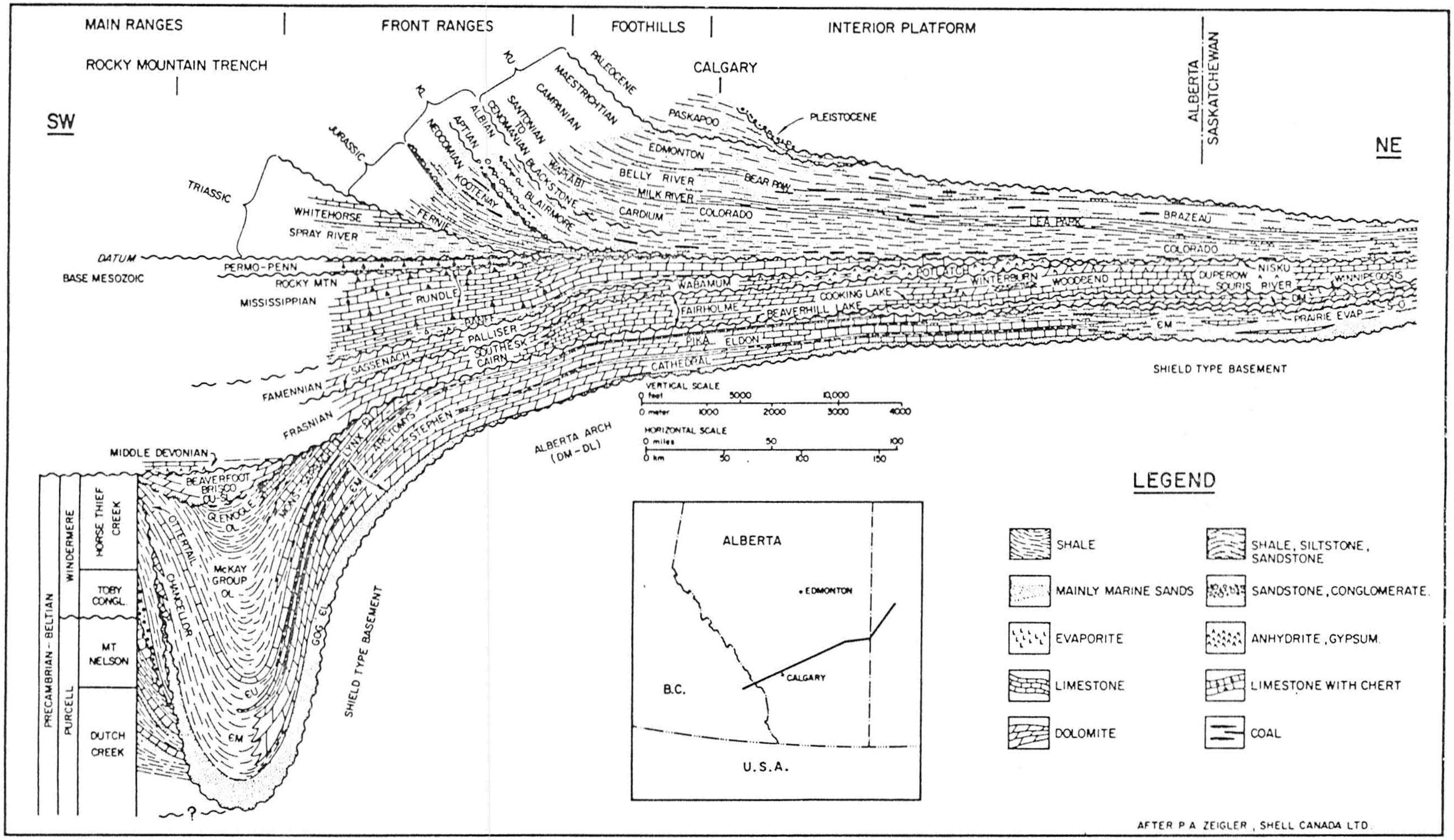
The remainder of Paleozoic sediments are characterized by thick successions of southwesterly thickening miogeosynclinal carbonates and clastics truncated by major transgressive unconformities (Price et al., 1972). Price and Mountjoy (1970) report a total miogeosynclinal thickness of 6000 feet (1829 m) beneath the western plains, thickening to 40,000 feet (12,192 m) in the western Rocky Mountain region.

The major influx of siliclastic sediments during the Mesozoic was separated from the Paleozoic by a major transgressive event which involved a duration from Pennsylvanian to Middle Jurassic times. The Mesozoic and early Tertiary were dominated by two major regressive cycles which resulted in the deposition of two "clastic megacycles" (Eisbacher et al., 1974). Both of these cycles (the Kootenay-Blairmore Assemblage and the Belly River-Paskapoo Assemblage) were deposited as a result of major orogenic uplifts in the Cordillera, the Columbian and Laramide orogenies, respectively.

The first record of sedimentation within the developing foreland basin (Price et al., 1972) was the shales of the marine Fernie Formation which coarsened upward into the fluvial sandstones of the northward progradational Kootenay Formation (Hamblin and Walker, 1979). A period of major uplift, resulting in the culmination of the Kootney-Blarimore Assemblage (Schulthies and Mountjoy, 1978), then led to the subsequent deposition of coarse conglomerates of the Cadomin Formation followed by coarse clastic sediments of the Blairmore Group (Taylor, 1981).

An interim period of relative orogenic stability (Douglas et al., 1970) led to the deposition of a thick sequence (up to 800 m) of predominantly black marine shales and sandstones which make up the Blackstone, Cardium, and Wapiabi Formations in ascending order. The Chungo sandstone of the upper Wapiabi Formation may reflect the first sedimentary response to the Laramide orogeny which resulted in

Figure 2-5. Stratigraphic diagram showing lithology and stratigraphy from the Interior Platform to the Rocky Mountain Trench. From Gordy, 1972.


MAIN RANGES
FRONT RANGES
FOOTHILLS
INTERIOR PLATFORM
ROCKY MOUNTAIN TRENCH
CALGARY
ALBERTA
SASKATCHEWAN
SW
NE
PALEOCENE
MAESTRICHTIAN
CAMPANIAN
SANTONIAN TO CENOMANIAN
ALBIAN
APTIAN
NEOCOMIAN
KU
KL
JURASSIC
TRIASSIC
PASKAPOO
PLEISTOCENE
EDMONTON
BEAR PAW
BELLY RIVER
MILK RIVER
CARDIUM
COLORADO
BLACKSTONE
BLAIRMORE
KOOTENAY
FERNIE
WHITEHORSE
SPRAY RIVER
DATUM
BASE MESOZOIC
PERMO-PENN
ROCKY MTN
MISSISSIPPIAN
RUNDLE
BANFF
PALLISER
SOUTHESK
CAIRN
SASSENACH
FAMENNIAN
FRASNIAN
MIDDLE DEVONIAN
LEA PARK
BRAZEAU
COLORADO
WABAMUM
NISKU
WINTERBURN
WOODBEND
DUPEROW
SOURIS RIVER
WINNIPEGOSIS
FAIRHOLME
COOKING LAKE
BEAVERHILL LAKE
PRAIRIE EVAP
PIKA
ELDON
CATHEDRAL
ЄM
SHIELD TYPE BASEMENT
ALBERTA ARCH (DM - DL)
VERTICAL SCALE
0 feet 5000 10,000
0 meter 1000 2000 3000 4000
HORIZONTAL SCALE
0 miles 50 100
0 km 50 100 150
BEAVERFOOT
BRISCO
OTTERTAIL
CHANCELLOR
McKAY GROUP
ЄU
ЄM
ЄL
GOG
LYNX
ARCTOMYS
STEPHEN
SHIELD TYPE BASEMENT
PRECAMBRIAN - BELTIAN
PURCELL
WINDERMERE
HORSE THIEF CREEK
TOBY CONGL.
MT NELSON
DUTCH CREEK
ALBERTA
EDMONTON
CALGARY
B.C.
U.S.A.
LEGEND
SHALE
SHALE, SILTSTONE, SANDSTONE
MAINLY MARINE SANDS
SANDSTONE, CONGLOMERATE
EVAPORITE
ANHYDRITE, GYPSUM
LIMESTONE
LIMESTONE WITH CHERT
DOLOMITE
COAL
AFTER P.A. ZEIGLER, SHELL CANADA LTD.

deposition of the Belly River-Paskapoo Assemblage. This second megacycle was characterized largely by the non-marine and deltaic sediments of the Belly River, Edmonton, and Paskapoo Formations (Upper Cretaceous to Paleocene) in ascending order (Eisbacher, 1979).

The remaining portion of the Tertiary, following deposition of the Paleocene Paskapoo Formation, was a time of general erosion. No Eocene to Pliocene deposits have been encountered in the central Foothills area. However, glacial deposition during the Pleistocene covered almost the entire Cordillera and Interior Plains regions (Douglas et al., 1970).

CHAPTER 3

STRATIGRAPHIC NOMENCLATURE FOR CENTRAL AND NORTHERN FOOTHILLS

Although considerably easier and more straightforward than the recently reviewed stratigraphy for the Wapiabi Formation in southern Alberta (Rosenthal, in prep), the stratigraphic nomenclature for the Wapiabi Formation from north of Calgary to Grande Cache is highly inconsistent and redundant. Two main reasons have led to this situation: 1) workers in separate areas developed independent nomenclature schemes that were either miscorrelated or not correlated at all and 2) subsurface lithologic nomenclature, largely developed by the petroleum industry (which in many cases is still not formally accepted), was not correlated with outcrop equivalents; thus two completely separate schemes were introduced.

To help understand present stratigraphic relationships, an evolution of the Alberta Foothills' nomenclature is included. Because the study area was divided into two general sections that were studied independently of each other, the central and northern portions, different stratigraphic names were often proposed for correlatable beds. Moreover, equivalent and correlatable formations within the Upper Cretaceous had been assigned to different groups across an arbitrary line that divided the central area from the northern area, the Alberta Group and the Smoky Group, respectively (Webb and Hertlein, 1934; McLearn and Kindle, 1950). This segregation within the study area has prompted the following discussion to be divided into three major segments: the first concerns the central Foothills north of Calgary as far as the Athabasca River, the second concerns the northern section which extends to the northern most part of the study area and continues into northwest Alberta and northeast British Columbia (Stott, 1963) and the last portion which discusses the nomenclature that has been adopted for application in this study.

Central Foothills

One of the first descriptions of the Cretaceous rocks in the Central Foothills was prepared by Dawson (1884) from observations made south of the Bow River. In this description, Dawson (1884) proposed the names Fox Hills, Pierre, and Belly River Formations for rocks which he mistakenly correlated to their namesakes in the United States. He also suggested the term "Dark Shales" for a thick sequence of shales below the Belly River Formation.

Cairnes (1907) first divided the Dark Shales of Dawson (1884) into upper and lower shale units to which he gave formational status and named the Claggett and Niobrara-Benton Formations, respectively. The Cardium sandstone was included in the upper portion of the Niobrara-Benton Formation. Again, the names Clagett and Niobrara-Benton were derived through miscorrelation with beds of the same name in the United States.

Present day terminology is based largely on the work of Malloch (1911) who first introduced and applied the names Blackstone, Bighorn, and Wapiabi Formations to a sequence of shales and sandstones lying between the Brazeau Formation (Belly River) on top and the Dakota Formation (Blairmore Group) below. His field area was near the North

Saskatchewan River. No correlations were suggested between Cairnes' (1907) and Malloch's (1911) areas.

The next twenty years saw various attempts at reintroducing the usage of the term "Benton Formation" in place of the Wapiabi, Bighorn, and Blackstone Formations (Rose, 1920; Slipper, 1921; Rutherford, 1927). However, Hume (1930) stated that usage of "Benton Formation" was inappropriate as these beds were not correlatable with the Fort Benton Formation in the United States. Furthermore, correlations suggested that the beds were not of Colorado age as Rutherford (1927) had proposed; the use of the term "Alberta Shale" was therefore applied.

A major regional study and nomenclature revision by Webb and Hertlein (1934) set the stage for the modern terminology. The Alberta Shale of Hume (1930) was raised to group status and was divided into three formations. In ascending order, the terms Blackstone, Cardium, and Wapiabi were applied. Within the Wapiabi Formation, four distinct zones were noted. From bottom to top: lower concretionary shale, platey shale, upper concretionary shale, and the transition zone into the overlying Belly River Formation. A sandstone unit in the upper portion of the Wapiabi Formation identified near Highwood River was designated as the Highwood Sandstone.

The present nomenclature has remained largely unchanged since a study by Stott (1957). While the name "Alberta Group" and the Blackstone, Cardium, and Wapiabi Formations within it were retained by Stott (1957), units within each formation were redefined and given member status. In the Wapiabi Formation seven members were recognized, in ascending order, the Muskiki, Marshybank, Dowling, Thistle, Hanson, Chungo, and Nomad Members. Each member will be discussed in detail at a later point.

Correlation of the central Foothills' Wapiabi stratigraphy with the stratigraphy of the southern foothills and the central and southern Alberta plains has recently been reviewed by Rosenthal (in prep). Within the central plains, where the Chungo Sandstone Member has pinched out towards the east, the Lea Park Formation has been named as the Hanson, Chungo, and Nomad Member equivalents. Below the Lea Park Formation, the names Alberta Formation or Colorado Formation have been used as equivalents to the lower Wapiabi Formation in the foothills. However, on drilling tickets the name "Lea Park Formation" is commonly used to describe all shales between the Cardium and Belly River Formations. Towards the south, where a sandstone equivalent of the Hanson and Chungo Members still persists, the name Milk River Formation is used. The overlying dark marine shale dividing the Milk River and Belly River Formations is termed the Pakowki Formation (Nomad Member equivalent). Stott (1963) has also shown the Chungo Member to be correlatable with the Eagle Sandstone in Montana.

Northern Foothills

Although not the first to describe the lithologies of the Cretaceous sediments in the northern Foothills, Dawson (1881), was the first to apply names to specific stratigraphic intervals. The Upper Cretaceous was separated into the following units, in ascending order, the Ft. St. John Shales (lower dark shales), the Dunvegan Sandstones, the Smoky River Shales (upper dark shales), and the Wapiti Sandstones.

McLearn (1919) proposed that the units in the Upper Cretaceous defined by Dawson (1881) be raised to formational status. He also divided the shales characterizing the Smoky River Formation into two shale units, the upper and the lower shales, separated by a 10 to 25 foot (3 to 8 meters) sandstone unit to which he applied the name "Bad Heart Sandstone". The lower shale unit was then assigned the name "Kaskapau" (McLearn, 1926) which was later raised to formational status when the Smoky River Formation was assigned group status (McLearn and Kindle, 1950). In the reclassification, the Fort St. John Formation was also given group status and the Bad Heart Sandstone and the Upper Shales were given formational status.

Correlation of the Upper Shale Formation with the Wapiabi Formation to the south led to the adoption of the name "Wapiabi Formation" in the north (Gleddie, 1949). Within the Wapiabi, Gleddie (1949) used the name "Chinook" to describe a 75 foot (23 meters) thick sandstone that lies approximately 100 feet (30 m) below the overlying Wapiti Formation. Stott (1963) states that, "This sandstone is probably equivalent to the sandstone at the top of the Wapiabi Formation farther to the south" to which he applied the name "Chungo Member". Gleddie (1949) also first applied the usage of "Cardium Formation" to the upper half of the formerly defined Kaskapau Formation. The Kaskapau Formation was restricted to beds lying below the Cardium Formation and above the Dunvegan Formation.

Miscorrelation of the Bad Heart Formation by Gleddie (1949) with the Cardium Formation to the south resulted in the Bad Heart being included in the Cardium Formation. Stelck (1955) realized the miscorrelation and reassigned the Bad Heart to the Wapiabi Formation.

Stott (1963) provided paleontologic evidence to substantiate Stelck's (1955) correlation.

Stott (1963) suggested that it was possible to identify most of the units he defined in the central Foothills in the northern Foothills. Both the Cardium and the Wapiabi Formations were shown to be completely correlatable from the central to the northern Foothills. Even the seven members of the Wapiabi Formation were identified by Stott (1963) in the northern Foothills. Because the stratigraphic nomenclature of Stott (1963) could, and was, applied to the northern portion of the study area, name changes were introduced. Firstly, the Chungo Member of the Wapiabi Formation replaced Chinook Sandstone of Gleddie (1949). The Chungo is also the apparent equivalent of the Solomon Sandstone further to the northwest into British Columbia (Stott, 1967). The formerly defined Bad Heart Formation was included by Stott (1963) with the Wapiabi Formation and was given member status as the Marshybank Member. However, Stott (1967) returned to the use of the Bad Heart Formation in the north where the unit was a distinct sandstone and restricted the use of Marshybank Member to equivalent siltstones further south. These siltstones reflected a facies change from the pinching out sandstone facies. This change takes place just south of Little Berland River in the study area. Where both the Cardium and Bad Heart Formations are recognized, the shales between the two formations are referred to as the Muskiki Formation, rather than the Muskiki Member of the Wapiabi Formation to the south. As well, Stott (1967) accepted a suggestion by Wall (1960) to replace the usage of "Wapiabi Formation" with "Paskwaskau Formation" where the Muskiki and the Bad Heart Formations are recognized. The Paskwaskau Formation would still include the Dowling, Thistle, Hanson, Chungo and Nomad Members recognized in the Wapiabi Formation further south. This proposal by Wall (1960) was based on the idea that the type Wapiabi Formation included the Muskiki and Marshybank Members. In the area where the two members were not recognized as part of the Wapiabi Formation, he believed that a new terminology should be introduced.

Table 3-1 outlines the stratigraphic history and evolution for both the central and northern study regions. A correlation chart of formations for the same foothills region is provided in Figure 3-1.

Nomenclature Adopted in this Study

The stratigraphic scheme adopted for this study is largely based on schemes presented by Stott (1963) and Irish (1965). Two diagrams outlining the division of the Upper Cretaceous into the formational nomenclature used here are presented in Figures 3-2 and 3-3.

In the central Foothills area, the stratigraphic terminology as proposed by Stott (1963) has been adopted. In this scheme, the Upper Cretaceous Alberta Group has been divided into three formations, in ascending order, the Blackstone, Cardium, and Wapiabi Formations. In turn, each formation was divided into several members. The Wapiabi Formation was segregated into seven such members that have been previously listed. However, recognition of the Marshybank Member (the equivalent of the Bad Heart Formation to the north) in some sections in the study area is highly dubious and speculative.

In the northern Foothills section north of the Athabasca Valley, the use of the term "Wapiabi Formation" is favoured over the adoption of who he ?

Table 3-1. Nomenclature of Alberta Group and Equivalent strata, Alberta Foothills. From Stout, 1963.

	SOUTHERN AND CENTRAL REGION											NORTHERN REGION						
HECTOR (WHITEAVES 1895)	DAWSON 1884	CAIRNES 1907	MALLOCH 1911	ROSE 1920	SLIPPER 1921	RUTHERFORD 1927	HUME 1930	WEBB & HERTLEIN 1934	HAKE, WILLIS, & ADDISON 1942	SCOTT 1951	STOTT 1957	STOTT 1957	STELCK 1955	GLEDDIE 1949	McLEARN & KINDLE 1950	McLEARN 1926	McLEARN 1919	DAWSON 1881
		BELLY RIVER	BRAZEAU	ALLISON	BELLY RIVER	BELLY RIVER	BELLY RIVER	BELLY RIVER		BRAZEAU	BELLY RIVER	BRAZEAU	WAPITI	WAPITI	WAPITI	WAPITI	WAPITI	WAPITI
				Benton formation	Benton formation		Alberta shale	Transition	Transition	Transition	Nomad	Nomad	Wapiabi formation					
								Wapiabi formation Upper concretionary shale	Wapiabi formation Blocky	Wapiabi formation Siltstone	Wapiabi formation Chungo	Wapiabi formation Chungo		Wapiabi fm. Chinook				
									Upper conc. shale	Upper conc. shale	Hanson	Hanson			"Upper shale" formation	Upper shale	Upper shale	
		Claggett formation	Wapiabi formation			Upper Benton formation	Upper	Platy shale	Platy shale	Platy shale	Thistle	Thistle						
									Lower concretionary shale		Dowling	Dowling						
								Lower concretionary shale		Lower concretionary shale	Marshybank	Marshybank	Bad Heart	Bad Heart	Bad Heart fm.	Bad Heart	Bad Heart sandstone	
									Striped		Muskiki	Muskiki	SMOKY GROUP Cardium formation Baytree	SMOKY GROUP Cardium formation Baytree	SMOKY GROUP	Smoky River formation	Smoky River formation	
		Niobrara Benton formation				COLORADO GROUP		ALBERTA GROUP Cardium formation Upper member	ALBERTA GROUP Bighorn formation Upper sandstone	ALBERTA GROUP Cardium formation Upper sandstone	ALBERTA GROUP Cardium formation Sturrock	SMOKY GROUP Cardium formation Sturrock						
									Upper shale		Leyland	Leyland						Smoky River shales
"Cardium" shales and also "Ostrea" shales	Dark shales	Cardium sandstone	Bighorn formation	Cardium sandstone	Lineham member	Cardium formation	Cardium bands	Middle member	Middle sandstone	Middle shaly member	Cardinal							
									Lower shale		Kiska	Moosehound						
											Moosehound				Kaskapau formation	Kaskapau	Lower shale	
								Lower member	Lower sandstone	Basal sandstone	Ram	Ram						
								Transition	Transition	Transition	Opabin	Kaskapau fm. Opabin	Kaskapau fm.					
								Blackstone formation Rusty shale	Blackstone formation Hydrogen sulphide	Blackstone formation Rusty shale	Blackstone formation Haven	Haven	Kaskapau	Kaskapau				
			Blackstone formation			Lower Benton formation	Lower	"I. labiatus"	"I. labiatus"	"I. labiatus"	Vimy	Vimy						
												Sunkay						
								Barren	Barren	Barren	Sunkay	Dunvegan fm.	Dunvegan	Dunvegan	Dunvegan	Dunvegan	Dunvegan	Dunvegan
												"Ft. St. John"	FORT ST. JOHN GROUP	FORT ST. JOHN GROUP	FORT ST. JOHN GROUP	St. John fm.	St. John fm.	Ft. St. John
		"Dakota" fm.	"Dakota" fm.	"Dakota" fm.	"Dakota" fm.	Blairmore fm.	Blairmore fm.	Blairmore fm.	Mtn. Park fm.	Blairmore fm.	Blairmore fm.	Mtn. Park fm?						

GSC

Table 3-1. Nomenclature of Alberta Group and Equivalent strata, Alberta Foothills. From Stott, 1963.

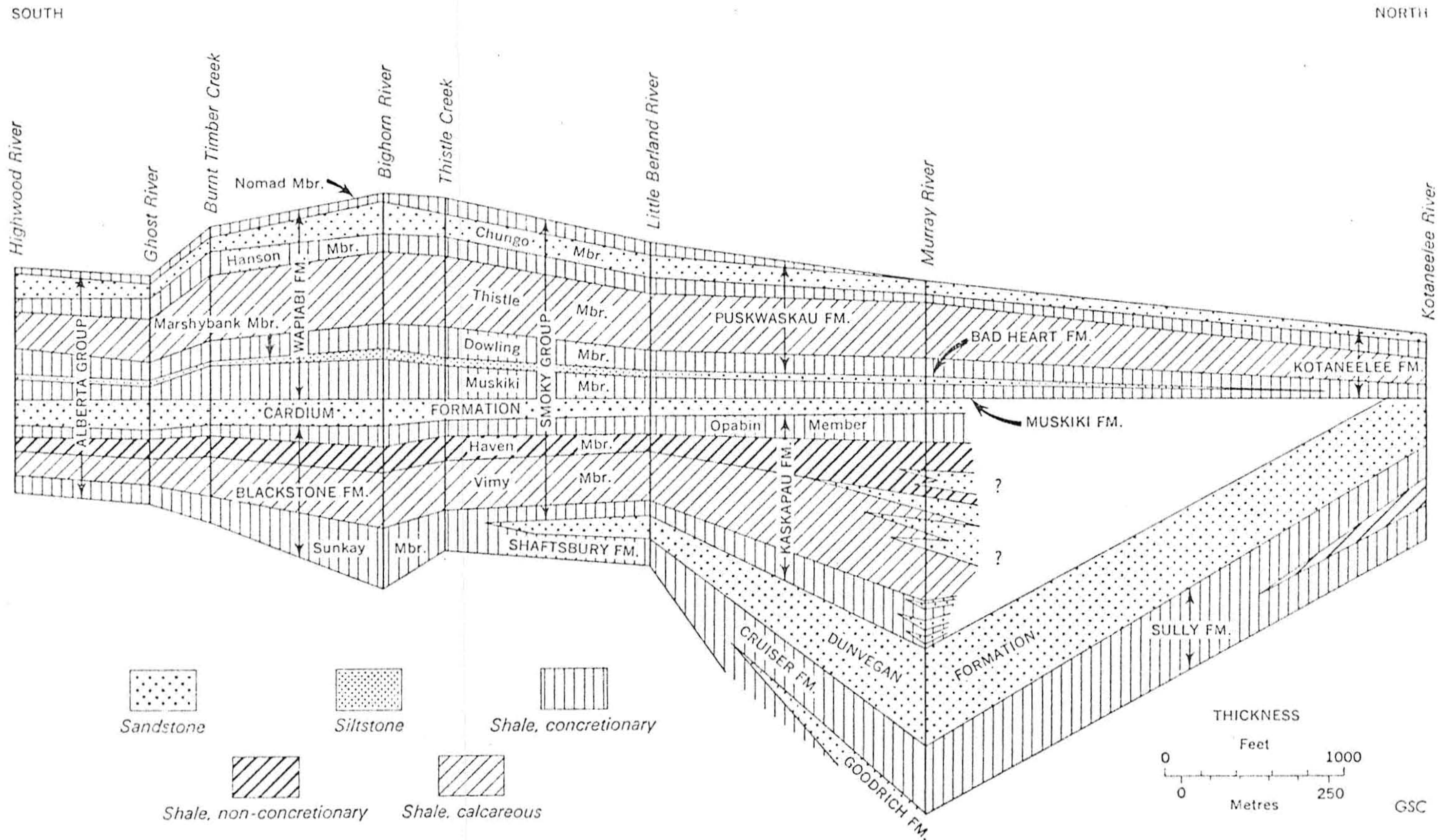

SOUTH
NORTH
Highwood River
Ghost River
Burnt Timber Creek
Bighorn River
Thistle Creek
Little Berland River
Murray River
Kotaneelee River
Nomad Mbr.
Hanson Mbr.
Chungo Mbr.
Thistle Mbr.
Marshybank Mbr.
WAPIABI FM.
ALBERTA GROUP
SMOKY GROUP
PUSKWASKAU FM.
BAD HEART FM.
Dowling Mbr.
Muskiki Mbr.
KOTANEELEE FM.
CARDIUM FORMATION
Opabin Member
MUSKIKI FM.
Haven Mbr.
Vimy Mbr.
BLACKSTONE FM.
KASKAPAU FM.
?
Sunkay Mbr.
SHAFTSBURY FM.
CRUISER FM.
DUNVEGAN FORMATION
SULLY FM.
GOODRICH FM.
Sandstone
Siltstone
Shale, concretionary
Shale, non-concretionary
Shale, calcareous
THICKNESS
Feet
0
1000
0
250
Metres
GSC

Figure 3-2. Diagram showing changes in stratigraphy and terminology of Upper Cretaceous formations from Athabasca River to Peace River. From Irish, 1965.

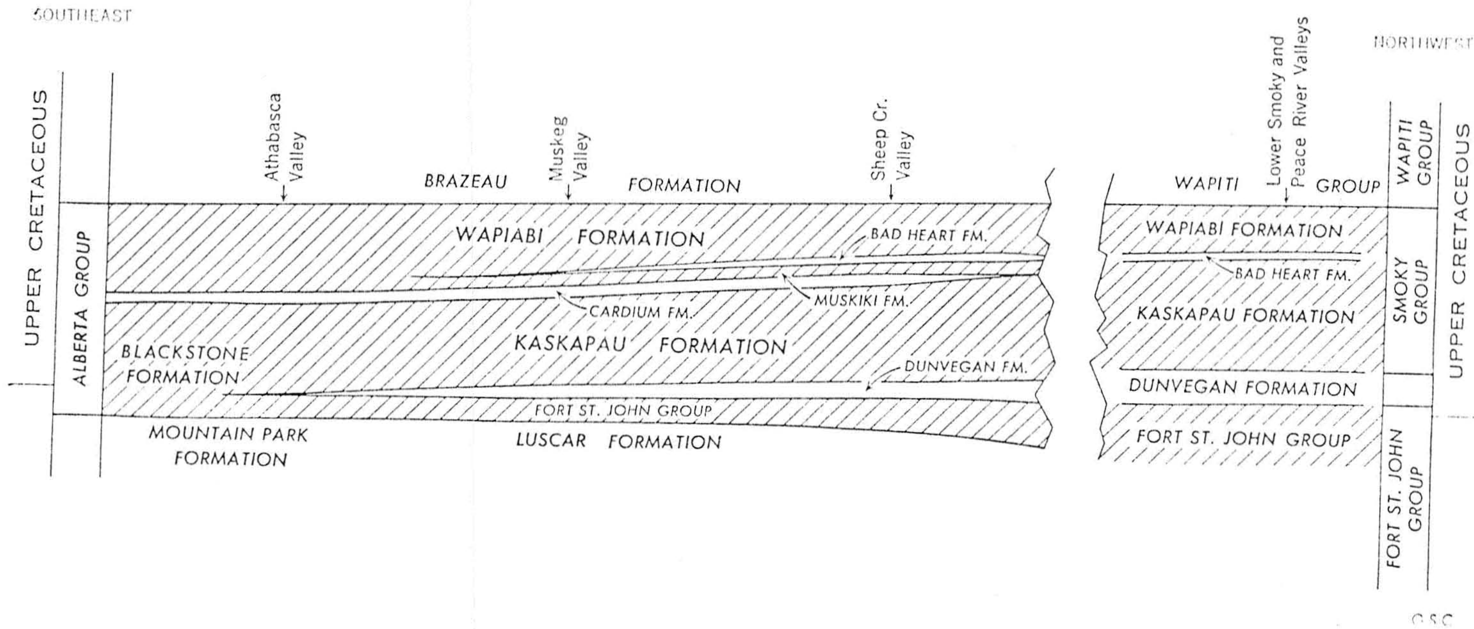
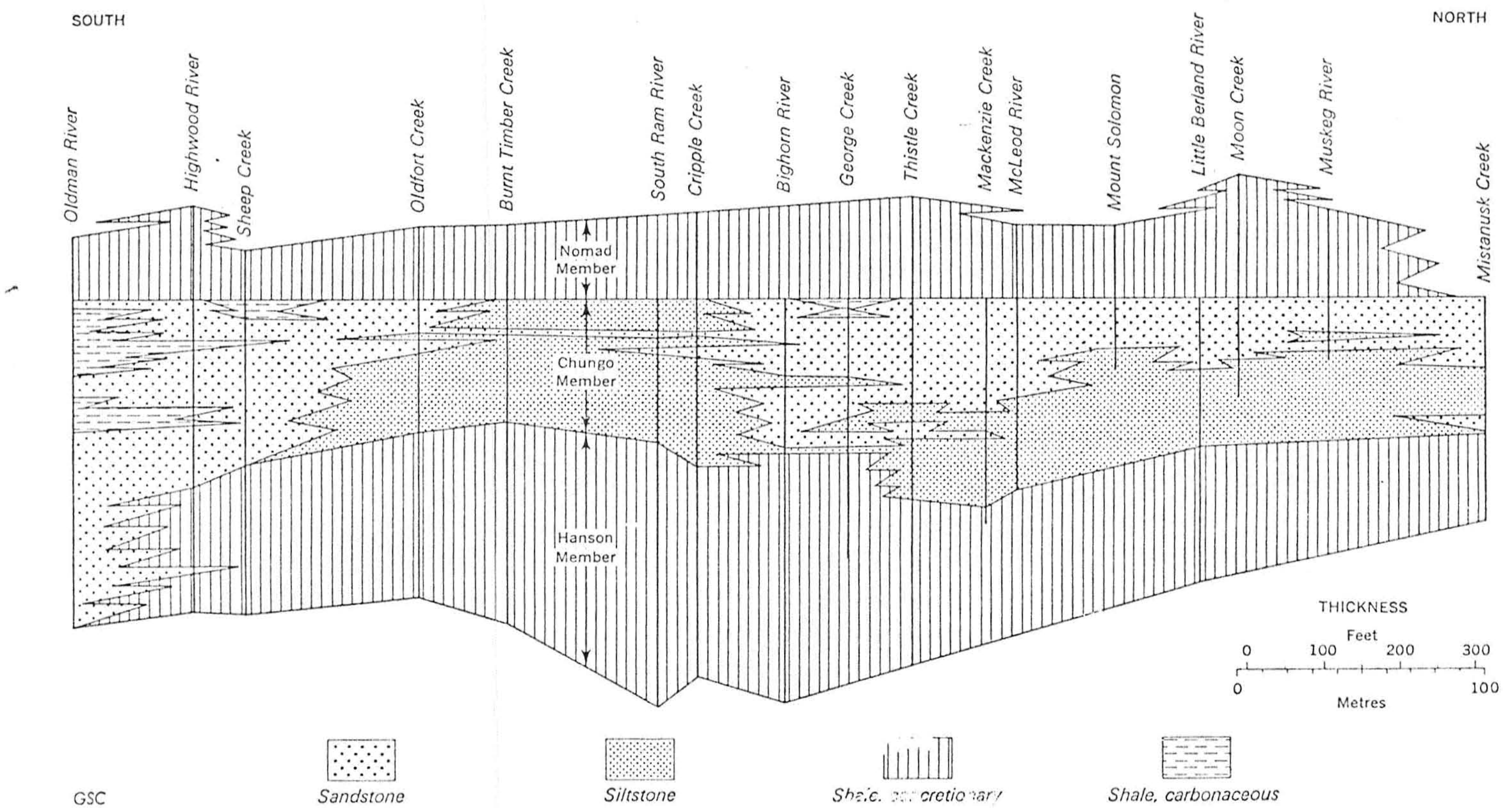

SOUTHEAST
NORTHWEST
UPPER CRETACEOUS
ALBERTA GROUP
Athabasca Valley
Muskeg Valley
Sheep Cr. Valley
Lower Smoky and Peace River Valleys
BRAZEAU FORMATION
WAPIABI FORMATION
BAD HEART FM.
MUSKIKI FM.
CARDIUM FM.
BLACKSTONE FORMATION
KASKAPAU FORMATION
DUNVEGAN FM.
FORT ST. JOHN GROUP
MOUNTAIN PARK FORMATION
LUSCAR FORMATION
WAPITI GROUP
WAPIABI FORMATION
BAD HEART FM.
KASKAPAU FORMATION
DUNVEGAN FORMATION
FORT ST. JOHN GROUP
WAPITI GROUP
SMOKY GROUP
FORT ST. JOHN GROUP
UPPER CRETACEOUS
GSC

Figure 3-3. Schematic diagram of Hanson, Chungo, and Nomad Members of the Wapiabi Formation, illustrating facies changes along the Foothills from southern to northern Alberta. From Stott, 1967.


SOUTH
NORTH
Oldman River
Highwood River
Sheep Creek
Oldfort Creek
Burnt Timber Creek
South Ram River
Cripple Creek
Bighorn River
George Creek
Thistle Creek
Mackenzie Creek
McLeod River
Mount Solomon
Little Berland River
Moon Creek
Muskeg River
Mistanusk Creek
Nomad Member
Chungo Member
Hanson Member
THICKNESS
Feet
0 100 200 300
0 100
Metres
GSC
Sandstone
Siltstone
Shale, carbonaceous

of the Paskwaskau Formation as proposed by Wall (1960) and favoured by Stott (1967). This is used mainly to avoid confusion of placing recognizably continuous units (i.e. the Thistle, Hanson, Chungo, and Nomad Members) under different formational names.

At the Little Berland River section, the thin lower sand unit which Stott (1963) refers to as the Marshybank Member is here taken as the most southerly extension of the Bad Heart Formation. This usage is supported by Irish (1965). The use of "Muskiki Formation" for the shales separating the Bad Heart Formation from the Cardium Formation (where the two are both recognized) is supported. The Little Berland River section is the only section in the study area where this terminology can be applied.

The vast amount of subsurface correlation in the central and northern plains of Alberta has resulted in common abuse of the present nomenclature system. The most blatant example in the Wapiabi Formation concerns the Chungo Member. From drilling tickets and well logs taken from the central Alberta region, four names were noted to have been assigned to a seemingly continuous and correlatable sand body: the Chungo, Solomon, Chinook, and the Milk River. The Solomon, Chinook, and Milk River terms are all legitimate Chungo equivalents in other areas. However, their application to the study area is inappropriate.

Other drilling terms can also be accommodated within the present nomenclature scheme. For example, the upper or first white specks zone is correlatable with the upper portion of the Thistle Member of the Wapiabi Formation (Stott, 1967).

It is concluded that the usage of "Wapiabi Formation" is appropriate for the entire study area and that within this area, the Thistle, Hanson, Chungo, and Nomad Members can easily be identified and traced. The Bad Heart Formation was identified only in the northern most section and was separated from the Cardium Formation by the Muskiki Formation which was lowered to member status where the Bad Heart Formation was not identified. The Dowling Member, although continuous throughout the study area, was difficult to discern from the Muskiki Member where it did not overlie the Bad Heart Formation. The use of "Marshybank Member" as an equivalent of the Bad Heart Formation to the south has been used cautiously in this study as recognition of the member at some sections was questionable.

CHAPTER 4 - PREVIOUS STUDIES OF THE WAPIABI FORMATION IN THE CENTRAL AND NORTHERN FOOTHILLS

Within the Wapiabi Formation there has historically been a distinct lack of studies of a strict sedimentological nature. This trend has begun to change in the last few years in the southern plains and southern Foothills areas. Several interpretations for specific outcrops of the Wapiabi Formation in the southern regions (particularly for the Chungo Member) have been proposed (Nelson and Glaister, 1975; Lerand and Oliver, 1975; Hunter, 1980; and Bullock, 1981). These studies have been reviewed and placed into a regional context in a recent sedimentological study of the Wapiabi Formation by Rosenthal (in prep.). Petrographic studies have also been undertaken in the southern region by Lerbekmo (1961) and Campbell and Lerbekmo (1963). However, these studies are largely oriented towards determining stratigraphic relations.

In the central and northern Foothills, the Wapiabi Formation has generally not had the attention of sedimentologically oriented studies. Little interpretive work has been done. Most of what is presently known in these areas is the result of classic stratigraphic studies by Stott (1963, 1967) and Irish (1965). Although these studies set the stage for the modern stratigraphic nomenclature, they shed little light on the sedimentological aspects of the Wapiabi Formation. Most of their interpretive work was based solely on lithology and paleontology. The only modern sedimentological study undertaken in the study area was by

Lerand (1982) who described and interpreted the Chungo Member at the Mt. Yamnuska section. His interpretation will be discussed at a later point.

One of the first environmental interpretations suggested for the Upper Cretaceous shale units was proposed by Dawson (1881) who worked in the Smoky River area. Dawson (1881) recognized the marine nature of the shales on the basis of the marine fauna. He also suggested that sandstone units identified within the shale sequence were deposited under shallow marine littoral conditions and were spread widely to the east. No evidence was given to substantiate the littoral interpretation.

In 1949, Gleddie made a specific reference to the depositional environment of the Chungo Member of the Wapiabi Formation (he used the name "Chinook Sandstone" in his study). Recognized was a littoral marine sandstone and sandy-shale member containing glauconite and having a thickness of 75' (23 m). The member occurred 90 to 100' (27 to 30 m) below the basal sandstone of the overlying Wapiti Formation (Belly River equivalent), the same stratigraphic level occupied by the Chungo Member. However, as in Dawson's (1881) interpretation, no evidence was presented to uphold the littoral interpretation.

The first thorough description of the Wapiabi Formation's lithology on a regional basis was presented by Stott (1963). Although the study was largely stratigraphic in its emphasis, it is also considered as a first attempt at a somewhat sedimentologically oriented study of the Upper Cretaceous Alberta Group assemblage. From this regional study, which extended from the southern Foothills north to Grande Cache, seven members were identified within the Wapiabi Formation which could be recognized as being regionally extensive. Two subsequent, but complementary studies of the Alberta Group equivalent in the north, the Smoky Group, were published by Irish (1965) and Stott (1967). The later study by Stott (1967) was slightly more interpretively oriented than the original Stott (1963) publication.

Table 4-1 gives brief descriptions of each of the seven members of the Wapiabi Formation identified by Stott (1963). A more comprehensive look at each member is presented here.

Muskiki Member

The Muskiki Member shale is the lowest depositional unit of the Wapiabi Formation. It has been identified across the entire study area and Stott (1963) suggests that the member thins in an overall south to east direction.

The Muskiki Member has been defined as containing all sediments between the underlying fine-grained sandstones of the Cardium Formation and the overlying fine-grained sandstones to massive siltstones and shales representative of the Marshybank Member. The member largely consists of dark shales with a varying proportion of terrigenous clastic material of silt to sand size. The shales often show a well developed fissile nature. However, localized intervals are notably rubbly or blocky and somewhat sideritic. Sideritic concretions have been observed at all measured sections and may make up a substantial portion of the member (up to 5%). Bentonite intervals are also characteristic of the Muskiki Member.

Table 4-1. Brief descriptions of each of the seven Wapaibi Formation members. From Stott, 1963.

Table of Formations

Southern and Central Foothills				Description	Northern Foothills			
Series	Group	Formation	Member		Member	Formation	Group	Series
			Nomad 90′–130′	Rusty weathering, rubbly shales, grading upwards into greenish grey shales and fine-grained, thinly bedded sandstones. Base is marked by band of pebbies.	Nomad 100′ ±			
			Chungo 135′–416′	Fine-grained, thickly bedded, light brown weathering sandstones (lithic arenites to quartz wackes), and dark grey siltstone with reddish brown weathering concretions.	Chungo 205′ ±			
			Hanson 0–232′	Dark grey, rusty weathering, blocky to rubbly shales, with reddish brown weathering sideritic concretions.	Hanson ?			
		Wapiabi 1,043′–2,146′	Thistle 384′–778′	Dark grey to black, calcareous, platy to fissile shales, weathers grey to light grey, with thin, dense, bluish grey dolomitic beds.	Thistle 650′ ±	Wapiabi 1,550′ ±		
			Dowling 101′–351′	Dark grey, rubbly to platy shales, weathers rust, with reddish brown weathering sideritic concretions.	Dowling 250′ ±			
Upper Cretaceous	Alberta 2,000′–4,100′		Marshybank 41′–104′	Dark grey, massive, argillaceous siltstone, with large reddish brown concretions; siltstone grades into sandstone.	Marshybank ?		Smoky 2,900′	Upper Cretaceous
			Muskiki 144′–325′	Dark grey, rubbly to platy shales, weathers rust and has banded or striped appearance, some reddish brown sideritic concretions. Bed of coarse-grained, pebbly sandstone or pebble-conglomerate at base.	Muskiki 250′–275′			

The lower contact of the Muskiki Member with the Cardium Formation is considered by Stott (1963) to be unconformable and represents a sharp break in sedimentation. This break possibly represents the beginning of the Wapiabi transgression (Stott, 1963). The contact with the overlying Marshybank Member, however, is gradational with the Muskiki Member shales grading upwards into massive siltstones and laminated sandstones.

The basal 1 to 3 m of the Muskiki Member has been noted to be somewhat gritty in nature. Stott (1963) has identified sandstone and chert pebbles in this zone and considers the deposits to be associated with the advancing Wapiabi Sea rather than with the last phases of the regressing Cardium Sea. Stott (1967) states that these transgressive deposits are marked by siderite and glauconite which suggest a shallow offshore depositional environment with mildly oxidizing to mildly reducing conditions.

The earliest deposits of the Muskiki Member have been considered by Stott (1963, 1967) to be of late Turonian age. However, the bulk of Muskiki deposition took place during the Coniacian. Upper Muskiki sediments in the southern portion of the study area have been dated as young as early Santonian (Stott, 1963).

Marshybank Member

Thin shales and structureless siltstones or laminated sandstones with interbedded shales characterize the Marshybank Member. The facies change from sandstone to siltstone occurs slightly south of the Little Berland River section. The siltstone facies then continues to the southern portion of the study area.

The lower contact of the Marshybank Member with the Muskiki Member is gradational in nature and is difficult to define on some sections. The upper contact is normally distinct with the overlying Dowling Member where a facies change from siltstone to dark shale occurs. Chert pebbles have been noted by Stott (1963) to define the contact at a few sections.

On a regional basis along the foothills margin, the Marshybank Member varies little in thickness. However, a general thinning trend has been reported towards the south and east (Stott, 1963).

Lithologically, structureless, buff coloured siltstones dominate the Marshybank sediments in the central Foothills. The siltstones are argillaceous and contain rusty coloured sideritic concretions in varying proportions. The buff coloured sandstones to the north are very fine to fine-grained and show well developed laminations. Sandstone beds are separated by thin beds of structureless silty-shale or shale.

Deposition of the Marshybank Member has been suggested by Stott (1967) to have occurred during an early Santonian regression that took place along the entire length of the Foothills region. Identification of _C. pauperculum_ (usually found in nearshore deposits) by Stott (1963) prompted an interpretation of the depositional environment for the Marshybank Member. Deposition within "relatively shallow water" was suggested (Stott, 1967).

Dowling Member

Thinly laminated and dark gray to rusty coloured shales are the major constituents of the Dowling Member. Only minor amounts of terrigenous siltstone and sandstone are recognized. Bands of sideritic concretions along bedding planes are quite common and may make up a few percent of the member.

The lower contact of the Dowling Member with the Marshybank Member is generally distinct with a sharp lithologic change from siltstone or sandstone to shale. However, Stott (1963) suggests that the contact is conformable. The upper contact is indistinct and occurs where the sideritic and concretionary Dowling Member shales grade into the calcareous, non-concretionary shales of the Thistle Member.

Faunal studies in the Dowling Member by Stott (1963) showed that the member was probably deposited during the late-early Santonian. The identification of Scaphites depressus in the lower portion of the member and Scaphites vermiformis in the upper part suggested to Stott (1967) that the receiving basin was deepening with time. However, Stott (1967) placed the entire Dowling Member within a coastal environment of deposition.

Thistle Member

The Thistle Member of the Wapiabi Formation is by far the thickest depositional unit within the formation, reaching a thickness of over 220 m on the Wapiabi type section of Thistle Creek. The member is largely characterized by platey calcareous shales which are given a banded appearance by a significant proportion of well defined, thin (less than 5 cm) siltstone beds. Sandstones are also locally recognized. As well as being a dominant lithologic unit in the Wapiabi Formation, the Thistle Member is one of the more regionally persistent units, ranging from well into the foothills of British Columbia in the north (where it is recognized as a member of the Puskwaskau Formation) to the international border in the south.

The lower contact of the Thistle Member with the underlying Dowling Member was suggested to be gradational and conformable (Stott, 1963). The upper contact with the Hanson Member was also suggested to be gradational and conformable. Both contacts are characterized by a change from calcareous and non-concretionary shales of the Thistle Member to sideritic and concretionary shales of the Dowling and Hanson Members.

Shales of the Thistle Member are generally fissile to platey with localized zones of shales which appear rubbly or blocky. They are black to dark gray in colour and are calcareous. Stott (1963) has reported many occurrences of lensoid bodies of grayish-yellow dolomitic limestone within the shales. These bodies are of limited lateral extent and average 0.6 m by 1.3 m. Stott (1967) considers these deposits to be indicative of periods of little terrigenous input.

Siltstones within the Thistle Member weather a gray colour, are well laminated, and rarely exceed 5 cm in thickness. Bases are quite sharp, while tops are less well defined. These siltstones make up, volumetrically, a very significant portion of the Thistle Member as localized horizons contain over 50% siltstone. Correlative coarsening upward sequences recognized by Stott (1963) within the Thistle Member tend to be capped by these high siltstone proportion zones.

Stott (1967) has proposed a mid-basin environment of deposition for the Thistle Member. This interpretation is based on the occurrence of calcareous shales which Stott believes to be indicative of relatively deep depositional environments. A reducing bottom environment is also suggested by an occurrence of pyrite.

Faunal studies of the Thistle Member indicate that deposition took place during the middle to late Santonian. These studies also show the upper portion of the Thistle Member to be of equivalent age across the entire Foothills region.

Hanson Member

The upper concretionary shale unit of the Wapiabi Formation, known as the Hanson Member, is a regionally variable unit which tends to thin out in a easterly direction. Lithologies which characterize the member range from totally structureless, rusty to buff coloured, blocky to rubbly weathered sideritic shales and argillaceous siltstones to fissile laminated, buff to dark gray silty-shales and siltstones with well laminated discrete sandstone beds up to 10 cm thick. Sideritic concretions are a significant constituent of this member.

The upper contact of the Hanson Member with the overlying Chungo Member is considered to be at the first discrete sandstone bed greater than 10 cm thick. However, this definition is merely arbitrary and the lithologies suggest a gradational and conformable transition. Because of this definition, the stratigraphic level of the upper boundary and the thickness of the member is variable from section to section. The lower contact with the Thistle Member is indistinct and occurs where calcareous and non-concretionary Thistle Member shales grade into the sideritic and concretionary Hanson Member shales.

Citing the concretionary nature of the Hanson sediments as supportive evidence, Stott (1967) suggested that the member was deposited within a well oxygenated environment of nearshore muds.

Identification of minor amounts of glauconite was also used as evidence in favour of this interpretation. An overall coarsening upwards of Hanson sediments from silty-shales to siltstones and sandstones suggested that the Hanson Member marked the beginning of the major, regionally extending, regressive cycle that occurred towards the end of Santonian time. Fossil evidence by Stott (1963) places the Hanson Member as late Santonian in age.

Chungo Member

The sedimentology of the Chungo Member of the Wapiabi Formation forms the emphasis of the present study. This member is defined as those beds of shale, siltstone, and sandstone lying between the concretionary siltstones of the Hanson Member and the dark marine shales of the overlying Nomad Member (Stott, 1963). The Chungo deposits are generally marine in origin, but non-marine beds are present in the uppermost part of the member in several sections.

The lower boundary with the Hanson Member is gradational and conformable and exists at various stratigraphic levels. The upper contact with the Nomad Member is quite distinct, especially where the member is well defined sandstone. This boundary is often marked by a regionally persistent, clast supported to matrix supported, chert pebble conglometate bed which separates the distinct facies change from the Chungo Member to the Nomad Member. The contact is unconformable and marks a major late Santonian or early Campanian marine transgression (to be discussed later in detail).

The Chungo Member is regionally very extensive. It has been identified well into northeastern British Columbia in the north and extends far to the south where it is correlated with the Milk River Formation of southern Alberta and the Eagle Formation of Montana.

The marine sediments represented in the Chungo Member vary on a regional basis and range from buff to rusty coloured, very-fine to medium grained sandstone beds with well developed sedimentary structures to massively weathered or structureless sandstone, to buff coloured, argillaceous, concretionary siltstone. Minor amounts of coal and carbonaceous coarse sandstones are typical of the non-marine facies. The sediments in the Chungo Member have generally been deposited in the form of a single coarsening upward sequence and range from the concretionary sandy-siltstones near the base of the member to the fine to medium grained sandstones at the upper limit. The pebble conglomerate lying at the top of the coarsening upwards sequence is not suggested as the culmination of the regressional coarsening upward sequence. It will be proposed later that the genesis of the conglomerate is related to other depositional processes.

The overall thickness of the Chungo Member is variable from section to section. Where the unit is dominated by siltstone deposition rather than sandstone deposition (e.g. Burnt Timber Creek and Ghost Dam), the member is relatively thin. The sections of maximum thickness (e.g. Thistle) are characterized by well developed sandstones facies and coarsening upward sequences.

Paleontological dating by Stott (1963) indicated that deposition in the Chungo Member took place at least until the end of Santonian time. Evidence was inconclusive as to whether or not sedimentation continued into the early Campanian. In an environmental interpretation, Stott (1963) proposed that Chungo deposition climaxed during a late Santonian regressional period. The Chungo sediments that were deposited during this regression were deposited in proximal nearshore environments, barrier island and lagoonal environments, and non-marine environments (Stott, 1967). Glauconite and siderite were suggested as indicators of shallow marine deposition and carbonaceous material (particulary coal) was proposed as the main non-marine indicator.

The present study will show that while the Chungo Member was deposited largely within a shallow marine environment, a barrier island-lagoonal system as proposed by Stott (1967) was probably not developed. Instead, the Chungo coast can be described in terms of a storm dominated, progradational shoreline with some local tidal influence.

Nomad Member

The uppermost member of the Wapiabi Formation, the Nomad member, contains all sediments between the top of the Chungo Member and the base of the Belly River (Brazeau) Formation. The member is characterized as an interval of platey to rubbly shales with a regionally variable proportion of interbedded discrete siltstone beds.

The lower boundary with the Chungo Member is quite distinct and is defined at the top of the laterally extensive pebble conglomerate bed. The upper contact is drawn at the base of the Belly River (Brazeau) Formation. This contact is gradational. Stott (1963) suggested that the contact generally is somewhat higher stratigraphically in the east than in the west. Stott (1963) also suggested that the lower contact is unconformable (based on a lithologic change from non-marine and shallow marine sediments to deep marine sediments) and the upper contact is conformable.

The shales of the Nomad Member are typically dark gray to black in colour and exhibit fissile lamination. Localized horizons of rubbly or blocky shales occur. Discrete siltstone beds and rare discrete sandstone beds characterize the upper portion of the member in some sections. These beds are generally rusty coloured, sharp based, well laminated, and range from 1 cm to greater than 10 cm. They occur within a coarsening and thickening upward sequence.

The Nomad Member is the least variable Wapiabi Formation member in terms of thickness. In the study area, measured sections are all between 30 m and 40 m. This also tends to hold for the entire Foothills region. However, the member does tend to thicken towards the east as the result of a thinning of the Chungo Member and a facies change at the base of the Belly River (Brazeau) Formation (Stott, 1963).

Deposition of the Nomad Member presumably took place during early Campanian times (Stott, 1967). Stott (1967) suggested that this time was characterized by a brief marine transgression that spread westward beyond the Smoky River region and southward into the southern Foothills. Apparently, this sea did not advance as far west as the British Columbia Foothills. Marine conditions did not last long as tectonic events to the west caused vast amounts of coarse clastic material to be shed eastward over the Nomad Basin.

Although the lithostratigraphic studies of Stott (1963, 1967) should be considered as classic works on the Upper Cretaceous molasse sediments of Alberta, the studies should also have been considered as an impetus for modern sedimentologically oriented studies. However, such studies were not undertaken on the Wapiabi Formation in the present field area until Lerand (1982) thoroughly described and interpreted the Chungo Member at Mt. Yamnuska. The measured section of the Chungo Member by Lerand (1982) is provided in Figure 7-1.

Lerand (1982) has described the Mt. Yamnuska section in terms of two possible environments of deposition: a beach environment and an offshore bank environment. In the beach interpretation, the trough cross bedding in unit 3 is said to "record longshore currents generated by dominantly northerly onshore winds, with the reversals accounted for by either occasional southeasterly storms or seasonal changes in wind regime." This unit overlies a sequence of storm dominated features. However, no foreshore sands were noted by Lerand to support a beach environment. He proposed that the absence of foreshore sands could be accounted for by an erosional degradational interval.

In the offshore bank model, Lerand (1982) envisions an analgous situation to the present North Sea where the trough cross-bedding of unit 3 may represent linear, tidally induced, sand banks separated by troughs. These tidal processes were suggested to have been mixed with storm processes to produce unit 2b. All deposition below this level was proposed to have been purely storm generated.

The present sedimentologically oriented study on the Wapiabi Formation is an attempt to expand on the basework provided by Stott (1963, 1967). The main intent of the project is to propose depositional environment and process interpretations for each of the measured sections in the study area. Individually interpreted sections will then be reviewed and incorporated into an overall regional perspective. The Mt. Yamnuska interpretations by Lerand (1982) will also be reviewed.

It is hoped that the results of this study will shed some light on the depositional environments and processes that were acting during deposition of the Wapiabi Formation, basinal geometries during deposition, and sediment dispersal patterns influencing sedimentation.

CHAPTER 5 - DESCRIPTIONS OF FACIES

The application of the term "sedimentary facies" to a particular horizon or level within a stratigraphic unit is based on the identification of several criteria which distinguish that horizon or level from other parts of the stratigraphic unit (Moore, 1949). These criteria include lithology, sedimentary structures, colour, bedding type, texture, and fossils (Reading, 1978). Once a facies has been identified on the basis of these characteristics, inferences can be made concerning the specific conditions of sedimentation under which the facies were deposited. In this study, use of the term "facies" is applied only in a descriptive manner rather than in an interpretive manner.

Facies Descriptions

Facies in this study were defined on the basis of four main characteristics: sedimentary structures (i.e. trough cross-bedded facies), composition (lithology), bedding (i.e. shale interbedded with discrete sandstone/siltstone beds less than 5 cm thick), and the degree of bioturbation. As a rule, if the finer grained shales dominated, facies based on lithologies and discrete sandstone bed thicknesses were assigned. The coarser grained material was divided on the basis of internal sedimentary structures.

Definition and division of the facies presented in this study were largely determined in the field. For division of the vast thicknesses of shale in the Wapiabi Formation, sand-shale ratios were visually estimated in the field and expressed in terms of percentages. The shale was then divided into three arbitrary facies based on variations of the silt and sand content. The coarser facies were all defined from direct field observations of sedimentary structures.

This chapter is intended to be purely descriptive in nature and discusses the characteristics of each facies separately. No hydrodynamic, genetic, or environmental interpretations are presented at this point.

The grain size of sandstones that are referred to in this study were recorded according to the Canstrat grain size scale. For convenience, the conversion from the Canstrat scale to a millimeter scale is provided here:

very coarse upper (vc_u)	-	1.41 - 2.00 mm
very coarse lower (vc_l)	-	1.00 - 1.41 mm
coarse upper (c_u)	-	.710 - 1.00 mm
coarse lower (c_l)	-	.500 - .710 mm
medium upper (m_u)	-	.350 - .500 mm
medium lower (m_l)	-	.250 - .350 mm
fine upper (f_u)	-	.177 - .250 mm
fine lower (f_l)	-	.125 - .177 mm
very fine upper (vf_u)	-	.088 - .125 mm
very fine lower (vf_l)	-	.062 - .088 mm

FACIES 1 Shale with various proportions of discrete siltstone and sandstone beds less than 5 cm thick

The Muskiki, Marshybank, Dowling, Thistle and Nomad Members of the Wapiabi Formation are generally characterized as being made up of thick sequences of black to gray shale with minor amounts of interbedded siltstone and sandstone. The variation in colour from black to light gray reflects the relative proportion of shale to siltstone and sandstone. The shale dominated zones are black to dark gray while zones with interbedded siltstones and sandstone become lighter gray with an increasing percentage of sand and silt sized material. Diagenetic alterations have imposed a rusty colour on some silts and sands while bentonite gives a white speckled appearance to several shale horizons in the lowest three members.

The more resistant-weathering siltstones and sandstones are normally deposited in sharp based beds less than 5 cm thick. The change from siltstone to very fine sandstone in these beds was difficult to determine in the field. Therefore, they were grouped together under most circumstances. Small amounts of silt sized material were also noted disseminated within shale horizons. This caused the shale to be slightly gritty in nature.

The Wapiabi shales, which may make up to 80% of the entire Wapiabi Formation, have been broken down into three subfacies based on the percentage of interbedded siltstone and sandstone. These subfacies are herein designated as subfacies 1a, 1b, and 1c. In some cases, the change from one subfacies to another occurs over stratigraphic intervals of only a few tens of centimeters. However, the contacts are normally gradational over several meters and are very difficult to determined.

Subfacies 1a Shale with less than 10% interbedded discrete siltstone and sandstone beds less than 5 cm thick

The sediments which make up subfacies 1a are characteristically black to dark gray shales with fewer than 10% discrete siltstone and sandstone beds individually less than 5 cm thick. This subfacies occupies a volumetrically significant portion of the lowest three members of the Wapiabi Formation and is a minor constituent of the Nomad Member (Figure 5-1).

Lithologically, the shales which dominate this subfacies are generally devoid of silt and sand sized material with only rare occurrences of scattered silt sized material incorporated in with the shales. These horizons may form thick (up to tens of meters) continuous sections. Contacts are generally gradational over a few meters. However, bottom contacts have been noted to change over several centimeters. Typically, the shales are quite fissile and are easily broken into thin flakes as small as 1 mm thick. This fissile nature can change abruptly into localized zones of a micro-blocky texture which are very rusty in colour. Shales in these bioturbated zones tend to break into angular fragments which are up to 1 cm across. Although no burrows were noted, bioturbation plays a moderate role in subfacies 1a. Body fossils, which include several examples of ammonites shells, have also been found in this subfacies.

The discrete beds of lighter gray to buff siltstone and sandstone within this subfacies are usually less than 3 to 4 cm thick and average approximately 2 cm. Silt sized material dominates

Figure 5-1. <u>Subfacies 1a</u>. Typical fissile to rubbly weathered black shale. Note the small proportion (less than 10 percent) of interbedded discrete siltstone beds. The thick "bed" in the middle of the photo is a sideritic concretion band. From Cripple Creek section.

significantly over sand size sediment. The beds are always sharp based and often show well developed parallel lamination. Some of the thicker beds show current ripple lamination development on top of parallel lamination. Amplitude of the ripples ranges from 0.4 cm to 1.1 cm. Wavelengths average on the order of 10 cm to 12 cm.

Often noted within this facies is the development of sideritic concretions which have been measured up to 20 cm thick by 70 cm in length. These diagenetic features are normally disk shaped and form in distinct bands. However, examples of randomly distributed concretions have been found.

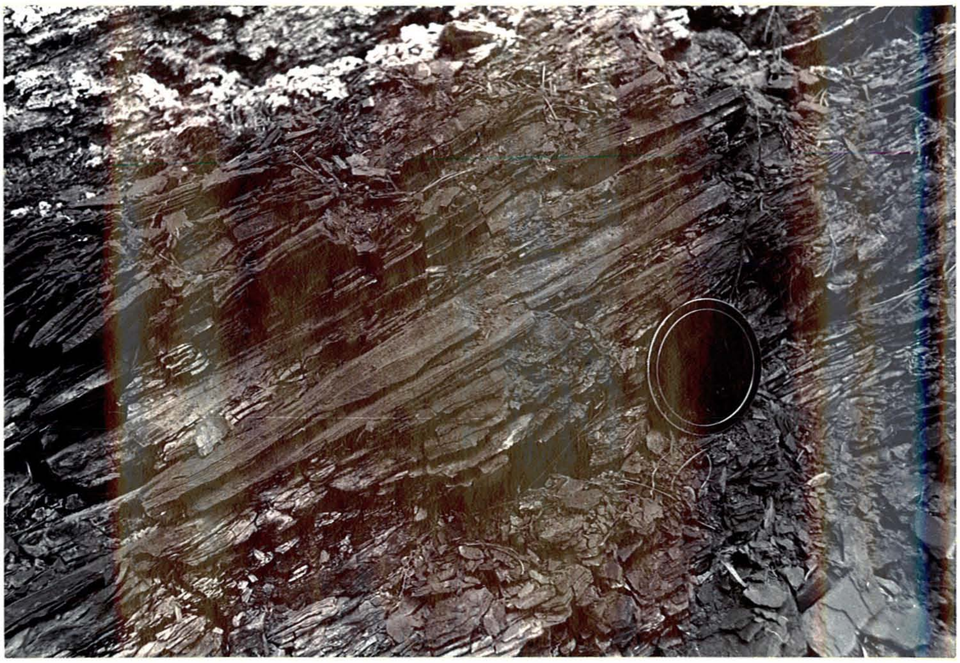
Subfacies 1b Shale with 10 to 50% discrete siltstone and sandstone beds less than 5 cm thick

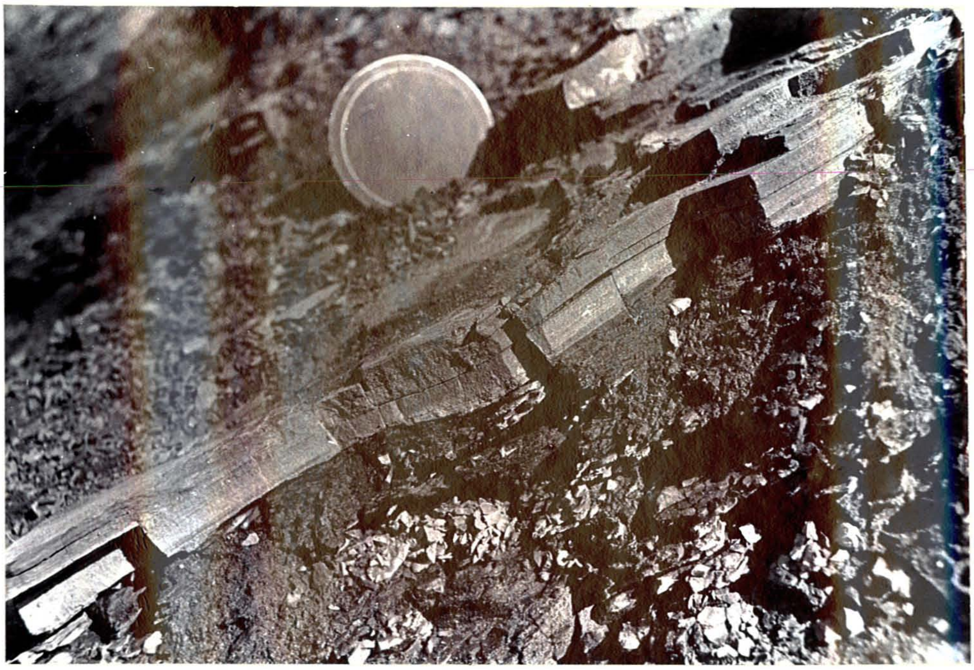
This subfacies makes up, volumetrically, the major portion of the Thistle Member. It is characterized by shale to silty-shale with 10 to 50% discrete siltstone and sandstone beds less than 5 cm thick (Figure 5-2). These discrete beds provided the bulk of paleocurrent data measured in this study (Figure 5-3).

The shales of this subfacies tend to vary from quite "clean" (no coarse fragment within) and black to silty and light gray to buff coloured. Horizons of the latter are, in places, up to 18 m thick and are generally associated with discrete siltstone and sandstone beds. Weathering of the shales normally produces a fissile nature with the shale breaking into plates 1 mm to 2 mm thick. Bioturbation in this subfacies (particularly in the Thistle Member) is generally negligible. However, rare local horizons do show massive appearing, totally bioturbated shales. Although no trace fauna were identified, several examples of ammonite shells were found and one unidentifiable clam

Figure 5.2. Subfacies 1b Typical unit of subfacies 1b. Note that the sharp bases of many of the more resistently weathered siltstone beds scour into the shales below. Lens cap is 67 mm across.

Figure 5-3. Subfacies 1b Closeup of a single 2 to 3 cm thick siltstone turbidite. Note sharp base. Laminations at the base are parallel (Bouma division b). Upper laminations are undulatory to current rippled (Bouma division c). From Muskiki Member, Cripple Creek section.





fossil was noted.

The interbedded, sharp-based, 2 to 5 cm thick, siltstone and sandstone beds of subfacies 1b vary from showing well developed internal stratification with no bioturbation to beds which show poorly defined lamination and a low degree of bioturbation. Gyrochorte, Ophiomorpha, and Rhizocorallium have been identified from the bases of these beds. The major sedimentary structure exhibited by beds is parallel lamination. Several light buff coloured 5 cm thick silty beds showing nothing but parallel lamination were noted. Current ripple lamination on top of parallel lamination was also very common, but was restricted largely to light gray coloured, 2 to 3 cm thick beds of slightly more sandy nature. The sandstone to siltstone ratio has increased in this subfacies to appoximately 1:3 or 1:2. Overall, sand sized sediment was found only within the smaller 2 to 3 cm thick beds while silt sized material was contained in both the 4 to 5 cm beds and the smaller beds.

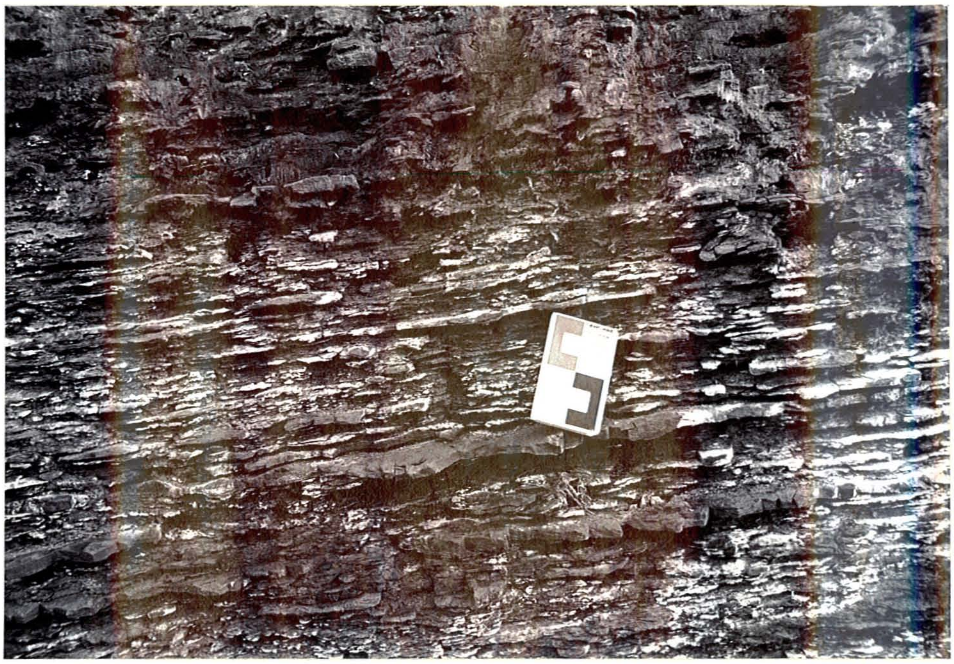
Subfacies 1c   Shale with greater than 50% discrete silt and sandstone beds less than 5 cm thick

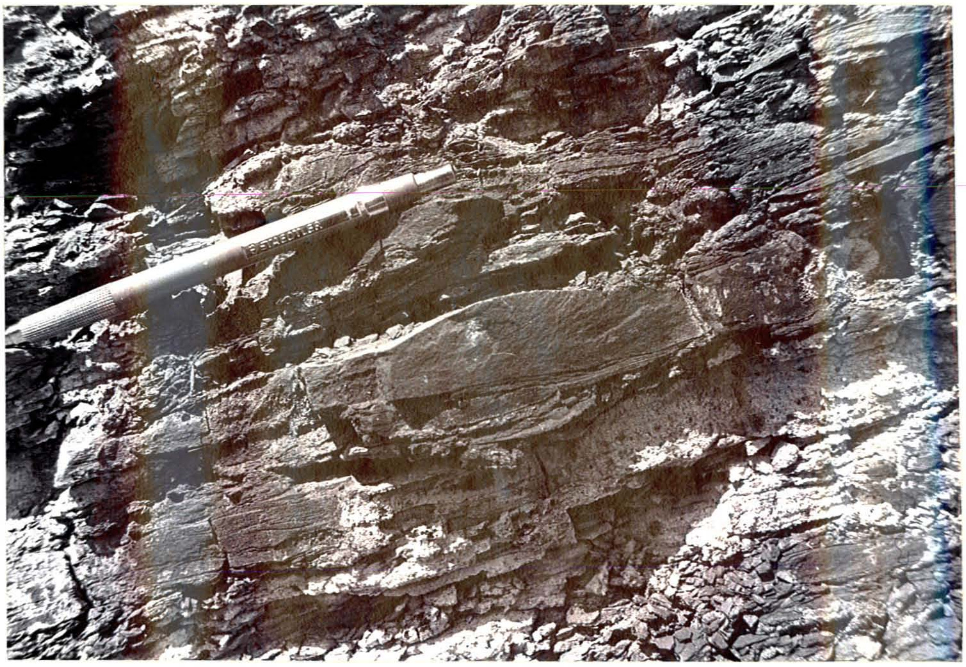
The deposits of subfacies 1c are largely composed of moderately bioturbated siltstone and sandstone beds (less than 5 cm thick) with less than 50% interbedded moderately to well bioturbated shales. The overall light gray to buff colour is an indicator of the increased sand and silt content. This subfacies is commonly found in the upper Thistle and Hanson Members (Figures 5-4 and 5-5).

The shales are largely gritty in nature and are gray to rusty in colour. Disseminated silt particles may make up to 10 or 20% of the shale. Weathering is normally blocky and structureless due to

Figure 5-4. Subfacie 1c Black shales with greater than 50% interbedded discrete siltstone beds. Differential weathering of the shales and the siltstone turbidites results in the protruding nature of the siltstones. Note the generally well defined outline of the siltstone beds. This suggests that bioturbation was not effective at this particular locality. From Bighorn River section.

Figure 5-5. Subfacie 1c Closeup of an individual siltstone turbidite. Note current ripple lamination at top of bed. Faint internal laminae suggest that flow was to the left. "Popcorn" textured material in lower right of photo is bentonite.





bioturbation. Diagenetic sideritic concretions are common in subfacies 1c in the Hanson Member.

The sandstone and siltstone beds of subfacies 1c show a much larger, and slightly more diversified faunal content than the previously described subfacies. The trace fossils Gyrochorte, Ophiomorpha, and Rhizocorallium have been identified along with noted examples of disarticulated body fossils which include brachiopods and pelecypods. As a result of increased biogenic reworking, the internal structure of many beds is poorly defined although bed discreteness has been maintained. Several cases show the top 0.5 to 2 cm of beds to be bioturbated while leaving parallel lamination at the base of the beds intact. On the Blackstone River section, several examples of oscillation rippled sand beds were found. The overall sandstone to siltstone ratio in subfacies 1c has increased to approximately 1:1. The maximum grain size recorded was very fine (upper) sandstone.

Although considered a separate subfacies defined by the above criteria, subfacies 1c contains within it several examples of sandstone and siltstone beds greater than 5 cm and thick zones of bioturbated mudstone. These features have been defined in this study to be part of facies 2 and facies 3. Therefore, the distinction between these facies is often vague and they are often interbedded with one another.

FACIES 2 Bioturbated Siltstone/Mudstone

The thoroughly bioturbated siltstones and mudstones that characterize facies 2 are generally found in the lower Chungo and Hanson Members. Colour varies little from buff to rusty with only a few light gray shales. Some horizons within the facies can be highly concretionary (sideritic concretions), thus giving an overall rusty coloured stain to the rocks. Most of these concretions formed in distinct layers up to 10 cm thick rather than in a random distribution throughout the facies. Outside morphologies vary from spheroid to disk to amorphous. Internal growth is generally concentric about a middle, occasionally an ammonite shell.

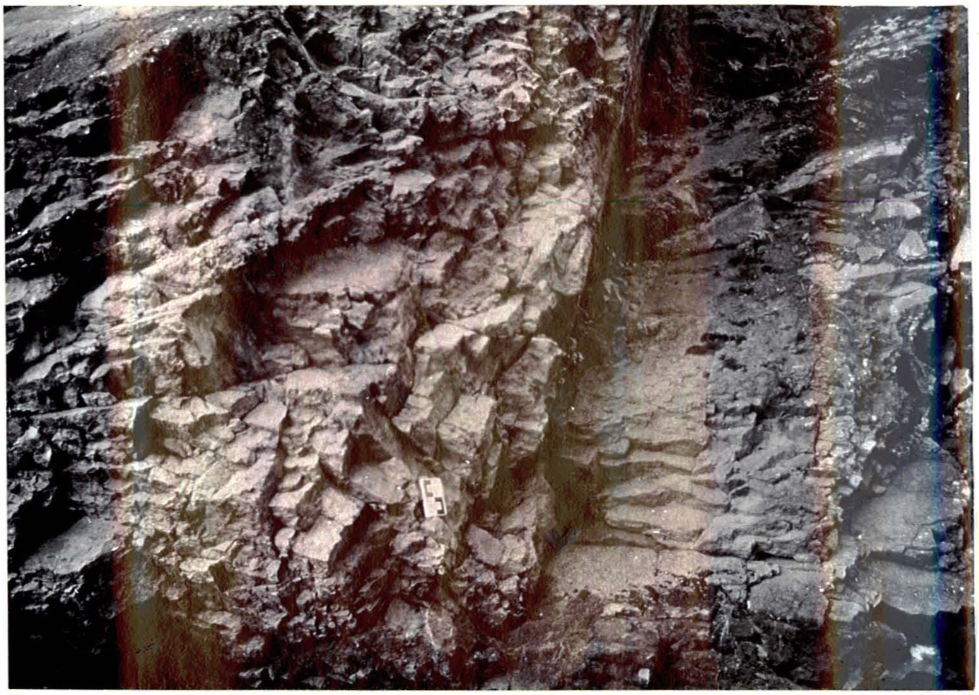
Weathering of the silts and muds yielded both blocky and massive/structureless appearing zones (Figure 5-6 and 5-7). Fractures in the blocky material often appeared as imitation bedding planes. Strike and dip measurements were next to impossible to obtain in this facies.

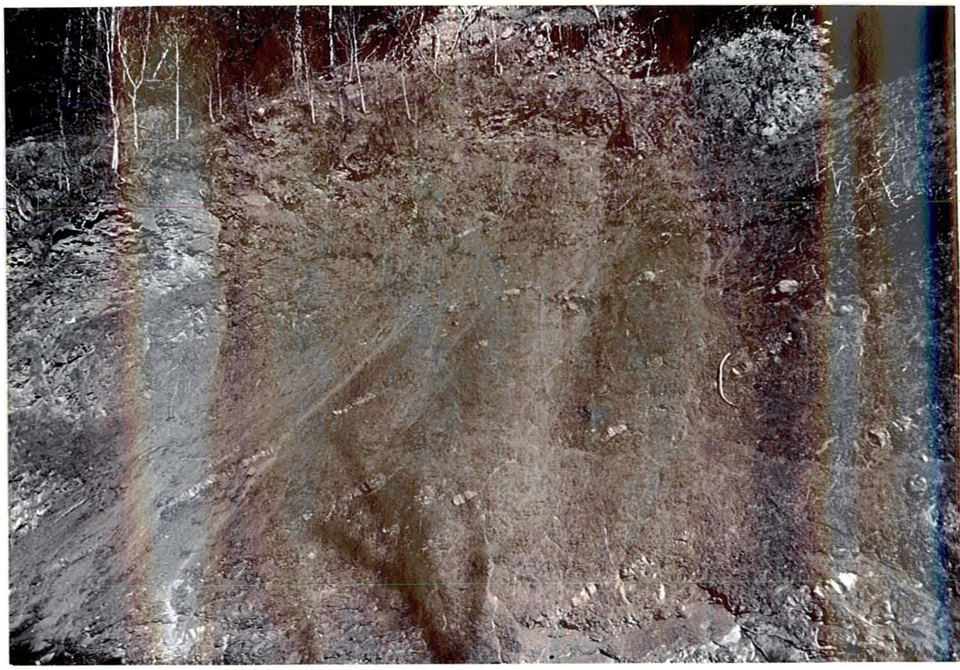
By definition, this facies contained a large and active fauna that completely destroyed any primary bedding features. Because of this complete bioturbation, identification of trace or body fossils was quite difficult. Trace fossils identified were Ophiomorpha, Gyrochorte, and Rhizocorallium. As well, several ammonite shells and one complete belemnite shell were recovered.

Although facies 2 is regarded to be a completely distinct facies, it is sometimes associated with facies 1c, 3, and 5. On Blackstone River, one well developed HCS bed of facies 5 was noted within facies 2. This suggests that the upper contact is poorly defined

Figure 5-6. Facies 2 Thoroughly bioturbated, rubbly to blocky weathered, buff coloured siltstones and mudstones. Absolutely no laminations were preserved within this particular example. The nearly vertical plane to the right of the notebook was a product of weathering. It was not a bedding plane. From Hanson Member, Blackstone River section.

Figure 5-7. Facies 2 Well bioturbated mudstones from the Hanson Member on Cripple Creek. Banded appearance is due to siderite concretion layers.





and very gradational with the above facies. The lower contact is also gradational and often found to be associated with facies 1c and 3.

FACIES 3 Interbedded Sandstone/Siltstone and Mudstone (discrete sandstone and siltstone beds greater than 5 cm)

Facies 3 is defined to be interbedded sandstone/siltstone and mudstone with the sandstone and siltstone being in discrete beds greater than 5 cm thick (Figure 5-8). Because the definition of facies 3 states that discrete beds must be greater than 5 cm, the lower contact of the facies is normally gradational with and often interbedded with facies 1b and 1c. The upper portion of this facies is commonly gradational with and not independent of facies 5.

The interbedded, recessively weathered shales of facies 3 vary from light to dark gray and average 15 to 20 cm thick between discrete beds. A changing percentage of silt content accounts for the colour fluctuations. A rusty colour and sideritic cement can be found in lower portions of the facies where sideritic concretions are abundant. Good fissility is developed in some zones with the shales breaking into plate-like chips 1 to 2 mm thick. However, localized bioturbation often obscures all primary sedimentary features.

The discrete beds of sandstone and siltstone commonly display both a coarsening upward and a thickening upward sequence (Figure 5-9). The lowest beds in these sequences are sometimes less than 5 cm thick and are therefore part of facies 1b or 1c. They then grade upwards into rocks defined as facies 3. The beds display sharp bases with well developed planar lamination with current ripples on top. The beds are also largely very silty in composition with only infrequent examples of very fine (lower) grained sandstone beds. This silty nature gives the beds an overall rusty to buff colour. Beds continue to thicken upwards

Figure 5-8. Facies 3 Slightly overturned section of facies 3 on left and facies 2 on right. Stratigraphic top is to the left. Discrete siltstone turbidites are up to 10 cm thick. Interbedded shales and mudstones are quite thin. Note the gradual increase in bioturbation in the downsection direction. From Hanson Member, McLeod River section.

Figure 5-9. Facies 3 A reasonably well defined example of a thickening and coarsening upward sequence. Stratigraphic top is to the right. Note the overall increase in turbidite bed thickness from left to right. Some of the thinner beds to the left of centre are classified as subfacies 1b. The sandstone bed on the extreme right is the base of the Brazeau Formation. From Nomad Member, McLeod River section.

and coarsen upwards until upper beds are as thick as 25 cm and as coarse as very fine (upper) sandstone. An average thickness is 10 cm. These upper beds are sharp based with rare bases showing a thin veneer of micaceous and carbonaceous material. One example from McLeod River has a wood fragment on its base. Some bases also expose well developed scours, lineations, and flute sole markings. Others show a diversified trace fauna which includes Ophiomorpha and Gyrochorte. In several cases, the upper 2 to 3 cm of a bed has been destroyed by such faunal activity.

Beds in the upper portion of facies 3 may also display two features which indicate the facies is in an intermittant stage between facies 3 and facies 5. The first feature is a slight undulating nature of the normally planar laminated lower portion of a bed. The second is the development of symmetrical crested wave ripples on top of a bed instead of the characteristic asymmetrical current ripples of the lower part of the facies. While both of these characteristics are rare, they certainly are indicators of a changing environment.

In general, facies 3 is characterized by recessively weathered shales deposited between sharp based, resistently weathered siltstone/sandstone beds. This generated an overall striped appearance for the facies. Bioturbation is not a specific criteria for this facies, but localized areas may be intensely bioturbated.

FACIES 4 Bioturbated Sandstone Facies

Facies 4, characterized by moderately to thoroughly bioturbated sandstone, is restricted to the Hanson and Chungo Members and appears in three different settings within these members. No trace fauna have been identified from facies 4.

Subfacies 4a

Subfacies 4a is generally defined as the coarser grained equivalent of facies 2 bioturbated mudstones. The two facies are, therefore, associated and gradational with one another. However, discrete units of subfacies 4a were also noted in association with facies 5, 6 and 11 at McLeod River.

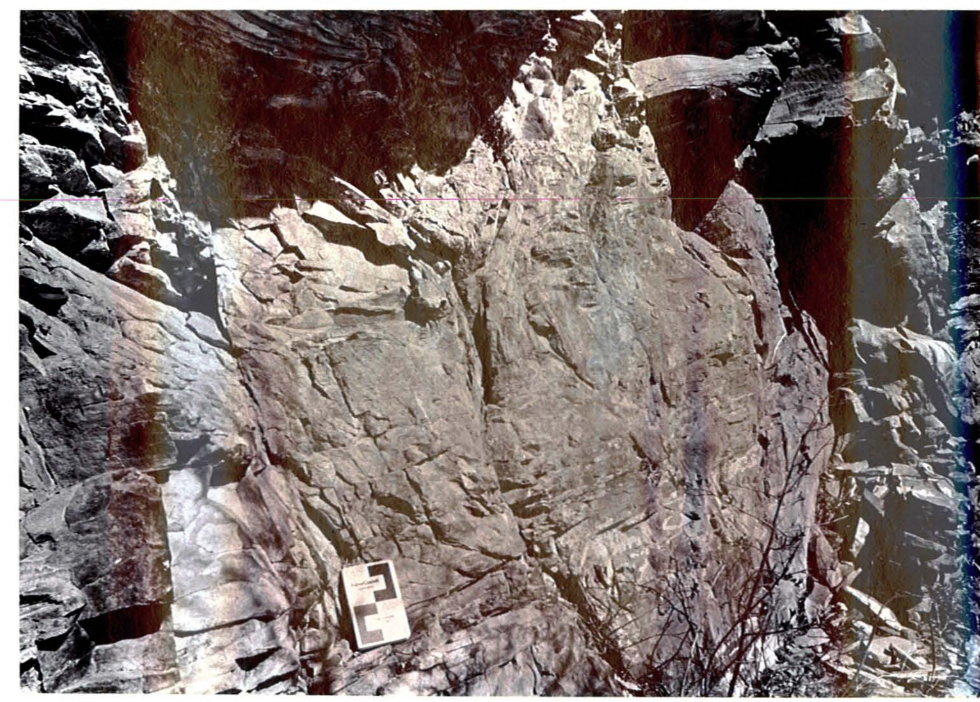
These very fine (lower) grained sands, usually found in the upper Hanson and lower Chungo Members, are very light buff to rusty coloured and show an irregular blocky to massive weathering texture (Figure 5-10). Rare, sharp based, discrete beds of sandstone up to 5 cm thick are preserved. However, internal lamination within these beds was very difficult to distinguish. Many of these beds are laterally restricted and discontinuous due to varying degrees of burrowing intensity.

Subfacies 4b

The second type of bioturbation recognized was very localized and was identified only at the Mt. Yamnuska section. Laterally from well developed areas of facies 6 were isolated zones of highly bioturbated fine (lower) to fine (upper) sandstone (Figure 5-11). These

Figure 5-10. Subfacies 4a These units are very similar to facies 2 units (see facies 2 photo). Only a slight increase in grain size separates the two. In this photo an increase in the intensity of bioturbation is noted upsection. The lower beds have maintained their morphologies although internal laminations were destroyed. The upper beds have been totally bioturbated. From Chungo Member, Bighorn River section. Stratigraphic top to left.

Figure 5-11. Subfacies 4b A 2 m x 4 m local unit of total bioturbation. Primary bed laminations, both laterally from and above and below this unit, were perfectly preserved. This unit had a very mottled appearance. *Ophiomorpha* is suggested to have been responsible. From Chungo Member, Mt. Yamnuska section.



zones were measured to average 2 to 3 m thick by 7 to 10 m laterally along the outcrop. The sands have a very mottled appearance and are coloured brown with some yellowish areas. No discrete sandstone beds occur in facies 4b.

Subfacies 4c

Facies 4c has been identified only in the upper Chungo Member at the Thistle Creek section. The 2.5 m to 3 m thick zone is very carbonaceous-appearing and is both overlain and underlain by coals of facies 10 (Figure 5-12). A thinner (0.5 to 1 m) bed of facies 4c is exposed directly below the first coal seam. A few single, straight, carbonaceous, root-like structures 1 to 2 mm wide and 2 to 3 cm long have been noted extending down from the coal seam into this bed.

The fine (upper) grained sands of the subfacies are poorly indurated, very friable , and therefore very recessive in nature. Weathering has produced a very mottled look and colour varies from yellow to light green to white. Although no internal laminations were noted for either of the above discribed beds, both bioturbated units seemed to be stacked into discernible, individual layers 5 cm to 10 cm thick.

Figure 5-12. Subfacies 4c This subfacies was associated with facies 10 coals, seen in the upper portion of the photo. No laminations were identified. The sandstone was very friable, mottled, and light green to yellowish in colour. Carbonaceous stringers are shown within the sandstone. From Chungo Member, Thistle Creek section.

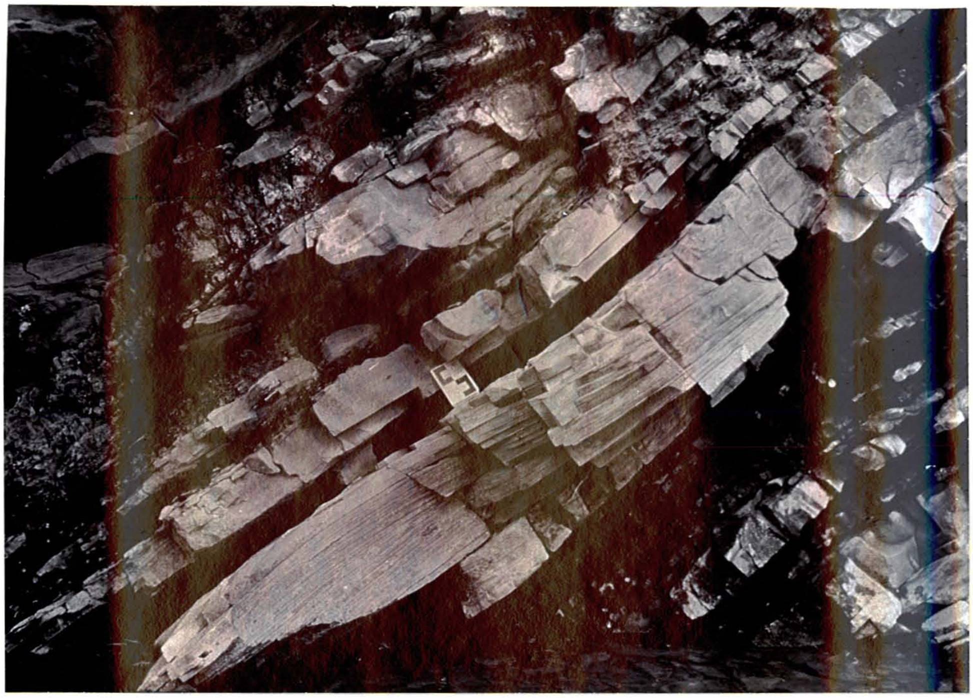
FACIES 5 Hummocky Cross-Stratified Facies

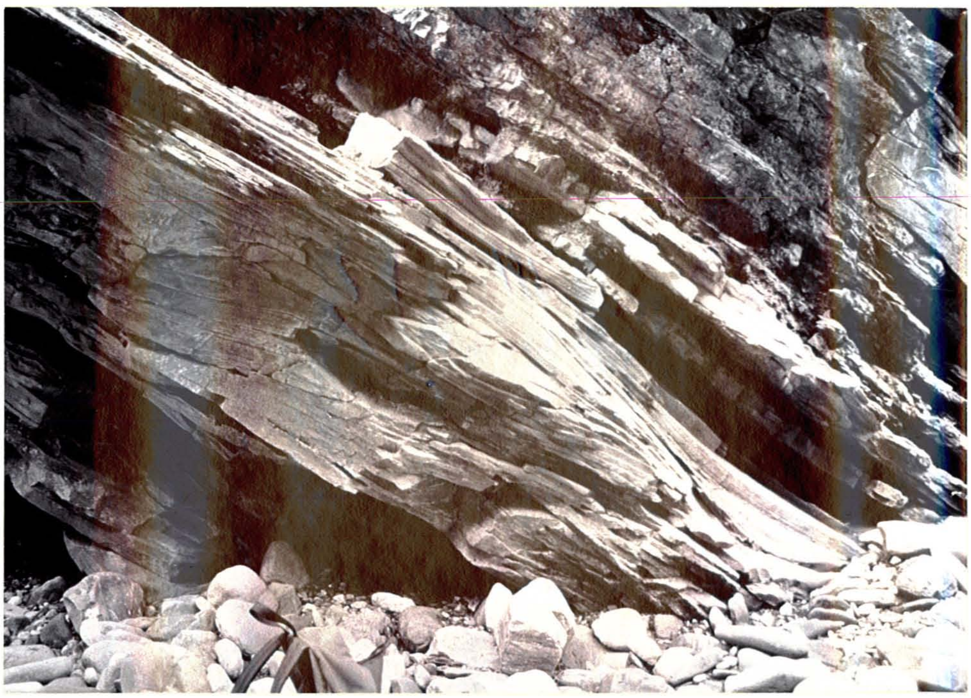
Facies 5 is a gradational unit from the underlying sediments of facies 2 or facies 3 and is restricted to the Chungo Member. It is made up of sharp based and erosive, very fine (lower) to very fine (upper) sandstone beds interbedded with mudstones which are normally bioturbated. The discrete sandstone beds display the bedform hummocky cross-stratification (HCS) which was first defined by Harms et al. (1975)(Figures 5-13 and 5-14). Several recent publications have provided a more thorough insight into this sedimentary structure (Hambler & Walker, 1979; Bourgeois, 1980; and Wright & Walker, 1981). Dott & Bourgeois (1983) proposed a "Bouma-type" sequence of sedimentary structures for HCS which has subsequently been modified by Walker et al. (1982).

On the Blackstone River section, facies 5 was developed to its maximum measured thickness of 20 m. Individual discrete beds were noted to range from 9 cm to 1 m in thickness with the average thickness at approximately 25 cm. An overall thickening upward trend was recorded at each occurrence. The discrete beds were noted to be laterally extensive (up to tens of meters) within the confines of the outcrop. The interbedded bioturbated mudstones ranged from black shales to buff siltstones. Lateral thickness changes and continuity varied due to the erosive nature of the discrete sandstone beds lying above. These mudstones tend to thin upsection until discrete HCS beds become "amalgamated" together by erosion or non-deposition of the mudstones (Leckie and Walker, 1982). In this study, amalgamated HCS beds are included in facies 6. The transition between these two facies was not

Figure 5-13. Facies 5 Well preserved example of HCS. Base of bed is very sharp. Note parallel lamination towards the base of the bed which "grows" upward to produce the dome morphology. A second dome is shown in the upper right. Symmetrical ripples are developed on the top of the HCS bed approximately 50 cm to the left of the notebook. From Chungo Member, Blackstone River section.

Figure 5-14. Facies 5 HCS sandstone beds. Note several individual beds which have subsequently cut into one another. The lack of shale horizons between beds was due to either a very short interval of time between subsequent beds resulting in non-deposition of shales or the erosion of thin shales during the event which eventually deposited the HCS bed. A backpack is in the lower portion of the photo for scale. From Chungo Member, Blackstone River section.





documented due to poor outcrop exposures.

Discrete HCS sandstone beds characteristically have sharp bases. Abundant trace fauna and some paleoflow indicators were noted on these bases. Gyrochorte and Ophiomorpha traces were identified along with shell fragments of unidentifiable pelecypods. Sole marks on the bases included flutes, grooves, and lineations. Occasional occurrences of wood fragments were also recorded along with very rare occurrences of thin, patchy zones (1 mm thick or less) of carbonaceous and micaceous material.

Stratification within individual discrete beds changes in both lateral and vertical directions. Typically, zones over the sharp bases are parallel laminated. This parallel lamination may "grow" upward into convex upward hummocks producing divergent laminations dipping at angles up to 15°. Average dips are generally less than 10°. Erosional truncational surfaces are formed between the parallel and dipping strata. The overall morphology produced is of undulating nature. Amplitudes between successive hummocks or swales range from 10 cm to 50 cm with the average at approximately 20 cm. Commonly capping the HCS sequence is the occurrence of straight crested and bifurcating oscillation ripples. Average wavelength is around 15 cm and an average amplitude is approximately 0.5 cm to 1 cm. Often, these ripples cover entire exhumed surfaces of HCS beds. The preserved condition of these ripples varied with the intensity of local bioturbation in the shales deposited above the HCS beds. Generally, bioturbation is of minor consequence within the discrete sandstone beds. However, biological activity in the overlying mudstones may affect the upper 2 to 3 cm of

HCS beds, the rippled zones.

The weathering of facies 5 produces jutting out discrete sandstone beds between the recessive interbedded mudstones. Internally, the HCS beds show a flaggy nature that is particularly well defined in the laminations in the transition zone from hummocks into adjacent swales. Individual plates defined from the weathering are 1 to 3 cm thick. Parallel laminated zones appear relatively much more massive in nature with only hints of developed lamination.

FACIES 6 Swaley Cross-Stratified Facies

Swaley cross-stratification (SCS), the defining criteria for facies 6, is a bedform which has only recently been introduced into the geologic literature (Leckie and Walker, 1982). The feature, largely made up of very fine to fine grained sand, is well developed within the study area, but is restricted to the Chungo Member where it may make up a volumetrically significant portion of the member (30% of Chungo Member at the Mt. Yamnuska section). Facies 6 is always underlain by HCS beds of facies 5. However, due to poor exposures, the contact was never seen. Leckie and Walker (1982) and Rosenthal (in prep.) note that the transition from facies 5 to facies 6 is always gradational over several meters. The upper contact is indistinct and very gradational into sands of facies 8. Because of the nature of this transition zone (to be discussed), it has been given a separate facies status (Facies 7).

Internally, the SCS beds consist of broad (1 to 2 m), low (average 15 cm, maximum 30 cm) scours which have cut down into preexisting swales and/or associated parallel lamination (Figure 5-15). Stratification follows the contour of the erosive base. Maximum dip angles average less than 10°, but they have been measured up to 17°. However, SCS laminations never approach angle of repose. This is one criteria in distinguishing SCS from trough cross-bedding (Leckie and Walker, 1982). The main distinguishing feature between the two structures is, however, the 3-dimensional symmetry of SCS. As well as broad low swales, SCS beds sometimes exhibit a very slight upward doming nature; not nearly as well developed as in HCS beds. Divergent angles in these low swells were not seen to exceed 4 to 5°. Amplitudes were

Figure 5-15. Facies 6 This photo shows the typical flaggy, almost indicative, weathering of SCS. Note the parallel lamination at the base and the very low angle truncating lamination just above the hammerhead. A slight doming upwards can be seen in the uppermost of the bed. From Chungo Member, Mt. Yamnuska section.

Figure 5-16. Facies 6 Another example of the superbly displayed floggy nature of SCS. Note the parallel to undulating character of the lamination. This particular example has been subjected to a slight amount of soft sediment deformation. Observe the small, tight fold just to the right of the hammerhead. From Chungo Member, Bighorn River section.

never greater than 4 cm.

SCS beds tend to weather into a very characteristic flaggy nature (Figure 5-16). Parting planes average 1 to 2 cm thick and give the feature a very "slabby" look. This "look" is particularly well developed in the parallel lamination. Colours varied between rusty or sideritic brown and buff.

Bioturbation played a very insignificant role in this facies. Although rare occurrences of Gyrochorte were noted, the effectiveness of biological activity was negligible.

FACIES 7 Interbedded SCS and TXB

Facies 7 is a transitional unit between SCS beds of facies 6 and TXB of facies 8. It is given facies status in this study because it makes up a volumetrically significant portion of the Chungo Member in some sections. A maximum measured thickness of 19 m crops out on Thistle Creek.

Trough cross-bedding generally makes up the lesser percentage of the two bedforms in this facies. However, the 4 m thick zone measured at Mt. Yamnuska shows a strong domination of TXB. Set thicknesses average approximately 20 to 25 cm with rare occurrences of thicknesses up to 80 cm. Sandstone grain sizes range from very fine to medium. Lerand (1982) noted localized concentrations of ironstone pebbles and conglomerates in TXB scour surfaces at the Mt. Yamnuska section.

Swaley cross-stratified beds of facies 7 generally fit the description of SCS beds presented under facies 6. The characteristic low angle (less than 10°) truncations scour into both previously deposited SCS and TXB laminations. Rare occurrences of broad, upward doming laminations growing out of parallel lamination were noted (Figure 5-17). Maximum amplitudes of these convex up features were up to 10 cm with typical wavelengths of 3 to 5 meters.

Weathering of facies 7 normally produces the typical flaggy nature of SCS beds. Trough cross-bedded zones are also flaggy, causing the internal laminations to be well defined. However, some examples (i.e. Bighorn River) appear structureless and show only slight hints of laminations. Only close inspections of these zones discounted an interpretation of complete bioturbation. Colours typical of facies 7 are rusty brown to light buff.

Figure 5-17. Facies 7 This photo shows a unit in which two alternating methods of sediment transport processes were working. The lowest flaggy weathered, parallel laminated unit is suggestive of SCS. Above this is a longitudinal section of a TXB. Flow was to the right. Then, above the hammer is a doming upwards bed, again suggestive of SCS. From Chungo Member, Mt. Yamnuska section.

FACIES 8 Trough Cross-Bedded Facies

Always lying above the sediments of facies 7 in the mid to upper Chungo Member are the massive appearing, trough cross-bedded, fine to medium grained sandstones characteristic of facies 8 (Figure 5-18). The thickest section measured was at the Mt. Yamnuska outcrop which showed 11 m of the facies (Figure 5-19). However, the top of this example was sharply and erosionally overlain by facies 6 deposits. The measured thickness may therefore be considerably less than the original thickness. Only on Thistle Creek was the upper contact of facies 8 not truncated. In this example, facies 8 was overlain by parallel lamination of facies 9.

The TXB of facies 8 generally shows a variation in size in the vertical direction. Measurements shows that TXB low in the section rarely exceeded 18 to 25 cm in thickness. A gradual increase upsection was noted until maximum thicknesses of 80 cm were recorded at the top. Average bed thickness was on the order of 30 to 40 cm. A slight increase in grain size from fine to medium sands accompanied the increase in bed thicknesses in some sections. The bases of these beds were sharp and erosive. Lerand (1982) noted occurrences of shale ripup clasts, concentrations of ironstone pebbles and large wood fragments along several examples from Mt. Yamnuska.

One exception to the thickening upward trend was recorded from Mt. Yamnuska. In a laterally discontinuous zone, smaller scale TXB that weathered dramatically differently and decreased slightly in thickness in an upward direction was found. Average bed thickness was approximately 2 cm. The colour was medium gray. Weathering of this area produced a very flaggy appearence with parting planes averaging 1 to 2 cm thick. Laminations were very well defined. Also observed in this area was a continuous, 50 cm thick bed of parallel lamination which exhibited a sharp base and a scoured upper surface.

Weathering of facies 8 typically produced very resistant and massive appearing horizons which varied in colour from buff to sideritic to greenish-brown. Laminations were commonly very poorly defined.

From well exposed examples of TXB, several paleoflows were measured. These measurements were taken from end, oblique, and longitudinal views. Rare exhumed plan views of TXB gave excellent axial directions. Occassional examples of truncated foreset laminae yielded rib and furrow paleoflow measurements.

Figure 5-18. Facies 8 This photo shows part of an 11 m thick, thoroughly TXB unit. The important thing to note is the multidirectional nature of the TXB. The two lowest beds show a longitudinal and an oblique view, both suggesting flow to the left. Directly below the foot is a transverse view, recording flow either into or out of the exposure. The previous photo (same location) shows flow to the right. From Chungo Member, Mt. Yamnuska section.

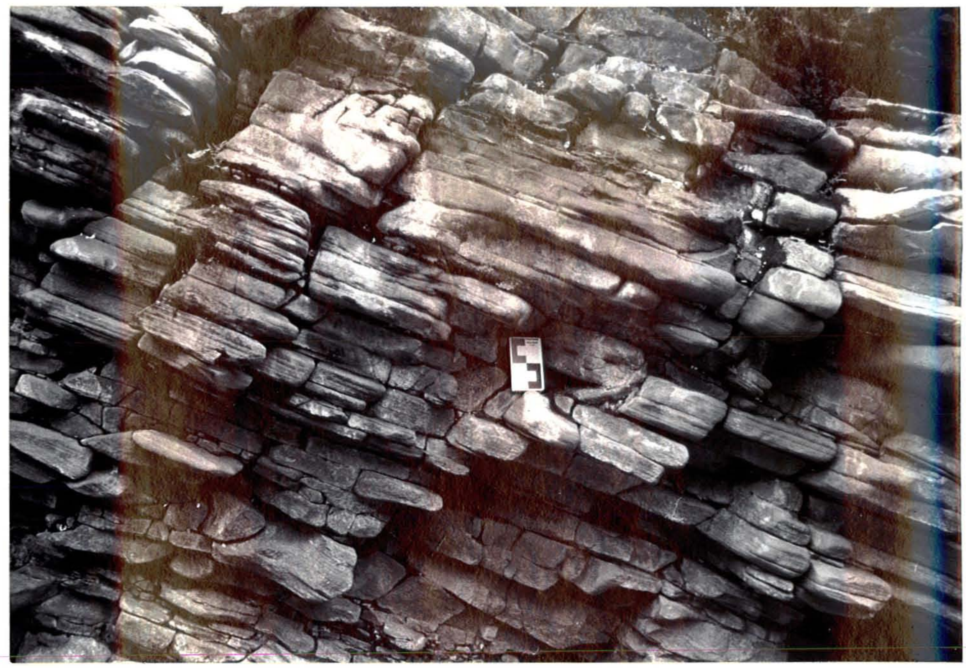
Figure 5-19. Facies 8 A more general view of a portion of the 11 m thick TXB unit shown in the previous photo. From Chungo Member, Mt. Yamnuska section.

FACIES 9 Parallel Laminated Sandstone Facies

The parallel laminated sandstones of facies 9 are herein defined to be parallel laminated sandstone deposits not genetically related to or intimately related with the parallel laminated sandstones associated with HCS and SCS of facies 5 and facies 6 respectively. With this definition, the only parallel laminated sands that fit the facies 9 criteria are exposed in the upper Chungo Member of Thistle Creek. Lying gradationally (over 3 to 4 cm) above sands of facies 8 lies a 3.5 to 4 m thick zone of entirely parallel laminated, fine (upper) grained sandstones (Figure 5-20). The laminations in the lower 0.5 m are slightly undulating with no specific wavelentgh. Maximum amplitude of the undulations is approximately 0.5 cm. Above the basal zone the laminations are very parallel and make no angle with regional bedding. No internal scouring or divergence of lamination was noted. However, weathering of this facies yielded a flaggy nature with the sediments divided up into "slabs" averaging 7 to 8 cm in thickness.

Overall colour of facies 9 is light brown. The top of the unit is gradational over 3 to 4 cm with the overlying bioturbated sands of facies 4c.

Figure 5-20. Facies 9 A 3 m thick unit of parallel lamination is shown in this photo. Rather than showing flaggy weathering (typical of parallel lamination associated with SCS) this unit is more "slabby" appearing. No examples of diverging laminations were noted. Beds at the base may be very slightly undulating. This unit is interpreted as a beach environment. From Chungo Member, Thistle Creek section.



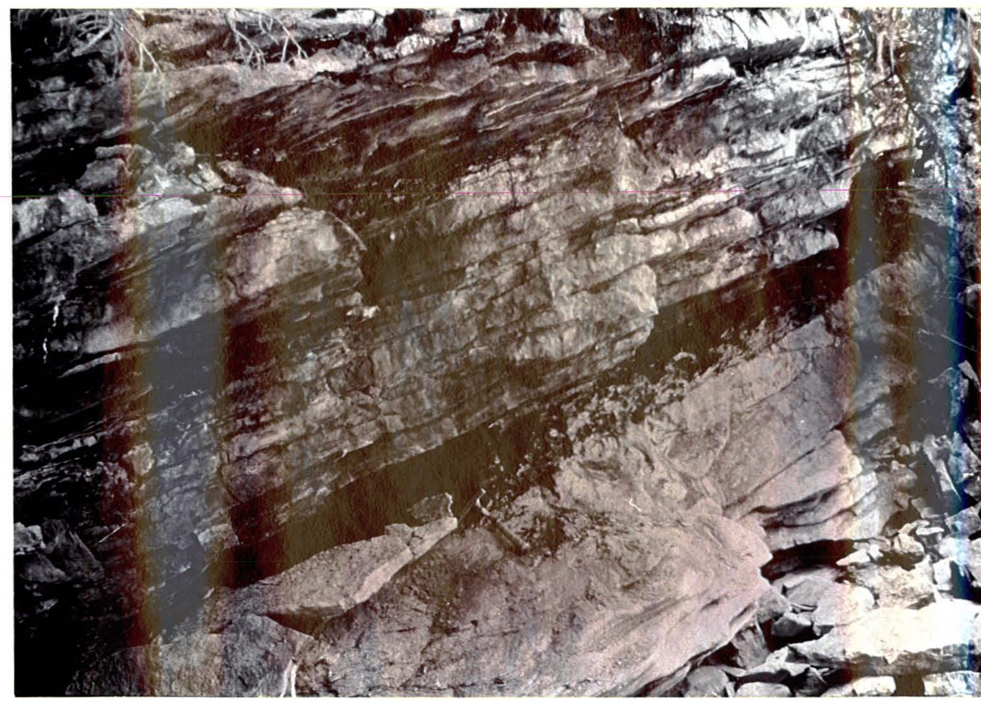
FACIES 10 Coal Facies

Individual thin coal seams, characteristic of facies 10, made up only a very minute proportion of the Wapiabi Formation sediments studied. Two occurrences were noted, both of which cropped out in the upper Chungo Member on the Thistle Creek section. The lower of the two examples is 18 cm thick and is both underlain and overlain by sandstones of facies 4c (Figures 5-21 and 5-22). Upper and lower contacts are generally sharp and distinct. The coals show no internal structure and are quite friable. Pieces tend to break into small blocks with maximum sizes of approximately 1 cm^3.

The upper coal seam, only 275 cm above the top of the lower seam, is much thicker at 150 cm. The lower contact with facies 4c is indistinct and has an irregular, undulatory nature. The top of the seam has been eroded by the subsequent deposition of a pebble conglomerate bed of facies 11. The coals are very recessive in nature and break easily into small blocky pieces.

Figure 5-21. Facies 10 An 18 cm thick coal bed of facies 10 is shown in this photo. The lighter colour in the coal was caused by sulfur. This sulfur can also be noted in the very mottled subfacies 4c bed below the coal. From Chungo Member, Thistle Cr. section.

Figure 5-22. Facies 10 This photo shows an overall view of the coal seam in the previous photo. The very sharp contact with the lower thoroughly bioturbated subfacies 4c bed is clearly shown. The subfacies 4c bed above the coal seam is not nearly as thoroughly bioturbated. From Chungo Member, Thistle Cr. section.



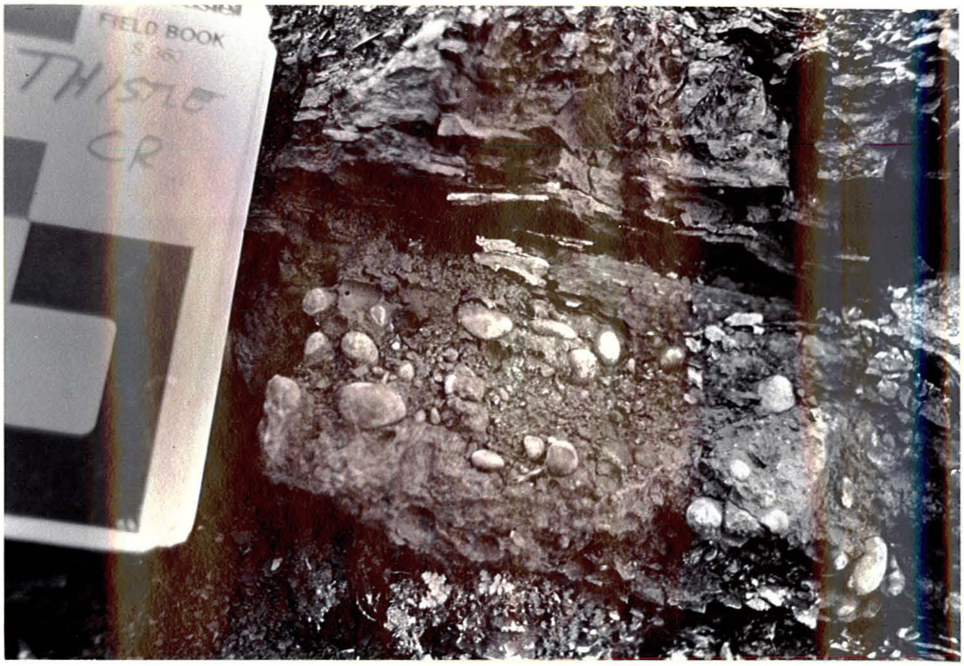
FACIES 11 Pebbly Conglomerate Facies

The pebbly conglomerates characteristic of facies 11 lie in beds (3 cm to 2 m thick) that have been traced regionally between six of the eight measured sections (Mt. Yamnuska, Cripple Cr., Bighorn River, Blackstone River, Thistle Cr., and McLeod River). The beds all lie at the same stratigraphic level and cap the Chungo Member coarsening upward sequence. Because of this, these "marker" beds have been used as the datum for measured sections. Bases of the pebble conglomerates were very sharp and erosive in nature. Upper contacts were more gradational with the overlying strata.

Pebbles within the facies, made up of black chert and lighter coloured quartz, are contained within a matrix that varies from a gray coarse (lower) sandstone to a very sideritic silty-mudstone. On Cripple Creek, the matrix is a sideritic mudstone with approximately 50% floating coarse chert grains (Figure 5-24). Sorting of the pebbles is poor with the maximum recorded size being 4 cm (Figure 5-23). Average size was on the order of 0.5 to 0.8 cm. Pebble content within the conglomerate ranges from 5 to 60% by volume and is arranged in both clast and matrix supported beds. Matrix supported conglomerates make up the largest percentage of the facies with clast supported conglomerates being deposited in randomly distributed zones through the whole facies. The two support types are gradational with one another. Roundness and the shape of the pebbles were visually estimated in the field to be well rounded and disk to equant, respectively. In most cases, the conglomerate beds were found to be completely structureless; this included no grading and no imbrication. In only one section (Bighorn

Figure 5-23. Facies 11 This example of facies 11 was the thinnest measured in the study area. Note the well rounded, poorly sorted, and matrix supported nature of the pebbles. There is a visual lack of grading, imbrication, and long axis orientation. From Chungo Member, Thistle Creek section.

Figure 5-24. Facies 11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River) was there a slight orientation of the two long axes to lie in the horizontal plane.

In two cases (Mt. Yamnuska and Cripple Cr.), three layers and two layers of pebble conglomerates were noted, respectively. In each, only short (< 1 m) stratigraphic intervals separated the conglomerates. In these sections, and the remaining four that showed facies 11, facies 11, being deposited over both marine and non-marine strata, truncated and marked the top of the Chungo coarsening upward sequence. Deposits above the conglomerates were largely part of facies 1; a dramatic change in sedimentation had occurred across facies 11. Only on the Bighorn River section did the underlying deposits of facies 6 cross the conglomerate. At that location, 8 more meters of facies 6 were deposited before truncation occurred. The pebble conglomerates of facies 11 therefore suggest some sort of sedimentation "break".

CHAPTER 6

INTERPRETATION OF FACIES

Facies 1 Interpretation

Because of gradational and conformable contacts between each of the three subfacies of facies 1, from an intermixing of the subfacies, and because each subfacies contains differing proportions of the same lithologies (interbedded silts and shales), it is suggested that similar but variable depositional processes were acting within each case. This genetic and intimate relationship between the subfacies has resulted here in a single environmental interpretation for facies 1.

Paleontological studies of facies 1 unquestionably yield an interpretation of deposition within a marine environment. Body fossils found included several unidentified ammonite shells (found within the shales) along with several examples of disarticulated brachiopod and pelecypod shells which were found on the bases of discrete sandstone and siltstone beds. The identified trace fauna included Ophiomorpha, Rhizocorallium, and Gyrochorte. Seilacher (1967) has suggested that Ophiomorpha is characteristic of offshore marine to fluvial environments. However, in light of the associated marine body fossils, the former environment is adopted here. Seilacher (1967) has also placed Rhizocorallium in a shallow marine shelf environment. Gyrochorte, although suggestive of a marine setting, has been recorded from deep marine to brackish eustuarine environments (Hallam, 1970).

The discrete siltstone and very fine grained sandstone beds that were observed in facies 1 were generally characterized by having sharp bases and plane lamination overlain by current ripple lamination. Some beds did not show current ripple development. These types of stratification are indicative of Bouma bc and b turbidites, respectively (Bouma, 1962). The sharp bases suggest that the sediment was introduced rapidly into a normally quiet basin.

Stow and Shanmugan (1980) have discussed and classified fine grained turbidity flow deposits on the basis of internal stratification. Their sequence begins with T_0 (Bouma c) which includes current ripple lamination of the same scale observed in facies 1. A proposal of suspension deposition from very thick, low concentration, slow moving (10-15 cm/sec) muddy turbidity currents was suggested. While this accounts for current ripple lamination development, it does not explain the upper plane bed produced below ripple lamination. Middleton and Southard (1977) have shown that a minimum velocity of 60 cm/sec must be attained to produce upper plane bed in very fine grained sand. Ripples will be produced with velocities less than 60 cm/sec. It is suggested here that while the discrete siltstone and sandstone beds of facies 1 may have been deposited from suspension from low concentration muddy turbidity currents (as proposed by Stow and Shanmugan, 1980); speeds reached by these flows were at least 60 cm/sec.

The generation of the turbidity currents which deposited the silt and sandstone turbidites in facies 1 could have resulted from several mechanisms, eg. storms, slumping, earthquakes, etc. As the Wapiabi Formation was deposited near an orogenically active area (the rising Cordillera), earthquakes may at first appear attractive as the generating mechanism. However, facies evidence, to be discussed at a later point, highly suggests that storms were very influencial in the generation of turbidity currents in the Alberta Basin. It is therefore suggested that turbidity currents were induced by storm events that eroded and suspended nearshore sediments.

Because the uppermost portion of the turbidite beds were not reworked into wave ripples (except in a few subfacies 1c horizons), it is probable that deposition took place below effective storm wave base. However, no particular depth or distance from shore can be assigned to such deposits. Hayes (1967) describes similar laminated silt layers in depths of 40 m as far as 30 km off the Texas coast.

The vast thicknesses of black shales within facies 1 are characteristically thinly laminated and very fissile (except for the bioturbated horizons). This strongly suggests that deposition took place from suspension in a quiet marine setting below storm wave base. Silty material, commonly found to be disseminated within the shales, possibly resulted from suspension deposition during waning stages of storm events. Storm events could have entrained the slightly coarser sediment and carried it out into the depositional basin.

The varying proportions of siltstone and sandstone interbeds within the shales of facies 1, facies 1a through 1c, can be accounted for by two mechanisms. An increase in interbeds may firstly record a slight progradation of the shoreline which would bring the sediment source in closer proximity to the depositional basin. The second possibility is that the increased proportion of silt and sand recorded an increased sediment supply rate. Evidence gathered during this study was not decisive as to whether either or both mechanisms caused the increased coarser size fraction.

Facies 2 Interpretation

The thoroughly bioturbated siltstone and mudstones of facies 2 show the same trace faunal assemblage as identified in facies 1, verifying the depositional environment as marine. A belemnite fossil and several ammonite shells substantiated the marine interpretation.

Stratigraphically, the deposits of facies 2 generally lie between facies 1 and facies 5 in the Chungo progradational sequence. Within this context (facies 5 will be interpreted later as storm generated, lying above storm wave base), it is suggested that deposition occured from slightly below storm wave base to slightly above. Any reworking by storm waves was obliterated by bioturbation.

Because of the total degree of bioturbation, sedimentation must have occured only at a moderate rate. As well, the environment must have been well oxygenated from circulating water. The intensity and thoroughness of the bioturbation also places restrictions on the thickness of any one depositional event. Due to the stratigraphic position of facies 2 between two facies characterized by differing degrees of storm influenced deposits, it is proposed that the siltstones of facies 2 may also have been emplaced by storm generated density currents. Each depositional event must have been restricted to depositing beds only a few tens of centimeters thick for complete bioturbation to occur between subsequent storms. Shales deposited from suspension during quiet interim periods between storm events were as well bioturbated and incorporated within the siltstones. No estimate of the original siltstone/shale ratio is possible to postulate.

Overall, facies 2 was deposited in a marine setting at or slightly below storm wave base. Similar deposits from the storm dominated Oregon shelf have been found in the mid to outer shelf area (Kulm et al., 1975). Frequent storm events supplied silt size material to the area and kept the waters well circulated and high in dissolved oxygen content.

Facies 3 Interpretation

The characteristic sharp based nature of siltstone and sandstone beds interbedded with marine shales in facies 3 is suggestive of periodic introductions of coarser sediment into a depositional basin. Many occurrences of broken and disseminated pelecypod and brachiopod shells, wood fragments, and carbonaceous material found on the bases of these beds supports the hypothesis of fast introduction of sediment. The interbedded shales record periods of quiet and slow "normal marine" sedimentation. The thorough bioturbation of the shales attests to relatively low sedimentation rates.

The marine nature of the sediments was verified by the well developed trace faunal assemblage noted within facies 3. Ophiomorpha, Gyrochorte, and Terebellina were all identified. Each of these are associated with shallow shelf conditions. As well, pelecypods and brachiopods from sharp based siltstone and sandstone beds indicate normal marine conditions.

The deposits of facies 3 are often found to be interbedded and transitional with deposits of subfacies 1b and 1c. In each occurrence, facies 3 is found to overlie subfacies 1b and 1c within the overall progradational sequence. An intimate and genetic relationship between the two is implied. It is suggested here that the thicker discrete beds within facies 3 indicate that deposition took place in a more proximal environment relative to subfacies 1c and 1b. A similar relationship was suggested by Hamblin and Walker (1979) for beds within the Fernie - Kootenay transition.

Although the actual mechanisms of flow for the entrainment and deposition of the discrete siltstone and sandstone beds have been argued (Hayes, 1967; Morton, 1981), beds of similar morphology, size, and internal structure to that found in facies 3 have been the result of storm generated processes. The storm emplaced processes suggested for facies 3 provides the previously implied genetic link between facies 3 and subfacies 1b and 1c. This proposal also acounts for interbedding of the storm interpreted HCS bedform of facies 5 and facies 3.

The parallel lamination overlain by ripple lamination within discrete beds can be described by the Bouma (1962) T_b and T_{bc} sequence, which in turn suggests deposition by turbidity current. Also, because the Bouma sedimentary structure sequence has not been reworked by storm effects, it is probable that deposition took place below storm wave base. However, an environment more proximal to the paleoshoreline than that proposed for facies 1 is suggested here as the approximate depositional site for facies 3.

Facies 4 Interpretation

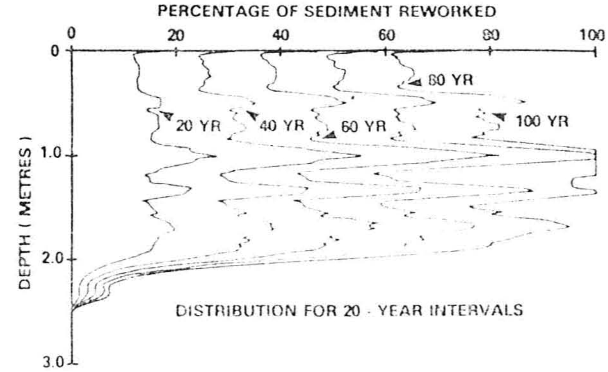
Because the bioturbated sandstones of subfacies 4a, 4b and 4c occur at different stratigraphic and environmental levels within the Hanson and Chungo Members, the generation of each type of bioturbation was influenced by unique conditions. For this reason, each subfacies will be discussed separately.

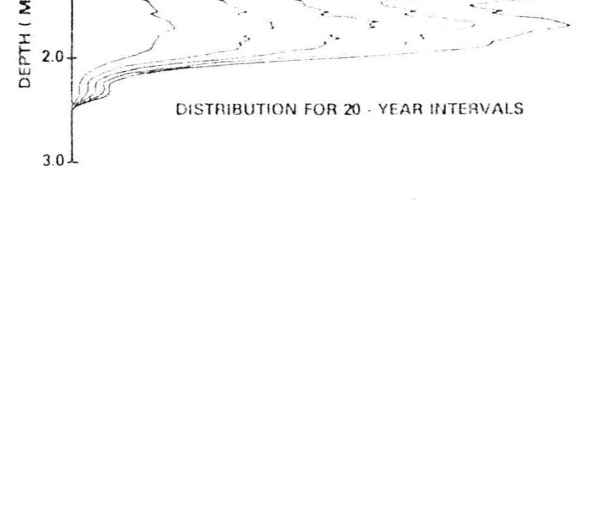
Although no trace fauna had been identified to verify a marine depositional environment for subfacies 4a, marine conditions were inferred from its stratigraphic relationship and the extensive bioturbation. Facies both above (facies 5) and below (facies 2) subfacies 4a have been interpreted as marine and no unconformable surfaces were noted. Because of this relationship, it is proposed that subfacies 4a was deposited slightly below storm wave base with moderate sedimentation rates to accommodate the complete extent of the bioturbation.

The only occurrence of subfacies 4b, located at Mt. Yamnuska, has been described by Risk (per. comm.) as "patchy" bioturbation and was probably produced by burrowing Arthipods. Risk (per. comm.) also ? suggested that Ophiomorpha could well have produced such a feature. Pemberton et al. (1976) and Risk et al. (1978) described Axius serratus, a burrowing crustacean from Nova Scotia. This shrimp was noted to produce burrows up to 2.5 m in depth and live at depths greater than 3 m. Burrow openings were on the order of 1.5 to 3 cm wide, very similar in size to Ophiomorpha trails. From Fig. 6-2 it can be shown that the horizon noted from Mt. Yamnuska could have been produced in approximately 50 years by such an organism.

Figure 6-1. Results of multiple simulation of the cumulative effect on sediment mixing of a population density of <ins>Axius serratus</ins> of $9/m^2$. Contour lines indicate cumulative years of burrowing. From Risk et al. (1978).

Figure 6-2. Results of multiple simulation of cumulative effect on sediment mixing of a population density of <ins>Axius serratus</ins> of $18/m^2$. Contour lines indicate cumulative years of burrowing. From Risk et al. (1978).


PERCENTAGE OF SEDIMENT REWORKED
0
20
40
60
80
100
0
1.0
2.0
3.0
DEPTH (METRES)
80 YR
20 YR
40 YR
60 YR
100 YR
DISTRIBUTION FOR 20 - YEAR INTERVALS


PERCENTAGE OF SEDIMENT REWORKED
0
20
40
60
80
100
0
1.0
2.0
3.0
DEPTH (METRES)
10 YR
20 YR
30 YR
40 YR
50 YR
DISTRIBUTION FOR 10 - YEAR INTERVALS

The interpretation of subfacies 4c is quite problematic. Due to the stratigraphic level of subfacies 4c, between two coal beds of questionable origin (facies 10 interpretation) in the upper Chungo Member, it was not possible to establish whether subfacies 4c was deposited within a marine or a non-marine environment. If marine in origin, the sands were probably deposited as a washover deposit in a quiet lagoonal setting where bioturbation could have been intense due to low sedimentation rates. A non-marine origin of the sandstone could imply several conditions. If eolian in nature, rooted, well sorted sands could give the impression of bioturbation. Bioturbated fluvial overbank or splay deposits could easily result from rooting, burrowing worms, or pedogenic activity. While an exact depositional environment is not suggested here, it is likely that deposition took place very close to a shoreline sequence.

Facies 5 Interpretation

The characteristic sharp-based and fine grained sandstones of facies 5 have been regarded by Harms et al. (1975, 1982) to have been deposited during oscillatory flows produced by large waves. Harms et al. (1975) further proposed that such flows may have resulted from storm events. Shales deposited between HCS beds were probably deposited during hydraulically quieter times when sediment supply rates were much lower.

The stratigraphic position of facies 5 within the overall progradational sequence in the study area, between the storm emplaced deposits of facies 2 or 3 and facies 6 (to be discussed), supports this mechanism.

It is suggested here that initial deposition of sand from storm generated, high sediment concentrated density currents (Hamblin and Walker, 1979) produced upper flow regime plane bed. This implies deposition from flows with velocities greater than 60 cm/sec. Subsequent deposition was affected by the oscillatory flows produced by large storm waves into the characteristic doming upward nature of HCS (Walker et al., 1983; Dott and Bourgeois, 1972). This would account for the occurrence of plane bed lamination at the base of most of the observed HCS beds and the appearance of plane bed "growing" upward into the three-dimensional hummocky morphology. As the oscillatory velocity of waves then decreased from the diminishing storm intensity, two-dimensional symmetrical wave ripples were generated on the upper bed surface. Preservation of these ripples was controlled by the degree of bioturbation during subsequent quiet periods of shale deposition.

Because the processes generating HCS beds are directly related to storm activities working on the bed and because normal fairweather processes do not rework HCS beds, it was proposed (Harms et al., 1975) that this bedform is produced below fairweather wave base and above storm weather wave base.

Hamblin and Walker (1979) have recently proposed that turbidites, like those found in facies 3, are the down slope equivalents of HCS beds. Rare occurrences of individual HCS beds within the turbidites of facies 3 therefore reflect the deepest penetration of storm influence on the bed surface.

Duke (in press) is suggesting that HCS is largely restricted to occurrences within two specific paleolatitudinal regions. Development of HCS within equatorial waters was probably generated by tropical storms, i.e. hurricanes. Large winter storms are proposed to account for high latitudinal occurrences. As the paleolatitude of the Cretaceous of Alberta was approximately 55 degrees, winter storms may have been the generating factor of HCS in the study area.

Facies 6 Interpretation

Swaley-cross stratification (SCS), always found above HCS beds of facies 5 in the progradational coarsening upward sequence of the Chungo Member, was interpreted by Leckie and Walker (1982) as a storm emplaced sand deposit deposited above fairweather wave base. Rare convex upward domes are suggested to show a genetic link between SCS and HCS beds.

In general, the lack of a well developed "hummocky" morphology and the prevalence of parallel lamination and low angle swaley sets perhaps suggest that orbital velocities of storm waves acting on the bed were higher than those acting on HCS beds below fairweather wave base (Harms et al., 1982). This evidence is suggestive of a shallower depositional environment than that proposed for HCS. As well, the lack of any preserved structures representative of waning storm stage (i.e. wave ripples) or interstorm finer grained deposits is indicative of a higher energy and erosional environment. However, the fact that no fairweather deposits were noted at all may indicate that deposition took place slightly below fairweather wave base.

Although hurricane or winter storm generated density currents are proposed for HCS beds (Duke, in press), Leckie and Walker (1982) suggested that SCS beds may not have necessarily required storms of that magnitude. In contrast, they proposed that more frequent storms of a smaller proportion would more easily rework and remove any record of fairweather processes. Walker et al. (1983) were more specific by suggesting that SCS may be emplaced by bottom rip current, storm ebb surges, wind-forced currents, or turbidity currents.

Within the context of the Chungo Member, it is proposed here that SCS was deposited in a proximal marine setting slightly below to just above fairweather wave base.

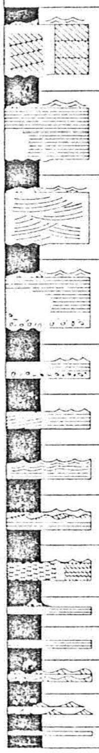
Facies 7 Interpretation

As previously interpreted, SCS is the result of storm processes and possibly records deposition from below to slightly above fairweather wave base. In a progradational system such as the Chungo Member, any fairweather influenced SCS deposits would therefore occur within the upper portion of the SCS stratigraphic level. The interbedded SCS and TXB sandstone beds of facies 7 are in such a stratigraphic position.

Two distinct types of facies 7 have been noted within the study area, a TXB dominated type at Mt. Yamnuska and a SCS dominated type at the remaining sections. It is felt that the differences are due to two distinctly different mechanisms responsible for the generation of the TXB. The SCS dominated facies 7 is discussed first.

Because deposition within the Chungo Member was dominated by storm produced sedimentary bedforms already discussed, it is reasonable to suggest that relatively high energy conditions prevailed in the shallow marine environment. This is also supported by the general lack of biogenic structures and reworking normally associated with low energy environments. Reading (1978) described several "typical beds" produced in high energy sand dominated shelf environemnts (Fig. 6-3). Low angle TXB with bed thicknesses of 20 to 150 cm was shown to occur in close association with upper flow regime plane bed lamination, lamination which is also closely related to SCS beds. A transitional flow regime characterized by deposition from bedload and suspension in a waning flow regime is therefore proposed as one mechanism which could produce TXB interbedded with SCS. However, the sharp, but not deeply scouring contacts that have been observed in the Chungo Member between TXB and

Figure 6-3. Summary of the internal stratification sequences and their interpretation from some typical sublittoral, inferred storm-generated sheet sandstones, from Reading (1978).



TYPICAL BED	BED THICKNESS	INTERNAL STRUCTURES	PROCESS INTERPRETATION	ENVIRONMENTAL INTERPRETATION	High energy shelf	Low energy shelf
	20 – 150 cm.	Cross-bedding (trough, tabular and climbing). Discontinuity or reactivation surfaces.	Dune migration in response to storm-enhanced tidal currents. Fluctuations in flow power.	Extension of dune field down the tidal current transport path when intense storm conditions enhance normal tidal currents.	PROXIMAL STORM DEPOSITS	ABSENT
	20 – 150 cm.	Parallel lamination with parting lineation.	Upper flow regime plane bed with sediment movement conditions. Sand deposited as bed load, but introduced mainly from suspension currents commonly of decreasing strength.	Combination of (1) exceptionally high energy conditions associated with intense wave action and storm-generated currents, and (2) abundant sand availability in shoreline and subtidal zones. Forms a facies belt immediately down current/offshore from extensive tidal dune fields. Thick sheet sandstones also characterize upper shoreface deposits (chapter 7).		
	20 – 150 cm.	Low-angle trough cross-bedding with parting lineation on foresets.	Transitional flow regime period or washed-out dune phase. Combination of bed load and suspension deposition of sand. Currents of decreasing strength.			
	ca. 20 cm.	Graded units occasionally passing upwards into parallel lamination.	Deposition from suspension by a current of decreasing strength.			
	5 – 10 cm.	Graded: massive or parallel laminated.	Deposition from suspension by a decelerating current.	Depending on the overall sequence this could represent more distal storm deposits to the thicker beds or they may represent less intense energy conditions, depending on strength of tidal, storm and wave processes.	INTERMEDIATE STORM DEPOSITS	PROXIMAL STORM DEPOSITS
	5 – 10 cm.	Parallel or low-angle lamination with wave ripples on top.	Upper flow regime conditions, followed by wave reworking.			
	5 – 10 cm.	Parallel to wave-ripple lamination.	Deposition by a decelerating current, followed by progressive oscillatory wave action.			
	5 – 10 cm.	Parallel to current-ripple lamination.	Deposition by a decelerating current: plane bed with movement to ripples.			
	5 – 10 cm.	Current ripple cross-lamination, occasionally climbing.	Lower flow regime current ripple phase.			
	0.5 – 3 cm.	Graded to parallel lamination.	Deposition from suspension.	Interpretation depends on position within the sequence but characterizes the distal parts of tidal current transport paths and low energy shelf deposits.	DISTAL STORM DEPOSITS	DISTAL STORM DEPOSITS
	0.5 – 3 cm.	Parallel lamination.	Deposition from suspension.			
	0.3 – 3 cm.	Current ripple cross-lamination (continuous layers).	Migration of current ripples with moderate sand supply.			
	0.3 – 3 cm.	Current ripple cross-lamination (discontinuous lenses).	Migration of current ripples with deficient sand supply.			
	1 – 5 mm.	Flat sand layers.	Sand incursions from suspension.			

SCS do not support an interpretation of deposition from a single waning flow regime event. Moreover, because occurrences of TXB within SCS are relatively few, thin, and isolated and because the contacts are not deep scouring surfaces (indicative of high energy events), a storm ebb surge event as a mode of TXB generation as proposed by Banks (1973b) has not been considered seriously. Banks (1973b) describes surge deposits as being up to 10 m wide, 1.5 m deep, and channel shaped. The TXB from facies 7 in the Chungo Member does not exemplify these features.

A second cause for generation of TXB within SCS may have resulted from fairweather processes acting between high energy events. Paleocurrent data from TXB (to be discussed later in the text) suggest that in the Chungo nearshore environment, a strong longshore current flowed towards the southeast. It is proposed that this current was strong enough to cause the active migration of dunes over previously deposited storm influenced sands (SCS). During subsequent storm events, most of the fairweather deposits were eroded, leaving only rare examples of fairweather generated deposits to be preserved in the rock record. If this interpretation is correct, the change from SCS into interbedded SCS and TXB may record the approximate level of fairweather wave base.

As previously noted, facies 7 at the Mt. Yamnuska section is very different from facies 7 observed at other sections. In this case, TXB is by far the dominant structure. As well, the TXB is unique in that bidirectional paleoflows (approx. 160° apart) were recorded at both equivalent stratigraphic levels and levels slightly higher (Lerand, 1982). Reading (1978) notes that flow directions of these sorts are definite indicators of a tidal influence. Such a tidal interpretation for the TXB would be consistent with the Mt. Yamnuska facies 8 interpretation which is to be discussed in the subsequent section.

The fact that only rare occurrences of SCS are noted within facies 7, that TXB is of possible tidal origin, and that facies 7 overlies storm generated SCS of facies 6, may indicate that during progradation of the shoreline, tidal influence began to overshadow the effects of shore domination. The mechanism by which both tidal and storm produced structures were generated together is discussed under the section interpretations chapter.

Anderton (1976) provides a hypothetical model for the Jura Quartzite shelf, which modified, may in part be applied to the Chungo Member at Mt. Yamnuska. During times of moderate and normal storm conditions, active dune migration could be tidally controlled with the result of TXB and planar tabular cross-bedding. A storm of greater than normal intensity may result in SCS development within the normally tidally influenced area. Normal, fairweather tidal conditions would generate the observed bimodal flow distribution.

It is therefore concluded that Mt. Yamnuska shows evidence of both storm and tidal influence which worked together to produce a facies 7 that differs significantly from the totally storm generated facies 7 of the remaining measured sections. The TXB of the latter sections possibly resulted from a strong longshore current working in the nearshore environment during Chungo times.

Facies 8 Interpretation

Because of its conformable stratigraphic relationship with the underlying marine facies, from the lack of evidence suggesting emergence, and because facies 8 continues the progradational coarsening upward sequence, it is proposed here that the TXB characteristic of facies 8 was deposited in a shallow, nearshore marine environment.

Reading (1978) states that two different processes acting in a nearshore environment are capable of producing migrating sinuous crested bedforms which would result in TXB. The first process is storms; the second is tides. At first thought, the idea of storm generated TXB is appealing in light of the fact that other facies lower in the sequence have been interpreted as storm-generated. However, on closer inspection, an interpretation of tidally generated TXB may be reasonable, particularly for the TXB at Mt. Yamnuska section.

Three criteria from which tidal influence can be inferred were proposed by Reading (1978): 1) bidirectional and bimodal paleocurrent patterns which reflect a reversing in flow directions, 2) multimodal paleocurrent patterns which suggest temporal fluctuations in direction or the rotary nature of tidal currents, and 3) the abundance of cross-bedding (both trough and/or planar tabular) which signifies the migration of dunes and sandwaves.

The TXB of facies 8 at Mt. Yamnuska shows each of these tidally indicative features. From the Mt. Yamnuska section, Lerand (1982) has measured both bidirectional paleoflows and multidirectional flows which are oriented in almost every compass direction (these will be discussed in more detail at a later point). As well, the preservation potential of an 11 m thickness of TXB, as recorded at Mt. Yamnuska, would be quite low if the environment were completely storm dominated. Such storm domination at Mt. Yamnuska is suggested by the facies lower in the section. Migrating dunes, generated under fairweather or "normal" marine conditions, would undoubtedly be destroyed at storm stage.

If the TXB of unit 8 at Mt. Yamnuska is to be considered as tidal in origin, one problem still remains. How is 11 m of tidally produced TXB preserved in an environment that is proposed to be storm-dominated? This question will be addressed at a later point.

The facies 8 TXB identified at sections other than that at Mt. Yamnuska has specific characteristics which separates it from the Mt. Yamnuska section. Most of the TXB from these occurrences are stratigraphically thin (usually not more than a few meters thick) and show a strong association with the storm induced sediments of facies 7. It is quite possible that TXB zones on this scale could have been preserved within a storm-dominated environment. The association of facies 8 with facies 7 is also suggestive that deposition took place in a high energy environment. Furthermore, paleoflow measurements from TXB (although being few in number and therefore probably inconclusive) do not show any significant evidence of multidirectional sediment transport, a criteria for a tidal interpretation.

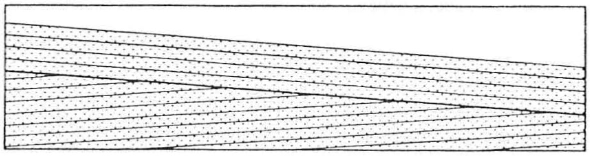
This evidence, taken together, suggests that facies 8 sediments, other than those at Mt. Yamnuska, were deposited under conditions of a moderate to high energy nearshore environment.

Facies 9 Interpretation

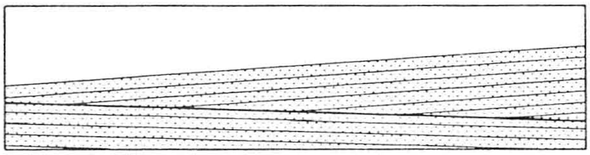
Within its stratigraphic context, the parallel lamination of facies 9 probably records foreshore or beach processes. This interpretation is justified as facies 9 conformably overlies the mid to upper shore face deposits of facies 8, continues the Chungo coarsening upward progradational sequence, and is overlain by coals of probable lagoonal, marsh, or floodplain origin.

Features characteristic of the foreshore environment are largely controlled by swash-zone processes (Reading, 1978). These processes commonly result in a rhythmic topography of asymmetric, parallel to coast, ridges being produced within the beach face. Davidson-Arnott and Greenwood (1976) have illustrated a modern intermediate wave energy beach profile from Kouchibouguac Bay, New Brunswick. In this profile, TXB, landward dipping planar tabular XB, and various combinations of planar tabular XB, rippling, and TXB are represented. This profile, as well as others which describe modern high wave energy conditons (Clifton et al., 1971) or low wave energy conditions (Howard and Reineck, 1972b), does not represent the facies recognized in facies 9. However, parallel lamination produced by storm events and described by Thompson (1937) is much more likely to be preserved in the geologic record (Reading, 1978). Thompson (1937) recorded four types of low angle parallel lamination from modern wave dominated beaches in California, types A through D (Figure 6-4). Type C most reflects that seen in facies 9. In type C, all laminations, including truncating surfaces, are seaward dipping at 1 to 3 degrees in a section normal to beach orientation. Any section at an oblique angle to the beach would therefore show lamination angles

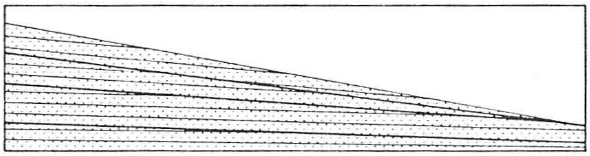
Figure 6-4. Schematic illustrations showing four major types of cross-bedding in beach deposits. The sea is toward the right. (Modified after Thompson 1937).



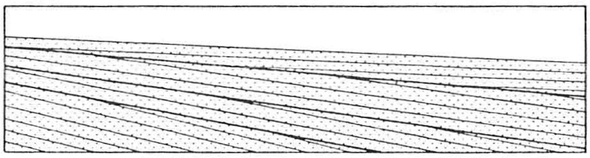
Type A



Type B



Type C



Type D

less than 1 to 3 degrees. Under such conditions, any dipping laminations would be very difficult to detect. Such a case is proposed for the exposure of facies 9 on Thistle Creek where dipping laminations were not recorded. Although laminations appeared perfectly parallel, this may have been only an apparent view. In Thompson's (1937) diagram of type C laminations, no scale between truncation surfaces were inferred. It is suggested here that the "slabby" weathering noted in facies 9 may have been produced as a reflection of the original truncation surfaces.

Facies 10 Interpretation

The coals of facies 10 are of problematic origin. Being deposited directly above a foreshore environment, the coals may represent deposition within either a marine back-island lagoonal environment or a fluvial floodplain environment. However, in neither case is there a sedimentary record of the transition from beach environment to a lagoonal or fluvial system. No direct evidence of backshore dunefields, vegetated eolian flats, or washover sands indicative of barrier island lagoons was noted (Hayes, 1967; Dickinson, 1971). As well, roots commonly associated with both environments were not present. The total lack of any internal stratification could be indicative of either environment.

Because of the lack of evidence for an unconformable surface between the foreshore deposits of facies 9 and the coal sequence of facies 10, it is postulated here that the coals may have formed in a backshore marsh environment. The 1 m thick bioturbated sandstone bed which separates the two coal beds may be a record of backshore dunes, eolian dunes, or a large storm generated washover deposit breaking into the marsh environment.

Facies 11 Interpretation

The chert pebble conglomerates representative of facies 11 have been noted by Stott (1963) to truncate or cap the coarsening upward sequence of the Chungo Member on a regional basis. Similar relationships between conglomerates and progradational sequences have been noted to exist within the Alberta Basin, namely in the Cardium and Viking Formations.

In five of the six sections measured which had occurrences of facies 11 (Bighorn River being the exception), the end of Chungo sedimentation was represented by the erosive deposition of a pebble conglomerate. The Mt. Yamnuska and Cripple Cr. sections yielded three and two conglomerate beds with thin interbedded sandstone beds, respectively. Truncation and deposition occured over both marine and non-marine sediments. This suggests a genetic relationship between facies 11 and regional transgression which resulted in deeper marine Nomad Member shales overlying nearshore and non-marine sediments of the upper Chungo Member.

The source of the pebble size material in the Chungo is problematic in origin. However, it is proposed here that during the regional transgression, gravelly, fluvial systems, equivalent of the marine Chungo further in towards the rising Cordillera to the west, were inundated. Smith and Smith (1980) and Smith and Putnum (1982) suggested that within the structural context of the rising Cordillera, conditions could have been right for gravel accumulation within braided or anastomosing environments. As well, thick (greater than 5 m) beds of pebble conglomerates were noted at the base of the Belly River

Formation at at least two of the measured sections, suggesting that pebble size material was available from fluvial systems. The ultimate source of chert pebbles for the Mesozoic clastics was suggested by Rapson (1965) to be Mississippian and Pennsylvanian cherty dolomites and limestones. However, petrographic analyses by Rapson (1965) proved to be inconclusive.

As transgression proceeded, fine-grained material was largely eroded and winnowed away, leaving coarser pebble and sand size material as a relict lag. This coarser grained fraction was then introduced into the basin on a region wide scale. The processes involved in such a dispersion of coarse sediment are questionable. Nevertheless, the internal structure and general morphology of pebble conglomerate beds are very analogous to those described by Middleton and Hampton (1976) as representative of beds resulting from debris flow mechanisms. Although coarse material within debris flows is generally supported by a clay-water matrix (of which there is little evidence of in facies 11), Hampton (1972a) demonstrated that submarine debris flows could be low in clay and that sandy debris flows should generally be common in ocean environments. Several criteria indicative of debris flows were proposed by Middleton and Hampton (1976): irregular top contact (large grains projecting upwards), massive, poor sorting, random fabric, and poor grading (if any). Each of these characteristics are displayed in facies 11. Several other lines of evidence are also supportive of a debris flow process. Firstly, initiation of debris flow movement can occur on slopes of almost zero. This is attractive in an eperic sea environment where basin slopes are generally quite low. Secondly, pebble conglomerates of facies 11 appear to have a sharp upper grain size boundary of approximately 4 cm. No anomalously large pebbles were noted. Middleton and Hampton (1976) suggested that debris flows should show this abrupt upper grain size. Grains too large to be supported by the flow would be left at the source area.

Although evidence is largely supportive of a subaqueous debris flow mechanism for the dispersion of pebbles across the basin, it is not suggested that debris flows were the only mechanism working. It is probable that several gravity mechanisms worked together in generating the deposits of facies 11.

CHAPTER 7

SECTION DESCRIPTIONS AND LATERAL RELATIONSHIPS

The following chapter briefly describes each of the eight measured sections in the study area. Emphasis has been placed on facies relations within individual sections and features which are unique to particular sections. For reasons of contrasting lithologies, the upper four members of the Wapaibi Formation, the Thistle, Hanson, Chungo, and Nomad Members, are discussed separately under each section description. The lowest three members, the Muskiki, Marshybank, and Dowling Members are discussed together as they are generally characterized by similar lithologic relationships.

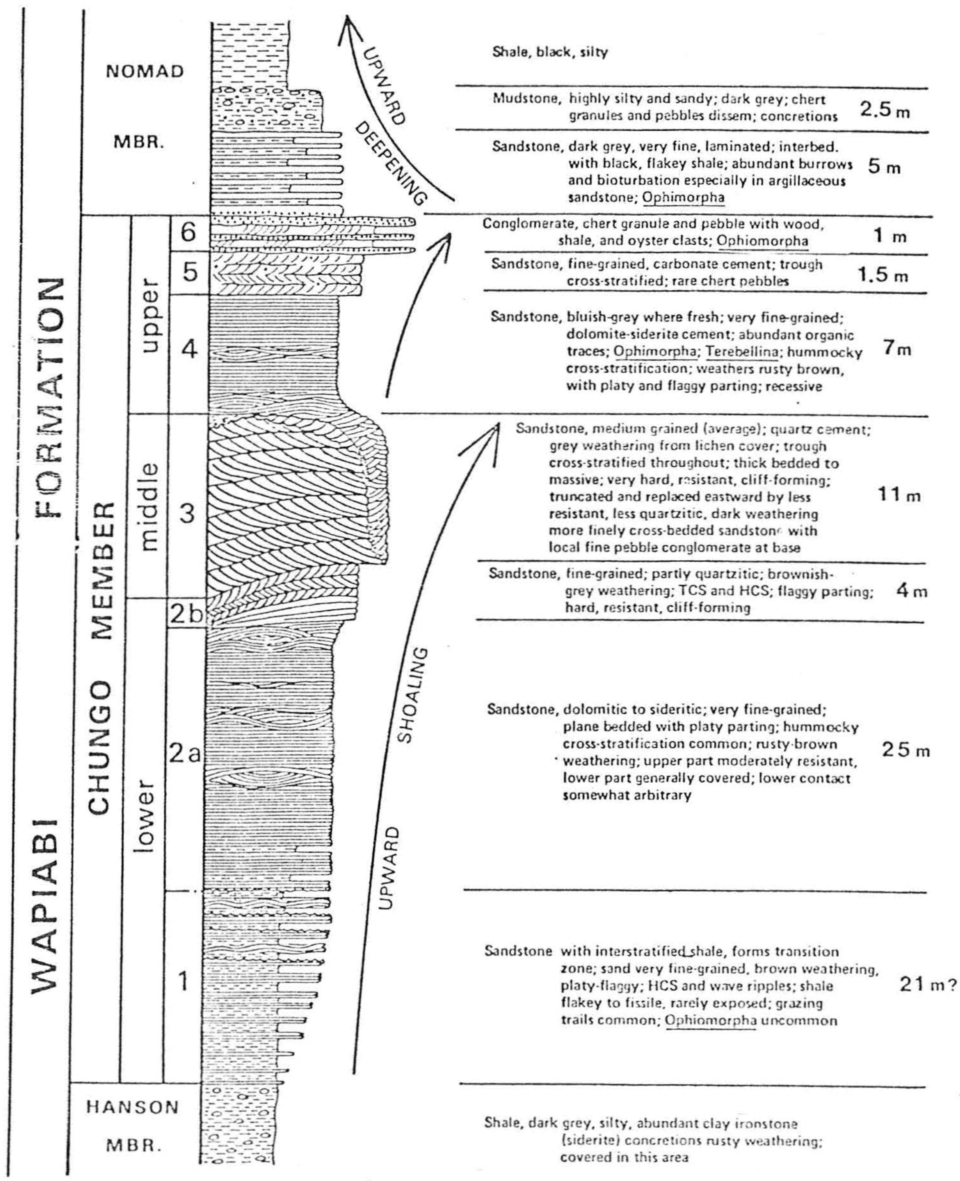
For convenience, the sections will be described in a systematic order from south to north.

Mt. Yamnuska Section

The Chungo Member is the only exposed member of the Wapiabi Formation at Mt. Yamnuska, the most southerly studied section. It has recently been described in detail by Lerand (1982) who divided the member into six units (Fig. 7-1). The six units are easily recognizable and are therefore retained for the purposes of this discussion. However, a seventh unit is included in this study.

Unit 1, the lowest stratigraphic unit which records the transition from the Hanson Member to the Chungo Member, is largely dominated by 5 to 50 cm thick, very fine grained, HCS sandstone beds (facies 5) interbedded with shale intervals up to 10 cm thick. As the

Figure 7-1. Schematic composite secton, Chungo sandstone, Mt. Yamnuska (from Lerand, 1982).


WAPIABI FORMATION
CHUNGO MEMBER
NOMAD MBR.
HANSON MBR.
upper
middle
lower
6
5
4
3
2b
2a
1
UPWARD DEEPENING
UPWARD SHOALING
Shale, black, silty
Mudstone, highly silty and sandy; dark grey; chert granules and pebbles dissem; concretions 2.5 m
Sandstone, dark grey, very fine, laminated; interbed. with black, flakey shale; abundant burrows and bioturbation especially in argillaceous sandstone; Ophimorpha 5 m
Conglomerate, chert granule and pebble with wood, shale, and oyster clasts; Ophiomorpha 1 m
Sandstone, fine-grained, carbonate cement; trough cross-stratified; rare chert pebbles 1.5 m
Sandstone, bluish-grey where fresh; very fine-grained; dolomite-siderite cement; abundant organic traces; Ophimorpha; Terebellina; hummocky cross-stratification; weathers rusty brown, with platy and flaggy parting; recessive 7m
Sandstone, medium grained (average); quartz cement; grey weathering from lichen cover; trough cross-stratified throughout; thick bedded to massive; very hard, resistant, cliff-forming; truncated and replaced eastward by less resistant, less quartzitic, dark weathering more finely cross-bedded sandstone with local fine pebble conglomerate at base 11 m
Sandstone, fine-grained; partly quartzitic; brownish-grey weathering; TCS and HCS; flaggy parting; hard, resistant, cliff-forming 4 m
Sandstone, dolomitic to sideritic; very fine-grained; plane bedded with platy parting; hummocky cross-stratification common; rusty-brown weathering; upper part moderately resistant, lower part generally covered; lower contact somewhat arbitrary 25 m
Sandstone with interstratified shale, forms transition zone; sand very fine-grained, brown weathering, platy-flaggy; HCS and wave ripples; shale flakey to fissile, rarely exposed; grazing trails common; Ophiomorpha uncommon 21 m?
Shale, dark grey, silty, abundant clay ironstone (siderite) concretions rusty weathering; covered in this area

base of the unit was not exposed, a complete measured thickness was not obtainable. However, 7.5 m was measured. The sandstone beds in this unit are generally discrete, but bioturbation commonly obscures the upper 2 to 5 cm. Trace fossils are common.

Lying above the HCS of unit 1, the 24 m thick unit 2 shows the typical rusty-brown colour and flaggy weathering indicative of SCS (facies 6). The contact between the two units is poorly exposed, but occurs over 25 to 50 cm. Sandstones within unit 2 range from very fine-grained in the basal and central portions to fine-grained in the upper portion and are quite sideritic in nature. The coarsening from very fine to fine-grained sandstone takes place in the upper 4 to 8 m of the unit.

Unit 3 (Lerand's unit 2b) is characterized by facies 7, a transitional unit between SCS of facies 6 and trough cross-bedding of facies 8. The sandstones are fine-grained and are part of a coarsening upward sequence at Mt. Yamnuska. Both upper and lower contacts are gradational over less than 50 cm.

Unit 4 (Lerand's unit 3) is 9 to 11 m thick, is dominated by facies 8, and is the most prominent unit of the Mt. Yamnuska section. It forms a very resistant cliff, is thickly bedded, and appears massive. Grain size has increased to medium sand, thus continuing the coarsening upward sequence. However, a sharp erosional upper contact separates unit 4 from the very fine to fine-grained sandstones of unit 5. Trough cross-bed sets in unit 4 range from 20 to 60 cm thick and show a consistent thickening upward trend through the unit.

A 7 to 9 m thick unit of facies 6 sandstones, characteristic of unit 5, begins a second coarsening upward cycle at Mt. Yamnuska. Although the second cycle is similar to the first, it is considerably thinner (8.5 m) and lacks the transitional facies 7 between unit 5 (facies 6) and unit 6 (facies 8). The unit 6 trough cross-bedding consists of fine to medium-grained sandstone with a few small, scattered chert pebbles. Unit 5 is 1.5 m thick and is truncated by the 1 m thick chert pebble conglomerate of unit 7. This unit, which divides the Chungo Member from the overlying Nomad Member, is characterized by three discrete conglomerate layers (averaging approximately 20 to 30 cm each) separated by fine-grained sandstone beds. Trough cross-bedding was observed in these sandstone beds and was gradational above conglomerate beds. The conglomerate beds are clast to matrix supported and show no visual evidence of imbrication, orientation, or grading. The average pebble size is approximately 0.5 to 1 cm in diameter. The maximum recorded size was 3 cm.

The Chungo Member at Mt. Yamnuska is generally characterized by two coarsening upward cycles. The first, and most complete cycle, contains a facies sequence of facies 5, 6, 7 and 8, in ascending order. The second cycle is characterized by facies 6 and 8 only. Chungo sedimentation culminated with the deposition of facies 11 conglomerates over the second coarsening upward cycle.

Burnt Timber Creek

Muskiki, Marshybank, and Dowling Members

The Muskiki Member on Burnt Timber Creek, which sharply overlies the Cardium Formation, is relatively thin compared to other Muskiki sections. A thickness of 46 m was recorded. Shales were observed to be very fissile, dark, concretionary, and generally devoid of a significant silt content (facies 1a). Although siltstones deposited as discrete beds locally make up to 10% of local stratigraphic horizons, no local horizons contain enough silt size material to be classified as facies 1b.

The Marshybank Member is fairly easy to recognize at Burnt Timber Creek. It consists of a 20 m interval characterized by interbedded siltstones and shales. In the basal 12 m of the member, the siltstones are relatively well indurated, buff to rusty coloured, and contain faint examples of parallel lamination. Some sharp bases were observed. Generally, however, base and tops of beds are quite obscure.

The upper part of the Marshybank Member is not nearly as well defined. This portion is highly concretionary and is considerably more bioturbated. Siltstone and shale beds are much thinner. Both upper and lower contacts of the Marshybank Member are gradational over several meters.

The Dowling Member is largely composed of thick sequences of black, fissile shales of facies 1a. Sideritic concretions up to 75 cm long by 20 cm thick were noted throughout. Only two occurrences of facies 1b were recorded. The first is 2 to 4 m thick, while the second occurrence (which caps the Dowling Member) is approximately 7 m thick. This example may be part of a coarsening upward cycle as it marks the transition from facies 1a shales in the Dowling Member to facies 1b shales in the Thistle Member. The change from the Dowling Member to the Thistle Member is gradational and is characterized by a change from sideritic and concretionary shales to platey, calcareous and non-concretionary shales, respectively.

Thistle Creek Member

Studies by Stott (1963) suggest that the approximately 250 m thick Thistle Member at Burnt Timber Creek can be divided into five coarsening upward sequences. Five cycles were also identified in the present study.

The beginning of each coarsening upward cycle is generally characterized by a unit of facies 1a. This unit is normally less than 10 m thick, but 18 m was measured at the beginning of one cycle. Shales are very fissile, black to rusty coloured, and are locally characterized by an abundance of bentonite. These shales grade upwards into facies 1b shales.

The vast majority of the Thistle Member is dominated by the platey shales of facies 1b. The platey appearance is due to resistantly weathered siltstone and sandstone turbidites which protrude out of the recessively weathered shales. As well, the turbidites are light gray, a contrast from the dark gray to black interbedded shales. While the majority of turbidites are less than 5 cm thick, rare occurrences of buff to rusty coloured siltstone beds up to 10 cm thick are present. These thicker beds are not, however, concentrated enough to be classified as facies 3.

Overall, facies 1b dominates the Thistle Member and caps four of the five coarsening upward cycles. Facies 1c is only a minor constituent of the Thistle Member at Burnt Timber Creek. It caps only one coarsening upward cycle.

Hanson Member

Two of the thickest measured sections of the Hanson Member in the study area were obtained from Burnt Timber Cr. and Cripple Cr. (the adjacent section to the north). This was due largely to the general lack of a well defined overlying Chungo Member. The Hanson Member essentially replaces the Chungo sandstone. Both sections yielded a thickness of approximately 110 m.

Bioturbated mudstone of facies 2 was, by far, the dominant lithology observed in the buff to rusty coloured, recessively weathered Hanson Member at Burnt Timber Cr. However, a diverse faunal assemblage was not observed. A progressive coarsening upward sequence of facies 2, from shale at the base into siltstone and sandy siltstone, was noted. Neither the upper nor the lower contact of the Hanson Member is particularly well defined. The lower contact is placed at the first sideritic concretion layer above the non-concretionary Thistle Member. The upper contact is defined at the base of the first (and only) discrete sandstone unit of the Chungo Member.

Chungo Member

The Chungo Member at Burnt Timber Cr. is virtually non-exisent. It is limited to a single meter of very fine-grained, moderately bioturbated sandstone beds approximately 10 to 15 cm thick, interbedded with thin shales and siltstones. No laminations were observed in these beds. Surrounding beds are well bioturbated sandy-siltstones. Both upper and lower contacts are obscure.

Nomad Member

The Nomad Member at Burnt Timber Cr. is 34 m thick, the lower 25 m of which is poorly exposed. The section is dominated by fissile, black silty-shales. Sideritic concretion bands were noted throughout. Localized horizons are quite bentonitic.

The basal contact of the Nomad Member is placed directly above the 1 m thick sandstone bed previously described as the Chungo Member. The upper contact is placed at the first discrete, prominent and carbonaceous sandstone bed of the overlying Belly River Formation. A more complete description of the Nomad transition into the Belly River Formation is described in a Bachelors Thesis by Reich (1983).

Cripple Creek

Muskiki, Marshybank, and Dowling Members

The lowest three members of the Wapiabi Formation at Cripple Cr. are very difficult to separate as discrete members. A slight increase in the degree of bioturbation is the only criteria which identifies the Marshybank Member from the Muskiki or Dowling Members.

The Muskiki Member is approximately 98 m thick and is poorly exposed at this location. The lowest 23 m, lying directly above the Cardium Formation, is not exposed at all. What is exposed suggests that the member is highly concretionary, slightly bioturbated, and dominated by facies 1b (Figure 7-2). The discrete siltstones of facies 1b are relatively recessive, thus they do not give the platey appearance characteristic of the Thistle Member. Some bentonite was recorded. The upper contact with the Marshybank Member is gradational.

Figure 7-2. Muskiki Member, Cripple Cr. Section. This photo shows a laminated nature within the shales. This is caused by thin siltstones of facies 1b. The cut bank is approximately 3 m high.

The 22 m thick Marshybank Member was largely identified on the basis of an increased content of facies 2 bioturbated mudstones. Fissile shales of facies 1a and 1b were noted to be interbedded with facies 2 mudstones. Colour is still dark gray to black.

The Dowling Member of Cripple Cr. is dominated by black to dark gray fissile shales of facies 1a. Locally within the member are blocky weathered and rusty coloured shale units with a moderate degree of bioturbation. Sideritic concretions are quite abundant. Thickness of the member is approximately 60 m. The upper contact with the Thistle Member is gradational, but is marked at the last concretionary band of the Dowling Member.

Thistle Member

The cyclical nature of the Thistle Member at Cripple Cr. is not nearly as well defined as it is at Burnt Timber Cr. Only two coarsening upward sequences can clearly be recognized. This is in contrast with Stott (1963), who recorded five coarsening upward cycles. Stott (1963) suggested that these cycles were completely correlative with those defined at Burnt Timber Cr.

Shales of facies 1b completely dominate the lithology at the Thistle Member at Cripple Cr. The well indurated nature of the thin siltstone turbidites gives the member the typical and characteristic platey appearance. Occurrences of facies 1a and 1c are quite rare. Where facies 1c was noted, it indicated the top of a small coarsening upward sequence.

As the top of the Thistle Member at Cripple Cr. was not exposed, a complete thickness was not obtained. However, a minimum thickness of approximately 237 m was measured.

Hanson Member

The Hanson Member at Cripple Cr., being approximately 110 m thick, is one of the thickest measured sections of the member in the study area. As at Burnt Timber Cr., the thickness of the member is due largely to a significantly reduced overlying Chungo Member.

The Hanson Member at Cripple Creek consists mainly of buff to rusty coloured mudstones characteristic of facies 2 (Figure 7-3). However, local units of discrete sandy-siltstone beds of facies 3 up to 35 cm thick occur. These beds show well defined parallel lamination. Facies 3 beds are particularly abundant in the upper portion of the member.

The upper contact with the Chungo Member, although transitional, is placed at the base of the first facies 5 HCS unit. This contact is difficult to discern as exposure of the transitional units is poor. The lower contact with the Thistle Member is not exposed.

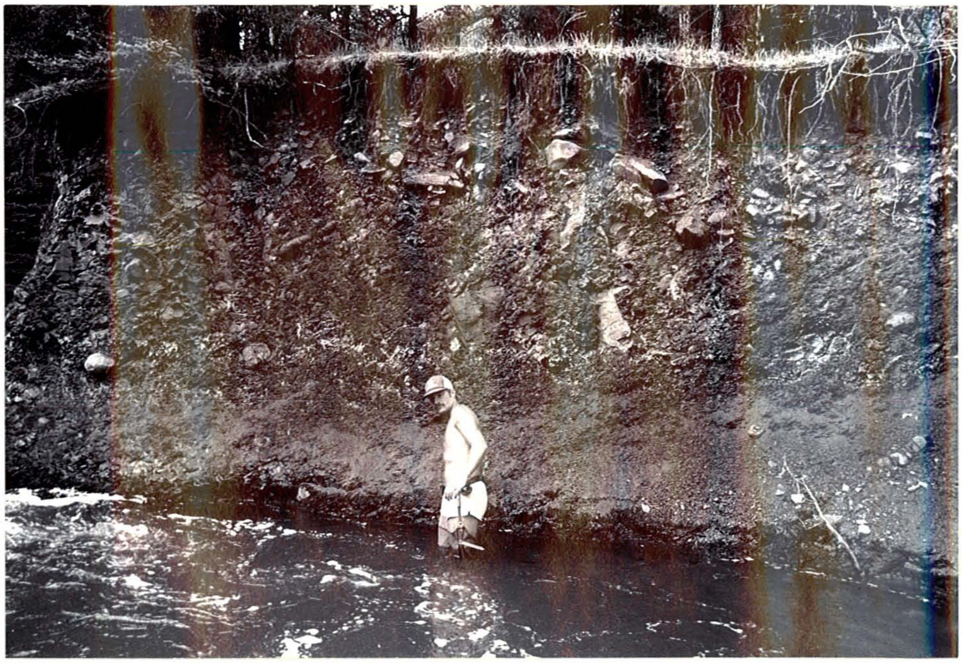
Chungo Member

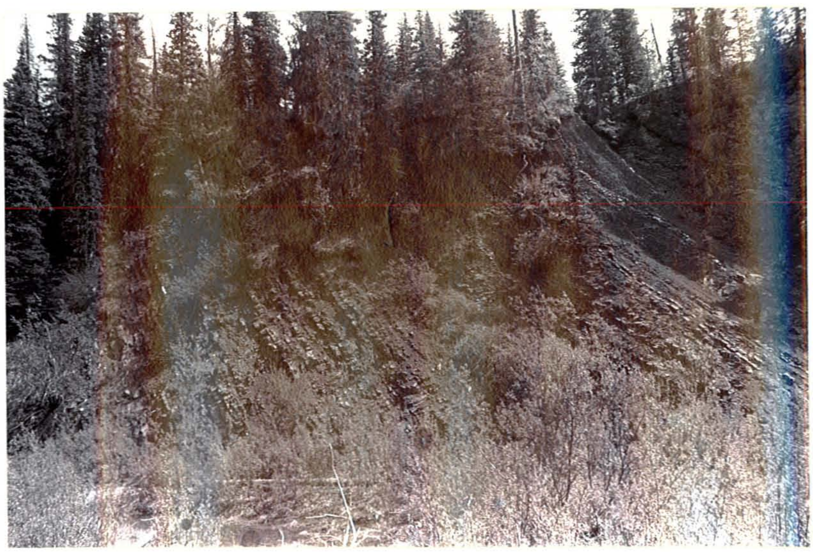
The Chungo Member at Cripple Cr. is 30 m thick and is very poorly exposed. The base of the member is defined at the first occurrence of facies 5 (HCS) above the bioturbated siltstones of facies 2. The facies 5 unit is approximately 5 m thick and is characterized by very fine-grained sandstone beds up to 25 cm thick interbedded with bioturbated shale to silty-shale. Scour marks were measured from bed bases, but ripples were not developed on bed tops.

Above the facies 5 unit is a 2 m thick unit of alternating

Figure 7-3. Hanson Member. This Hanson Member exposure, although poorly exposed, shows the thoroughly bioturbated nature of the black shales. The two well rounded "boulders" in the shale at the left of the photo are actually sideritic concretions. Stratigraphic top is to left. From the Cripple Creek section.

Figure 7-4. Hanson Member. A slightly overturned Hanson Member with stratigraphic top to the left is shown in this photo. This example was quite anomalous compared to other sections in the study area; bioturbation was generally lacking. Note the gradually thickening upwards nature of the discrete siltstone turbidites of facies 3. From the McLeod River section.





fissile shales and thinly bedded very fine-grained sandstones. Laminations are slightly subparallel to undulatory. *Ophiomorpha* trails are common. This unit grades up into a 2 to 5 m thick, well defined, massive and resistant sandstone unit of flaggy weathered facies 6 (SCS). All of the above units contain local horizons of complete bioturbation.

The next 15 m is characterized by completely bioturbated sandy-siltstones of facies 2. *Ophiomorpha* and *Rhizocorallium* are common. Occasional sideritic concretionary bands were observed. This unit is overlain by a 2 m thick, very sideritic unit which is characterized by a coarsening upward grain size to medium-grained sandstone. No internal laminations were observed from this unit. This is directly overlain by a 0.5 m thick, matrix supported chert pebble conglomerate bed (facies 11). The matrix is made up of very sideritic medium-grained sandstone. Maximum pebble size is approximately 1 cm. No imbrication, grading or orientation of the pebbles was observed. Following a 1.5 m thick cover above the facies 11 bed is a second 0.5 m thick pebble conglomerate. This bed is almost identical to the first and terminates the Chungo Member.

Nomad Member

The Nomad Member of Cripple Cr. is not exposed. However, a thickness of 24 m was measured from the top of the upper conglomerate bed to the base of the first carbonaceous and trough cross-bedded sandstone of the Brazeau Formation.

Bighorn River

Muskiki, Marshybank and Dowling Members

As with the lower three members at Cripple Cr., the Muskiki, Marshybank and Dowling Members at Bighorn River are difficult to separate. Only a slight coarsening in grain size to silty-shale and siltstone and an increase in bioturbation marking the Marshybank Member differentiates the three members.

The dominant lithology of the 118 m thick Muskiki Member is the shale of facies 1a. Only localized and thin units containing greater than 10% discrete siltstone beds (facies 1b) were recorded. Shales are normally black to dark gray coloured, fissile to blocky weathered, and quite concretionary. Moderate bioturbation was recorded through this member. Contact with the underlying Cardium Formation is sharp and distinct.

The Marshybank Member (18 m thick) is characterized by an increase in grain size to siltstone (buff coloured) and thorough bioturbation (facies 2). No primary laminations were noted. Sideritic concretions are abundant. Both upper and lower contacts are gradational.

Lying above the Marshybank Member are the black coloured, fissile to blocky facies 1a shales of the Dowling Member. This member, being 48 m thick, is relatively thin compared to other measured examples. However, this thickness is not exact as major portions of the member had to be estimated due to high water. Facies 1a is the only recognized facies in the member. Moderate amounts of bioturbation occured locally. Concretions are abundant throughout. Contact with the overlying Thistle Member is gradational, but is marked at the last sideritic concretion.

Thistle Member

The approximately 270 m thick Thistle Member at Bighorn River is dominated by shales and mudstones of facies 1a, 1b, 1c, and 3. There was no apparent pattern to the distribution of these facies in relation to one another. Only one clear example of a coarsening upward sequence was observed. This facies sequence changed upwards from facies 1b to 1c and finally into facies 3.

When compared to the measured section of Stott (1963), the measured section from the present study differs significantly. Stott (1963) suggested that five coarsening upward sequences could be identified at Bighorn River. He further suggested that the five sequences were correlative with those identified at Burnt Timber Cr. and Cripple Cr. While portions of the Bighorn River Thistle Member are noted to be coarser than others in the present study, it is not suggested that they cap coarsening upward sequences as Stott (1963) suggests. Neither is it proposed that slightly coarser portions can be correlated over tens or hundreds of kilometers to adjacent sections. This will be explained in greater detail in a subsequent portion of this chapter.

Hanson Member

Although the Hanson Member was well exposed along Bighorn River, nearly horizontal bedding allowed only the basal 28 m of the member to be measured. Facies 1a and 1b are the dominant lithologies, a direct contrast to the thoroughly bioturbated shales and mudstones of previously discussed sections. Thin (less than 4 cm thick) siltstone beds, interbedded with thick units of black fissile shales and siltstones occasionally show well developed ripple and parallel lamintion. Sideritic concretion horizons are abundant throughout the member. Neither upper nor lower contact is exposed.

Chungo Member

The complete thickness of the Chungo Member on Bighorn River could not be determined. This was due to the fact that the Chungo exposure forms a steep and prominent cliff high on an adjacent hill to the Bighorn River. Thus, several portions of the member are inaccessible. The basal portion of the member is particulary inaccessible. Nevertheless, the upper 65 m of the section was studied. Stott (1963) recorded at least another 30 m of Chungo below the 65 m that was measured in this study.

The lowest 9 m of the measured portion of the Chungo Member is dominated by facies 6 (SCS). The flaggy and sideritic nature of SCS is well displayed in this unit. The sandstone is very-fine grained.

A second 9 m thick sandstone unit overlies the facies 6 unit. The sandstone coarsens upwards from very fine-grained at the base to fine-grained at the top. The unit is buff coloured and appears massive and completely structureless. The reason for the massive and structureless nature is not clear. Bioturbations seems likely, but evidence from similar units upsection suggests that weathering may be the cause. In the upsection examples very faint laminations can be traced over several meters. As well, trace fossil or body fossil evidence of an active infauna was not found.

The next 12 m of the Chungo section is not exposed except for two, 1 m thick, very fine-grained sandstone beds which display facies 6 SCS. The beds are quite sideritic and show the characteristic flaggy weathering of SCS.

Four meters of facies 7, predominantly SCS with one example of a trough cross-bed set 45 cm thick, overlies the covered section. Grain size coarsens upwards slightly from very fine-grained (lower) to very fine-grained (upper) or fine-grained (lower) at the top. Laminations are very faint, but a slight flaggy nature is developed. Another 2.5 m of cover overlies this unit.

A second unit of buff to rusty coloured, massive appearing sandstone dominates the next 14 m. However, faint laminations were observed. The geometry and the undulatory nature of the laminations suggest that they are characteristic of facies 6 SCS. But, because laminations are observed only locally within the 14 m section, an interpretation of SCS for the entire unit is not proposed. A single example of a 35 cm thick trough cross-bed was identified in the very upper portion of the unit. Grain size is quite consistent at very fine-grained sandstone.

Directly overlying and truncating the afore mentioned trough cross-bed set is a 1 m thick, very sideritic, chert pebble conglomerate. This example of facies 11 is matrix supported. Individual clasts measured up to 3 cm. The average clast size is approximately 1 cm. This pebble conglomerate is one of the few in the study area to show an orientation of the pebbles. Almost all long axes are in the horizontal plane. No actual compass orientations of the long axes were recorded. No imbrication was observed.

The occurrence of facies 11 generally marks the top of the Chungo Member. At Bighorn River, however, 5 m of well developed, flaggy weathered, and sideritic SCS overlies the pebble conglomerate. After 4 m of cover, another 3 m of SCS was observed. This upper swaley unit marks the top of the exposed Chungo Member.

Nomad Member

The Nomad Member is not exposed at the Bighorn River section.

Blackstone River

The Blackstone River section is somewhat anomalous in the fact that its measured thickness is approximately 453 m. This contrasts with measured thicknesses of 620 m to the south on Bighorn River and 700 m to the north on Thistle Cr. The initial impression was that a major fault through a covered interval shortened the formation by several hundred meters. On closer inspection, however, it was noted that five of the seven members showed significant shortening, nullifying the single major fault hypothesis. The remaining two members (Marshybank and Nomad) showed no significant thickness changes. The possiblity that several smaller faults caused the shortening is also rejected because some of the shortened members are completely exposed and show no signs of faulting.

The lack of evidence for a structural solution to the problem suggests that local tectonics may have been affecting sedimentation rates in the area. However, any such tectonic activity must have begun immediately after deposition of the Cardium Formation as no shortening of the Cardium Formation was recorded in this area (Duke, per. comm.). Even if local tectonic activity was working in the Blackstone area, a 200 m to 250 m local shortening in a normally 600 m to 700 m thick formation is difficult to explain. The solution remains problematic.

Muskiki-Marshybank-Dowling

The lowest three members of the Wapiabi Formation on Blackstone River are easily identifiable and together make up approximately 120 m of section. The Muskiki Member, approximately 45 m thick, is dominated by facies 1a in the lower portion, but shows a steady increase in siltstone content upsection. The basal beds, lying sharply above the Sturrock Member of the Cardium Formation, are generally rubbly, dark gray to rusty coloured, and contain abundant small sideritic concretions.

In the middle portion of the Muskiki Member, siltstone content increases to approximately 30% (facies 1b). Some bentonite can be observed. The upper beds contain 50% to slightly greater than 50% interbedded discrete sandstone beds (facies 1c) and are coloured a much lighter gray. The whole member appears as a single coarsening upward sequence.

The Marshybank Member on Blackstone River has a measured thickness of 25 m. It gradationally overlies the Muskiki Member and is largely composed of blocky weathered siltstones and mudstones of facies 2. Bioturbation is quite extensive, resulting in structureless siltstone beds. No trace fauna were identified, however. Sideritic concretions are quite abundant. Colour ranges from dark gray in shales and silty-shales to light gray and buff in siltstones. Considering the facies relations between the Marshybank Member and the Muskiki Member, the Marshybank may actually be a continuation of the coarsening upward sequence referred to in Muskiki Member section.

The Dowling Member at Blackstone River is similar to most other Dowling sections in the study area as it is dominated by facies 1a. The entire member is devoid of a significant siltstone content. Shales are generally quite fissile, concretionary, and dark gray to black in colour. A few small local horizons showing minor bioturation were recorded. No cycles were noted with this member.

Thistle Member

The Thistle Member at Blackstone River, the thinnest measured example of the Thistle Member at 195 m, is made up of facies 1a, 1b, 1c, and 3. The most abundant of these facies, facies 1b, is responsible for the overall platey appearance of the member. Arrangement of the facies yields only one coarsening upward sequence, approximately 50 m thick. It consists of facies 1b, 1c, and 3 in ascending order. The facies 3 unit provides a considerable amount of paleoflow data from wave ripples. The remaining portion of the member largely consists of facies 1b with local occurrences of facies 1a.

As a whole, the Thistle Member is characterized by black, fissile, calcareous shale with varying proportions of light gray coloured discrete siltstone beds less than 5 cm thick. Bioturbation is generally minimal. Concretions are absent.

Hanson Member

The 100 m thick Hanson Member on Blackstone River is characterized by a changing nature in an upsection direction. The lowermost 9 m of the member is transitional from the underlying Thistle Member. It is dominated by facies 1a shales and siltstones. Some siltstones show well defined ripple and parallel lamination. The beds are quite sideritic with an abundance of sideritic concretion bands. The upper 1 to 2 m of this unit show a marked change, becoming bioturbated.

The second unit of the Hanson Member is 18 m thick and is characterized by thoroughly bioturbated black shale. Silt content appears minimal, but may be disseminated within the shales. Sideritic concretions are very abundant. Weathering of the shales results in an overall blocky appearance. Approximately 20 m of cover then follows.

Above the covered interval is a 5 m transitional zone which shows an increased silt content and a much lighter colour. Shales are still completely bioturbated (facies 2), but are much more sideritic and appear more massive than blocky.

The next 17.5 m is characterized by thoroughly bioturbated siltstone which is buff to light gray coloured and appears very blocky. Strike and dip measurements are impossible to obtain in this unit. Sideritic concretions are common.

A slight decrease in grain size and a large decrease in the degree of bioturbation is observable above the siltstone unit. This 30 m thick unit gradually coarsens upwards to the base of the Chungo Member. While the basal portion of the unit is characterized by black fissile shales and discrete siltstones of facies 1a and 1b, the upper portion is dominated by parallel and ripple laminated facies 3 siltstone beds up to 10 cm thick. Grain size and bed thickness continues to coarsen upwards into the Chungo Member. The top of the Hanson is defined at the base of a unit of facies 5 HCS beds. Interbedding of facies 5 and facies 3 is restricted to a 1 to 2 m thick unit.

Chungo Member

The 19 m thick Chungo section on Blackstone River is completely dominated by very fine-grained facies 5 HCS beds. Individual beds range from 15 cm to 40 cm thick and are interbedded with bioturbated silty-shale beds averaging 10 cm thick. No coarsening or thickening upward trend in HCS beds can be noted. Well defined wave ripple trains are developed on the tops of virtually all HCS beds, providing an abundance of paleoflow data.

The upper 1 m of the Chungo Member differs significantly from the gray coloured and non-concretionary facies 5 unit below. It consists of very sideritic silty-shale that is dominated by sideritic concretions. Directly above this concretionary unit lies a 12 cm thick chert pebble conglomerate of facies 11. Pebbles are well rounded and up to 3 cm in diameter. Both matrix and clast supported conglomerates were observed. No imbrication or pebble orientation was noted. This bed caps the Chungo Member on Blackstone River.

Nomad Member

A completely exposed example of the Nomad Member is observable at the Blackstone River section. The section is 32 m thick and is dominated by two sequences. In the first sequence (23 m thick), black shales show a coarsening upward trend from facies 1a in the basal 14 m into facies 1b, 1c, 3 and finally into facies 5. The facies 5 HCS unit is 2 m thick, gray to buff coloured, and is made up of very fine-grained sandstone. Ripples and trace fauna are common.

The upper 9 m sequence consists of facies 1b. The beginning of the sequence is marked by the sharp lithologic change from the very fine-grained sandstones of facies 5 to block shales and siltstones of facies 1b. The relative proportion of silt varies from approximately 10% to 25%. The top of the Nomad Member is defined at the base of the first massive, resistantly weathered, carbonaceous, medium-grained, and trough cross-bedded sandstone bed of the Brazeau Formation.

Thistle Creek

Muskiki, Marshybank, and Dowling Members

The three members on Thistle Cr. which make up the lowest 225 m of the Wapiabi Formation, the Muskiki (approximately 110 m), the Marshybank (17 m), and the Dowling (98 m) are virtually indistinguishable. Except for a very slight coarsening of shale to silty-shale in the Marshybank Member, no differences can be noted. All three members are highly concretionary, dark gray to rusty coloured, and are largely devoid of a significant siltstone content. Only towards the top of the Dowling Member do discrete siltstone beds make up greater than 10% of the section locally (facies 1b). This slight coarsening towards the top of the Dowling Member makes the contact between Dowling and the Thistle Members gradational. However, the contact is defined at the top of the last sideritic concretion.

Thistle Member

The Stott (1963) study suggests that five coarsening upward sequences of shale grading into siltstone (correlative with the five Thistle Member cycles Stott identified to the south) can be recognized in the Thistle Member of Thistle Cr. Although local occurrences of coarser silt size material were recognized in the present study, facies relations did not suggest that they capped coarsening upward sequences. In fact, only one sequence was identified. This sequence is characterized by facies 1b, 1c, 1c-3, and 3 in an upsection direction. It may be correlative with the single coarsening upward sequence developed at the Blackstone River section.

The Thistle Member, in general, is dominated by facies 1b. Shales are calcareous, fissile to moderately bioturbated, dark gray coloured, and non-concretionary. Areas with resistantly weathered discrete siltstone beds tend to give the member an overall banded to platey appearance. These discrete beds are lighter gray in colour and are normally less than 3 cm thick.

The top of the Thistle Member is drawn at the reappearance of the first sideritic concretion band. The transition into the overlying Hanson Member is gradational from the non-concretionary, calcareous Thistle shales to the concretionary and lighter coloured shales and siltstones of the Hanson Member.

Hanson Member

Although the basal contact of the Hanson Member with the Thistle

Member is gradational over approximately 5 m, the base of the Hanson Member at Thistle Cr. is placed at the base of the first occurrence of sideritic concretions above the underlying Thistle Member. Because of this designation, the basal 5 m of the Hanson Member is characterized by facies 1b containing approximately 25% siltstone, a lithology more characteristic of the Thistle Member.

Above the lowest 5 m is a 20 m unit of black to rusty coloured, very fissile shale of facies 1a. Bioturbation is minimal. Virtually no silt size material can be noted within this unit. Sideritic concretions are abundant throughout.

Approximately 25 m of cover separates unit 2 from unit 3. This third unit, 15 m thick, consists of gray to rusty coloured, very sideritic, silty-shale. This unit is classified as facies 2 because of extensive bioturabation. Weathering produces a very blocky appearance. Sideritic concretions are again common.

After a further 9 m of cover, 22 m of bioturbated and relatively massive appearing siltstone occurs. A few minor shale or silty-shale beds of a darker colour were observed within this unit. Ophiomorpha and Gyrochorte are common. Fourteen meters of cover then follows.

The upper contact of the Hanson Member with the Chungo Member is gradational over approximately 7 m. Massive appearing facies 2 siltstones grade up into very fine-grained sandstone beds which exhibit facies 5 HCS. Laminations are faint and not well defined.

Chungo Member

Although forming a prominent cliff along Thistle Cr., the Chungo Member is only moderately well exposed and only partially accessible. A thick rubble blanket covers several intervals of the member, making facies relations difficult to define. To further complicate matters, laminations are very faint, thus making primary structures difficult to identify.

The first 2 to 3 m of the 58 m thick Chungo section are characterized by poorly defined facies 5 HCS beds with interbedded bioturbated siltstones. Lying above this unit is 5 to 6 m of what appears to be facies 6 SCS beds with a few rare examples of trough cross-beds. The sandstone is very fine-grained, buff coloured, and weathered to a massive appearance. Laminations are quite faint.

After 7 m of rubble cover, another 5 to 6 m of facies 6 with horizons of facies 7 occurs. The swaley stratification in this interval is very low angle, parallel in many instances. Grain size is very fine sandstone. Ophiomorpha and Gyrochorte are common.

In a gradational manner, facies 6 and 7 of the previous unit change upwards into 4 m of facies 7 trough cross-bedding. This unit is flaggy weathered and is buff to rusty coloured and exhibits a few concretion layers. Sixteen meters of an inaccessible interval follows upsection.

The next 4 m is characterized by facies 7 trough cross-bedding and SCS grading into 2 m of TXB (facies 8). This unit, which shows a slight coarsening upwards in the upper 2 m to fine-grained sandstone, appears massive to blocky, and is very sideritic. Parallel lamination associated with SCS occurs throughout. Trough cross-bed sets average approximately 30 cm in thickness and make up roughly one third of the unit.

Lying above the facies 7-8 units are the only recorded examples of facies 9, 4c, and 10 in the study area. Being the only examples, these particular units are more thorougly described in the facies descriptions chapter. The facies 9 unit is approximately 3 to 4 m thick and is dominated by parallel lamination. It forms a very resistant and prominent ridge, accentuated by the very friable nature of the facies 4c unit above. This facies 4c unit is 1 m thick and appears extremely mottled. The fine-grained sandstone is very poorly indurated and is yellowish in colour (due to a high sulfur content). Carbonaceous stringers, 1 to 2 mm wide and 2 to 3 cm long, extend from the overlying facies 10 coal unit vertically downwards into this unit. Whether or not these "stringers" are actually roots has not been confirmed.

Above the facies 4c unit is a 14-21 cm thick coal unit. The facies 10 example is very friable and is in turn overlain by another 2.5 m of yellowish-green friable sandstones of facies 4c. Another 1.5 m of coal then occurs.

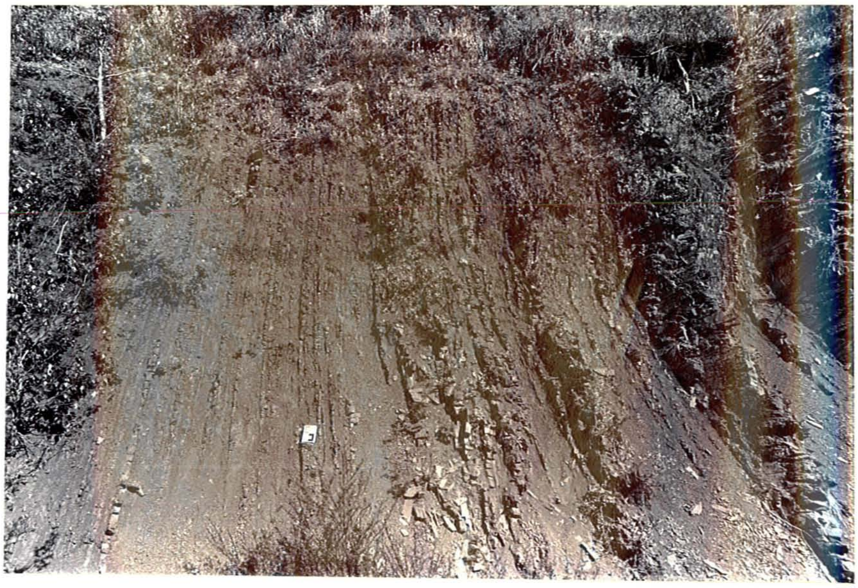
The final unit of the Chungo Member is a 6 cm thick bed of facies 11 chert pebble conglomerate. This unit appeared to be matrix supported by medium grained-sandstone, but close inspection was hampered due to the precarious position of the exposure near deepwater. Nomad shales are deposited directly on top of the pebble conglomerate.

Nomad Member

Due to poor exposure of the upper portion of the Nomad Member, the true thickness of the member was not determiend. However, the lowermost 21 m was exposed (Figure 7-5). The first 5 m is characterized by thoroughly bioturbated shales in the basal 2 m coarsening upwards

Figure 7-5. Nomad Member at Thistle Creek. This photo shows approximately 20 m of section.

Figure 7-6. Nomad Member at McLeod River. This photo shows an excellent example of a thickening upward sequence. This sequence is dominated by facies 3.



into sandy-siltstone in the top 2 m. A sharp lithologic change into subfacies 1b shales then occurs over a sideritic concretionary layer. This subfacies 1b unit is 14 m thick and is sharply truncated by a 2 m thick, very fine-grained sandstone unit of facies 6 SCS or amalgamated HCS.

The upper portion of the exposed Nomad Member begins with 6 m of cover overlying the facies 6 unit. Above the cover, 1 m of medium-grained, dark brown, sandstone of facies 8 occurs. Whether this unit is part of the marine Nomad or the non-marine Brazeau Formation is not known. After another 4 m of cover, 2 m of medium-grained, structureless sandstones occur. Five meters of cover then separates this unit from the first massive and resistant appearing, completely trough cross-bedded sandstones of the Brazeau Formation. Chert pebbles are common at the base of troughs.

McLeod River

Hanson Member

The section on McLeod River is restricted to the upper three members of the Wapiabi Formation. Portions of the Thistle Member are exposed, but tight folding prevented any stratigraphic measurement. Measuring of the McLeod River section therefore begins in the basal part of the Hanson Member. The lower Hanson contact is not exposed.

The first 38 m of the Hanson Member is dominated by very fissile, black shale (facies 1a) with abundant sideritic concretion bands. Bioturbation and silt content then gradually increases upsection (facies 2). Colours correspondingly change to gray while weathering produces a blocky texture. This second unit is approximately 51 m thick.

Nearly 90 m upsection bioturbation decreases and thin (1 to 2 cm thick) siltstone beds of facies 1c begin to dominate (Figure 7-4). Bed thicknesses then increase upsection to a maximum of 15 cm (facies 3). The resistant nature of these siltstone beds give this portion of the section a striped appearance. Laminations within siltstone beds are very faint, but range from parallel to subparallel.

Approximately 117 m above the first measured bed, the first facies 5 HCS bed was encountered. This bed is discrete and is contained within facies 3 beds. Three more meters of facies 3 was recorded above the HCS bed. High water then prevented the measuring of the next 10 m (estimated).

Chungo Member

The 39 m thick Chungo Member on McLeod River forms a very prominent and resistant, rusty coloured ridge (Figure 7-7). The lowest 6 m of the member (above the 10 m of cover) consists of very sideritic, very fine-grained sandstone that is thoroughly bioturbated (subfacies 4a). This is overlain by 10 m of well defined facies 5 HCS. Shell fragments, trace fauna, and ripples are common throughout this unit. Interbedded shales and siltstones are no greater than 10 to 15 cm thick.

Stratigraphically above the facies 5 unit lies 4 m of massive, structureless, very sideritic, and very fine-grained sandstone. This unit is classified as subfacies 4a bioturbation. This unit is in turn overlain by 1.5 m of massive to slightly flaggy weathered facies 6 SCS.

Figure 7-7. Chungo Member. The Chungo Member at McLeod River forms a very prominent and resistively weathered ridge. In this photo the Chungo is slightly overturned with stratigraphic top to the right.

Grain size increases slightly to fine-grained sandstone. Six more meters of subfacies 4a then follow.

The next two meters are dominated by facies 7 SCS and TXB and some sandstone beds which exhibit well developed wave ripple trains. Interference wave ripple patterns can also be noted within this interval. An 8 m thick unit of subfacies 4a was then the last Chungo unit to be deposited before the introduction of the chert pebble conglomerate. This unit is highly sideritic and has a highly mottled appearance. A single layer of well rounded chert pebbles occurs within this unit, approximately 3 m above its base. Abruptly truncating the top of the facies 4a unit is a 2 m thick chert pebble conglomerate, the thickest single bed thickness in the study area. Pebbles are generally less than 3 cm in diameter, well rounded, and are matrix supported. The medium-grained sandstone matrix is extremely sideritic. No imbrication was noted, however orientations were not measured. Deposition of the pebble conglomerate again marked the end of Chungo deposition.

Overall, the Hanson and Chungo Members on McLeod River can be characterized as a single coarsening upward progradational sequence. Excluding the bioturbated sandstone units, the sequence is made up of facies 1a, 2, 3, 5, and 6 in ascending order.

Nomad Member

Although not completely exposed, the 27 m thick Nomad Member is an excellent example of a thickening and coarsening upward shale and siltstone sequence (Figure 7-6). The lower 5 m is characterized by subfacies 1a with discrete siltstone beds 2 to 4 cm thick. The basal 1 m is quite bentonitic. Following 8 m of cover, the dominant lithology is that of subfacies 1a and facies 3. Individual siltstone to very fine-sandstone beds were measured up to 20 cm thick. Most show well defined parallel lamination. Ripples are scarce. Corresponding with the increase in siltstone and sandstone bed thicknesses is the systematic decrease in bed thicknesses of interbedded shales.

The last 8 m of the Nomad Member is completely dominated by facies 3. Although not significantly thicker, beds are consistently made up of very fine-grained sandstone. Trace fauna and shell fragments are abundant on bed bases. Poorly defined ripples dominate bed tops. This unit is conformably overlain by carbonaceous, trough cross-bedded sandstones of the Brazeau Formation.

Little Berland River

Muskiki and Bad Heart Formations, Dowling Member

As a result of a sandstone facies being developed at the stratigraphic level of the Marshybank Member on Little Berland River, the term "Bad Heart Formation" is used here. This also results in the adoption of "Muskiki Formation" in place of "Muskiki Member". The Dowling Member is maintained to describe the shale beds overlying the Bad Heart Formation.

Measurement of the Muskiki Formation is restricted to 7 m of poorly exposed shales lying immediately below the Bad Heart Formation. The majority of the formation is covered. The exposed shales are black, fissile, concretionary, and contain minor amounts of thin silty beds. They are classified as subfacies 1a.

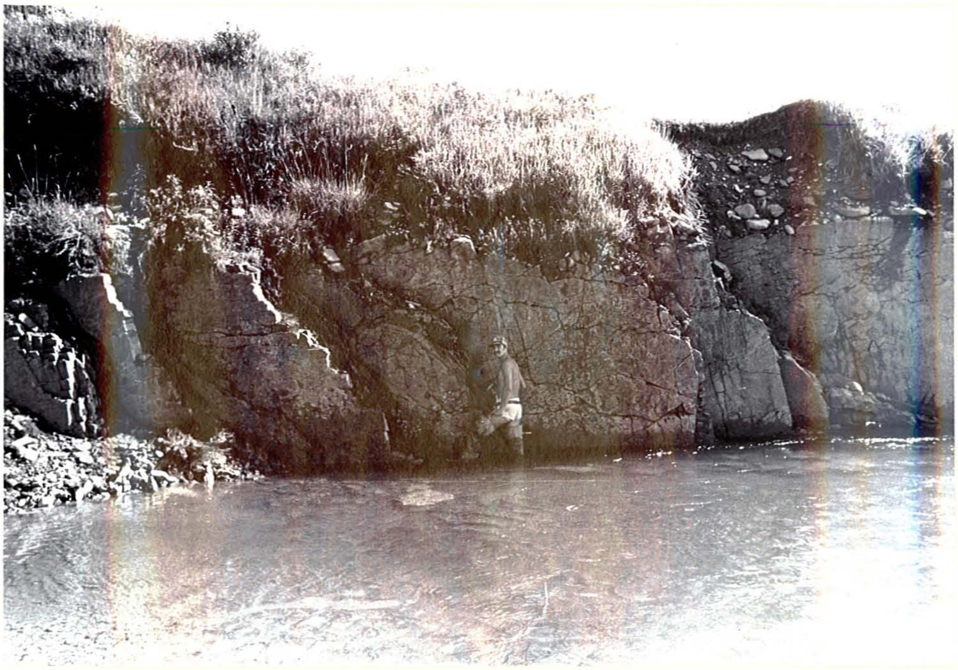
A 1 m thick unit of fissile weathered, brown to rusty coloured silty-shale characterizes the initial deposition of the 5.5 m thick Bad Heart Formation. The next 4.5 m of the formation are dominated by sharp based facies 3 sandstones with interbedded siltstones and shales (Figure 7-8 and 7-9). Sandstone beds are up to 20 cm thick and show well developed parallel and current ripple lamination. Sole marks are beautifully displayed on bed bases. Grain size is very fine sandstone. On a second Bad Heart Formation exposure immediately across the Little Berland River, but not accessible during the present study due to high water, Duke (per. comm.) measured a 14 cm thick chert pebble conglomerate bed at the top of the formation. Duke (per. comm.) also recorded rare discrete pebbles disseminated in the lower several meters of the Dowling Member.

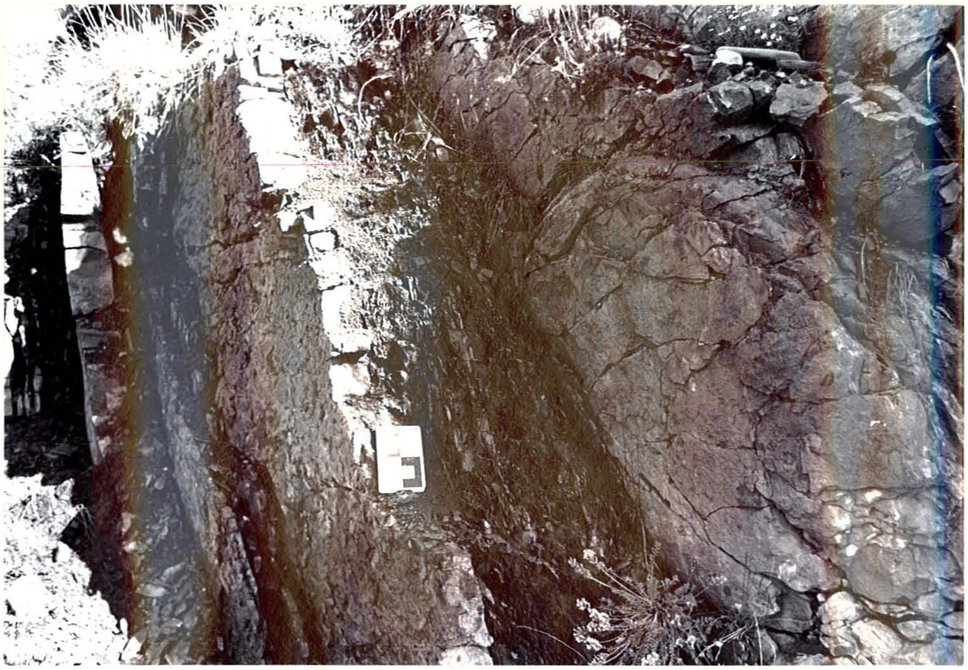
Although not well exposed, the Dowling Member at Little Berland River was measured to be anomalously thick. The upper contact of the member with the Thistle Member, defined at the point where concretionary shales changed into non-concretionary, platey, calcareous shales, is not exposed, but concretionary shales are still exposed approximately 125 m above the top of the Bad Heart Formation.

The first 17 m of the Dowling Member is characterized by shales completely devoid of siltstone (subfacies 1a). Minor amounts of discrete siltstone beds occur above the first unit, but rarely do they make up greater than 10% of any particular stratigraphic horizon. Thus, the Dowling Member is dominated by subfacies 1a with only rare occurrences of subfacies 1b. No sedimentary cycles were identified in the Dowling Member.

Figure 7-8. Bad Heart Formation. This photo shows the only exposure of the 5 m thick Bad Heart in the study area. The bases of the facies 3 very fine-grained sandstone beds are shown here. Sole marks were very abundant on these bases. From the Little Berland River section.

Figure 7-9. Bad Heart Formation. The sharp bases of facies 3 turbidites and the interbedded shales are shown here. Stratigraphic top is to the right. From the Little Berland River section.





Thistle Member

The Thistle Member at Little Berland River shows the typical black fissile shales and the platey nature described in previous sections (Figure 7-10). Subfacies 1b dominates the lithology, with only small amounts of subfacies 1a and 1c being identified locally.

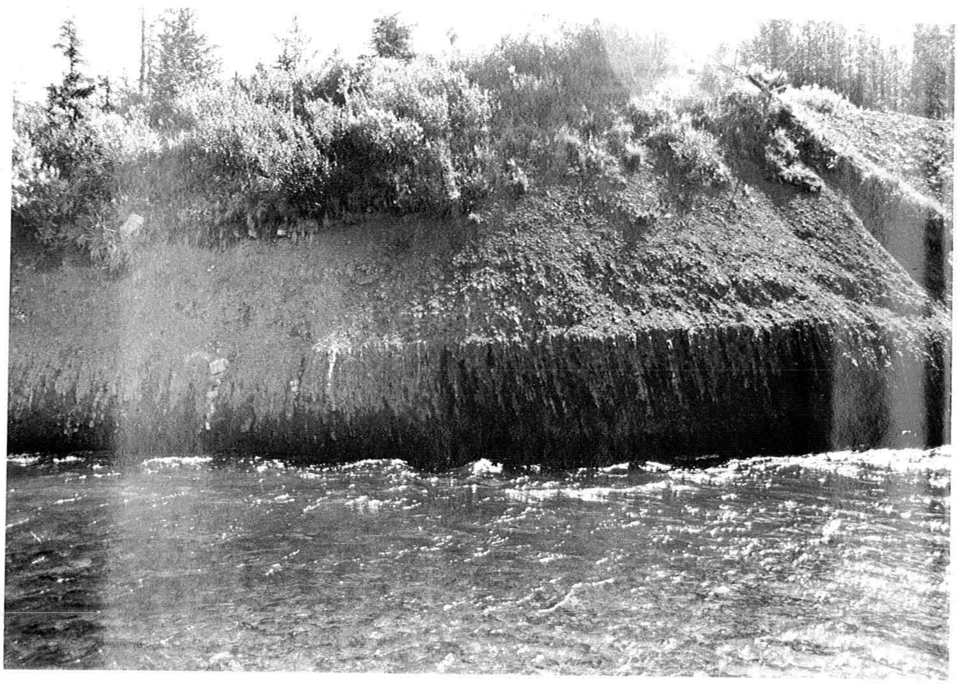
Two cycles were identified in the Thistle Member, the first of which was tentative. The first cycle began with 2 m of subfacies 1a at the base of the member which graded up into a continuous section of subfacies 1b. Across 7 m of cover, the second, and more clearly defined cycle was identified. It is 42 m thick and consists of subfacies 1a grading into subfacies 1b grading into subfacies 1c in the last several meters. These two cycles agreed with those noted by Stott (1963). However, Stott (1963) also identified two additional cycles. The fact that major portions of the Thistle Member were estimated due to high water may account for this discrepancy.

The upper contact of the Thistle Member with the Hanson Member is not exposed.

Hanson Member

A distinct thinning of the Hanson Member occurs at Little Berland River. A maximum thickness of approximately 70 m was recorded between the last Thistle Member exposure and the contact with the overlying Chungo Member. However, only the uppermost 7 m of the Hanson Member is exposed. These beds are characterized as massive to blocky appearing, buff to rusty coloured, bioturbated siltstones (facies 2). The upper contact of the member is defined at the base of a unit of poorly defined facies 5 HCS beds.

Figure 7-10. Thistle Member. This vertically dipping example of the Thistle Member clearly shows the platey nature of subfacies 1b shales. The lighter coloured beds are the discrete siltstone turbidites. Stratigraphic top is to the left. From the Little Berland River section.



Chungo Member

The Chungo Member sandstones on Little Berland River are very recessively weathered, very fine-grained, and contain very poorly defined lamination. Facies are discernible, however.

The first unit of the Chungo Member overlying the Hanson Member is characterized by 9 m of sandy-siltstones to very fine-grained sandstone beds of facies 5 HCS. Beds are rarely thicker than 15 cm. Ripples, pelecypod shells, and trace fauna are abundant on the sharp bases. Laminations are faint and poorly defined.

Following a 5 m covered section, 9 m of buff coloured siltstone is exposed. A few discrete beds within the siltstone unit are recognizable, but no laminations are exhibited. The tops of these beds are not clearly defined, probably a result of intense bioturbation. The very blocky texture of the unit, characteristic of other bioturbated units at other Chungo sections, also suggests bioturbation.

Immediately above the bioturbated siltstone unit is 1 m of well developed, very fine-grained HCS beds. Individual beds are approximately 30 cm thick and are interbedded with shales 2 to 5 cm thick. Shale content quickly diminishes upwards so discrete beds become amalgamated (facies 6). This facies 6 unit is approximately 10 m thick. This includes 2 m of cover.

A gradational interval of approximately 50 cm separates the facies 6 unit from the overlying 2 m of parallel lamination. The origin of this lamination is somewhat problematic. No suggestion of low angle diverging lamination (typical of beach lamination) was noted. As well, this lamination does not appear to be associated with facies 6.

The parallel laminated unit is the last exposure of the Chungo Member on Little Berland River. Overall, the measured thickness of the Hanson and Chungo Members is relatively short at this section. A single coarsening upward progradational sequence was recorded.

Lateral Relationships Within The Wapiabi Formation

The original stratigraphic study on the Wapiabi Formation by Stott (1963) suggests that the Wapiabi Formation can be divided into seven members which are regionally continuous from at least the Highwood River in the south to the Little Berland River in the north (Stott's Fig. 10). Thickness variations of the members from section to section are shown to be generally consistent with only few exceptions. From observations during this study, most members and member boundaries are easily recognizable at individual outcrops as well as on a regional basis. Therefore, the Wapiabi stratigraphy as set forth by Stott seems quite reasonable.

The only problem with the stratigraphic scheme concerns the Hanson-Chungo boundary. The Hanson-Chungo boundary has been placed by Stott (1963, 1967) at a transition from concretionary shales into concretionary siltstones. As this transition is very gradational (sometimes over tens of meters) and quite difficult to pin-point, it is proposed here that the contact be placed higher stratigraphically, at the much sharper siltstone to sandstone transition. It should be pointed out that the division of the Hanson and Chungo Members is largely arbitrary as the two members make up a single and continuous upward progradational sequence. Therefore, the contact should be placed at the most convenient and recognizable marker, i.e. the siltstone-sandstone transition. This division also has the effect of dividing the more proximal coarser grained material from the more distal finer grained material.

On a regional basis, the consequence of redefining the Hanson-Chungo boundary is that the Chungo Member becomes virtually non-existent at some outcrops while remaining largely unchanged at others. This effect is best shown in Figure 3-3. In the area between Burnt Timber Cr. and Cripple Cr. a siltstone facies dominates the lithology, i.e. the Chungo Member is largely absent. In areas such as Bighorn River and Thistle Cr., sand sized material is much more abundant, thus the Chungo is quite thick. It is felt that this redefinition of the Hanson-Chungo boundary gives a much more reasonable perception of the proximal-distal relationships of the members. It also has the effect of breaking away from the older "layer-cake" idea of stratigraphy (which was vogue during the time of the initial stratigraphic studies) which proposes that what is present at one place is also present at another.

While Stott (1963) suggested correlations of members on a regional basis, Stott also proposed correlations within members on a regional basis. Most of the proposed intramember correlations were within the shaly members (Muskiki, Dowling, Thistle, and Hanson), although several correlations were shown within the Chungo Member.

The main basis for correlations within and between shaly members is the development of slightly coarsening upward sequences within members. In accordance with the idea of "layer-cake stratigraphy", Stott (1963) felt that a slightly coarsening upward sequence in one outcrop section must be equivalent to a similar sequence noted in another outcrop. While this may be true in some cases, correlations across distances of up to 100 km for coarsening upwards sequences which are only meters thick and are described as shales going into silty shales seem a bit unreasonable. In fact, several of the sequences from which Stott was correlating were so subtle that they were not even identified in the present study.

It is very probable that several of the coarsening upward sequences described by Stott resulted from localized conditions such as a localized increase in clastic sediment supply. Therefore, it is suggested that with the limited amount of outcrop control, it is not justifiable to correlate minor cycles on a regional basis.

CHAPTER 8 - PALEOCURRENT DATA

In an attempt to reconstruct the regional pattern of sediment dispersal and the basin geometry during the deposition of the Wapiabi Formation, orientations of paleoflow indicators were measured in the field wherever possible. Paleoflow directions were measured from trough cross-bedding, flutes, lineations, symmetrical ripples, asymmetrical ripples and rib and furrow structures. All measurements were corrected for regional dip in the field by a technique outlined in Appendix III. Measurements from symmetrical ripples were recorded as ripple crest trends, 90° to the implied flow direction of the water.

This chapter is generally intended to be descriptively oriented. However, some broad interpretations have been made. More specific interpretations concerning these data will be presented in a subsequent chapter.

In order to note changing dispersal patterns throughout the deposition of the Wapiabi Formation, the recorded paleoflow data were divided up to show particular sedimentation patterns during the deposition of each Wapiabi Formation member. Within each member, the mean value vector ($\bar{\theta}$), the magnitude of the resultant vector (R), the magnitude of the resultant vector in terms of percent (L), and the probability that the distribution is random (P), were calculated for each measured section. These results are presented in Table 8-1. From these data, figures have been constructed which show dispersal patterns during each of the seven members (Figs. 8-1 through 8-9). Note that

Table 8-1. Summary of all paleocurrent data recorded in the present study. θ = mean value vector, R = the magnitude of the resultant vector, L = the magnitude of the resultant vector in terms of percent, and P = the probability that the distribution is random.

Section / Member	Burnt Timber Creek	Cripple Creek	Big Horn River	Blackstone River	Thistle Creek	McLeod River	Little Berland River
Nomad	NIL	NIL	NIL	asy ripples $\bar{\theta}$=20°, R=1 L=100 P=3.7x10^{-1}	NIL	asy ripples $\bar{\theta}$=15°, R=1 L=100 P=3.7x10^{-1}	NIL
Chungo	NIL	r&f, flutes $\bar{\theta}$=24°, R=38 L=95.7 P=2.5x10^{-2} lineations $\bar{\theta}$ =5/185°,R=1 L=100 P=3.7x10^{-1}	TXB,asy ripples, r&f $\bar{\theta}$=145°,R=6.96 L=99.4 P=9.9x10^{-4}	r&f $\bar{\theta}$=43.11, R=2.76 L=91.88 P=7.9x10^{-2} sym ripples $\bar{\theta}$=114/294° R=25.56 L=98.8 P=1.2x10^{-11}	TXB, r&f, flutes $\bar{\theta}$=153°, R=9.17 L=53.93 P=7.1x10^{-3}	asy ripples,r&f,TXB, flute $\bar{\theta}$=113°, R=8.43 L=60.24 P=6.2x10^{-3} sym ripples $\bar{\theta}$=115/295° R=4.65 L=93.03 P=1.3x10^{-2}	asy ripples,r&f $\bar{\theta}$=18°,R=10.94 L=99.48 P=1.9x10^{-5}
Hanson	NIL	NIL	asy ripples $\bar{\theta}$=48°, R=2.99 L=99.65 P=5.1x10^{-2}	asy ripples $\bar{\theta}$=38°, R=2.3 L=99.95 P=5x10^{-2}	NIL	asy ripples $\bar{\theta}$=10°, R=1 L=100 P=3.7x10^{-1} lineation $\bar{\theta}$=40/220° R=1 L=100 P=3.7x10^{-1}	asy ripples $\bar{\theta}$=21°, R=1 L=100 P=3.7x10^{-1}

Table 8-1. (continued)

Section	Burnt Timber Creek	Cripple Creek	Bighorn River	Blackstone River	Thistle Creek	McLeod River	Little Berland River
Member							
Thistle	sym ripples $\bar{\theta}$=112/292°, R=6.43 L=64.32 P=1.6x10^{-2} lineations $\bar{\theta}$=43/223°, R=5.83 L=97.18% P=3.5x10^{-3} Flutes,asy ripples $\bar{\theta}$=346°, R=9.56 L=86.94 P=2.4x10^{-4}	asy ripple $\bar{\theta}$=34° R=34.33 L=98.1 P=3.7x10^{-1} lineations $\bar{\theta}$=152/332°, R=1 L=100 P=3.7x10^{-1}	asy ripple,r&f $\bar{\theta}$=77°, R=58.96 L=75.58 P=4.4x10-20 sym ripple $\bar{\theta}$=114/294°, R=10.04 L=99.65 P=5.1x10^{-2}	asy ripples $\bar{\theta}$=26°,R=12.80 L=98.45 P=3.4x10^{-6} sym ripple $\bar{\theta}$117/297°, R=27.37 L=88.30 P=3.2x10^{-11}	asy ripple $\bar{\theta}$=36°,R=68.82 L=86.02% P=1.9x10^{-26} lineation $\bar{\theta}$=10/190° R=1 L=100% P=3.7x10^{-1}	NIL	asy ripple,r&f $\bar{\theta}$=18° R=5.83 L=97.18 P=3.5x10^{-3}
Dowling	NIL	asy ripples $\bar{\theta}$=40°, R=1 L=100 P=3.7x10^{-1}	asy ripples $\bar{\theta}$=40°, R=22.24 L=96.68 P=5x10^{-10}	asy ripples $\bar{\theta}$=48°, R=1 L=100 P=3.7x10^{-1}	asy ripples $\bar{\theta}$=90°, R=3.81 L=95.33 P=2.6x10^{-2}	NIL	NIL
Marshybank/Bad Heart	NIL	NIL	NIL	NIL	NIL	NIL	asy ripples, flutes $\bar{\theta}$=99° R=4.15 L=83.03 P=3.2x10^{-2} lineations $\bar{\theta}$=151/331° R=2.00 L=99.94 P=1.4x10^{-1}
Muskiki	lineation $\bar{\theta}$=75/255°, R=1 L=100 P=3.7x10^{-1}	asy ripples $\bar{\theta}$=50°, R=2.95 L=98.23% P=5.5x10^{-2}	asy ripples $\bar{\theta}$=58°, R=2.91 L=96.89 P=5.9x10^{-2}	NIL	asy ripples $\bar{\theta}$=60°, R=1 L=100 P=3.7x10^{-1}	NIL	

these figures feature palinspastically reconstructed positions of the measured sections. Rose diagrams have also been prepared to show paleoflow patterns during the deposition of each Wapiabi Formation member. These diagrams are shown on the measured sections which are located in the rear pocket.

A discussion of paleoflow data recorded from each member now follows.

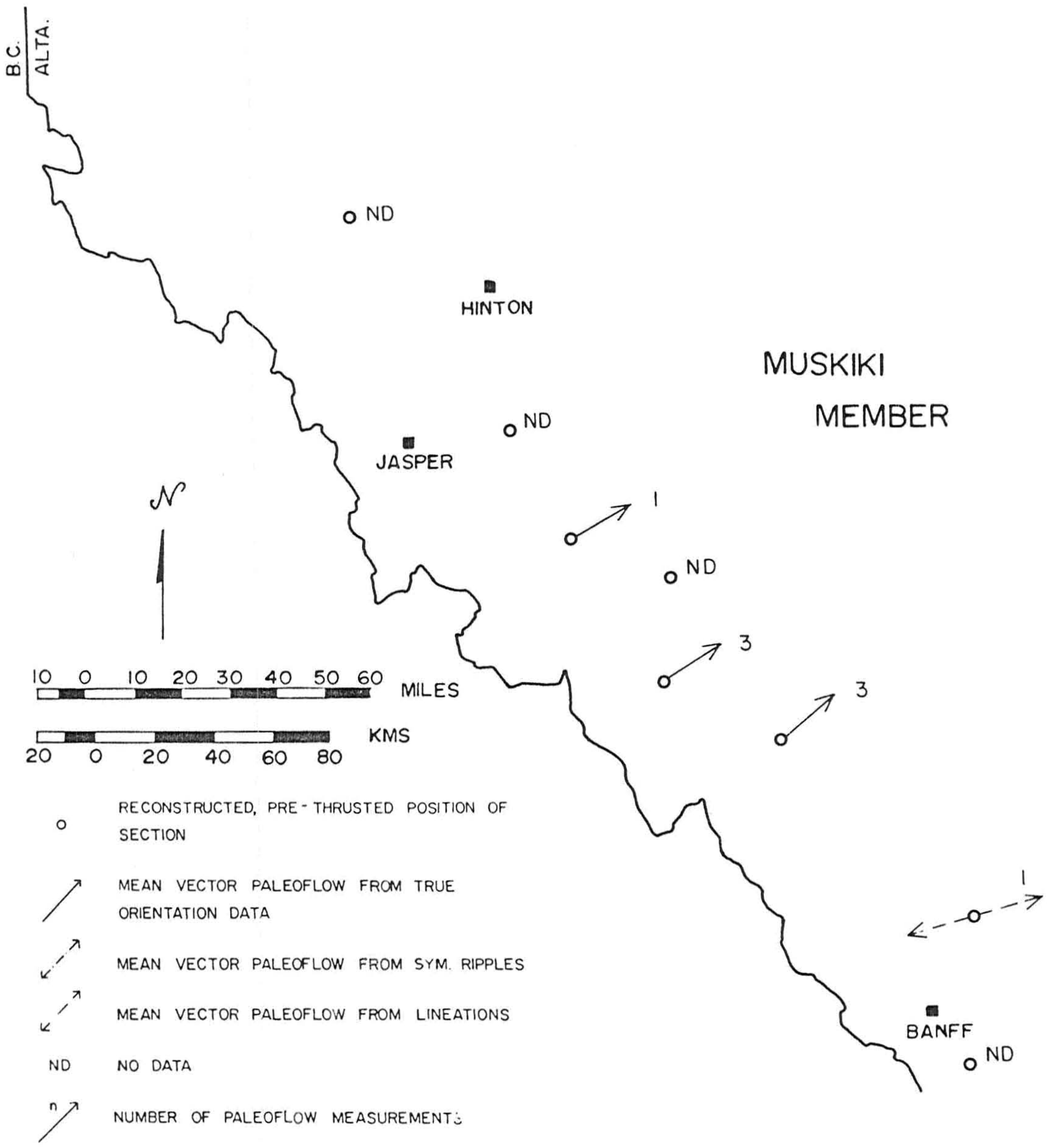
Muskiki, Marshybank, and Dowling Members

Due to the lack of paleocurrent data measured from the Muskiki, Marshybank, and Dowling Members, and because identification of these three members is questionable at some localities, the recorded paleocurrent data have been grouped into a single discussion. The grouping of the members seems justifiable when it is noted that there is generally little variation in flow orientation between members at any one outcrop (Figures 8-1, 8-2, and 8-3).

Paleoflow measurements collected from the lowest three Wapiabi Formation members were measured from asymmetrical ripples and lineations. In order to interpret these data, an assumption must be made. It is assumed that the lineations and current ripples were formed by turbidity currents. Thus, paleoflow indicators reflect a downslope and offshore direction. Paleoslope directions and basin geometries for the Fernie Formation suggested by Hamblin and Walker (1979) were based on a similar assumption.

Five of the eight measured sections in the present study yielded paleoflow data from the Muskiki, Marshybank, and Dowling Members, in a south to north direction: Cripple Creek, Bighorn River, Blackstone

Figure 8-1. Paleocurrent data from the Muskiki Member of the Wapiabi Formation in the study area.


B.C.
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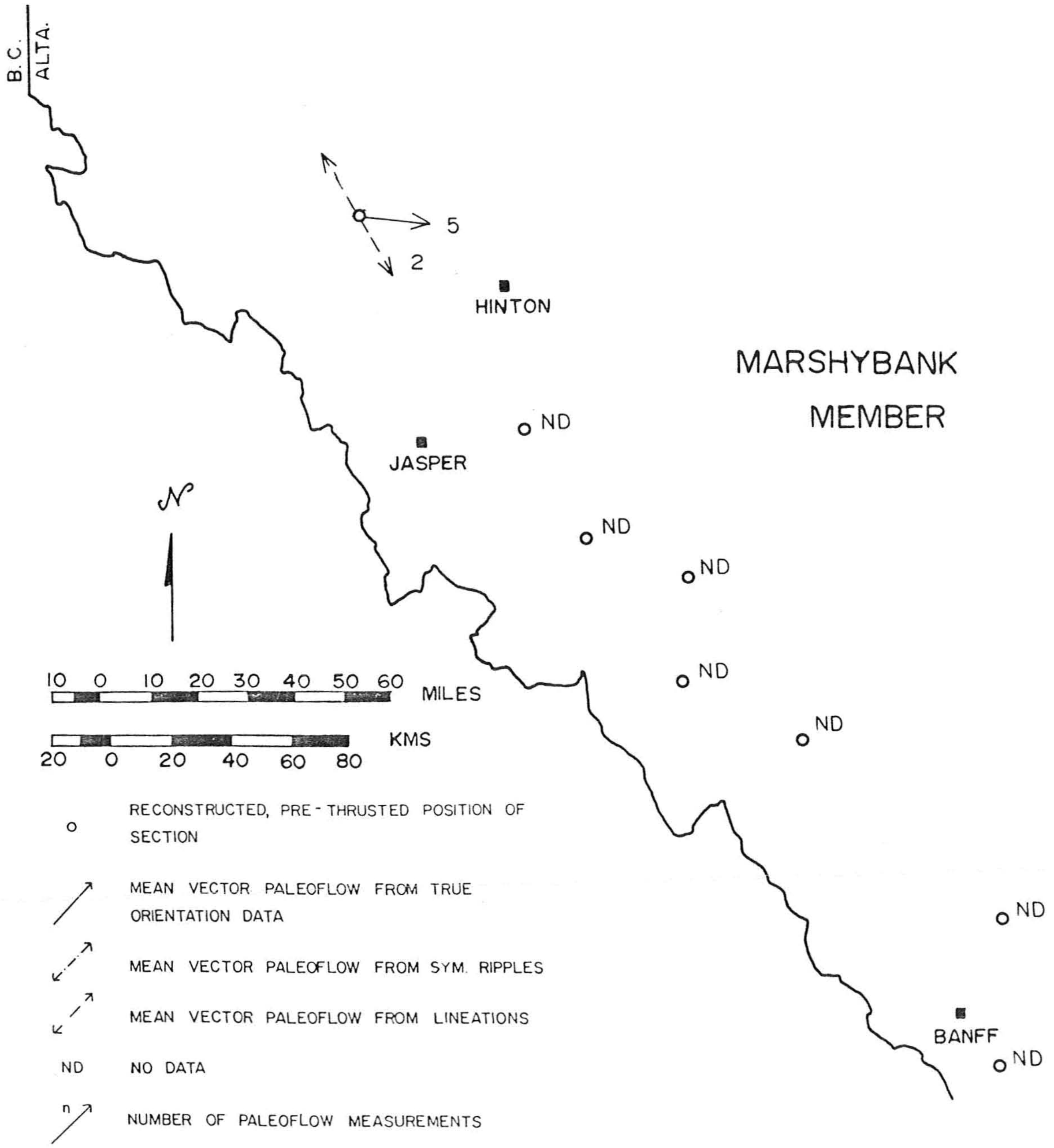
RECONSTRUCTED, PRE-THRUSTED POSITION OF SECTION
MEAN VECTOR PALEOFLOW FROM TRUE ORIENTATION DATA
MEAN VECTOR PALEOFLOW FROM SYM. RIPPLES
MEAN VECTOR PALEOFLOW FROM LINEATIONS
ND NO DATA
n NUMBER OF PALEOFLOW MEASUREMENTS

Figure 8-2. Paleocurrent data derived from the Marshybank Member in the present study area.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HINTON
MARSHYBANK
MEMBER
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KMS
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RECONSTRUCTED, PRE-THRUSTED POSITION OF SECTION
MEAN VECTOR PALEOFLOW FROM TRUE ORIENTATION DATA
MEAN VECTOR PALEOFLOW FROM SYM. RIPPLES
MEAN VECTOR PALEOFLOW FROM LINEATIONS
ND NO DATA
n NUMBER OF PALEOFLOW MEASUREMENTS
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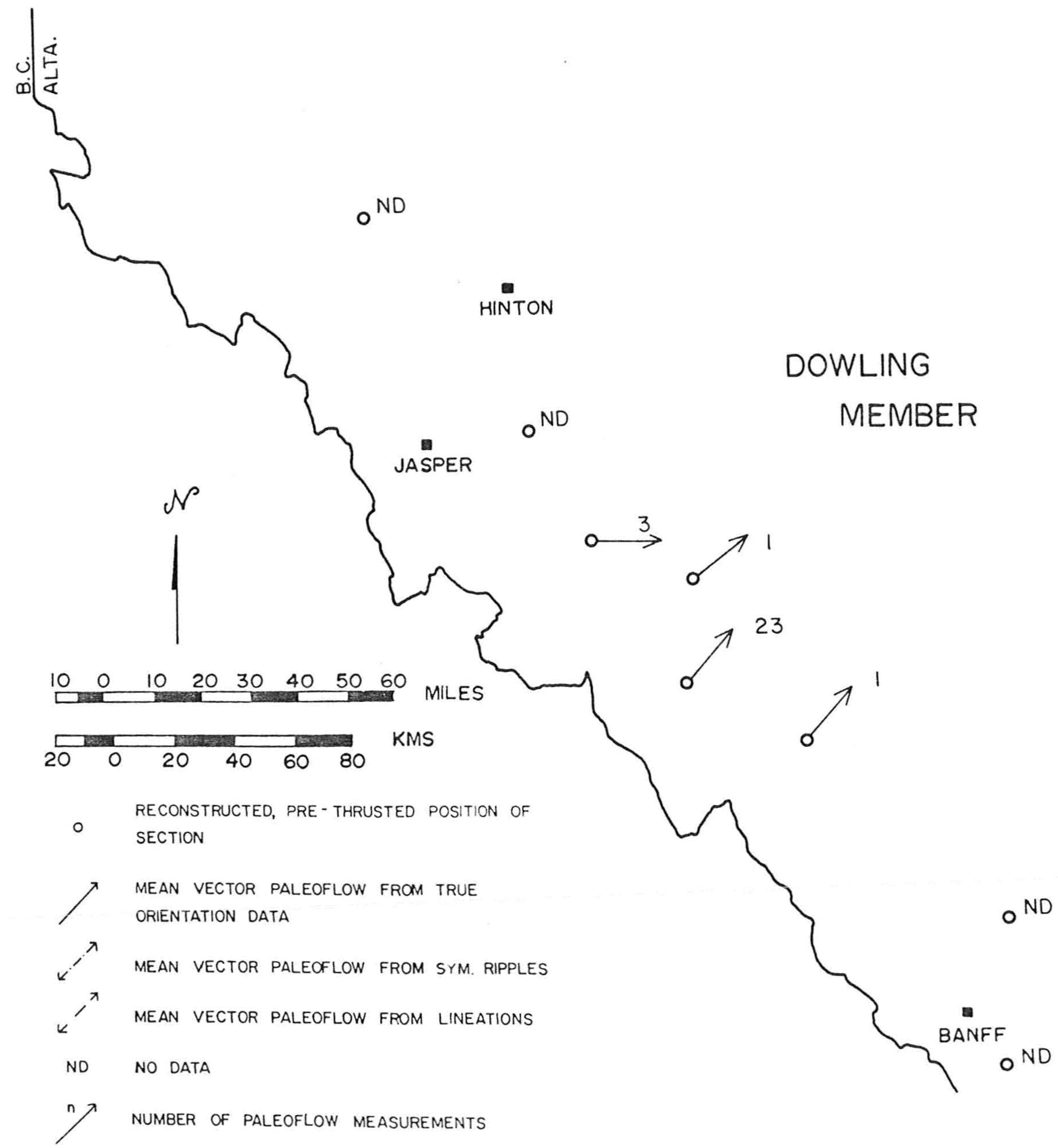
ND

Figure 8-3. Paleocurrent data from the Dowling Member in the present study area.



B.C.
ALTA.
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HINTON
DOWLING
MEMBER
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10 0 10 20 30 40 50 60 MILES
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RECONSTRUCTED, PRE-THRUSTED POSITION OF SECTION
MEAN VECTOR PALEOFLOW FROM TRUE ORIENTATION DATA
MEAN VECTOR PALEOFLOW FROM SYM. RIPPLES
MEAN VECTOR PALEOFLOW FROM LINEATIONS
ND
NO DATA
n
NUMBER OF PALEOFLOW MEASUREMENTS
ND
BANFF
ND

River, Thistle Creek, and Little Berland River. The three most southerly sections showed strongly consistent flow directions to the northeast. However, at Thistle Cr. an easterly oriented flow (mean vector = 90°) was recorded from the Dowling Member. On Little Berland River, during Marshybank time, the flow orientation had changed significantly. Seven paleoflows measured from sandy turbidite beds in the Bad Heart Formation (Marshybank equivalent) indicated the mean flow direction to be to the east-southeast. This coincides with a thickening and coarsening of the Bad Heart Formation sandstone facies in the opposite direction, northwest (Stott, 1967; Duke, pers. com.). This suggests that sandy fluvial and nearshore environments, which existed some distance to the northwest of the Little Berland River, shed detritus east-southeastward.

From these data, a northwest-southeast striking paleoslope is inferred to have existed in the southern study area during the late Turonian to early Santonian. In the central portion of the study area, paleoflow measurements indicated that the basin slope strike changed to a more north-south orientation. Basin dip was to the east. Finally, in the northern part of the study area, a basin slope strike with a northnortheast-southsouthwest to northeast-southwest orientation is indicated to have existed.

Thistle Member

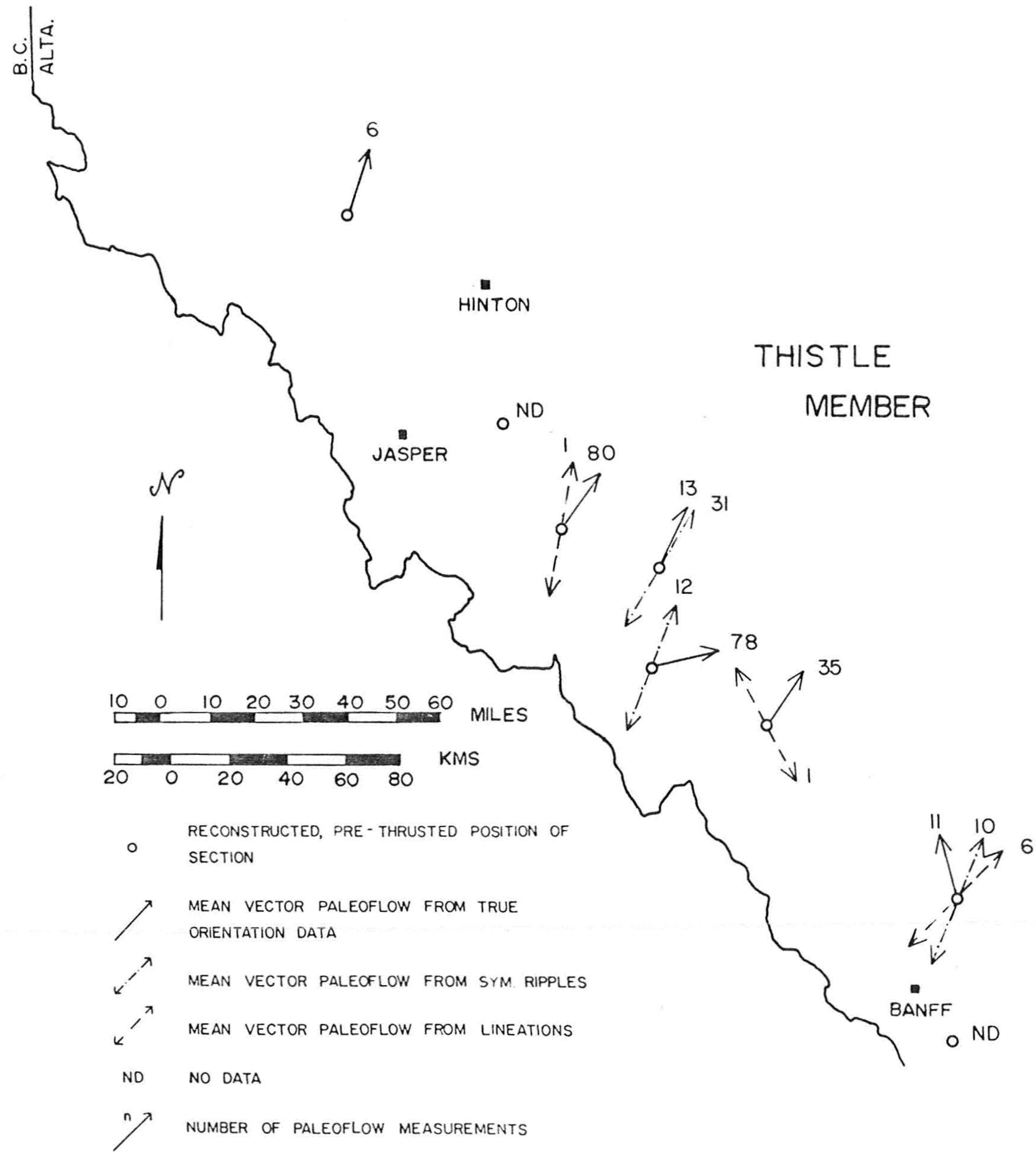
Most paleoflow measurements taken in this study were recorded from Bouma's division c in the abundant subfacies 1b turbidites of the Thistle Member. Several examples of symmetrical ripples were also measured in this member. However, only paleoflow measurements recorded from turbidite beds were used in calculating vector means for the member and determing basin geometries. This method is used through this chapter.

Although paleocurrent results from the Thistle Member are slightly variable from section to section in a south to north direction, the variability shows a systematic change rather than a random change. At Burnt Timber Cr. in the south, 3 flutes, 8 asymmetrical ripples, and 6 lineations (it is suggested that lineations show flow in the direction closest to orientations recorded from other turbidite flow indicators) yielded a vector mean paleocurrent direction of 007° (Figure 8-4). This measurement is interesting when compared to paleoflow measurements from the Thistle Member at Ghost Dam, a section approximately 60 kms to the south (Rosenthal, in prep.). Rosenthal (in prep.) recorded predominantly southerly oriented paleoflows.

At the Cripple Cr. section further to the north, the paleoflow vector mean was calculated to be 032°, a change to the east from the Burnt Timber Cr. orientation of 007°. A further shift to the east was noted at the Bighorn River section where a vector mean of 077° was obtained. In the northern half of the study area, relatively consistent flow directions to the northnortheast were recorded. Vectors means ranged from 018° (Little Berland River) to 036° (Thistle Cr.).

Disregarding the measurements from Burnt Timber Cr., paleoflow measurements strongly suggest a westnorthwest-eastsoutheast striking paleoslope in the northern portion of the study area and a northwest-southeast to northnorthwest-southsoutheast striking paleoslope in the southern portion of the study area. Both areas dip

Figure 8-4. Paleocurrent data recorded from the Thistle Member in the study area. Data from symmetrical ripples are shown as the direction of wave motion rather than the crest trend of the ripples.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ALTA.
THISTLE
MEMBER
HINTON
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20 0 20 40 60 80 KMS
RECONSTRUCTED, PRE-THRUSTED POSITION OF SECTION
MEAN VECTOR PALEOFLOW FROM TRUE ORIENTATION DATA
MEAN VECTOR PALEOFLOW FROM SYM. RIPPLES
MEAN VECTOR PALEOFLOW FROM LINEATIONS
ND NO DATA
n NUMBER OF PALEOFLOW MEASUREMENTS

towards the northeast or east. The shoreline during this time probably would have been a considerable distance to the west or southwest. The Burnt Timber Cr. results are somewhat anamolous and will be discussed more thoroughly in a subsequent section.

Hanson Member

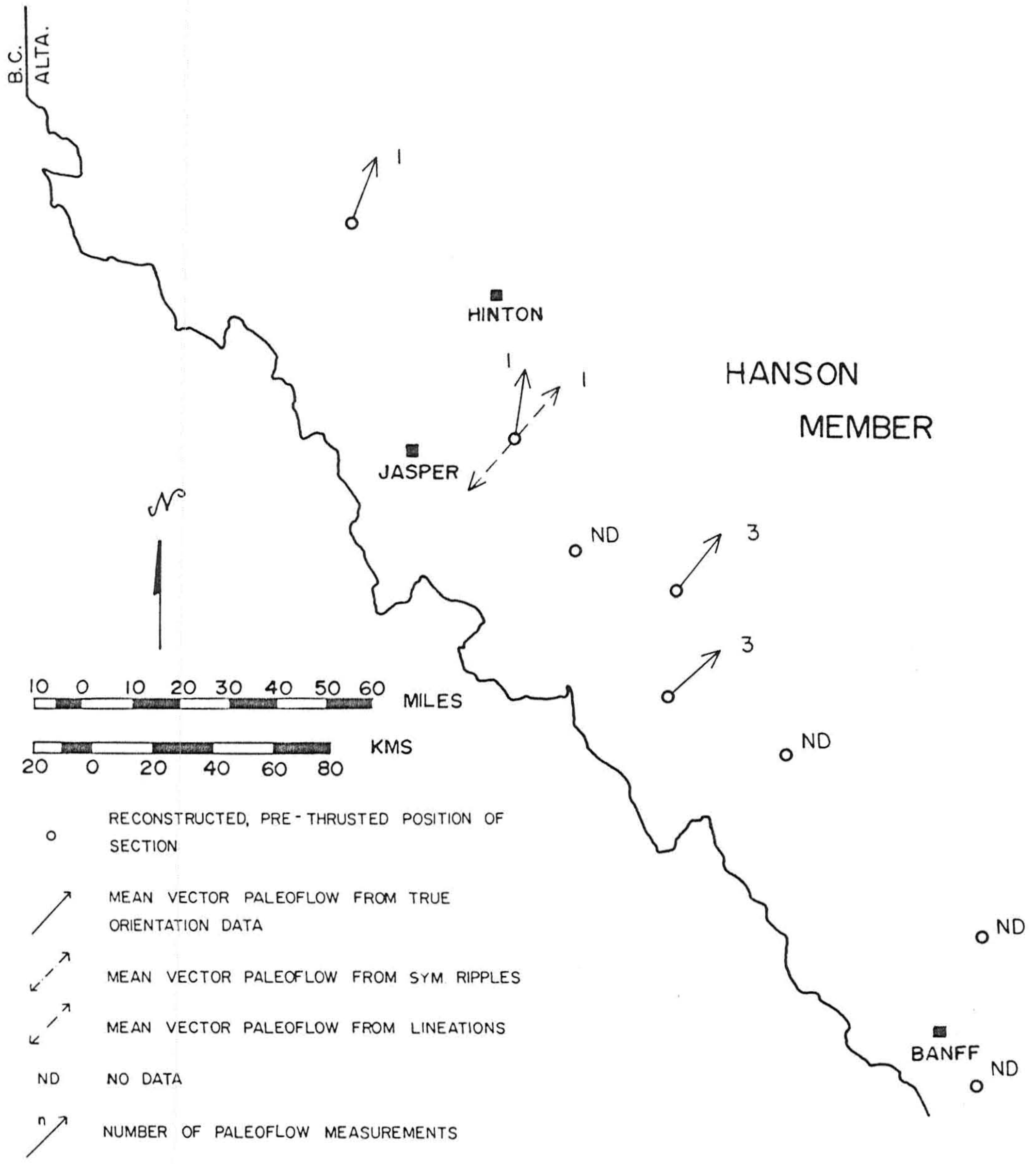
One of the characteristics of the Hanson Member is its high degree of bioturbation. Because of this bioturbated nature, only 9 paleocurrent measurements were recorded, three from Bighorn River, three from Blackstone River, 2 from McLeod River and one form Little Berland River. However, of the nine measurements, 8 showed flow directions oriented towards the northeast (Fig. 8-5). The one anomalous paleoflow, oriented at 130°/310°, was measured from a lineation at McLeod River. However, in a regional sense, the paleoslope strike during the Hanson time can probably be inferred to have been oriented approximately northwest-southwest. Slope dip was then to the northeast.

Chungo Member

Due to a wide variety of preserved sedimentary structures within the Chungo Member sandstones, paleoflow measurements were derived from an array of structures. These included trough cross-bedding, asymmetrical and symmetrical ripples, rib and furrow structures, flutes, and lineations. Overall, paleoflows measured from the Chungo Member showed a high degree of variability in preferred paleocurrent orientations between sections. Considerable variation was also noted within individual sections, particularly the Mt. Yamnuska section.

Lerand (1982) measured 129 paleocurrent directions from trough

Figure 8-5. Paleocurrents recorded from the Hanson Member in the present study area.



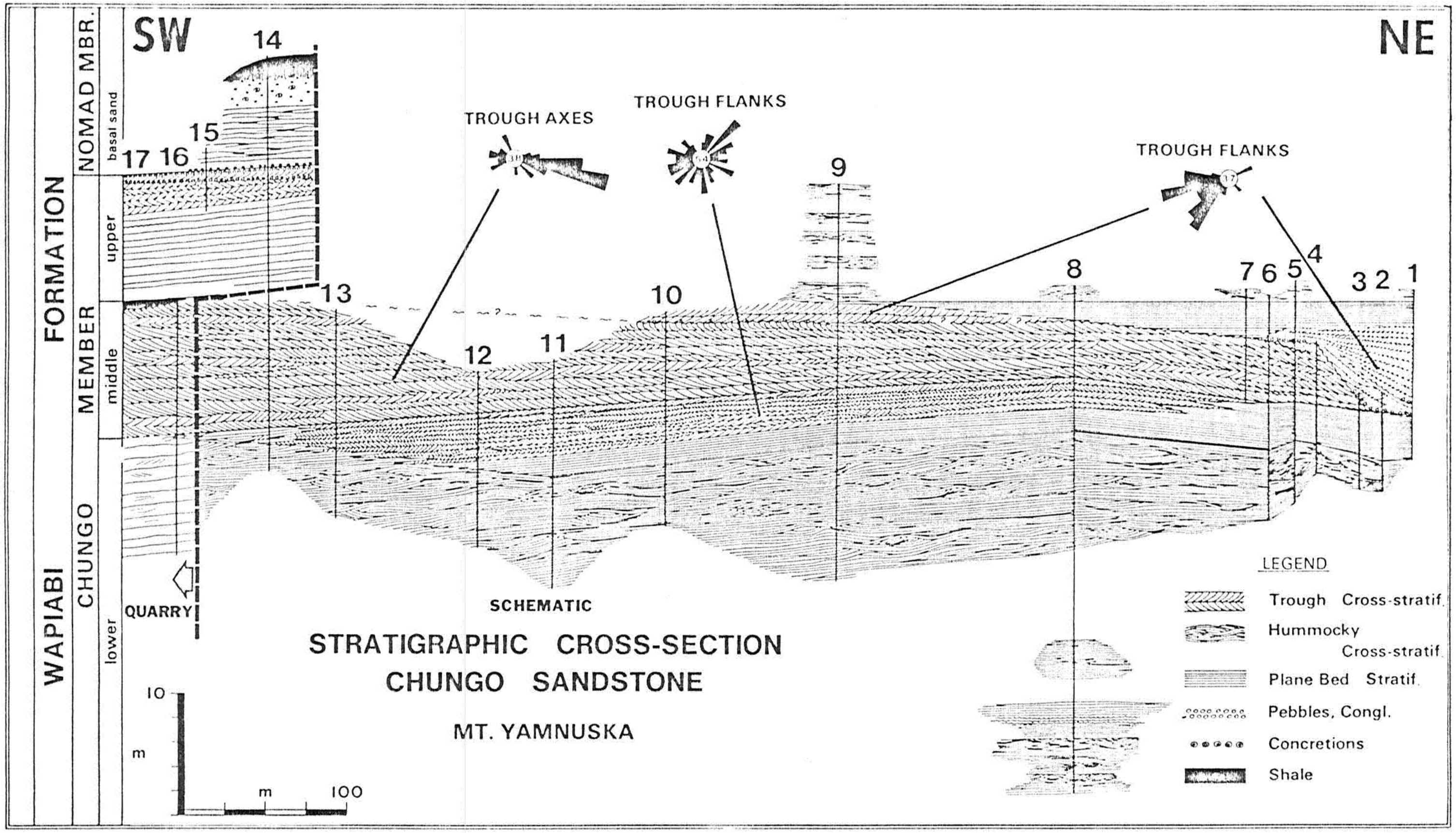
B.C.
ALTA.
HINTON
JASPER
BANFF
HANSON
MEMBER
N
1
1
1
ND
3
3
ND
ND
ND
10 0 10 20 30 40 50 60 MILES
KMS
20 0 20 40 60 80
RECONSTRUCTED, PRE-THRUSTED POSITION OF SECTION
MEAN VECTOR PALEOFLOW FROM TRUE ORIENTATION DATA
MEAN VECTOR PALEOFLOW FROM SYM. RIPPLES
MEAN VECTOR PALEOFLOW FROM LINEATIONS
ND
NO DATA
n
NUMBER OF PALEOFLOW MEASUREMENTS

cross-bedding at the Mt. Yamnuska section. The radial distribution of these measurements is shown in Figure 8-6. As can be seen, there is no overall preferred paleflow orientation. Rather, paleocurrents were multidirectional. However, in the middle and upper portions of unit 4 (see previous section), unimodal flows were developed. These flows were oriented at approximately 100° and 250°. Evidence therefore suggests that flows generally followed a radial pattern with locally developed preferred orientations. The fact that the two preferred orientations are roughly 150° apart (almost 180° bimodal) and that the other paleocurrents are radially oriented is highly suggestive of a tidally influenced area.

Although it is difficult to propose basin geometries from radially oriented paleoflow measurements, other lines of evidence suggest that the basin strike at Mt. Yamnuska was roughly oriented north-south; dip was to the east. Two current ripples from turbidite beds measured at the base of the Chungo section yielded flow directions to the east and southeast. As well, the Chungo section exposed on Old Fort Cr., approximately 3 kms to the east of Mt. Yamnuska, was considerably thinner than the Mt. Yamnuska section (Stott, 1963). Thinning sandstone bodies in shallow marine environments generally indicate the down basin and offshore direction; in this case approximately to the east.

The number of paleoflows measured from the seven remaining Chungo Member sections in this study was relatively small, a result of weathering, intense bioturbation, and poor exposure. At Burnt Timber Cr., no paleoflows were obtained at all due to the total degree of

Figure 8-6. Schematic stratigraphic cross-section of the Chungo Member at Mt. Yamnuska. From Lerand, 1982.



SW
NE
WAPIABI FORMATION
CHUNGO MEMBER
NOMAD MBR.
basal sand
upper
middle
lower
TROUGH AXES
TROUGH FLANKS
TROUGH FLANKS
17 16 15 14
13 12 11 10 9 8
7 6 5 4 3 2 1
QUARRY
SCHEMATIC
STRATIGRAPHIC CROSS-SECTION
CHUNGO SANDSTONE
MT. YAMNUSKA
10
m
m 100
LEGEND
Trough Cross-stratif.
Hummocky Cross-stratif.
Plane Bed Stratif.
Pebbles, Congl.
Concretions
Shale

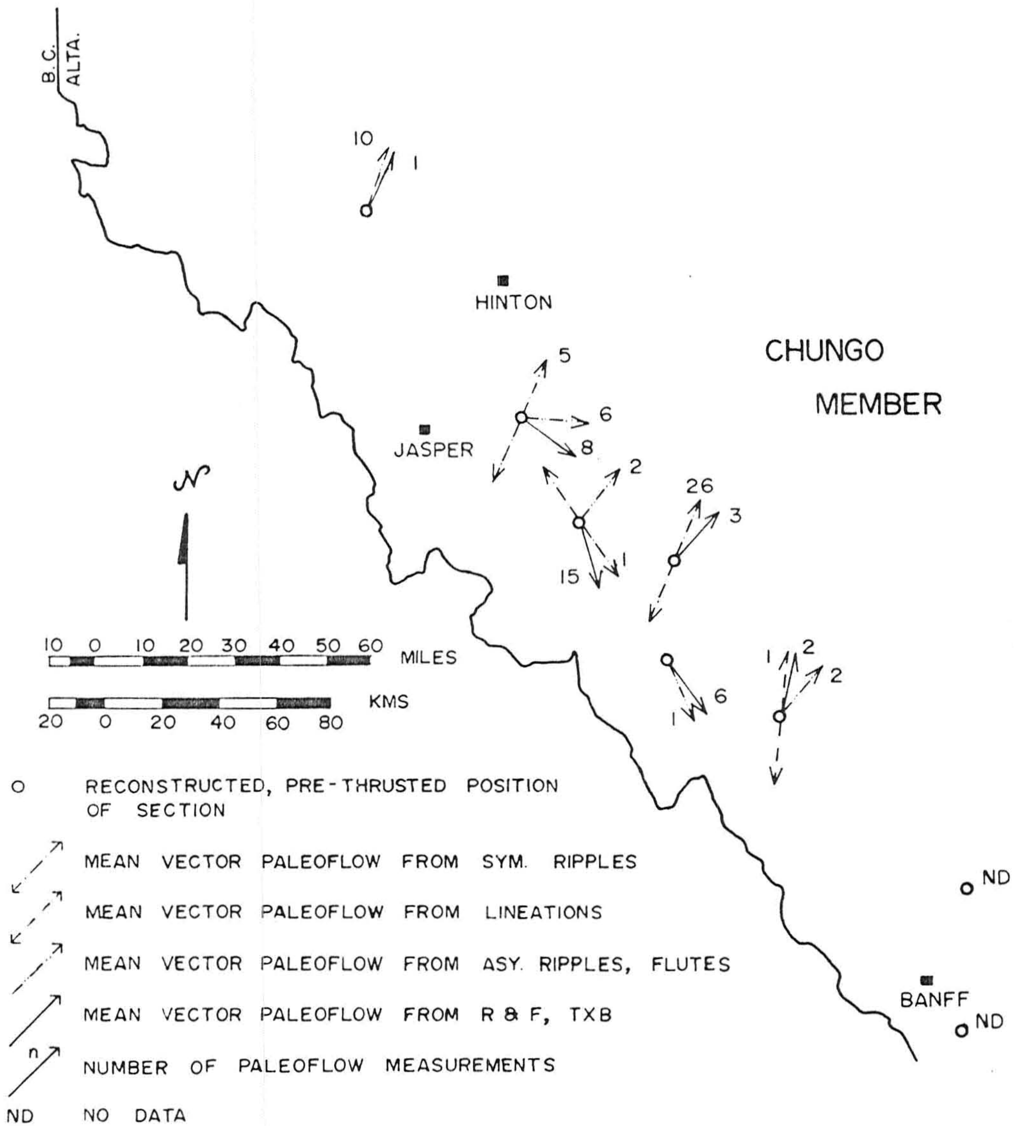
bioturbation. Three paleoflow measurements, obtained from one lineation and 2 flutes at Cripple Cr., yielded a vector mean of 028° (Figure 8-7). Rib and furrow structures gave a vector mean of 008°. However, slightly to the north on Bighorn River an asymmetrical ripple was oriented 155°. Trough cross-beds and rib and furrows averaged 143°.

Three rib and furrow structure measurements obtained from the Blackstone River section suggested that flow was to the northeast, $\bar{\theta}$ = 48°. This reflected an almost 90° change in flow orientation from the Bighorn River section to the south. No orientations from turbidites were measured from this section.

Slightly to the north on Thistle Cr., paleoflows showed a wide distribution of orientations. However, paleoflows measured from two flutes yielded flow directions of 35° and 40°, suggesting that the down basin direction was to the northeast. The remaining paleoflows from Thistle Cr., all from trough cross-bedding and rib and furrow structures, showed flows oriented from 20° to 202°. There was, however, a strong grouping of paleoflows oriented approximately 170° to 180°. The mean vector from trough cross-bedding and rib and furrow structures was 152°.

Paleoflows measured from rib and furrow structures, asymmetrical ripples, symmetrical ripples, flutes, and trough cross-beds at McLeod River showed that flow patterns were quite variable during Chungo deposition at that locality. Azimuths ranged from northeast to south-southwest in the eastern half of the compass. No consistent trend in paleoflow direction changes was noted in an upsection direction. Rather, paleoflows appeared to be randomly oriented throughout the section. However, paleoflow measurements from turbidite beds in the

Figure 8-7. Paleocurrents recorded from the Chungo Member in the present study area. Data from symmetrical ripples are shown as the direction of wave motion rather than the crest of the ripple.


B.C.
ALTA.
10
1
HINTON
CHUNGO
MEMBER
5
6
8
JASPER
2
26
3
15
1
10 0 10 20 30 40 50 60 MILES
KMS
20 0 20 40 60 80
1
2
2
6
1
RECONSTRUCTED, PRE-THRUSTED POSITION OF SECTION
MEAN VECTOR PALEOFLOW FROM SYM. RIPPLES
MEAN VECTOR PALEOFLOW FROM LINEATIONS
MEAN VECTOR PALEOFLOW FROM ASY. RIPPLES, FLUTES
MEAN VECTOR PALEOFLOW FROM R & F, TXB
n
NUMBER OF PALEOFLOW MEASUREMENTS
ND
NO DATA
ND
BANFF
ND

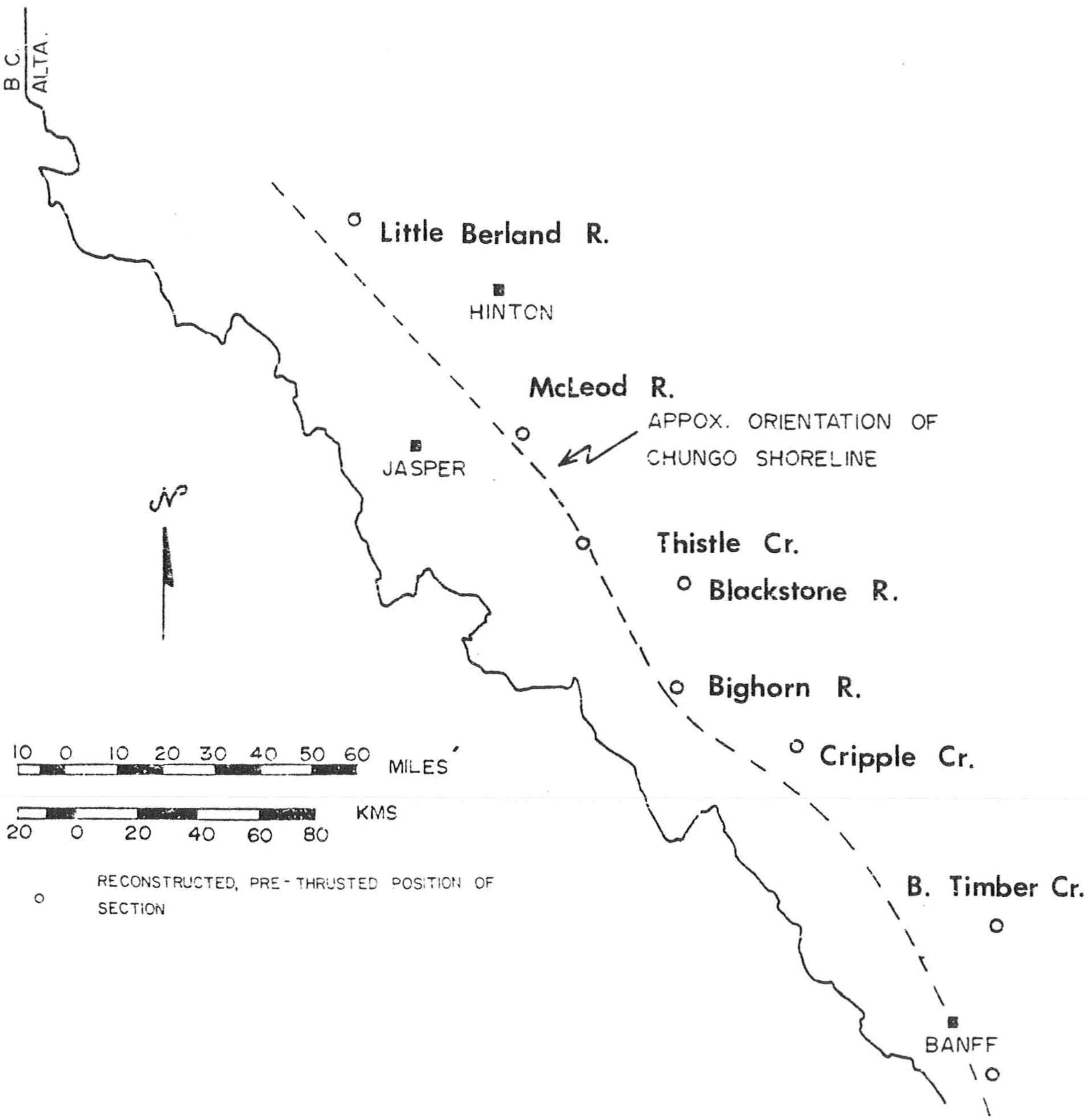
basal portion of the section (flutes and asymmetrical ripples) yielded a mean vector of 96°, suggesting that the basin slope was to the east. Trough cross-beds and rib and furrow structures had a mean vector of 125°.

Paleoflow measurements obtained from 10 Bouma division c asymmetrical ripples and one rib and furrow structure at the Little Berland River section, the northernmost section, showed a very consistent trend to the northeast. The vector mean orientation was 018°. This suggests that the paleoslope strike was oriented westnorthwest - eastsoutheast. The paleoshoreline was probably located a short distance towards the southwest.

Results of paleoflows measured from the Chungo Member tend to indicate that there was no single preferred flow orientation throughout the study area during the late Santonian. At Mt. Yamnuska, radially oriented paleoflows from trough cross-bedding with two locally preferred orientations suggests a tidal influence. The basin probably sloped to the east. At the remaining sections, no seemingly consistent orientations were noted. However, the general lack of multimodal or bimodal trough cross-bed paleoflows at these sections suggests that a significant tidal influence may have been restricted to the Mt. Yamnuska section.

To properly understand the distribution of paleoflows recorded from the Chungo Member, the mean flow directions from each measured section must be observed on a map showing the palinspastically reconstructed positions of the measured sections (Fig. 2-3). As can be seen, two sections, the Bighorn River section and the Thistle Cr.

Figure 8-8. Approximate orientation of Chungo shoreline as determined from paleoflow data and facies analysis.


B.C.
ALTA.
Little Berland R.
HINTON
McLeod R.
APPOX. ORIENTATION OF CHUNGO SHORELINE
JASPER
N
Thistle Cr.
Blackstone R.
Bighorn R.
Cripple Cr.
10 0 10 20 30 40 50 60 MILES
KMS
20 0 20 40 60 80
RECONSTRUCTED, PRE-THRUSTED POSITION OF SECTION
B. Timber Cr.
BANFF

section, are further to the southwest than the remaining sections. These two sections are also the sections which show paleoflow orientations to the southeast or southsoutheast rather than to the northeast as in the remaining sections. As well, the Thistle Cr. section was the only section to show non-marine strata, indicating that the Chungo shoreline was at or near the Thistle Cr. section during Chungo deposition. Although no non-marine units were recorded at Bighorn River, its position to the southwest may have been close to the paleoshoreline. These lines of evidence suggest that a southeasterly flowing longshore current was acting in close proximity to the Chungo shoreline. This longshore current may have also been acting at McLeod River where trough cross-beds and rib and furrow structures yielded a vector mean of 125°, or southeast. The sections which were further into the basin and out of the influence of the longshore current, Cripple Cr., Blackstone River, and Little Berland River, showed asymmetrical ripple, flute, and lineation orientations to the northeast, indicating that the basin strike was approximately northwest-southeast. This is generally consistent with basin orientations proposed to have existed during the deposition of the other Wapiabi Formation members. A proposed shoreline during the closing of the Chungo deposits is shown in Figure 8-8.

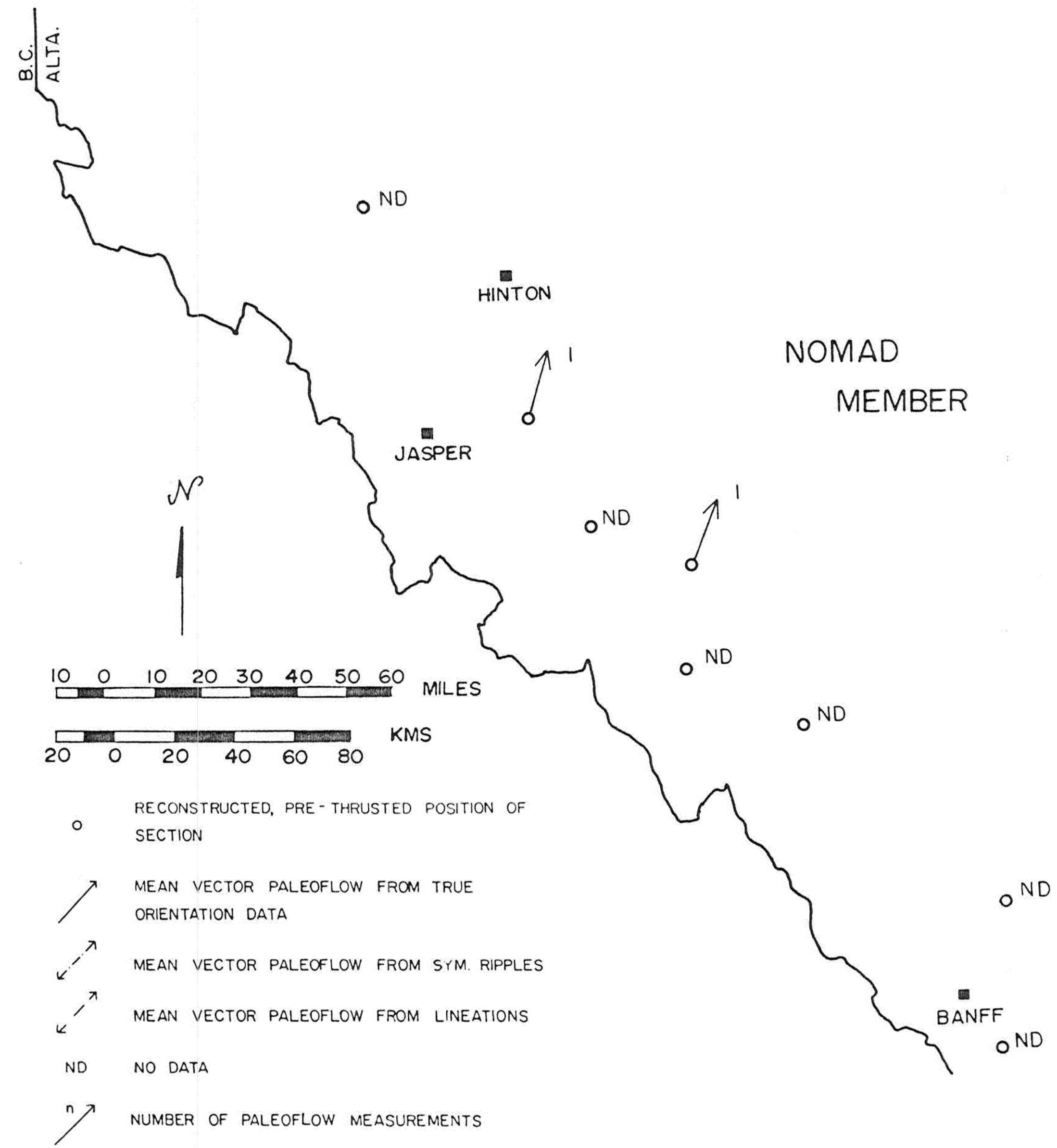
Because the Chungo Member has been interpreted as a shallow marine to beach sequence, the variations in the paleoflow measurements from trough cross-beds and rib and furrow structures from individual sections may have been due to local variations in the orientation of the paleoshoreline during deposition. These local shoreline orientations may have strongly influenced sedimentation patterns in the shallow water environment of the Chungo Member. However, once currents flowed further into the basin, more consistent regional dispersal patterns were probably set up.

Nomad Member

Although several siltstone and sandstone turbidite beds were observed within the Nomad Member, bioturbation effectively obscured most primary structures in the upper 2 to 3 cms of the turbidite beds. As a result, only two paleoflow directions were obtained from this member, one from Blackstone River and one from McLeod River. These orientations were 020° and 015°, respectively (Fig. 8-9).

Results of paleocurrent data suggest that throughout the deposition of the Wapiabi Formation, patterns of sedimentation tended not to change dramatically. An overall trend of paleocurrents to the northeast during Wapiabi time suggests a shoreline with a relatively consistent orientation of northwest-southeast. There are only two notable exceptions to this overall trend. Firstly, in the northern portion of the study area, southeasterly oriented paleocurrents measured from the Bad Heart Formation recorded the progradation of that formation from the northwest. Secondly, paleocurrents from the Thistle Member suggested a gradual change in basin strike orientation from almost east-west in the south near Burnt Timber Cr. to northwest-southeast in the central and northern areas. Interestingly, Rosenthal (in prep.) recorded paleocurrent directions of 180° in the Thistle Member at Ghost

Figure 8-9. Paleocurrent data recorded from the Nomad Member in the present study area.

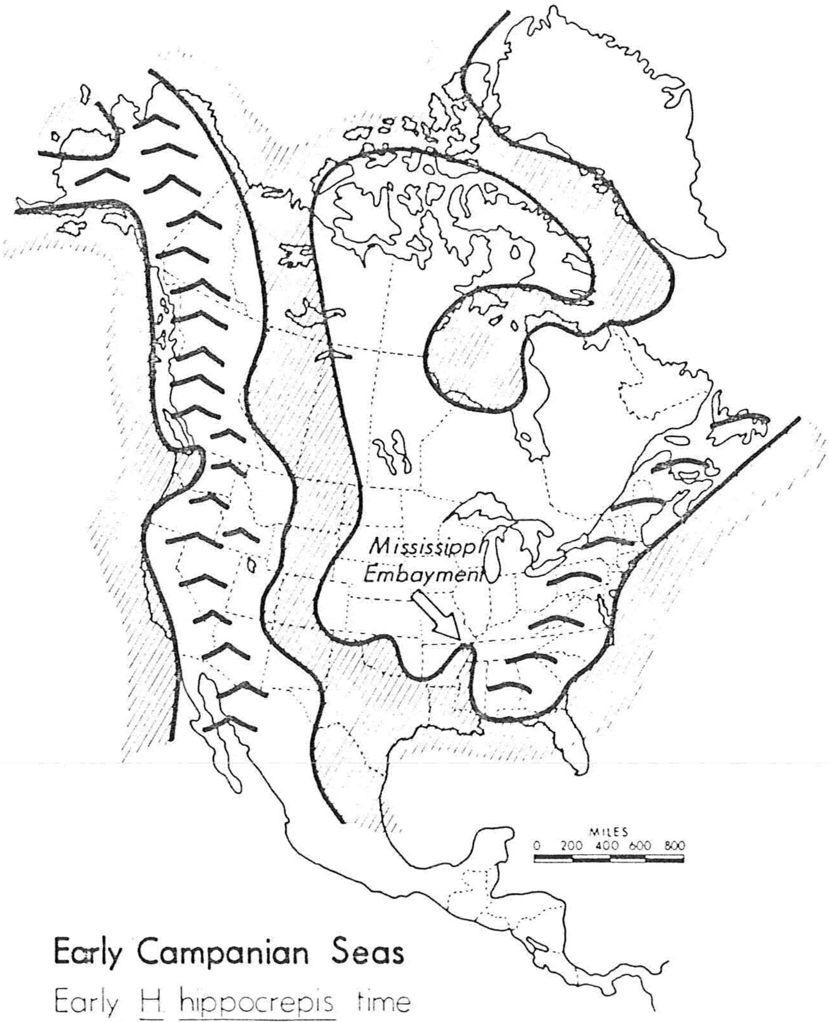

B.C.
ALTA.
NOMAD
MEMBER
ND
HINTON
1
JASPER
ND
1
ND
ND
ND
BANFF
ND
N
10 0 10 20 30 40 50 60 MILES
KMS
20 0 20 40 60 80
RECONSTRUCTED, PRE-THRUSTED POSITION OF SECTION
MEAN VECTOR PALEOFLOW FROM TRUE ORIENTATION DATA
MEAN VECTOR PALEOFLOW FROM SYM. RIPPLES
MEAN VECTOR PALEOFLOW FROM LINEATIONS
ND
NO DATA
n
NUMBER OF PALEOFLOW MEASUREMENTS

Dam, approximately 60 kms south of Burnt Timber Cr. A mean vector of 007° was recorded from Burnt Timber Cr. These results are problematic when attempting to reconstruct paleobasin geometries. However, two factors which could have influenced sediment dispersal patterns are as follows: 1) a change in provenance (to be discussed) occurred between the Ghost Dam and Burnt Timber Cr. areas, and 2) a slight submarine high between Ghost Dam and Burnt Timber Cr. could have effectively caused sediment to be diverted to the north and the south.

Although paleoflows from the Chungo Member are variable, plotting of the data onto palinspastically reconstructed section locations shows that a southeasterly flowing longshore current was acting very closely to the paleoshoreline. Proposing that such a current existed is not unreasonable, as the Cretaceous Seaway was open from the Arctic Ocean to the Gulf of Mexico during the Chungo regression (Williams and Stelck, 1975; Fig. 8-10). The overall basin geometry during the deposition of the Chungo member was, however, very similar to geometries proposed to have existed earlier in the Wapiabi Formation.

Figure 8-10. Approximate position of the Upper Cretaceous Seaway during the Early Campanian (from ~~Caldwell~~, 1975).

Williams & Stelck



Early Campanian Seas

Early H. hippocrepis time

CHAPTER 9 - PETROLOGY

The petrographic portion of this study is limited to the sandstones of the Chungo Member. It was undertaken to fulfill four basic objectives. First, to determine the general compositional makeup of the Chungo Member in the study area. Second, to note any changes in the compositional makeup within the study area and comment on the results or implications. Third, Rosenthal (in prep.) observes that the Chungo Member in the southern Foothills contains a significant amount of quartz, feldspar, and volcanic detritus. It is wanted to know if the Chungo Member of the central Foothills is compositonally similar or different. If it is different, where does the change occur? Finally, a few comments will be made on potential source areas for the Chungo Member of the central Foothills.

For the purpose of compositional analysis, 24 thin sections, cut perpendicular to bedding, were prepared and observed under a petrographic microscope. Of these, 14 were point counted using 300 point counts arranged in 6 traverses of 50 counts each. Results are shown in Table AIV-1, Appendix IV. Of the remaining ten slides, six were deemed unfit for compositional analysis due to either a high degree of alteration or a very fine grain size and four were cut from pebble conglomerate samples of facies 11. Raw data concerning the pebble conglomerates are shown in Table AIV-3, Appendix IV.

Compositionally, the sandstones from the Chungo Member in the central Foothills are classified as litharenites or sublitharenites according to the classification system of Folk (1968)(Figure 9-1). Within these sandstones there is a distinct lack of feldspar. Although rare plagioclase grains were observed during petrographic analyses, no feldspars were recorded during point counts. Staining for feldspar using the procedure outlined by Hutchinson (1974) was attempted, but proved unsuccessful.

In contrast to the strikingly low feldspar content in the Chungo sandstone, the quartz and lithic fragments content is high. Quartz grains are generally angular to subangular (Powers, 1953). However, grains commonly show well developed, euhedral quartz overgrowths (identified by distinct dust rims), suggesting that original grains may have been rounded to well rounded (Figures 9-4, 9-5). Diagenetic alteration of quartz is minimal. The lithic component is subdivided into rock fragments, chert, and detrital carbonates, of which chert makes up the greatest proportion. Rock fragments make up only a few percent of the total sandstone compositon and are generally restricted to sedimentary rock fragments. However, one volcanic glass fragment was identified from the Little Berland River section. Detrital carbonates, as well, make up only a small percentage of the total composition. These lithic fragments often show a high degree of alteration (Figures 9-6 and 9-7). In particular, chert grains are commonly altered to clay minerals (Figure 9-9). In many cases it was very difficult to distinguish chert from the clay mineral cement. However, in the interest of provenance, the bias during point counting was towards the original grain rather than towards authigenic replacements. As a result, point count results for thin sections which visually contained a

Figure 9-1. Classification of sandstones from the Chungo Member (this study). Q = quartz, F = feldspars, and RF = rock fragments plus chert. Classification system is from Folk, 1968.

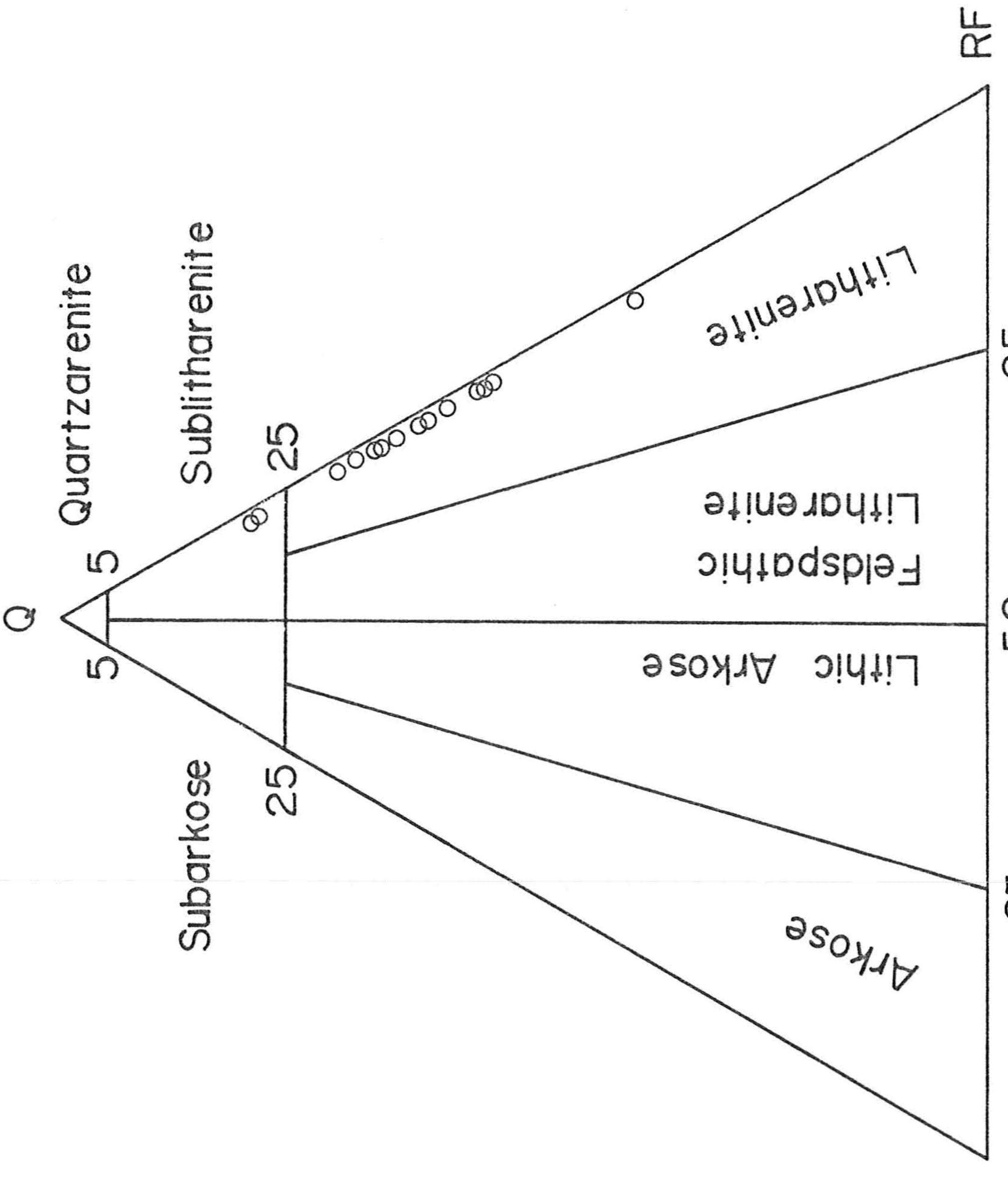
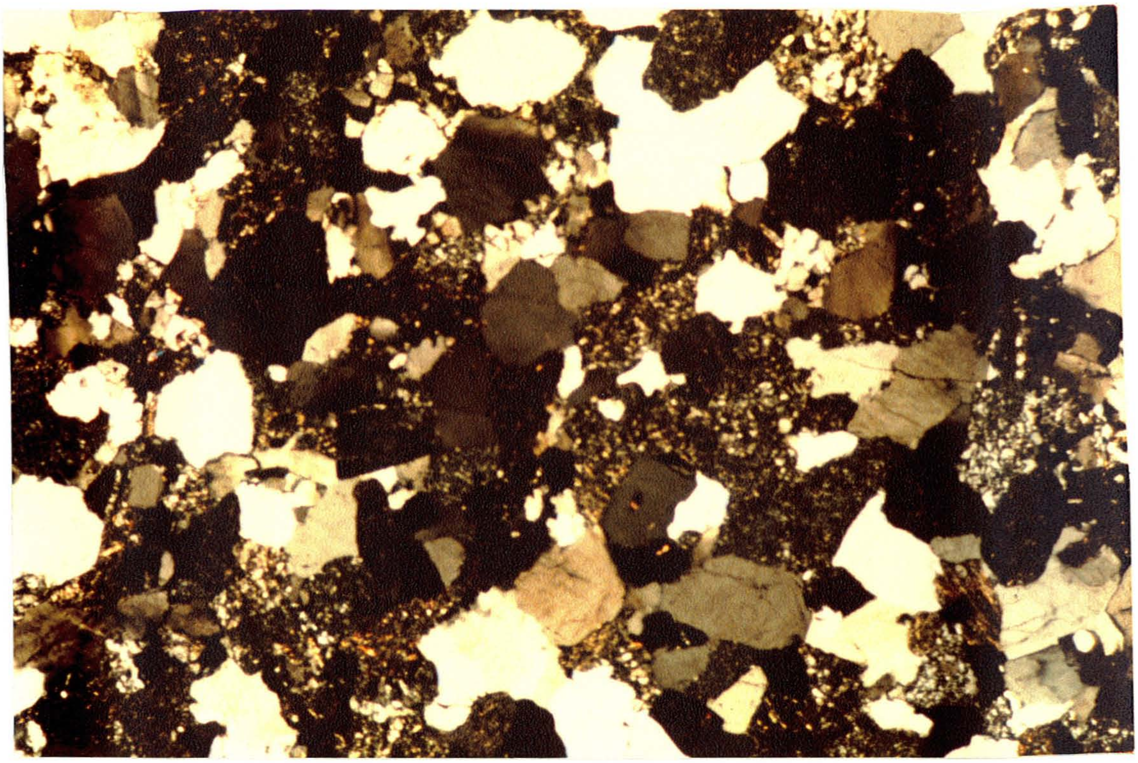

Q
RF
F
Quartzarenite
Sublitharenite
Subarkose
Litharenite
Feldspathic Litharenite
Lithic Arkose
Arkose
5
5
25
25
25
50
25

Figure 9-2. This photo shows the "typical" composition of the Chungo Member sandstones in the central Foothills. Note the complete domination of chert and quartz. Most quartz is monocrystalline. Sample MY-8, polarized light.

Figure 9-3. Calcite cement, as shown in this photo is quite common as primary cement in many Chungo samples. This particular sample has a high chert content. Note the highly altered chert just to the lower right of centre. One chalcedony grain is shown at the extreme right. Sample CC-16, polarized light.



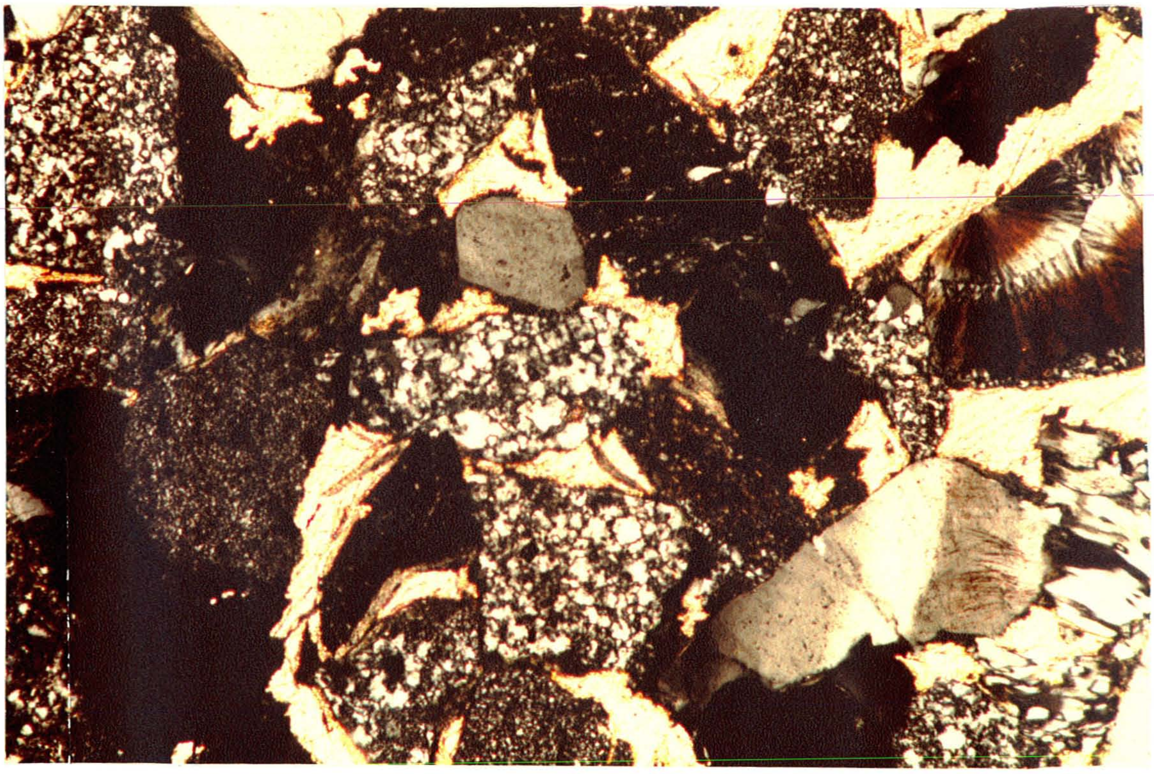
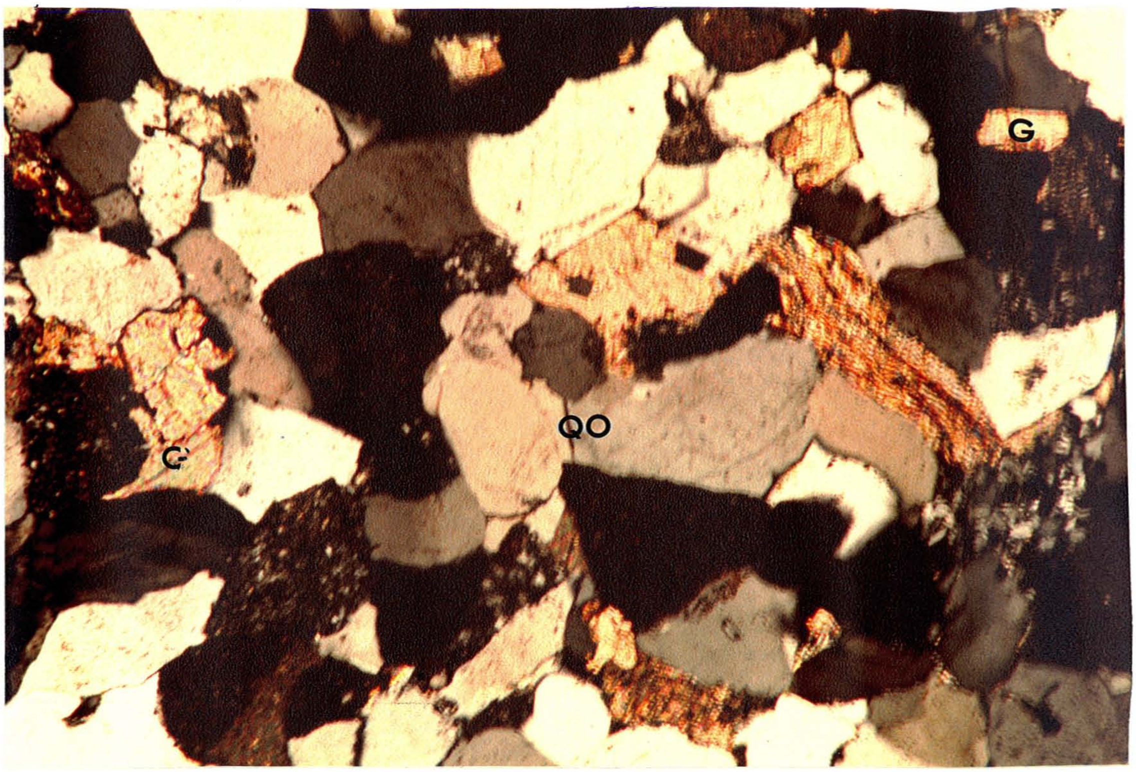


Figure 9-4. Two different types of cement can be observed in this photo. Calcite is the most obvious. However, quartz overgrowths (QO) can also be identified by dust rims in quartz grains. In this photo, calcite cement (C) and detrital calcite grains (G) are difficult to distinguish. Sample BHR-6, polarized light.

Figure 9-5. Same as previous photo with plane light. Quartz overgrowths are much more pronounced. Note clay minerals to right of center.


G
QO
G

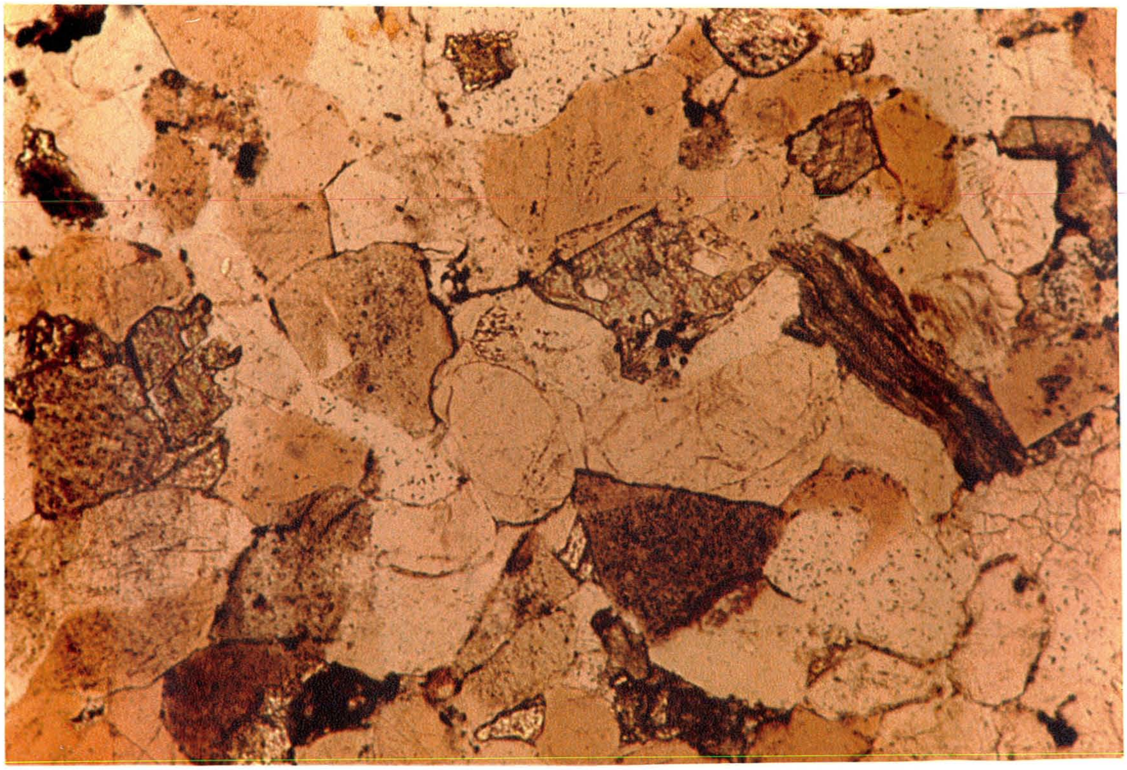
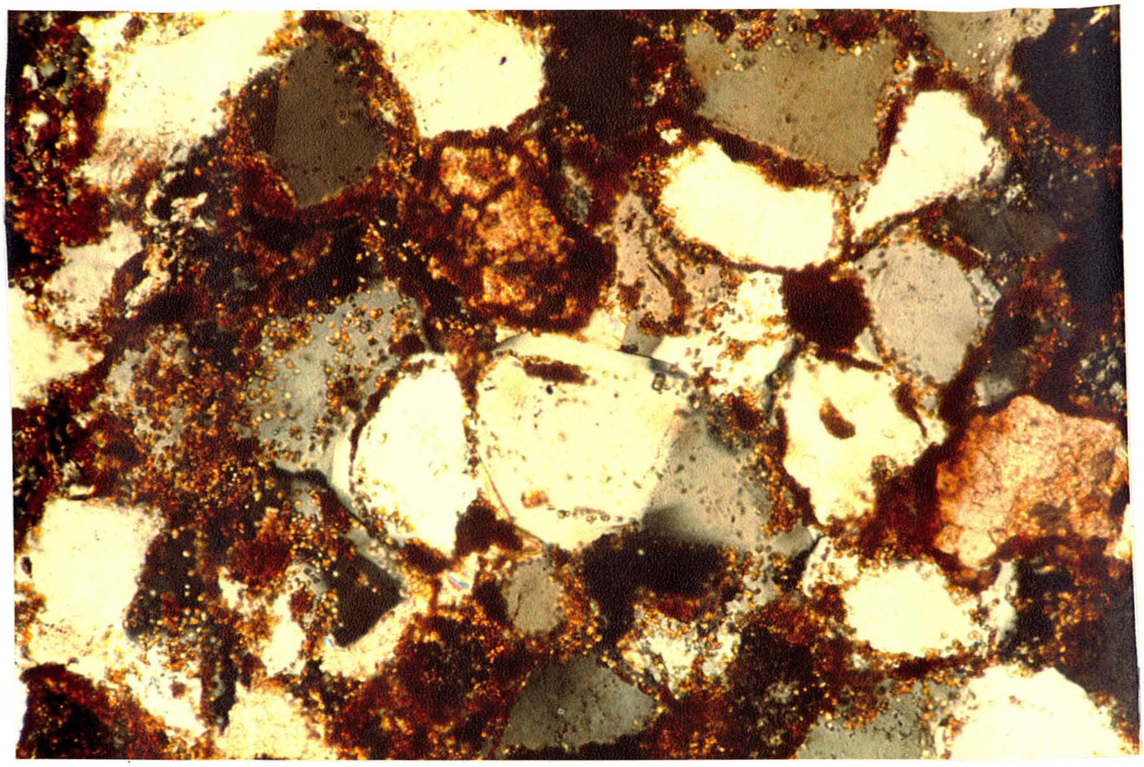


Figure 9-6. This photo shows a very high degree of cementation by siderite. Note the replacement of a previous calcite cement in the lower left portion of the photo. Detrital carbonate is also being replaced by the siderite (just above centre of photo). Sample MR-6, polarized light.

Figure 9-7. Same as previous photo in plane light. The iron-bearing cement is well displayed under plate light.

?



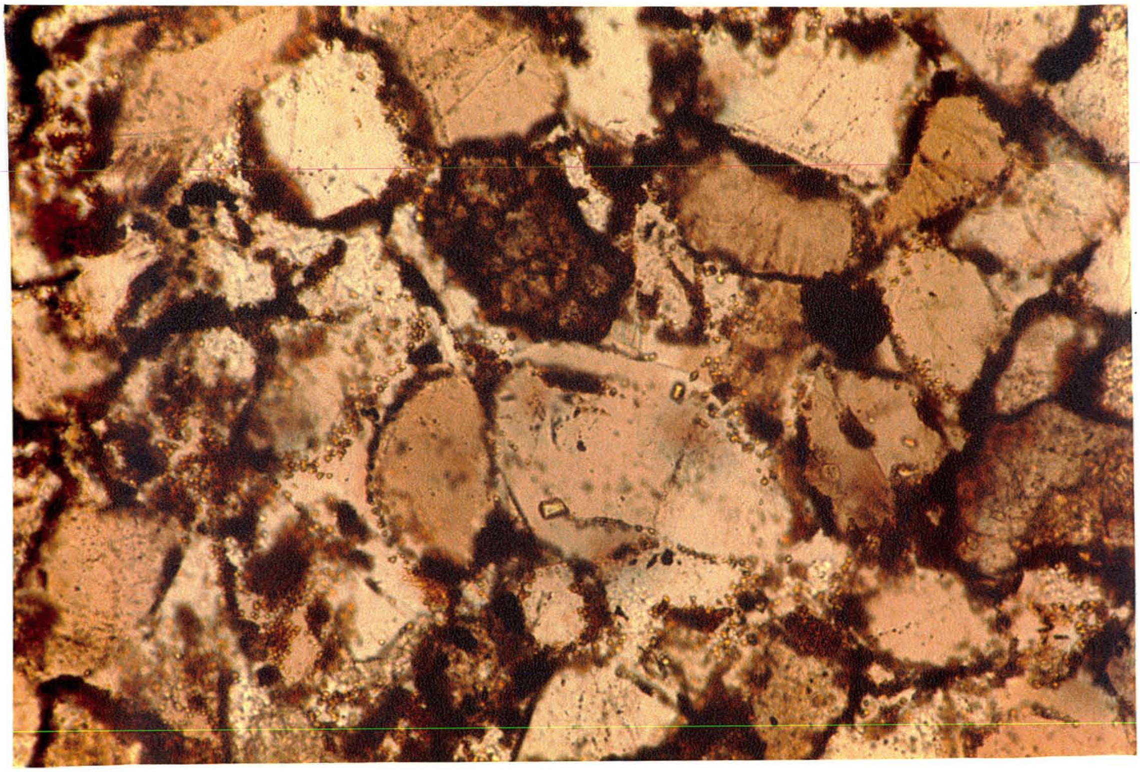
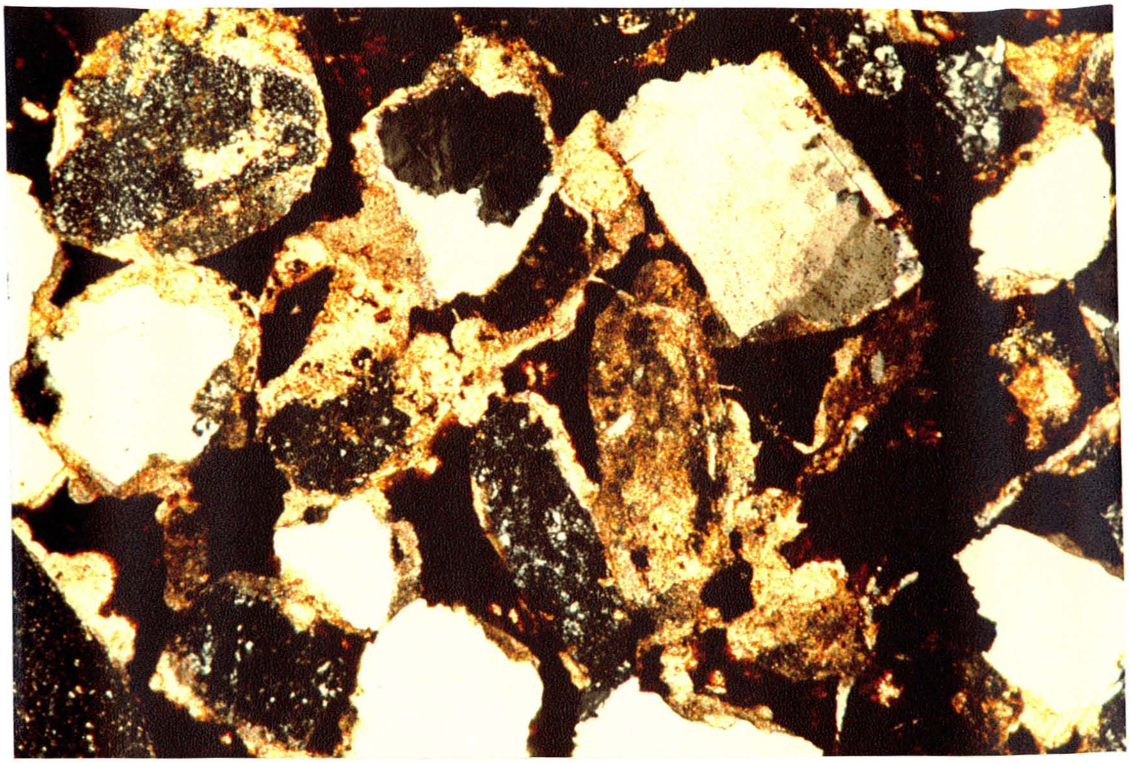
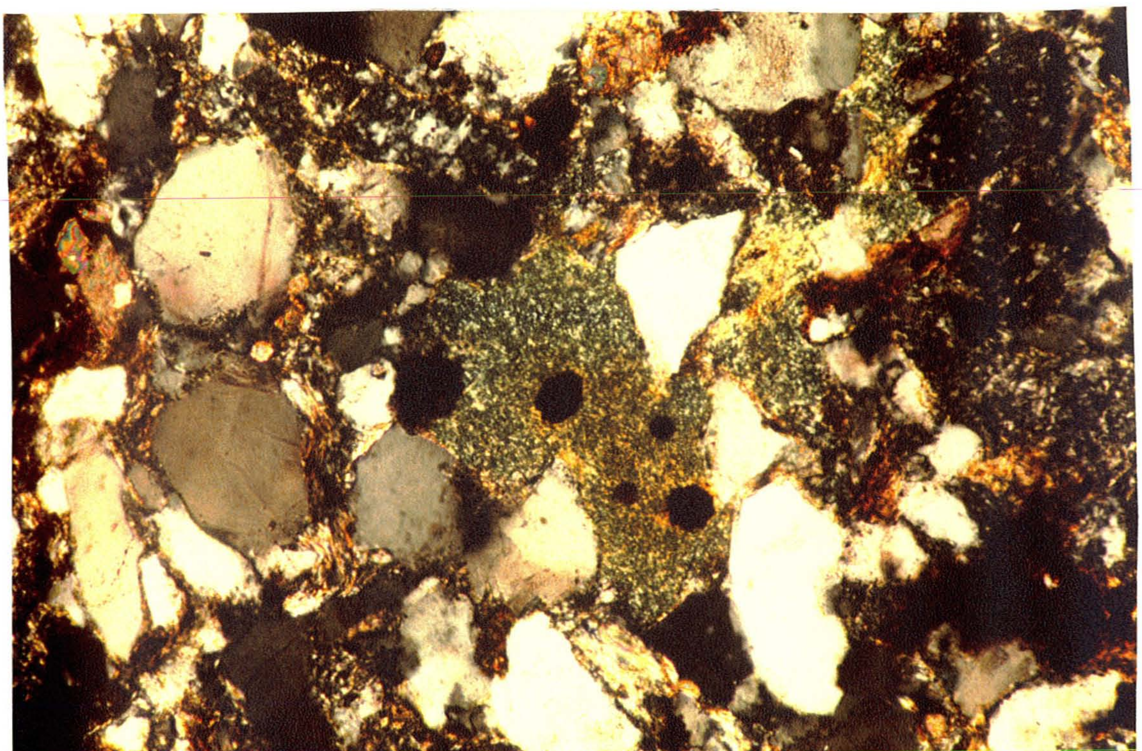


Figure 9-8. The matrix of a pebble conglomerate sample is shown in this photo. Lithology of the matrix is generally restricted to quartz and chert. However, two cements can be seen within the matrix, calcite and siderite. Calcite was the initial cement. Sample MR-5, polarized light.

Figure 9-9. The alteration of chert to clay minerals is very common in the Chungo samples. Here chert is altered to a green colour, possibly to chlorite. Note other alterations of chert. Sample LBR-9, polarized light.





high percentage of cement may suggest only a moderate amount of cement.

Using a Canstrat grain size card, the grain sizes of the Chungo samples were determined. All samples are in the range of siltstone/very fine grained sandstone to medium (lower) grained sandstone. The average is about very fine (upper) sandstone. Examples of medium grained sandstones are restricted to 6 thin section samples, of which 3 are from the matrix of pebble conglomerate beds.

Overall, the change in grain size vertically through to Chungo Member coarsening upward sequence, in all sections, is very subtle at best.

Compositional changes in the Chungo Member through the study area are very slight. As can be noted from Table AIV-1, Appendix IV, quartz and chert grains consistently make up 80 to 90 percent of the total sandstone composition. All other components make up only minor percentages. However, some very subtle changes were noted during the examination of all 24 thin sections.

From five thin sections examined from Mt. Yamnuska, a complete lack of detrital carbonate was observed. Evidence did not suggest that carbonates had been diagenetically altered. But, thin sections from six of the remaining measured sections (excluding Cripple Cr.), showed detrital carbonate percentages ranging from 1 to 7%. Although detrital carbonate was not included in point counts from Cripple Cr., several examples of detrital carbonate grains were noted during the normal petrographic examination. This implies that detrital carbonate was only being supplied to the seven northern most measured sections.

Accessory minerals also show a slight change through the study area. For the purposes of this study, accessory materials are defined as those minerals which are included under the "other" category in Table AIV-1, Appendix IV. Only two such minerals were identified, zircon and rutile. The occurrence of these minerals was restricted to the northern portion of the study area. From Bighorn River south, no accessory minerals were identified during the petrographic examination. However, from Blackstone River north, accessory minerals were identified in each thin section. Zircon was by far the dominant accessory mineral; rutile was only noted once. Although the number of accessory minerals counted during point counts was minimal (only one sample, MR-11, had 1% accessory minerals), it is still significant that there was a complete lack of such minerals in the south and several identified in the north.

Because the general makeup of the Chungo sandstone does not change from south to north, it is not suggested that completely separate source areas account for the identification of accessory minerals in the north and their absence in the south. Instead, accessory minerals in the north were probably derived from mixing of sediments from the main source with a more local source.

From the Chungo Member sandstones in the southern Foothills, Rosenthal (in prep.) reported a significant content of feldspar. Percentages of up to 29% were reported. As well, several lithic fragments of volcaniclastic origin were noted. This contrasts with the Chungo sandstones of the central Foothills where feldspars are strikingly absent and the only identified volcanic lithic fragment was from the northern most measured section. This data suggests that different source areas were supplying detritus to the area south of

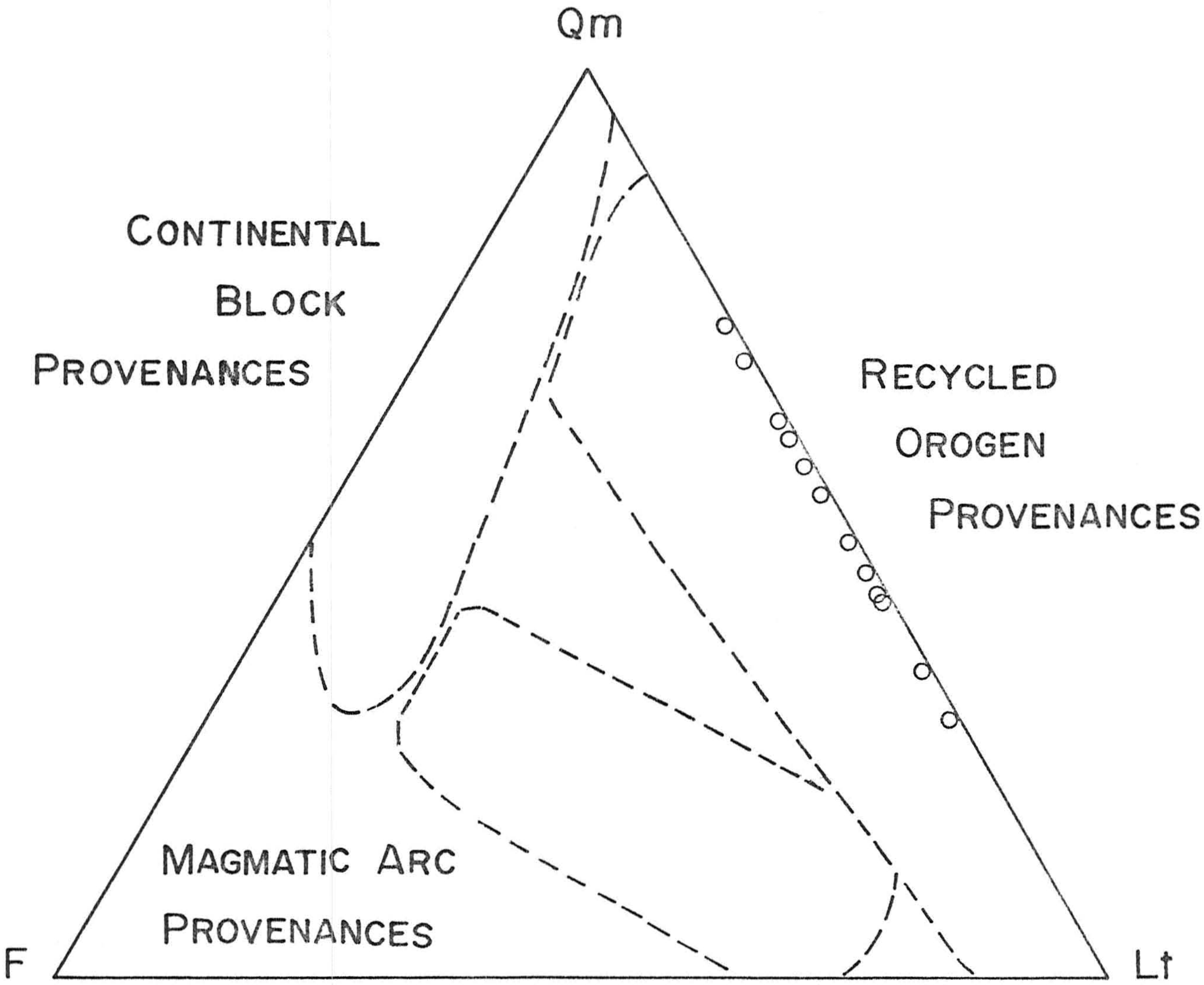
Ghost Dam (Rosenthal's northern most measured section) and the area north of Mt. Yamnuska (the most southerly section in this study).

Lerbekmo (1963) notes that the detrital feldspars from the southern Foothills have a volcanic affinity. This fits well with the results of Smedes (1966) who reports that volcanics of the Elkhorn Group in northern Montana have a characteristically high feldspar content. Rosenthal (in prep.) suggests that erosion and reworking of the volcanic source rocks to the south, mixed with a sedimentary source derived from the rising Cordillera to the west, could yield a composition similar to that of the Chungo Member in the southern Foothills.

The lack of feldspar and volcanic lithic fragments in the present study area suggests that the volcanic detritus shed from the Elkhorn Volcanic Group in Montana did not reach further north than Ghost Dam. Sediment supplied to the study area was therefore largely derived from the rising Cordillera. This is substantiated by plotting the composition of the central Foothills Chungo sandstones on a tectonic provenance QFL diagram proposed by Dickinson and Suscek (1979) (Figure 9-10). As can be noted, compositions plot firmly within the recycled orogenic province. This category includes sandstones derived from foreland basins which Dickison and Suscek (1979) suggest are characterized by very low feldspar content, moderately high quartz content, and a high content of recycled sedimentary material. The samples from the present study meet all of these conditions.

If the Chungo sandstone was indeed derived from a recycled sedimentary source, some speculation can be made as to the source beds. Rapson (1965) suggested that the most likely ultimate source of chert

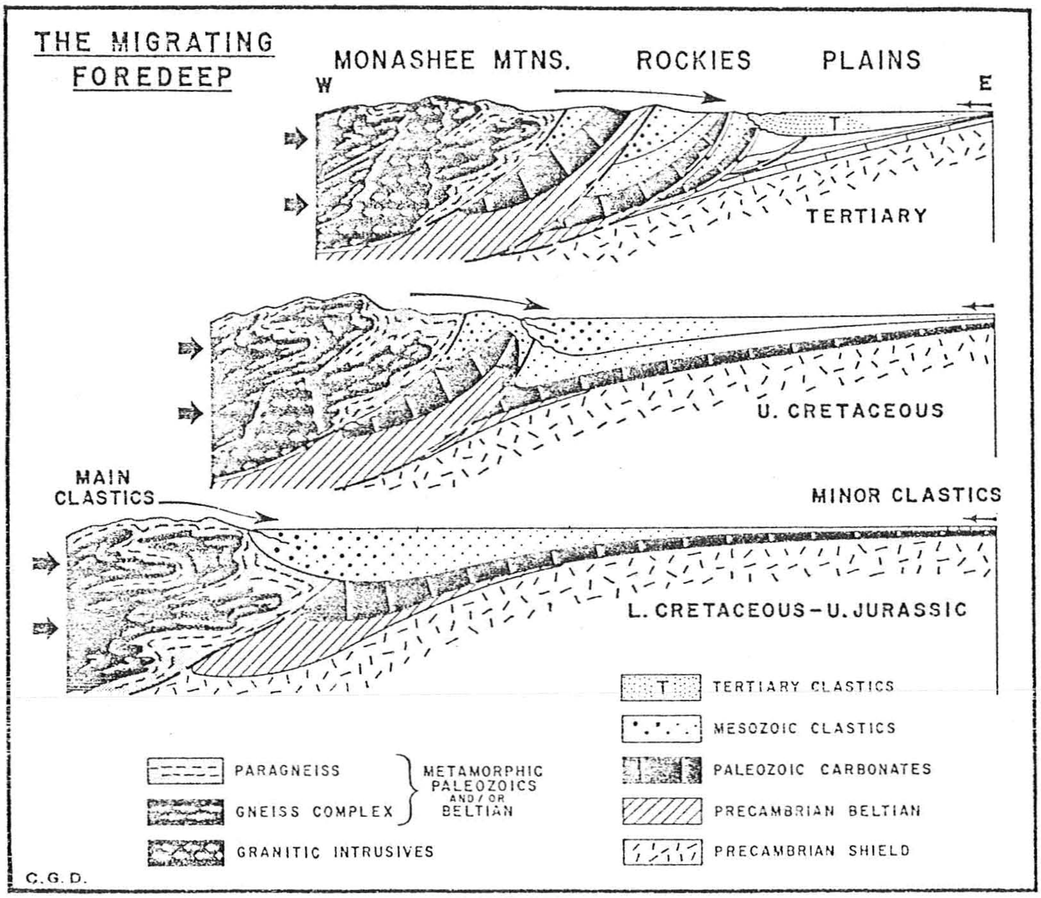
Figure 9-10. Classification of Chungo Member sandstones plotted on a tectonic province QFL diagram proposed by Dickinson and Suscek (1979). Qm = monocrystalline quartz, F = feldspar, and Lt = polycrystalline lithic fragments including chert and polycrystalline quartz.



Qm
CONTINENTAL
BLOCK
PROVENANCES
RECYCLED
OROGEN
PROVENANCES
MAGMATIC ARC
PROVENANCES
F
Lt

was from the Paleozoic carbonates. Any source from which these cherts were recycled would therefore have to be younger than late Paleozoic in age. Bally et al. (1966) have shown that at the time Upper Cretaceous beds were being deposited, foreshortening along thrusts was already occurring to the west (Figure 9-11). Uplift along these thrusts had exposed the Lower Cretaceous Kootenay and Cadomin Formations along with other formations of Lower Cretaceous age. As the Cadomin Formation in particular contains abundant chert, it is quite probable that this formation acted as a partial source for the Chungo Member. Coarse sediments from the Kootenay Formation may have also been supplied to the Chungo Member.

Figure 9-11. Migration of the Alberta foredeep basin (after Bally et al., 1966).


THE MIGRATING FOREDEEP
MONASHEE MTNS.
ROCKIES
PLAINS
W
E
T
TERTIARY
U. CRETACEOUS
MAIN CLASTICS
MINOR CLASTICS
L. CRETACEOUS - U. JURASSIC
TERTIARY CLASTICS
MESOZOIC CLASTICS
PALEOZOIC CARBONATES
PRECAMBRIAN BELTIAN
PRECAMBRIAN SHIELD
PARAGNEISS
GNEISS COMPLEX
METAMORPHIC PALEOZOICS AND / OR BELTIAN
GRANITIC INTRUSIVES
C.G.D.

CHAPTER 10 - REGIONAL INTERPRETATIONS AND DEPOSITIONAL SUMMARIES

This chapter discusses the regional relationships of each particular Wapiabi Formation member and draws on data previously presented in the text and elsewhere with the aim of developing depositional summaries for each member. As well, depositional summaries or "models" which have been previously proposed for the Wapiabi Formation and its members will be reviewed and discussed in light of the evidence gathered during this study.

Portions of the following discussion deal with the transgressive or regressive nature of the paleoshoreline. For the purposes of this study, transgression and regression will be used as defined in the A.G.I. Glossary of Geology. Transgression refers to an expansion of a sea resulting from sea level rises or land subsidence resulting in a deepening of the depositional basin. Regression refers to a gradual retreat of a sea due to sea level falls, land uplifts, or an increased sediment supply causing progradation of the shoreline and a shallowing of the depositional basin.

Muskiki, Marshybank, Dowling and Thistle Members

Specific discussions such as basin geometry, paleoflow direction, and cyclicity have previously been presented for the basal four members of the Wapiabi Formation in the study area. However, some broad generalizations and points will be discussed here. Most importantly, depositional environments and the transgressive or regressive nature of each member will be discussed.

Facies from the late Turonian to Coniacian aged Muskiki Member (facies 1 shales) suggest that the member was deposited a considerable distance from shore and below storm wave base in a quiet, deep shale basin. While the depositional environment can be reliably predicted, the transgressive or regressive nature of the Muskiki can not be assessed from evidence gathered in the study area. In all Muskiki exposures, the member overlies deep marine sandstones of the Cardium Formation's Sturrock Member (Duke, per. comm.). However, evidence from outside the study area does shed light on the problem. On Little Horn Cr. (near Lake Abraham west of Nordegg), Duke (per. comm.) has noted Muskiki shales overlying non-marine sediments of the Cardium Formation. This implies that a significant deepening of the depositional basin occurred subsequent to the deposition of the Cardium Formation rather than just a reduction in the influx of coarse clastic material. This transgressive event is also evidenced on a regional basis by a pebble conglomerate bed which separates the Cardium from the Muskiki (Stott, 1963; Duke, per. comm.). As this pebble conglomerate lies over the non-marine Cardium at Little Horn Cr. (Duke, per. comm.), it is suggested that the conglomerate is of probable transgressive origin. The mode by which the Chungo pebble conglomerate was deposited (previously discussed) may be applicable here.

From the above evidence, a suggestion by Stott (1963) which proposes that the Muskiki Member shales were deposited during the intitial transgressive stages of the Wapiabi Sea over the Cardium Formation must be accepted.

The gradational contact between the Muskiki Member and the

Santonian aged Marshybank Member (Bad Heart Formation) suggests that sedimentation was continuous between the two members. This, coupled with the facies 2 dominated nature of the Marshybank Member (interpreted as indicative of conditions below storm wave base) and the excellent preservation of facies 3 turbidites in the Bad Heart Formation (interpreted to have been deposited below storm wave base), indicates that the Marshybank-Bad Heart was deposited in a deep marine environment below storm wave base, but with a slightly increased sediment grain size over that of the Muskiki Member. However, this conflicts with an interpretation by Stott (1963) who suggested that the Marshybank-Bad Heart was deposited in "relatively shallow water". Stott's interpretation was largely based on the identification of C. pauperculum in the Bad Heart which indicates a nearshore environment. However, since the sandstones of the Bad Heart were deposited from turbidity currents, an environmental interpretation based on the identification of fauna is invalid.

From evidence in the study area alone it is not possible to predict as to whether the Marshybank and Bad Heart are of transgressive or regressive origin. However, supported by evidence from outside the study area, it is reasonable to propose that the Marshybank and Bad Heart were deposited during a marine regression. At Moberly Cr. (slightly south of Little Berland River) Stott (1963) recorded 3 m of Bad Heart Formation sandstones. At Little Berland River, 5 m of very fine grained facies 3 dominated sandstone (deposited below storm wave base) was measured in this study. Just northwest of Little Berland River on Muskeg River, Stott (1963, p. 44) reported 10 m of Bad Heart which Duke (per. comm.) noted was dominated by HCS beds deposited above storm wave base. Continuing to the northwest, Stott (1967, p. 44) reported 12 m of Bad Heart at Sheep Cr. and 20 m at Mistanusk Cr., the uppermost 10 m of which was identified as non-marine (on the basis of coal beds). The Bad Heart Formation, therefore, thickens, shoals upwards, and becomes emergent towards the northwest, a sequence which is indicative of marine regressions. This proposed progradation of the shoreline towards the southeast is also supported by paleocurrent data.

The early to middle Santonian aged Dowling Member is made up of dark shales which record deposition in a quiet, deep marine basin below storm wave base. As these shales overlie Marshybank and Bad Heart sediments deposited below storm wave base in the study area, it is not possible to determine if the Dowling Member was deposited during a transgression period or during a period of negligable sediment input. However, Stott (1967, p. 89, 94) has shown that the Dowling Member overlies non-marine Bad Heart Formation sandstones and coal to the northwest of the present study area. This indicates that the Wapiabi Sea did transgress subsequent to the deposition of the Bad Heart, and thus the Dowling shales are of transgressive origin. This interpretation is supported by a "transgressive" pebble conglomerate bed which regionally separates the Marshybank-Bad Heart from the Dowling (Stott, 1968, 1967; Duke, per. comm.) and has been identified in the study area at Little Berland River. The conglomerate is considered as transgressive as it overlies both marine and non-marine sands of the Bad Heart Formation. It is probably of similar origin to the Chungo and Muskiki pebble conglomerates.

The middle Santonian aged Thistle Member is a thick unit (up to 250 m) of non-bioturbated, deep marine shales with abundant interbedded thin Bouma b-c turbidites which were deposited below storm wave base. Within this member minor coarsening upward sequences, which end abruptly by a return to finer grained material, have been identified. These sequences possibly record periods of time which were characterized by a slight increase in sediment supply rate. However, the overall consistency of the turbidite facies throughout the entire thickness of the Thistle Member suggests that conditions in the shale depositional basin remained relatively static through Thistle time. It is impossible to predict if the Thistle was deposited during a transgression, a regression, or a long period of shoreline standstill.

Hanson and Chungo Members

Lithologic, paleontologic, and sedimentologic evidence presented earlier in this text strongly suggests that the Hanson and Chungo Members, together, comprise a single progradational and regressive coarsening upward sequence. This sequence, although relatively minor as compared to the overlying massive progradational sequence of the Belly River-Paskapoo Assemblage, may be the initial sedimentary response of tectonic activity leading to the deposition of the Belly River-Paskapoo Assemblage. Although the two members do represent a single coarsening upward sequence, they have been separated on the basis of a facies change from thoroughly bioturbated siltstones (Hanson Member) to moderately or non-bioturbated sandstones (Chungo Member).

The complete degree of bioturbation in the Hanson Member is probably the result of several factors. Some may include abundant food supply, well oxygenated waters, moderate rate of sediment supply, and a generally quiet depositional environment below storm wave base. Kulm et al. (1975) noted similar intense bioturbation in a middle shelf environment (below storm wave base) off the storm dominated coast of Oregon.

The characteristic silt size material of the Hanson Member was probably transported directly to the depositional site by turbidity currents. Some restrictions would have to be placed on this transport mode, however. Turbidite beds would have had to have been relatively thin and the time span between subsequent events must have been adequate to allow complete bioturbation. Facies evidence from the Thistle Member and the Chungo Member confirms that turbidity currents were indeed active within the depositional basin. Much of the sediment supplied to the environment below storm wave base was, therefore, probably transported by turbidity currents.

The Hanson Member exposed at the McLeod River section is somewhat anomalous when compared to other Hanson exposures. Rather than being completely bioturbated, the upper portion of the section is well laminated and dominated by turbidites of facies 1c and 3. These turbidite facies have been included in the Hanson Member rather than in the Chungo Member because of their siltstone composition. The Chungo is defined at the first sandstone beds. The lack of bioturbation was probably due to sedimentation conditions working specifically in the McLeod River area, such as a brief, locally increased sediment supply leading to a greater frequency of turbidites, a less active fauna, or both. The brief time interval is emphasized here as the Chungo Member at

McLeod River is only moderate in thickness and is well bioturbated at specific horizons, suggesting that a locally increased sediment supply did not work through the entire Hanson-Chungo progradational sequence. A greater frequency of turbidites suggesting a greater number of storm events concentrated in the McLeod area is probably unrealistic.

As progradation of the shoreline continued through Hanson time, the sea floor continued to aggrade. As storm wave base was reached bioturbation began to decline because of a higher degree of agitation at the sediment-water interface. As well, proximity to the shoreline was increasing, thus increasing the thickness and grain size of turbidite beds as well as the frequency of turbidite beds. As a result, the sandstones of the Chungo Member are characteristically non-bioturbated or only moderately bioturbated. Any bioturbation noted in the Chungo Member was, again, probably due to local sedimentation conditions.

Evidence from lithologic and facies sequences observed in the study area suggests that the seven northern most Chungo sections were deposited under somewhat similar storm influenced conditions. This is in contrast with the Mt. Yamnuska section where both tidal and storm influenced deposits have been observed. For these reasons, separate summaries have been generated to explain the differences. The conclusions for the seven northern most sections are discussed first.

The progradational sequence of the Hanson Member in each of the seven northern sections generally shows a consistent coarsening upward sequence of bioturbated shales of facies 2 grading up into turbidites of facies 1c or 3. A similarly consistent coarsening upward sequence is observed in the Chungo Member. However, the "completeness" of each

Chungo sequence is variable from section to section. One section, the Thistle Creek section, however, does show a "complete" progradational sequence from bioturbated shales and siltstones deposited below storm wave base up into non-marine coals.

The initial deposits of the Hanson-Chungo progradational sequence are those of facies 2. The conditions under which these sediments were deposited and became bioturbated have already been discussed. As progradation of the shoreline commenced, the size of the material supplied to the depositional basin increased. This resulted in the observed upsection increase in the siltstone-shale ratio from approximately 1:1 at the top of the Thistle to approximately 2:1 or greater in the Hanson. As bioturbation became less effective towards the top of the Hanson Member and events which supplied detritus became relatively more frequent, the preservation potential of turbidite beds became greater. As a result, turbidites of facies 1c and 3 began to be preserved in the upper portion of the Hanson Member. The thickness and grain size of beds continued to increase upsection. Eventually, the continued sediment input resulted in aggradation of the sea floor to storm wave base. At this point, storm effects began to rework turbidite deposits, resulting in HCS.

The base of the Chungo Member has generally been drawn at the base of a series of very fine-grained HCS beds. As previously interpreted, these beds indicate that deposition resulted from turbidity currents which were being affected by high energy storm events. Sand, suspended in the nearshore environment by storm surges and wave action, was transported offshore by turbidity currents and deposited as a sheet-like veneer below fairweather wave base. Reworking during deposition and subsequent to deposition by storm wave action "feeling the bottom" resulted in HCS.

A continued sediment supply, resulting in further progradation of the shoreline and aggradation of the sea floor, effectively enhanced the relative effect of each subsequent storm event. As proximity to the shoreline increased, shales deposited during quiet periods between storm events became thinner. This resulted from wave conditions which keep mud suspended and more effective erosion of shales during storm events. Storm deposits, therefore, gradually become "amalgamated" into SCS of facies 6 in an upsection direction.

The preservation of very fine to fine-grained SCS continued until the sea floor had aggraded to the point where fairweather influences began to affect the bottom sediments. This "fairweather wave base" has been estimated at approximately 10 meters depth for the present Atlantic shelf (Dietz and Menard, 1951; Dietz, 1963). The evidence for fairweather influence is the intermixing of storm generated SCS beds and TXB that has been interpreted as being of fairweather origin (facies 7). The TXB was possibly produced by longshore currents that have been shown to have been working in the nearshore environment during Chungo deposition. At one section, the Bighorn River section, the thickness of the facies 7 unit suggests that a delicate balance existed between sediment supply, basin subsidence, storm events, and fairweather conditions.

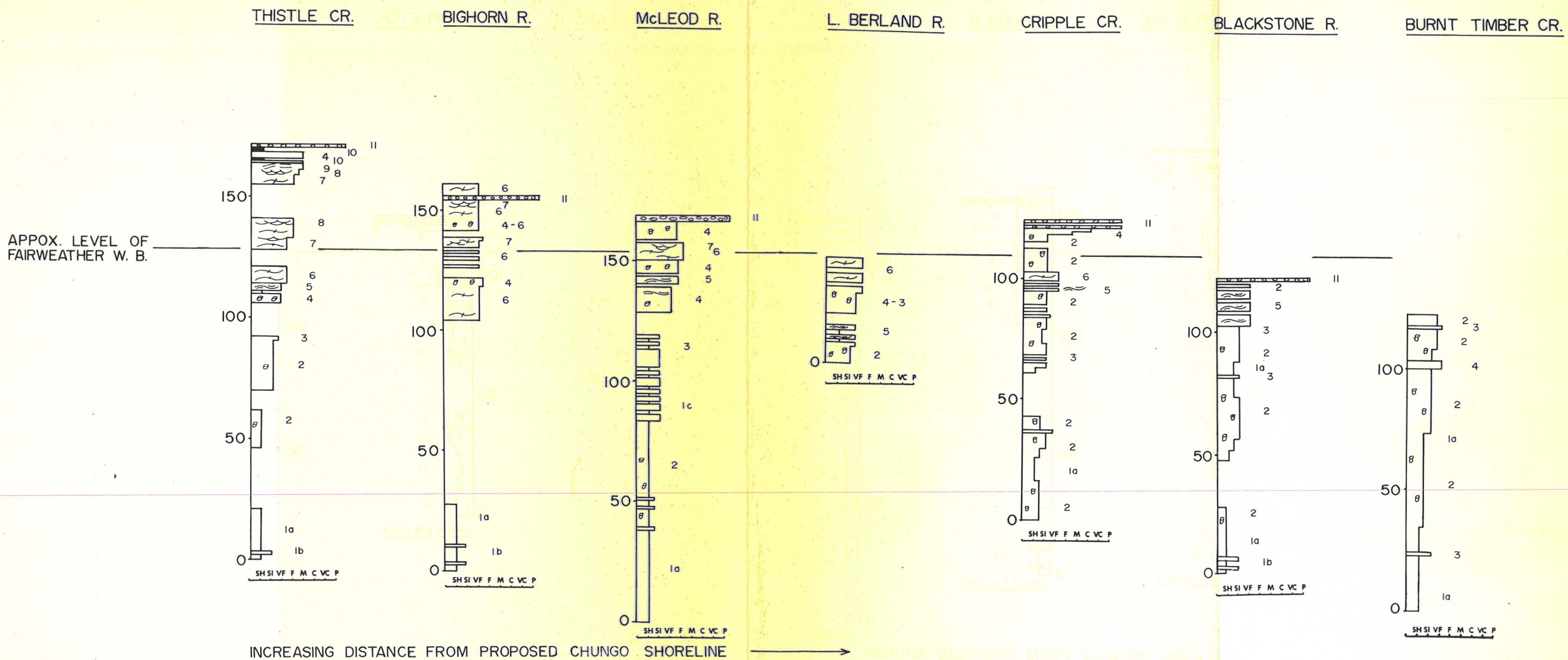
As the shoreline continued to prograde and the sea floor continued to aggrade, the relative effects of fairweather long shore currents began to dominate over storms. As a result, facies 8 TXB was deposited over facies 7. The only recorded example of facies 8 is 2 meters thick and is from the Thistle Creek section.

Directly overlying the facies 8 unit is the parallel laminated unit of facies 9 which has been interpreted as a beach deposit. Coal deposits were found directly above the beach.

The sequences of facies that have been observed in the seven northern examples of the Chungo Member strongly suggest that the Chungo can be interpreted in terms of a progradational, storm-dominated shoreline. Storms particularly affected sedimentation in the middle and lower shoreface environments, while fairweather longshore currents and storms affected the nearshore environment.

While the 6 remaining northern Chungo sections (other than the Thistle Creek section) do not show a "complete progradational sequence" of bioturbated siltstone up through non-marine coals, they do show the same order of facies; always beginning with bioturbated siltstone and terminating upsection somewhere within the "complete progradational sequence". On this basis, Figure 10-1 was constructed which shows the "complete" sequence on the left and the least "complete" sequence on the right, with the remaining sections in the middle placed according to their relative "completeness". For instance, the Burnt Timber Creek section shows only bioturbated siltstone and is therefore placed to the extreme right of the diagram. Blackstone River shows both bioturbated siltstone and HCS. It is therefore placed to the left of Burnt Timber Creek. Datum for the diagram has been chosen at the approximate level of fairweather wave base.

Figure 10-1. The seven northern Chungo sections shown in order of increasing distance from the paleoshoreline. Offshore is to the right of the diagram.


THISTLE CR.
BIGHORN R.
McLEOD R.
L. BERLAND R.
CRIPPLE CR.
BLACKSTONE R.
BURNT TIMBER CR.
APPOX. LEVEL OF FAIRWEATHER W. B.
SH SI VF F M C VC P
INCREASING DISTANCE FROM PROPOSED CHUNGO SHORELINE

In order to relate the facies within sections shown in Figure 10-1 to their relative distance from the paleoshoreline, an estimate of the relative distance from the paleoshoreline for each section was determined from the palinspastic map (Figure 2-3) and the proposed Chungo shoreline map (Figure 8-8).

The distance from paleoshoreline was determined to be exactly relatable to the facies, with the Thistle Creek section being essentially at the paleoshoreline and the Burnt Timber Creek section being the furthest offshore. The Bighorn River and McLeod River sections, which show facies sequences up through facies 7 ($\cong$ 10 m depth) can probably be placed within a few kilometers of the paleoshoreline. The only other reliable estimate of distance from shore is for the Blackstone River section, estimated at 20 km. This was possible because the Blackstone River section lies between Thistle Creek and Bighorn River for which reliable paleoshorelines and palinspastic reconstructions have been determined (Fig. 8-8). The estimate of 20 km from shore suggests that facies 5 HCS (the uppermost facies at Blackstone River) was deposited a minimum distance of 20 km offshore. A maximum distance offshore for HCS in the Chungo Member is difficult to suggest as a reliable distance from shore for the Burnt Timber Creek section (where no HCS was present) has not been determined. However, it is probably within 40 km.

The relations in Figure 10-1 show how the Chungo Member thins out and terminates in the offshore direction. Note that there is a consistent thinning of the sandstone from a maximum thickness of approximately 58 m at Thistle Creek (at the paleoshoreline) to virtually nil at Burnt Timber Creek (not greater than 40 km offshore). Likewise, sand/shale ratios consistently decrease in the offshore direction with Thistle Cr. yielding approximately 80-90% sandstone, Bighorn River approximately 80% sandstone, McLeod River approximately 80% sandstone, Little Berland River approximately 60-75% sandstone, Cripple Cr. approximately 50% sandstone, Blackstone River approximately 50% sandstone, and Burnt Timber Cr. approximately 5% sandstone. The above relationships shed a bit more light on interpretations by Stott (1967) who simply noted that the Chungo Member thins in the Burnt Timber Cr. and Ghost Dam (Rosenthal, in prep.) areas. This author feels that while the Chungo Member certainly thins in these areas, the thinning may be a function of the distance from the paleoshoreline. A thicker Chungo Member may be present further to the west. However, a precautionary note must be made about this suggestion. A major provenance change occurred in the area outlined by Mt. Yamnuska, Burnt Timber Cr. (this study), Barrier Lake and Ghost Dam (Rosenthal, in prep.). The effect of this change on sedimentation in the area is not well understood. It is possible that the Burnt Timber Cr. area received only a moderate sediment supply rather than being a long distance from the paleoshoreline.

As described in previous sections, the Chungo Member at Mt. Yamnuska differs significantly from the remaining sections in that it shows a strong tidal influence. As well, it is the only section which shows evidence of two coarsening upward cycles. For these reasons, the conclusions proposed in the previous discussion can not be directly applied to the Mt. Yamnuska exposure. Instead, two summaries, independent of the northern measured sections, have been generated. After the presentation of the two summaries, the "models" as proposed by Lerand (1982) for the Mt. Yamnuska section will be reviewed.

The first summary of deposition for the Chungo Member at Mt. Yamnuaka is described as a storm and tidally influenced prograding shoreline model. The second summary is termed the protected bay model. It is felt that this second summary more readily explains some of the Mt. Yamnuska characteristics and is therefore slightly favoured.

The basal beds of the Chungo Member at Mt. Yamnuska, which include facies 5 HCS and facies 6 SCS, indicate that deposition from turbidity currents was affected by high energy storm events. Methods by which the HCS of unit 1 and the SCS of unit 2 were deposited are the same as those proposed in the previous Chungo Model.

As aggradation of unit 2 continued towards fairweather wave base, fairweather conditions began to influence the dominantly storm influenced basin. This is evidenced in unit 3 where trough cross-bedding of probable tidal origin (see facies 7 interpretation) is found to be interbedded with SCS.

In the first Chungo summary, continued progradation of the shoreface environment resulted in the eventual domination of tidal effects over storm effects. The massive tidally generated TXB of unit 4 was the result. However, in order for tidal effects to completely overshadow storm effects in a region which is considered to be highly storm influenced (refer to previous section), the tidal range must have been quite significant. It is suggested that such a large tidal influence working in a region of simple shoreline progradation should be recognizable on a regional scale. But, evidence from the seven northern sections does not support an interpretation of tidal influence on a large scale. Interpretations by Rosenthal (in prep.) for the Chungo Member in the south also do not support a tidal influence.

In the second Chungo summary, the trough cross-bedding of unit 4 is again considered as tidal in origin. However, rather than being deposited in an open shoreface environment, an environment somewhat protected from storm effects is proposed. Such an environment may be a protected bay or an outer estuary. The accentuation of the tidal effects at Mt. Yamnuska could therefore be explained not only by suggesting an increased tidal range due to amplified tidal effects in a bay, but also by suggesting that tidal effects were relatively increased by the reduction of storm effects. Tidal currents acting within a somewhat protected environment could also explain the radial distribution of paleoflows measured in unit 4 by Lerand (1982). The two 160° bimodally oriented flows, also observed by Lerand (1982), probably indicate a domination by either flood or ebb currents at a specific point.

The very sharp contact between the medium-grained trough cross-bedded sandstones of unit 4 and the fine-grained SCS sandstone of unit 5 may be indicative of a minor transgressive event, a change in the rate of sediment supply, a change in the geometry of the protected environment, or a combination of the aforementioned factors. While a deepening of the basin can not be proven to have occurred during this event, it should be pointed out that SCS has always been observed at a stratigraphic level below trough cross-bedding in this study. It is felt that the event can most readily be explained by a locally reduced sedimentation rate coupled with continued basin subsidence.

The proposed slight basin deepening resulted in a re-exposure of the Mt. Yamnuska section to storm conditions from its protected environment. As a result, a second unit of storm generated SCS was deposited, beginning the second coarsening upward progradational sequence. With continued aggradation, the sea floor again reached fairweather wave base, resulting in deposition of the trough cross-bedding of unit 6. Because of a lack of measured paleoflow directions from the unit and the 1 m thickness of unit 6, the origin of the TXB is difficult to assess. Whether it was generated by tidal, storm, or normal marine conditions is not known. Rare pebbles in this unit suggest that it was deposited in close proximity to a source of pebbles (probably a non-marine floodplain).

As previously interpreted, the three pebble conglomerate beds lying above unit 6 were deposited from high energy events during the initial transgressive phase of the Nomad Sea. The trough cross-bedding interbedded with the pebble conglomerates was probably deposited during waning stages of each conglomerate event. This is evidenced by the gradational nature of the trough cross-bedding which overlies a specific conglomerate. On one occasion, oscillation ripples overlie the trough cross-bedding, indicating reworking by wave action.

Two depositional summaries for the Mt. Yamnuska section were presented by Lerand (1982), a beach "model" and an offshore bank "model". It should be mentioned that Lerand's beach model is somewhat misleading as no beach was identified. The major arguments against either of Lerand's models concern the trough cross-beds of unit 4.

In the beach model, Lerand (1982) proposes that the Chungo Member reflects the simple progradation of a wave dominated shoreface environment. Sediment was supplied from the northwest by longshore currents and driven offshore by storm surges and rip currents. It is felt that his interpretations for the basal SCS ad HCS facies are generally sound. However, he suggests that the unit 4 trough cross-bedding "records longshore currents generated by dominantly northerly onshore winds...". Even if wind directions were occasionally reversed, this model does not sufficiently explain the distribution of paleoflows recorded from unit 4. It has been concluded in this study that only a tidal influence can adequately explain the observed paleoflows.

In Lerand's (1982) second "model", the offshore bank model, the unit 4 trough cross-bedding is interpreted as tidal in origin. He suggests that the succession of units 1, 2, 3 and 4 represents a shoaling of a linear sand bank analogous to those described in the North Sea. In this model, it is not clear as to whether or not the basal HCS and SCS are considered as a portion of the tidal banks. A major piece of evidence against this bank model actually does not come from the Mt. Yamnuska section. If conditions similar to those in the North Sea existed in the Cretaceous Seaway, it is reasonable to suggest that evidence of tidal domination should be found on a regional basis. Such evidence has not been found in this study or in other studies, even for sections in close proximity to Mt. Yamnuska (eg. Barrier Lake; Rosenthal, in prep.). It is therefore concluded that a bank model for Mt. Yamnuska is not consistent with regional facies relationships.

Strong evidence for a tidal interpretation of unit 4 means that any overall interpretation of Mt. Yamnuska must explain both tidal and storm domination in a region which is dominated by storm effects. It is felt that the protected environment proposal most satisfactorily explains the local accentuation of tidal effects at the expense of storm effects.

Nomad Member

The Nomad Member sediments in the study area were deposited in deep basin to nearshore environments during a brief readvancement of the Wapiabi Sea in early Campanian times. As previously interpreted, the pebble conglomerate at the base of thr Nomad Member marks the initial phase of the transgressive event. It is therefore concluded that at least the basal portion of the member is transgressive.

Although evidence suggests that the lowermost beds of the Nomad Member are transgressive, the upper beds may in fact be progradational. The most supportive evidence for such an interpretation comes from the McLeod River exposure where the upper Nomad beds show a well defined coarsening and thickening upward sequence to the base of the Brazeau Formation. A similar sequence was reported by Reich (1983) from the transition of the Nomad Member into the Belly River Formation at Burnt Timber Creek. As well, the poorly exposed upper Nomad Member at Thistle Creek shows a coarsening upward from very fine-grained SCS into medium-grained TXB. Evidence, therefore, favours an interpretation of a marine regression for the upper Nomad beds.

In general, the regressive sequence of the upper Nomad Member, marking the transition from marine shales into non-marine sediments of the Belly River Formation, is stratigraphically incomplete, thin, and poorly defined as compared to the Chungo regressional sequence. The reason for this can probably be related to a very fast rate at which the Belly River Formation prograded from the west. Because of the quick progradation rate, the basin subsidence rate was initially thrown out of equilibrium, becoming too slow (relatively) to accommodate the first massive influx of coarse detritus. As a result, the sedimentary record of the first stages of the regression is quite thin and incomplete. As detritus continued to be supplied during Belly River time, basin subsidence rates probably increased to accommodate the clastic material.

CHAPTER 11 - CONCLUSIONS

Muskiki, Marshybank, Dowling and Thistle Member

The following conclusions have been reached for sedimentation in the Muskiki, Marshybank (Bad Heart Formation), Dowling and Thistle Members of the Wapiabi Formation:

1) The late Turonian and Coniacian aged Muskiki Member is interpreted to have been deposited in a deep marine shale basin formed after and/or during the intial inundation of the Wapiabi Sea. Limited paleocurrent data suggest that the depositional basin deepened to the northeast with a slope striking northwest-southeast.

2) The bioturbated siltstones of the Marshybank Member are interpreted as having been deposited in an environment below storm wave base, but above the shales of the Muskiki Member. The environmental interpretations of a deep shale basin for the Muskiki Member and a slightly below storm wave base environment for the Marshybank Member are substantiated by other evidence from Stott (1967) which shows that a shoaling of the depositional basin occured between the deposition of the Muskiki and Marshybank Members. The total degree of bioturbation in the Marshybank Member suggests that sedimentation rates must have been relatively moderate. Because of the total bioturbation, no paleoflows were recorded from this member.

3) The Bad Heart Formation (Marshybank equivalent) at Little Berland River is interpreted to have been deposited in an environment below, but within close proximity to storm wave base. Evidence from

outside sources coupled with evidence from the study area suggests that the Bad Heart is a progradational sandstone body (similar to the Chungo Member) which thickens and becomes more proximal to the northwest. Facies evidence, along with paleocurrent data (which show turbidity currents flowing downslope to the southeast), suggests the existence of a shoreline some distance to the northwest of the study area boundary.

4) A transgressive period at the close of Marshybank times (marked by a "transgressive pebble unit") caused a return to shale deposition within a quiet, deep basin environment well below storm wave base. The Dowling Member has been interpreted to have been deposited in a deep basin during times of little coarse clastic input.

5) The Thistle Member was deposited in a deep marine shale basin environment well below storm wave base. The thickness of the member, coupled with the consistency of the facies throughout the member, suggests that the sedimentation rate and basin subsidence may have been in relative equilibrium during deposition of the Thistle Member. Paleocurrents recorded from the Thistle indicate that in the southern portion of the study area flow was directed to the northwest and north. Further north the flows show more consistent directions to the northeast. This suggests a change in the orientation of the depositional basin strike from east-west in the south to northwest-southeast in the central and northern parts of the study area.

6) All four of the basal Wapiabi Formation members can be recognized on a regional scale and have been identified througout the study area. Except for some difficulties in placing the base of the Marshybank Member, most member boundary correlations were relatively

easy. However, correlations between sections of beds within members are somewhat speculative and for the purposes of this study have been considered as unreliable. More control between sections would be required to verify such correlations.

7) The long period of time represented by the shales and siltstones of the Muskiki, Marshybank, Dowling, and Thistle Members suggests that a long period of orogenic quiescence occurred during deposition of these units. Sedimentary evidence indicates that little uplift to the west occured until the Hanson-Chungo progradational sequence commenced in the late Santonian.

Hanson and Chungo Members

Sedimentological evidence derived from the Hanson-Chungo progradational sequence can be summarized as follows:

1) The Hanson Member sediments record the initial stages of a major regression of the Wapiabi Sea at the end of Santonian times. The sediments coarsen upwards from black, sideritic, and generally bioturbated silty shales into thin bedded (normally less than 10 cm) siltstone and sandstone Bouma b and bc turbidites interbedded with bioturbated shales. These sediments have been interpreted to have been deposited in an offshore environment below storm wave base.

2) The Chungo Member in the seven northern studied sections has been interpreted in terms of a storm dominated, progradational, shallow marine environment. These Chungo sections show a relatively consistent depositional sequence as follows (from stratigraphic bottom to top): hummocky cross-stratification, swaley cross-stratification, interbedded swaley cross-stratification and trough cross-bedding, trough

cross-bedding, beach lamination, and coal beds. These identified primary structures and beds suggest that the basal Chungo sediments (HCS and SCS) were strongly affected by storm influence. The interbedding of SCS and TXB in the middle of the sequence is suggestive of both storm and normal marine influences during sedimentation while the upper TXB is proposed to have been developed strictly under fairweather conditions. Paleocurrents recorded from the upper TXB unit indicate that a relatively strong longshore current was acting in the nearshore environment. Flow was from the northwest, towards the southeast. Paleocurrents from the lower Chungo beds verify that the Chungo shoreline had an orientation of northwest-southeast.

3) The stratigraphic level at which SCS and tidally or fairweather produced TXB become interbedded may possibly be used as an approximation of the maximum depth of tidal or fairweather processes, respectively.

4) Palinspastic reconstructions for and facies evidence from the measured sections suggest that the sections can be placed in an order of increasing distance from the furthest progradation of the Chungo shoreline. The Thistle Cr. section was deposited essentially at the paleoshoreline. The Bighorn River and McLeod River sections were probably deposited within a few kms of shore. The Blackstone River section was estimated to have been deposited approximately 20 kms from shore while the Burnt Timber Cr. section was probably deposited no greater than 40 kms from shore. With increasing distance from the paleoshoreline the sections show a consistent decrease in the sand/shale ratio (from approximately 9:1 at Thistle Cr. to approximately 1:20 at

Burnt Timber Cr.) and a general thinning of the Chungo Member from a 58 m maximum on Thistle Cr. to virtually nil at Burnt Timber Cr.

5) The Chungo Member at Mount Yamnuska is best interpreted in terms of a storm dominated shelf which was prograded over by a protected environment such as an outer eustuary or a protected bay. Within the protected environment tidal effects were relatively increased at the expense of the diminished storm effects. As a result, storm domination was recorded in the basal beds while tidal domination was recorded in the uppermost beds. This tidal domination is suggested to have influenced local sedimentation only as no evidence of tidal influence at other sections was found.

6) The pebble conglomerate beds on top of the Chungo Member mark the initial inundative phase of the Nomad Sea. As the sea began to trangress, pebbly, fluvial environments were encroached upon and eroded, winnowing away the fine-grained sediments. Then, through density current mechanisms, the pebbles became entrained and were transported offshore.

7) Petrographic evidence from Chungo Member samples suggests that detritus was derived from a recycled orogen province to the west. Samples are typically high in quartz and chert content and low in feldspar. A source from the Lower Cretaceous Kootenay and Cadomin Formations has been proposed. The lack of volcanic detritus in the Chungo sediments in the study area suggests that the supply of volcanics from the Elkhorn Volcanic Field in Montana, discussed by Rosenthal (in prep.), only reached as far north and west as the Ghost Dam area. A change in provenance is also indicated by the substantial change in the feldspar content. In the area north of Mount Yamnuska there is

virtually no feldspar while sections south of Mount Yamnuska show up to 15% feldspar (Rosenthal, in prep.).

8) The Chungo Member is recognizable in the Foothills on a regional basis from at least the most southerly portion of Alberta (Rosenthal, in prep.) to at least Grand Cache in the north. The seeminly thin and shaly Chungo Member in the Burnt Timber Cr./Ghost Dam region may merely be a reflection of distance from the paleoshoreline. However, this may also be related to a changing provenance which has been identified in that area.

Nomad Member

Deposition of the Nomad Member can be briefly summerized as follows:

1) The Nomad Member is interpreted to have been deposited in a deep shale basin during a brief readvancement of the Wapiabi Sea prior to the major Belly River - Paskapoo regression. The lowermost shale beds are transgressive while the uppermost turbidite beds record the fast progradation of the Belly River sedimentary pulse.

BIBLIOGRAPHY

Anderton, R., 1976. Tidal shelf sedimentation: an example from the Scottish Dolradian. Sedimentology, V. 23, pp. 429-458.

Bally, A.W., Gordy, P.L., and Stewart, G.A., 1966. Structure, seismic data, and orogenic evolution of the southern Canadian Rocky Mountains. Bulletin of Canadian Petroleum Geology, V. 14, pp. 337-381.

Banks, N.L., 1973b. Innerelv Member: Late Precambrian marine shelf deposit, East Finnmark. Norg. Geol. Unders., V. 288, pp. 7-25.

Blatt, H., Middleton, G.V., and Murray, R., 1980. Origin of sedimentary rocks, 2nd edition. Prentice-Hall, Inc., Englewood Cliffs, N.J. 782 p.

Bouma, A.H., 1962. Sedimentology of some flysch deposits: A graphic approach to facies interpretation. Elsevier, Amsterdam, 168 p.

Bourgeois, J., 1980. A transgressive shelf sequence exhibiting hummocky stratification: The Cape Sebastian Sandstone (Upper Cretaceous) southwestern Oregon. Jour. Sed. Petrology, V. 50, pp. 681-702.

Bozanich, R.G., 1979. The Bell Canyon and Cherry Canyon Formations, eastern Delaware Basin, Texas. In N.M. Sullivan (ed.), Soc. Econ. Paleont. Mineral. Special publication 79-18, pp. 121-141.

Bullock, A., 1981. Sedimentation of the Wapiabi-Belly River transition (Upper Cretaceous) at Lundbreck Falls, Alberta. B.Sc. thesis, McMaster University, Hamilton, Canada. 94 p.

Byers, C.W., 1977. Biofacies patterns in euxinic basins: a general model. In H.F. Cook and P. Enos (eds.), Deep-water carbonate environments. Soc. Econ. Paleont. Mineral. Special publication 25, pp. 5-17.

Cairnes, D.D., 1907. Moose Mountain district of southern Alberta, Geological Survey of Canada, Report 968.

Clifton, H.E., Hunter, R.E., and Phillips, R.L., 1971. Depositional structures and processes in the non-barred high energy nearshore. Jour. Sed. Petrology, V. 41, pp. 651-670.

Curray, J.R., 1956. The analysis of two-dimensional orientation data. Journal of Geology, V. 64, pp. 117-131.

Davidson-Arnott, R.G.D. and Greenwood, B., 1976. Facies relationships on a barred coast, Kouchibouguac Bay, New Brunswick, Canada. In R.A. Davis Jr. and R.L. Ethington (eds.), Beach and nearshore sedimentation. Soc. Econ. Paleont. Mineral. Special Publication 25, pp. 149-168.

Dawson, G.M., 1881. Report on an exploration from Port Simpson on the Pacific Coast to Edmonton on the Saskatchewan, embracing a portion of the northern part of British Columbia and the Peace River Country. Geologic Survey of Canada, Report Program 1879-80, pt. B, pp. 1-177.

____________, 1883. Report on the region in the vicinity of the Bow and Belly Rivers, N.W.T. Geological Survey of Canada, Rept. Prog. 1880-1882.

____________, 1984. Report on the region in the vicinity of the Bow and Belly Rivers, N.W.T. Geological Survey of Canada, Rept. Prog. 1882-1884.

Dickinson, W.R. and Suscek, C.A., 1979. Plate tectonics and sandstone compositions. Am. Assoc. Petrol. Geol. Bull., V. 63, pp. 2164-2182.

Dietz, R.S., 1963. Wave-base, marine profile of equilibrium, and wave build terraces: a critical appraisal. Geol. Soc. of Am. Bull., V. 74, pp. 971-990.

Deitz, R.S. and Menard, H.W., 1951. Origin of abrupt change in slope at continental shelf margin. Am. Assoc. Petrol. Geol. Bull., V. 35, pp. 1994-2016.

Dott, R.H. and Bourgeois, J., 1982. Hummocky stratification: significance of its variable bedding sequences. Geol. Soc. Am. Bull., V. 93, p. 663-680.

Douglas, R.J.W., 1950. Callum Creek, Langform Creek, and Gap map-areas, Alberta Geological Survey of Canada, Memoir 225. 124 p.

Douglas, R.J.W., Gabrielse, D.H., Wheeler, J.O., Stott, D.F., and Belyea, H.R., 1970. Geology of Western Canada. In R.J.W. Douglas (ed.), Geology and economic minerals of Canada. Geological Survey of Canada, Economic Geology Report No. 1, pp. 365-488.

Douglas, R.J.W., (ed.), 1970. Geology and economic minerals of Canada. Geological Survey of Canada, Economic Geology Report No. 1, 838 p.

Duke, W.D., in press. Hummocky cross-stratification, tropical hurricanes, and intense winter storms. Sedimentology.

Duke, W.D., in prep. Ph.D. Thesis, McMaster University, Hamilton, Canada.

Eisbacher, G.H., 1979. Late Mesozoic-Paleocene Bowser basin molasse and Cordilleran tectonics, western Canada. In A.D. Miall (ed.), Molasse sedimentation in North America. Geological Association of Canada, Special paper 23.

Eisbacher, G.H., Carrigy, M.A., and Campbell, R.B., 1974. Paleodrainage pattern and late orogenic basins of the Canadian Cordillera. In W.R. Dickinson (ed.), Tectonics and Sedimentation. Soc. Econ. Paleont. Mineral. Special publication 22, pp. 143-166.

Elliott, L., 1978. Clastic shorelines. In H.G. Reading (ed.), Sedimentary environments and facies, Elsevier, New York, 569 p.

Fairbridge, R.W., (ed.), 1968. The encyclopedia of geomorphology. Dowden, Hutchinson, and Ross, Inc., Straudsburg, Pa.

Folk, R.L., 1968. Petrology of sedimentary rocks. Hemphill's Book Store, Austin, Texas.

Gautier, D.L., 1982. Siderite concretions: indicators of early diagenesis in the Gammon Shale (Cretaceous). Jour. Sed. Petrology, V. 52, pp. 859-872.

Gleddie, J., 1949. Upper Cretaceous in Western Peace River Plains, Alberta. Am. Assoc. Petrol. Geol. Bull., V. 33, pp. 511-532.

Gordy, P.L., 1972. Rocky Mountains and Foothills. In J.W.H. Monger and V.A. Petrol (eds.), Geology of the southern Canadian Cordillera. XXIV International Geological Congress, Excursion A03-C03. pp. 12-21.

Hallam, A., 1970. Gyrochorte and other trace fossils in the Forest Marble of Dorset, England, In T.P. Crimes and J.C. Harper (eds.), Trace fossils, Seel House Press, Liverpool, England.

Hamblin, A.P. and Walker, R.G., 1979. Storm dominated shallow marine deposits; the Fernie-Kootenay transition, southern Rocky Mountains. Canadian Journal of Earth Sciences, V. 16, pp. 1673-1690.

Hampton, M.A., 1972a. The role of subaqueous debris flow in generating turbidity currents. Jour. Sed. Petrology, V. 42, pp. 775-7923.

Harms, J.C., Southard, J.B., Spearing, D.R., and Walker, R.G., 1975. Depositional environments as interpreted from primary sedimentary structures and stratification sequences. Soc. Econ. Paleont. Mineral. Short Course 2, Dallas, Texas, 161 p.

Harms, J.C., Southard, J.B., and Walker, R.G., 1982. Structures and sequences in clastic rocks. Soc. Econ. Paleont. Mineral. Short course 9, Calgary 1982.

Hayes, M.O., 1967a. Hurricanes as geological agents: case studies of Hurricanes Carla, 1961, and Cindy, 1963. Rep. Invest. Bur. Econ. Geol., Austin, Texas, V. 61, 54 p.

Hopkins, J.C., 1977. Production of foreslope breccia by differential submarine cementation and downslope displacement of carbonate sands, Miette and Ancient Wall buildups. In H.E. Cook and P. Enos (eds.), Deep-water carbonate environments. Soc. Econ. Paleont. Mineral. Special publication 25, pp. 155-170.

Howard, J.D. and Reineck, H.E., 1972b. Georgia coastal region, Sapelo Island, U.S.A.: Sedimentology and Biology, Senckenbergians Marit. 4, pp. 217-222.

Hume, G.S., 1930. The Highwood-Jumping Pound anticline. Geological Survey of Canada, Summary Report 1930b.

Hunter, D.F., 1980. Changing depositional environments in the Wapiabi-Belly River transition near Longview, Alberta. B.Sc. Thesis, McMaster University, Hamilton, Canada.

Hunter, R.E. and Clifton, H.E., 1982. Cyclic deposits and hummocky cross stratification of probably storm origin in Upper Cretaceous rocks of the Cape Sebastian area, southwestern Oregon. Jour. Sed. Petrology, V. 52, pp. 127-143.

Hutchinson, C.S., 1974. Laboratory handbook of petrographic techniques. John Wiley and Sons, New York, 527 p.

Irish, E.J.W., 1965. Geology of the rocky Mountain Foothills, Alberta. Geological Survey of Canada, Memoir 334, 241 p.

Johnson, H.D., 1978. Shallow siliciclastic seas. In H.G. Reading (ed.) Sedimentary environments and facies. Elsevier, New York, 569 p.

Kerr, P.E., 1977. Optical Mineralogy, 4th edition. McGraw-Hill Book Company, New York, New York, 492 p.

Leckie, D.A. and Walker, R.G., 1982. Storm and tide dominated shorelines in Cretaceous Moosebar-Gates interval; outcrop equivalents of deep basin gas trap in western Canada. Am. Assoc. Petrol. Geol. Bull., V. 66, p. 138-157.

Lerand, M.M., 1982. Chungo Sandstone Member, Wapiabi Formation at Mt. Yamnuska, Alberta. In R.G. Walker (ed.), Clastic units of the Front Ranges, Foothills and Plains in the area between Field, B.C. and Drumheller, Alberta. International Association of Sedimentology Eleventh Congress, Excursion 21A. pp. 96-116.

Lerand, M.M. and Oliver, 1975. Stratigraphy and sedimentology in the Lunbreck section. In M.S. Shawa (ed.), Guidebook to selected sedimentary environments in southwestern Alberta, Canada. Canadian Society of Petroleum Geologists, 56 p.

Lerbekmo, J.E., 1961. Stratigraphic relationship between the Milk River Formation of southern Alberta and the Belly River Formation of the southern Foothills of Alberta. Jour. of the Alberta Society of Petroluem Geologists, V. 9, p. 273-276.

___________, 1963. Petrology of the Belly River Formation, southern Alberta Foothills. Sedimentology, V. 2, pp. 54-86.

Malloch, C.S., 1911. Bighorn coal basin. Geological Survey of Canada, Memoir 9E.

McIlreath, I.A.. 1977b. Accumulation of a Middle Cambrian, deep-water, basinal limestone adjacent to a vertical submarine carbonate escarpment, Southern Rocky Mountains, Canada. In H.E. Cook and P. Enos (eds.), Deep-water carbonate environments. Soc. Econ. Paleont. Mineral. Special publication 25, pp. 113-124.

McLearn, F.H., 1919a. The Cretaceous of Peace and Athabaska Valleys. Geological Survey of Canada, Memoir 116.

___________, 1926. New species from the Coloradoan of the Peace and Smoky Valleys, Alberta. Geological Survey of Canada, Bulletin 42, pp. 117.

McLearn, F.H., and Kindle, E.D., 1950. Geology of northeastern British Columbia. Geological Survey of Canada, Memoir 259.

Middleton, G.V.and Hampton, M.A., 1976. Subaqueous sediment transport and deposition by sediment gravity flows. In D.J. Stanley and D.J.P. Swift (eds.), Marine sediment transport and environmental management. Wiley-Intersci. Publ., New York, pp. 197-218.

Middleton, G.V. and Southard, J.B., 1977. Soc. Econ. Paleont. Mineral. Short Course Notes 3, p. 7.37.

Moore, R.C., 1949. Meaning of facies, Geol. Soc. of Am. Mem. 39, p. 1-34.

Morton, R.A., 1981. Formation of storm deposits by wind-forced currents in the Gulf of Mexico and the North Sea. In S.D. Nio et al. (eds.), Holocene marine sedimentation in the North Sea basin. International Association of Sedimentologists, Special Publication 5, pp. 385-396.

Nelson, H.W. and Glaister, R.P., 1975. Trap Creek basal Belly River section or deltaic progradational sequence. In M.S. Shawa (ed.), Guidebook to selected sedimentary environments in southwestern Alberta, Canada. Canadian Society of Petroleum Geologists, 56 p.

Pemberton, S.G., Risk, M.J., and Buckley, D.E., 1976. Supershrimp deep bioturbation in the Strait of Canso, Nova Scotia. Science, V. 192, pp. 790-791.

Powers, M.C., 1953. A new roundness scale for sedimentary particles. Jour. Sed. Petrology, V. 23, pp. 117-119.

Price, R.A., Balkwill, H.R. Charleshouse, H.A.K., Cook, D.G., Simony P.S., 1972. The Canadian Rockies and tectonic evolution of the southeastern Canadian Cordillera. XXIV International Geological Congress, Excursion A15-C15, Montreal, Quebec. 129 p.

Price, R.A. and Mountjoy, E.W., 1970. Geologic structure of the Canadian Rocky Mountains between Bow and Athabasca Rivers - a progress report. In J.O. Wheeler (ed.), Structure of the southern Canadian Cordillera. Geological Association of Canada, Special Paper 6, pp. 7-25.

Rapson, J.F., 1965. Petrography and derivation of Jurassic-Cretaceous clastic rocks, Southern Rocky Mountains, Canada. Am. Assoc. Petrol. Geol. Bull., v. 49, pp. 1426-1452.

Reading, H.G., (ed.), 1978. Sedimentary Environments and Facies. Elsevier, New York, New York, 569 p.

Reich, B.M., 1983. Sedimentology of the Wapiabi-Belly River transition and the Belly River Formation (Upper Cretaceous) at Burnt Timber Creek, Alberta. B.Sc. Thesis, McMaster University, Hamilton, Canada. 121 p.

Reineck, H.E. and Singh, I.B., 1972. Genesis of laminated sand and graded rhythmites in storm-sand layers of shelf mud. Sedimentology, V. 18, p. 123-128.

___________, 1980. Depositional sedimentary environments, 2nd edition. Springer-Verlag, Berlin, West Germany, 549 p.

Risk, M.J., Venter, R.D., Pemberton, S.G., and Buckley, D.E., 1978. Computer simulation and sedimentological implications of burrowing by Axius serratus. Canadian Journal of Earth Sciences, V. 15, pp. 1370-1374.

Rose, B., 1920. Highwood coal field, Alberta. Geological Survey of Canada, Summary Report 1919c.

Rosenthal, L., in prep. M.Sc. Thesis, McMaster University, Hamilton, Canada.

Rutherford, R.L., 1927. Geology along Bow River between Cochrane and Kanaskis, Alberta. Sci. Ind. Research Council, Univ. Alberta, Report 17.

Schultheis, N.H. and Mountjoy, E.W., 1978. Cadomin Conglomerate of western Alberta - a result of Early Cretaceous uplift of the Main Ranges. Bulletin of Canadian Petroleum Geology, V. 26, pp. 297-342.

Seilacher, A., 1967. Bathymetry of trace fossils. Marine Geology, V. 5, pp. 413-428.

Shawa, M.S., (ed.), 1975. Guidebook to selected sedimentary environments in southwestern Alberta, Canada. Canadian Society of Petroleum Geologists, 56 p.

Slipper, S.E., 1921. Sheep River gas and oil field, Alberta. Geological Survey of Canada, Memoir 122.

Smith, D.G. and Putnam, P.E., 1980. Anastomosed river deposits: modern and ancient examples in Alberta, Canada. Canadian Journal of Earth Science, V. 17, pp. 1396-1406.

Smith, D.G. and Smith, N.G., 1980. Alluvial valleys near Banff. Jour. Sed. Petrology, V. 50, p. 157-164.

Spang, J.H., Brown, S.P. Stockmal, G.S., and Moffet, I., 1981. Structural geology of the foothills, and Front Ranges west of Calgary, Alberta. In Thompson, R.I. and Cook, D.G. (eds.), Field guide to geology and mineral deposits. Geological Association of Canada, Annual Meeting, Calgary, Canada.

Stelck, C.R., 1955. Cardium formation of the Foothills of northeastern British Columbia. Trans. Can. Inst. Min. Met., V. 58, pp. 132-139.

__________, 1975. Basement control of Cretaceous sand sequences in western Canada. In W.G.F. Caldwell (ed.), The Cretaceous system in the western interior of North America. Geological Association of Canada, Special Paper 13, pp. 427-440.

Stott, D.F., 1961. Summary account of the Cretaceous Alberta Group and equivalent rocks, Rocky Mountain Foothills, Alberta. Geological Survey of Canada, Paper 61-2, 34 p.

__________, 1963. The Cretaceous Alberta Group and equivalent rocks, Rocky Mountain Foothills, Alberta. Geological Survey of Canada, Memoir 317, 306 p.

__________, 1967. The Cretaceous Smoky Group, Rocky Mountain Foothills, Alberta and British Columbia. Geological Survey of Canada, Bulletin 132, 133 p.

Stow, D.A., and Shanmugan, G., 1980. Sequence of structure in fine grained turbidites, comparison of recent and ancient examples. Sedimentary Geology, V. 25, pp. 23-42.

Taylor, D.R., 1981. Sedimentology of the Moosebar tongue and bounding strata, Lower Cretaceous Blairmore Group, south-central Foothills, Alberta. M.Sc. Thesis, McMaster University, Hamilton, Canada. 239 p.

Thompson, W.O., 1937. Original structures of beaches, bars, and dunes. Geol. Soc. Am. Bull., V. 48, pp. 723-752.

Walker, R.G., (ed.), 1979. Facies Models. Geological Association of Canada, Reprint Series 1, 211 p.

Walker, R.G., Duke, W.D. and Leckie, D.A., 1983. Hummocky stratification: Significance of its variable bedding sequence. Geol. Soc. of Am. Bull., V. 94, pp. 1245-1249.

Wall, J.H., 1960. Upper Cretaceous Foraminifera from the Smoky River Area, Alberta. Research Council of Alberta, Bulletin 6.

Webb, J.B. and Hertlein, L.G., 1934. Zones in the Alberta shale (Benton Group) in Foothills of southwestern Alberta. Am. Assoc. Petrol. Geol. Bull., V. 18, pp. 1387-1416.

Wheeler, J.O., Campbell, R.B. and Reesor, J.E., 1972a. Structural style of the southern Canadian Cordillera. XXIV International Geological Congress, Excursion X01-A01, Montreal, Quebec. 118 p.

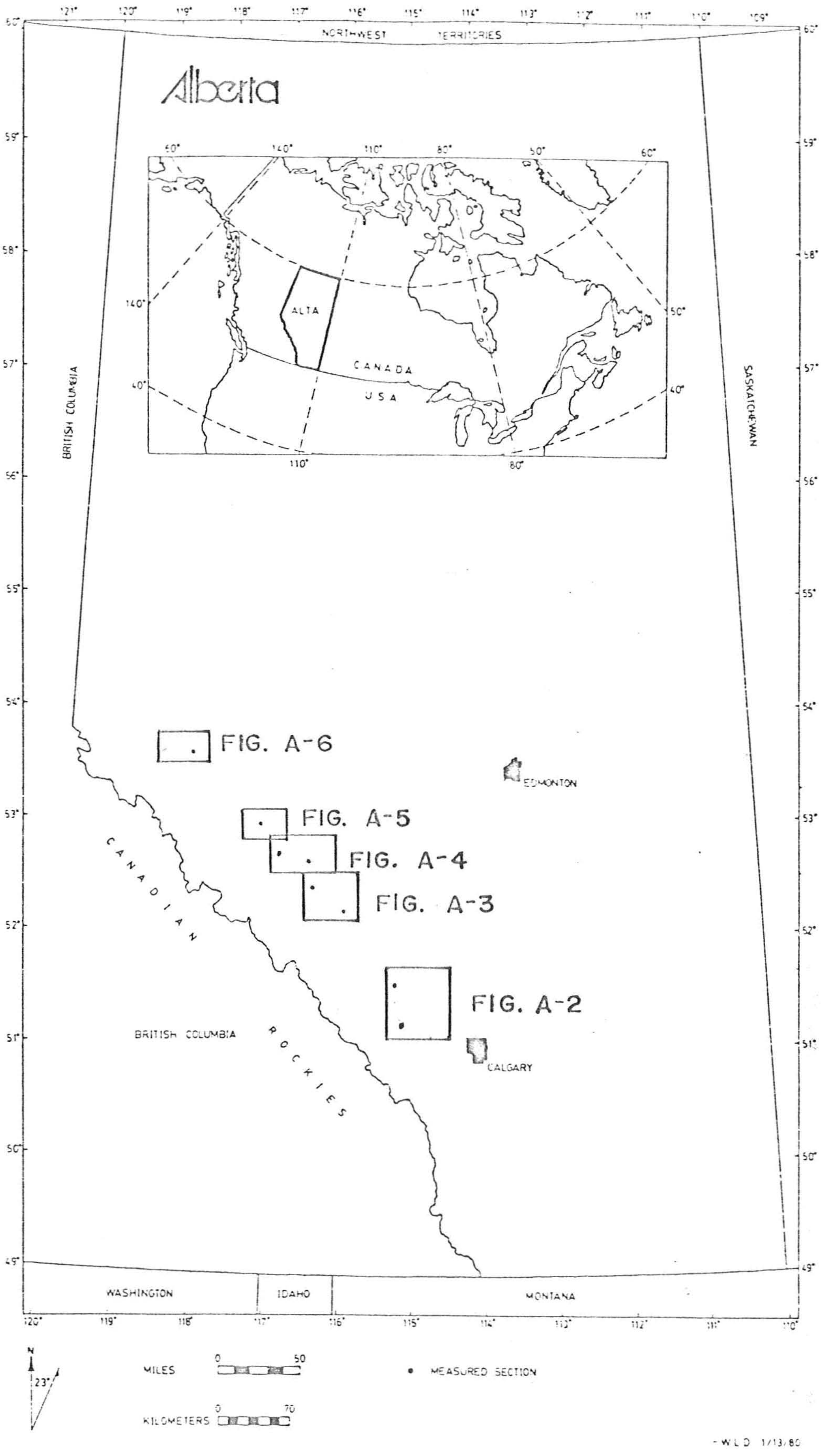
APPENDIX I

LOCATION OF SECTIONS

Introduction to Appendix I

During the summer of 1982, 8 sections of the Upper Cretaceous Wapiabi Formation were measured. All sections were located within the northwest-southeast striking structural Foothills Subprovince between Exshaw, Alberta and Grand Cache, Alberta, a distance of roughly 420 kms (Fig. AI-1). Of the 8 sections, 7 had previously been measured by Stott (1963). Very general locations of the sections were published by Stott (1963) in his G.S.C. Memoir 317. The location reference numbers used in the following portion of the text are from this publication (e.g. 5-38). The Mt. Yamnuska section, which was not studied by Stott (1963), has been designated as section MY-1.

Fig. AI-1. Location map of measured sections. Modified from Duke (in prep.)

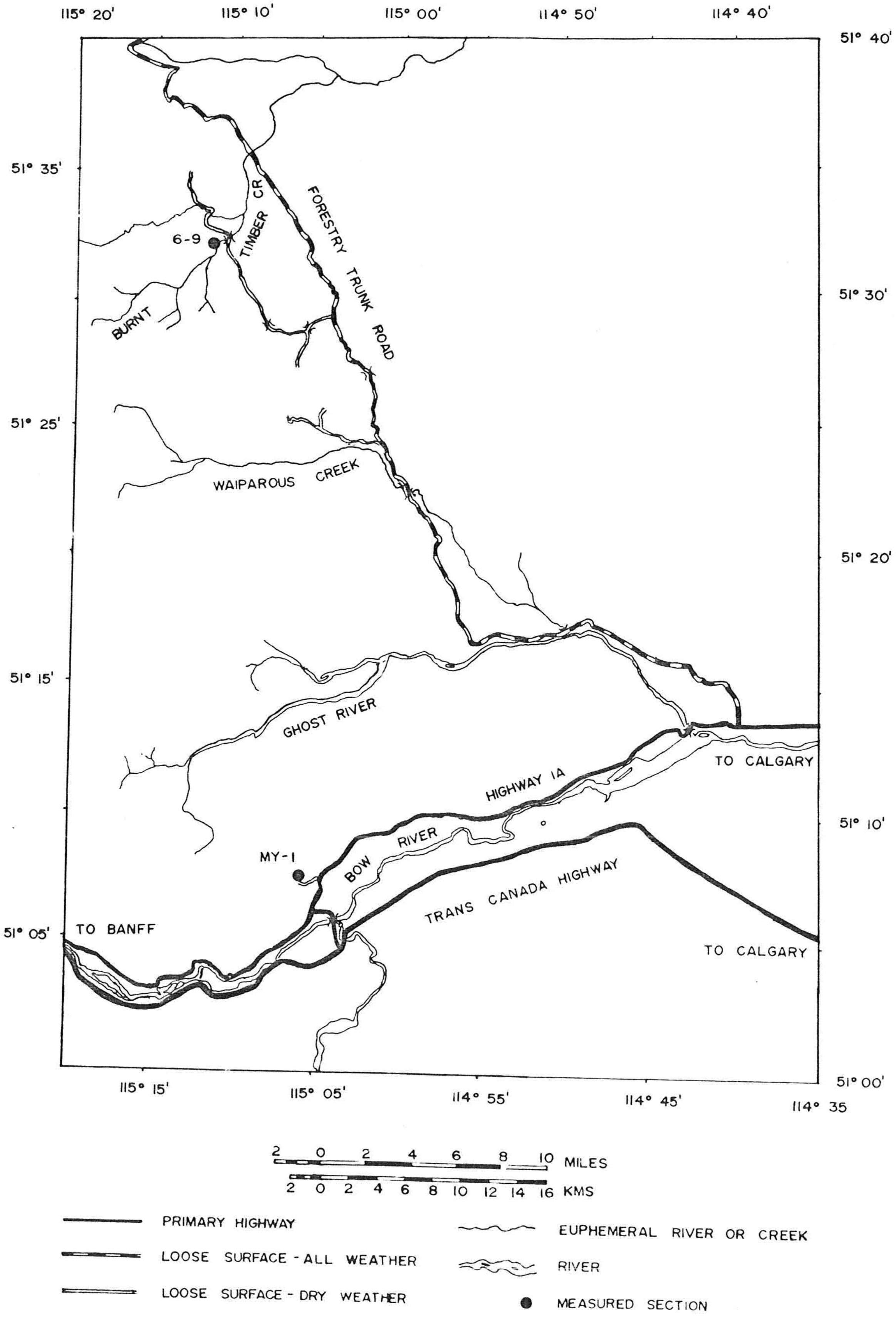

Alberta
NORTHWEST TERRITORIES
BRITISH COLUMBIA
SASKATCHEWAN
ALTA
CANADA
U.S.A.
FIG. A-6
FIG. A-5
FIG. A-4
FIG. A-3
FIG. A-2
EDMONTON
CALGARY
CANADIAN ROCKIES
BRITISH COLUMBIA
WASHINGTON
IDAHO
MONTANA
N
23°
MILES
0
50
KILOMETERS
0
70
• MEASURED SECTION
- W.L.D. 1/13/80

Mt. Yamnuska Section, Location MY-1

The Chungo Member sandstone is the only Member of the Wapiabi Formation exposed at Mt. Yamnuska. It is both well exposed and very accessible, located at the base of Mt. Yamnuska, west of Calgary (Fig. AI-2). The section is exposed in a cliff outcrop as well as in an adjacent quarry.

The Mt. Yamnuska section is quite accessible, regardless of weather. It can be approached from Calgary or Banff from highways 1 or 1A. If travelling along the Trans-Canada Highway, exit at the Seebe-Exshaw turnoff, approximately 65 km west of Calgary. Proceed 4 km to the T-intersection with highway 1A. After turning right and proceeding a further 2 km, veer left onto a gravel lane just before the highway turns east towards Calgary. If coming from Calgary along route 1A, the right turn onto the gravel lane is 3.3 km west of the Old Fort Cr. bridge. Once on the gravel road, continue for several hundred meters until an open area appropriate for parking is found. A five minute walk, continuing up the gravel road, will lead directly to the quarry exposure. The cliff section is exposed immediately to the east of the quarry.

Fig. AI-2. Location map for Mt. Yamnuska Section (MY-1) and Burnt Timber Creek (6-9).



115° 20'
115° 10'
115° 00'
114° 50'
114° 40'
51° 40'
51° 35'
51° 30'
51° 25'
51° 20'
51° 15'
51° 10'
51° 05'
51° 00'
115° 15'
115° 05'
114° 55'
114° 45'
114° 35'
TIMBER CR
FORESTRY TRUNK ROAD
6-9
BURNT
WAIPAROUS CREEK
GHOST RIVER
HIGHWAY 1A
TO CALGARY
BOW RIVER
MY-1
TRANS CANADA HIGHWAY
TO BANFF
TO CALGARY
2 0 2 4 6 8 10 MILES
2 0 2 4 6 8 10 12 14 16 KMS
PRIMARY HIGHWAY
LOOSE SURFACE - ALL WEATHER
LOOSE SURFACE - DRY WEATHER
EUPHEMERAL RIVER OR CREEK
RIVER
MEASURED SECTION

Burnt Timber Creek Section, Location 6-9 (tp. 39, rge. 8, W5)

The entire Wapiabi Formation is exposed in this superb continuous outcrop of the Blackstone, Cardium, Wapiabi and Belly River Formations along Burnt Timber Cr. (Fig. AI-2). Accessibility to the section is generally good. However, if one plans to travel to the section by car (rather than by truck) visits should be restricted to dry periods of the summer months.

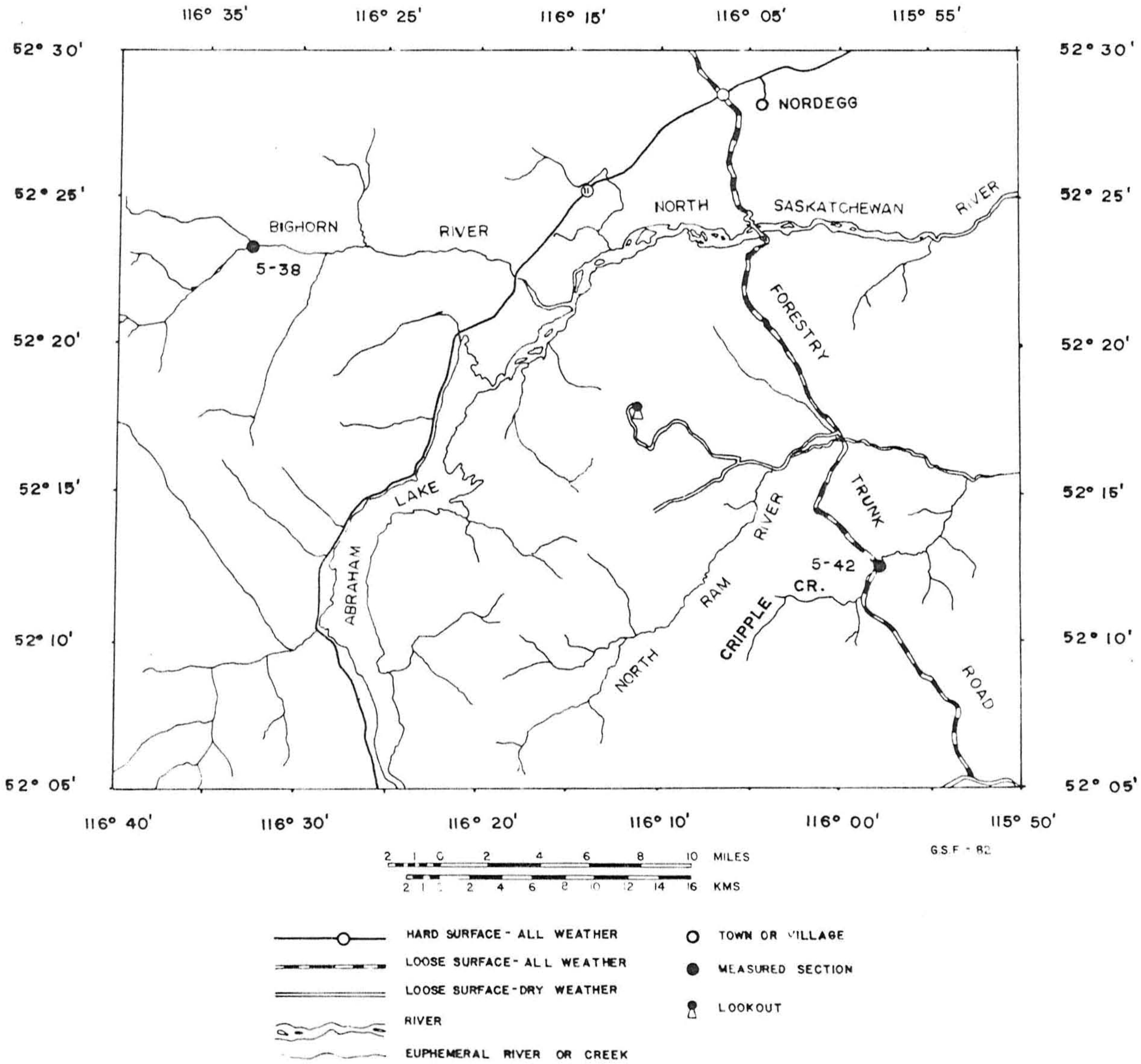
Those travelling from Calgary should begin by proceeding west along highway 1A towards Cochrane. Pass through Cochrane and continue approximately 9 km to the well marked Forestry Trunk Road intersection. Turn north, towards Nordegg, on the Forestry Trunk Road. From this intersection, continue for approximately 52 km. After the first 20 km, the paved surface will turn into an all weather loose surface road. At 52 km from the intersection, turn left onto a dry weather loose surface road marked by a "Texas-Gulf-Shell" lease road sign. Travel on this road by car during wet periods could have severe repercussions. Travel approximately 16 km on this road to the third bridge. This bridge crosses Burnt Timber Cr. The black shales noted while crossing the creek are those of the Blackstone Formation. The Wapiabi Formation is exposed 300 to 400 m upstream from this point. Continue over the bridge and proceed up the adjacent hill. Space for parking is available in an open field on a flat on the hill top.

Cripple Creek Section, Location 5-42 (tp. 37, rge. 14, W5)

Although very accessible, the Cripple Cr. section is only moderately well exposed. The Nomad Member is completely covered. Only approximately half of each of the remaining six members is exposed. The Chungo Member at Cripple Cr. is particularly difficult to observe. It is only partially exposed along both a very steep hillside adjacent to the creek and in the creek cut bank (Fig. AI-3).

To view this section, one should begin by travelling to Nordegg, approximately 80 kms west of Rocky Mountain House on highway 11. Continue west past the Nordegg Ranger Station for another 2 to 3 km to the Forestry Trunk Road intersection. Turn south onto the Forestry Road. This gravel road is all weather surfaced and therefore poses no problems during intense rainfalls. Travel as far south as the North Ram River. The outcrop is located directly along the road approximately 11± km past the river. A bridge which crosses Cripple Cr. is well marked and crosses the Thistle Member. The top of the member is upstream. A camping spot is located just before the bridge. A second Cripple Cr. bridge approximately ½ km further south crosses the well exposed Brazeau Formation. The Cardium Formation is also well exposed below the Wapiabi Formation.

Fig. AI-3. Location map for Cripple Cr. (5-42) and Bighorn River (5-38) sections.


116° 35'
116° 25'
116° 15'
116° 05'
115° 55'
52° 30'
52° 25'
52° 20'
52° 15'
52° 10'
52° 05'
116° 40'
116° 30'
116° 20'
116° 10'
116° 00'
115° 50'
NORDEGG
BIGHORN
RIVER
NORTH
SASKATCHEWAN
RIVER
5-38
FORESTRY
TRUNK
ROAD
LAKE
ABRAHAM
NORTH
RAM
RIVER
CRIPPLE
CR.
5-42
G.S.F. - 82
MILES
KMS
HARD SURFACE - ALL WEATHER
LOOSE SURFACE - ALL WEATHER
LOOSE SURFACE - DRY WEATHER
RIVER
EUPHEMERAL RIVER OR CREEK
TOWN OR VILLAGE
MEASURED SECTION
LOOKOUT

Bighorn River Section, Location 5-38 (tp. 39, rge. 18, W5)

The Bighorn River section is only accessible by helicopter. If wishing to view this section, it is most convient to fly out of Hinton where the closest permanent helicopter services are located. The oneway trip takes approximately 45 minutes. The location is shown in Figure AI-3.

First instruct the pilot to fly south (from Hinton) towards the Bighorn Dam on the North Saskatchewan River, some 20 kms southwest of Nordegg on Highway 11. The Bighorn River drains into the North Saskatchewan River approximately 4 kms downstream (northeast) of the damsite. Follow the Bighorn River west through a steep canyon and past Crescent Falls. Continue flying upstream. Roughly 7 kms upstream of Crescent Falls, the Littlehorn Cr. can be seen to drain into the Bighorn River from the south. the lower members of the Wapiabi Formation are located another 1½ to 2 kms upstream from this point. The area is quite flat, so landing the helicopter is no problem.

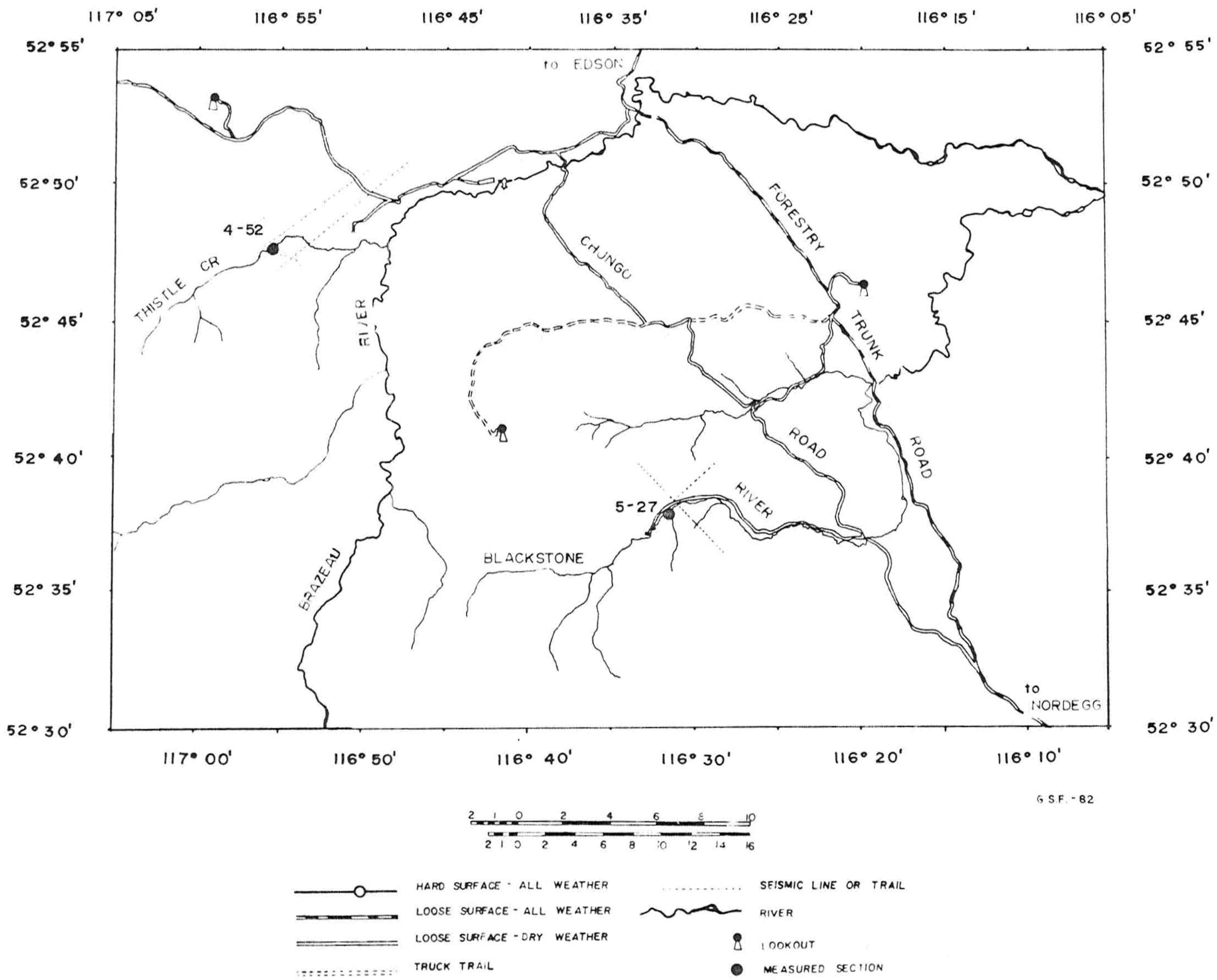
Only the Muskiki, Marshybank, Dowling, Thistle and Hanson Members are exposed directly along the Bighorn River. The Chungo Member is exposed high up on an adjacent hillside on the north side of the river. It forms a very prominent and resistant cliff which can easily be seen from the river. The hike up to the Chungo takes 1 to 2 hours, depending on physical condition. Persons with heart trouble would not be advised to make the climb. The Nomad Member is not exposed at this section.

Blackstone River Section, Location 5-27 (tp. 42, rge. 18, W5)

The Blackstone River section is a very well exposed example of the entire Wapiabi Formation. However, accessibility to the section is highly dependent on the vehicle and the weather. If attempting to view this section after or during a wet period, a 4-wheel drive truck with a winch is a must. During dry periods, a 2-wheel drive truck with high ground clearance would be sufficient. Cars are not suggested to attempt the trek to this section (Fig. AI-4).

To view the Blackstone River section, first travel to Nordegg. From Nordegg, turn north onto the Forestry Trunk Road and proceed approximately 8 kms until a gravel road is noted that veers off to the left. This road is clearly marked as the Chungo Road. Follow the Chungo Road until it comes to the Blackstone River, some 11 to 13 km. Cross the river. Roughly ¼ to ½ km past the bridge, turn left onto an unmarked road. This road is deeply rutted, which makes for slow travel. It would be smart to check the condition of this road at the Nordegg Ranger Station before attempting to view the section. Proceed along the road for approximately 12 to 15 km. An open area overlooking the Blackstone River and suitable for camping should be found on the left. Continue another 3 km until a seismic line transecting the road is found. Park the vehicle. Follow the seismic line south to the Blackstone River, about ½ km. The Wapiabi Formation is exposed about ½ km upstream (west). It is directly upstream of a well exposed example of the Cardium Formation. Stratigraphic top is to the west.

Figure AI-4. Location map for Blackstone River (5-27) and Thistle Creek (4-52) sections.


117° 05'
116° 55'
116° 45'
116° 35'
116° 25'
116° 15'
116° 05'
52° 55'
52° 50'
52° 45'
52° 40'
52° 35'
52° 30'
117° 00'
116° 50'
116° 40'
116° 30'
116° 20'
116° 10'
to EDSON
to NORDEGG
THISTLE CR
4-52
5-27
BRAZEAU RIVER
BLACKSTONE RIVER
CHUNGO
FORESTRY TRUNK ROAD
ROAD
G.S.F. - 82
HARD SURFACE - ALL WEATHER
LOOSE SURFACE - ALL WEATHER
LOOSE SURFACE - DRY WEATHER
TRUCK TRAIL
SEISMIC LINE OR TRAIL
RIVER
LOOKOUT
MEASURED SECTION

Thistle Creek Section, Location 4-52 (tp. 44, rge. 20, W5)

The Thistle Creek section is an excellent, fully exposed example of the entire Wapiabi Formation. It is easily accessible by automobile in any weather. As well, it is the thickest of the measured sections in the study area. Stott (1963) chose this section as the type section of the Wapiabi Formation. Location is shown in Figure AI-4.

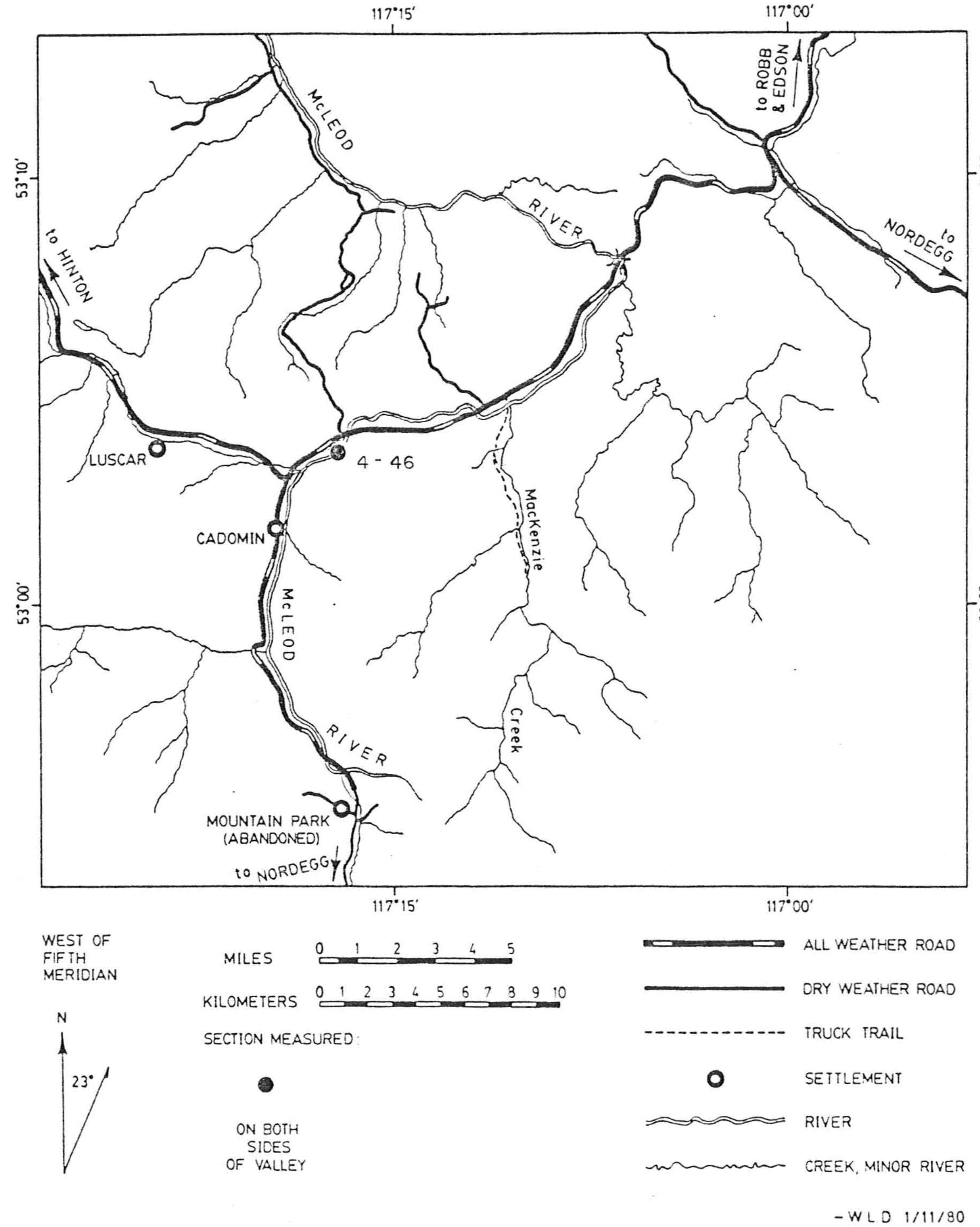
Although the Thistle Creek section crops out almost equidistantly from Hinton and Nordegg, it is easier to find if travelling via Nordegg. From Nordegg, travel north on the Forestry Trunk Road to the Brazeau River crossing. This is about a 1 hour drive. Proceed across the bridge and up the river bank. At the top of the bank, turn left onto another gravel road, the Cardinal River Road. Continue along this road approximately 16 km until you pass through an Indian village. One to two km past the village the road forks. Take the north fork. Roughly 3 km further, a dirt road should be noted on the left side of the road. Turn left and follow this road until you come to an uncrossable creek. This is Thistle Creek. The Thistle Member beds are well exposed at the crossing. The Chungo Member is exposed approximately ½ to 1 km upstream. An excellent example of the Cardium Formation can be viewed about ½ km to 1 km downstream from the crossing.

McLeod River Section, Location 4-46 (tp. 47, rge. 22, W5)

The McLeod River section is a superbly exposed and very accessible example of the Hanson, Chungo, and Nomad Members of the Wapiabi Formation. Lower members are exposed, but tight folding prohibited a complete measured section. Location shown in Figure AI-5.

To view this section, first travel to Hinton, approximately 80 kms east of Jasper on highway 16. Proceed 2 to 3 km west of Hinton on highway 16 to highway 40 (Luscar Road). Turn left (south) onto highway 40 and continue 48 kms until a T-intersection is reached. Turn left (east) at the T and proceed 2.5 km. At this point, park in the open area on the south side of the road. The section is exposed along McLeod River directly beside the parking area. The entire route is paved except for the last 2.5 kms, so accessibility is no problem in any weather. The Cardium Formation can also be viewed along the last 2.5 kms of the road.

Figure AI-5. Location map for the McLeod River (4-46) section. Modified from Duke (in prep.)

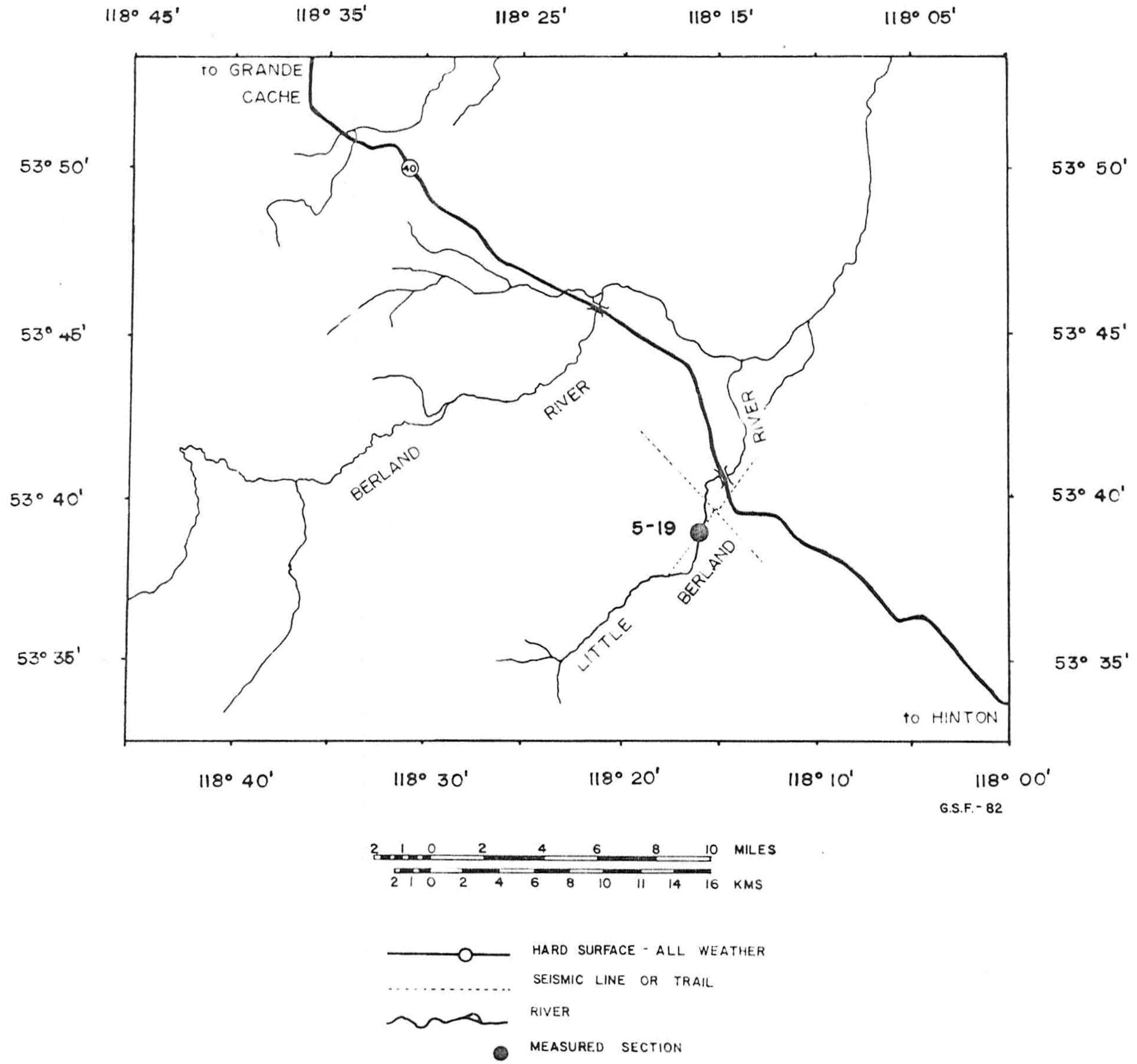

117°15'
117°00'
53°10'
53°00'
to ROBB & EDSON
McLEOD
RIVER
to NORDEGG
to HINTON
LUSCAR
4 - 46
CADOMIN
MacKenzie
Creek
McLEOD
RIVER
MOUNTAIN PARK
(ABANDONED)
to NORDEGG
WEST OF
FIFTH
MERIDIAN
N
23°
MILES
0 1 2 3 4 5
KILOMETERS
0 1 2 3 4 5 6 7 8 9 10
SECTION MEASURED:
ON BOTH
SIDES
OF VALLEY
ALL WEATHER ROAD
DRY WEATHER ROAD
TRUCK TRAIL
SETTLEMENT
RIVER
CREEK, MINOR RIVER
-W L D 1/11/80

Little Berland River Section, Location 5-19 (tp. 53, rge. 3, W6)

The Little Berland River section of the Wapiabi Formation is very short, poorly exposed and only accessible by helicopter or motorcycle. A four wheel-drive vehicle with a winch may get to the section during very dry periods only.

To view this section by truck or motorcycle, travel north from Hinton towards Grand Cache on highway 40. At 5.0 kms past the second railroad track crossing or 1.5 kms before the Little Berland River bridge, turn left onto a gravel lease road. Proceed up the road for approximately ½ km to an abandoned water well site beside the railroad tracks. Continue along the tracks a few hundred meters until a seismic cut line is noted on the right. Cross the railroad tracks and proceed up the cut line to the top of the hill where two cut lines intersect. Turn right and travel an estimated 4 to 6 km along this cut line to the Little Berland River. The Wapiabi Formation is exposed along the river at the crossing of the river with the cut line. Stratigraphic top is downstream. A very well exposed section of the Cardium Formation can be found by continuing on the cut line approximately 1 km after the river crossing.

Figure AI-6. Location of Little Berland River (5-19) Section.



118° 45'
118° 35'
118° 25'
118° 15'
118° 05'
to GRANDE CACHE
40
53° 50'
53° 45'
53° 40'
53° 35'
BERLAND
RIVER
RIVER
5-19
LITTLE
BERLAND
to HINTON
118° 40'
118° 30'
118° 20'
118° 10'
118° 00'
G.S.F.-82
2 1 0 2 4 6 8 10 MILES
2 1 0 2 4 6 8 10 11 14 16 KMS
HARD SURFACE - ALL WEATHER
SEISMIC LINE OR TRAIL
RIVER
MEASURED SECTION

APPENDIX II

MEASURED SECTIONS

Detailed measured sections with an accompanying legend are located in the rear pocket of this text. Paleocurrent data, presented in the form of rose diagrams, are included to the right of the measured sections. Thickness measurements (in meters) and formation and member names are shown on the left of the measured sections.

APPENDIX III

RAW PALEOCURRENT DATA

Due to the translational tectonic deformation in the study area (Price et al., 1972), sedimentary structures which yielded paleoflow measurements had to be restored to their original depositional orientation. This was done by rotating the structures about regional strike. All paleoflow measurements were rotated in the field by determining the rake angle (the angle between the paleoflow orientation and the horizontal within the place of the beds) and adding to it, or subtracting it from, the strike of the beds (Taylor, 1981).

The measured paleoflow data in the following pages has been recorded in the following manner: asymmetrical ripples, flutes and rib and furrow structures record the true flow direction; lineations show a sense of direction only; symmetrical ripples are recorded as ripple crest trends, 90° to the inferred direction of water movement. Symmetrical ripples are also represented this way in the rose diagrams in Appendix II.

Statistical analysis of these data have been performed using the methodology of Curray (1956) and a program written for a TI-58 calculator (Cheel, unpublished; Table AIII-1). The statistics calculated were: 1) vector mean ($\bar{\theta}$), a measure of the central tendency of the distribution; 2) vector length (R) - the magnitude of the resultant vector; 3) vector length in terms of percent (L) - the magnitude of the resultant vector as a percent; and 4) P - the probability that the distribution is random.

The fashion in which these statistics were applied needs some explanation. Independent values of $\bar{\theta}$, R, L, and P were calculated for symmetrical ripples, lineations, and true flow indicators (flutes, rib and furrow structures, and asymmetrical ripples) in each member at each measured section. The only exception to this was in the Chungo Member. In this member, flutes, asymmetrical ripples, and rib and furrow structures from asymmetrical ripples (from turbidite beds) are separated from trough cross-beds and rib and furrow structures from trough cross-beds (from non-turbidite beds). The intention is to separate structures indicating basin geomety from nearshore structures which may be quite variable.

RESULTANT VECTOR DETERMINATION FOR GROUPED DATA: TI-58

Formulae (from Curray, 1956)

$W = \Sigma n \cos\theta$

$V = \Sigma n \sin\theta$

$r = \sqrt{V^2 + W^2}$

$L = \frac{R}{\Sigma n} \times 100$

$P = e^{-(R^2/n)}$

Labels

A → enter θ, calc. $\Sigma\sin\theta$, $\Sigma\cos\theta$, $\Sigma\tan\bar{\theta}$

B → calculates $\bar{\theta}$

C → calculates R

D → calculates L

E → calculates P, the probability that the distribution is random.

User Defined Keys

x^2 → add 1 to mem. 7

$\sqrt{x}$ → add 1 to mem. 7

$1/x$ → add 1 to mem. 7

y^x → produce $\bar{\theta}$, where $\bar{\theta} = \bar{\theta}'$

÷ → $\bar{\theta}' + 360 = \bar{\theta}$

Memories

1 → θ

2 → $\Sigma\sin\theta$

3 → $\Sigma\cos\theta$

4 → $\Sigma\tan\theta$

5 → Σn

6 → $\bar{\theta}'$

7 → ≤ 3

8 → P

TABLE AIII-1

Program TI-58

000	76	2nd LBL	038	43	RCL	076	32	x≥t	114	33	x^2
001	11	A	039	02	2	007	43	RCL	115	85	+
002	42	STO	040	77	2nd x≥t	078	07	7	116	43	RCL
003	01	1	041	33	x^2	079	67	2nd x=t	117	03	3
004	38	2nd SIN	042	43	RCL	080	45	y^x	118	33	x^2
005	44	SUM	043	03	3	081	25	CLR	119	95	=
006	02	2	044	77	2nd x≥t	082	32	x≥t	120	34	$\sqrt{x}$
007	43	RCL	045	34	$\sqrt{x}$	083	43	RCL	121	52	STO
008	01	1	046	43	RCL	084	03	3	122	08	8
009	39	2nd COS	047	04	4	085	77	2nd x≥t	123	91	R/S
010	44	SUM	048	77	2nd x≥t	086	55	÷	124	76	2nd LBL
011	03	3	049	35	1/x	087	43	RCL	125	14	D
012	43	RCL	050	61	STO	088	06	6	126	43	RCL
013	01	1	051	00	7	089	85	+	127	08	8
014	30	2nd TAN	052	74	4	090	01	1	128	55	÷
015	44	SUM	053	76	2nd LBL	091	08	8	129	43	RCL
016	04	4	054	33	x'	092	00	0	130	05	5
017	25	CLR	055	01	1	093	95	=	131	65	x
018	01	1	056	44	SUM	094	91	R/S	132	01	1
019	44	SUM	057	07	7	095	76	2nd LBL	133	00	0
020	05	5	058	61	GTO	096	45	y^2	134	00	0
021	43	RCL	059	00	4	097	43	RCL	135	95	=
022	05	5	060	42	2	098	96	6	136	91	R/S
023	91	R/S	061	76	2nd LBL	099	91	R/S	137	76	2nd LBL
024	76	2nd LBL	062	34	$\sqrt{x}$	100	76	2nd LBL	138	15	E
025	12	B	063	01	1	101	55	÷	139	43	RCL
026	43	RCL	064	44	SUM	102	43	RCL	140	08	8
027	02	2	065	07	7	103	06	6	141	33	x'
028	55	÷	066	61	GTO	104	85	+	142	55	÷
029	43	RCL	067	00	4	105	03	3	143	43	RCL
030	03	3	068	46	6	106	06	6	144	05	5
031	95	=	069	76	2nd LBL	107	00	0	145	94	°
032	22	INV	070	35	1/x	108	95	=	146	95	=
033	30	2nd TAN	071	1	1	109	91	R/S	147	22	INV
034	42	STO	072	44	SUM	110	76	2nd LBL	148	23	Ln x
035	06	6	073	07	7	111	13	C	149	91	R/S
036	25	CLR	074	25	CLR	112	43	RCL			
037	32	x t	075	03	3	113	02	2			

Burnt Timber Creek

Muskiki Member

lineation 75/255 → $\bar{\theta}$=75/255, R=1, L=100%, P=3.7×10^{-1}

Marshybank Member

nil

Dowling Member

nil

Thistle Member

symmetrical ripples	141/321, 139/319, 140/320, 141/321, 50/320, 143/323, 51/231, 12/192, 80/260, 177/357 → $\bar{\theta}$=112/292, R=6.43, L=64.32%, P=1.6×10^{-2}
lineations	23/203, 33/213, 50/230, 38/218, 51/321, 65/245, → θ=43/223, R=5.83, L=97.18%, P=3.5×10^{-3}
flutes	320, 307, 313
asymmetrical ripples	352, 352, 350, 347, 353, 320, 35, 45 $\bar{\theta}$=346, R=9.56, L=86.94%, P=2.4×10^{-4}

Hanson Member

nil

Chungo Member

nil

Nomad Member

nil

Cripple Creek

Muskiki Member

asymmetrical ripples	60,55, 35 → $\bar{\theta}$=50°, R=2.95, L=98.23%, P=5.5×10^{-2}

Marshybank Member

nil

Dowling Member

asymmetrical ripples	40 → $\bar{\theta}$=40°, R=1, L=100%, P=3.7×10^{-1}

Thistle Member

asymmetrical ripples	40, 52, 43, 20, 15, 51, 22, 35, 35, 50, 42, 45, 20, 35, 30, 25, 46, 51, 41, 41, 55, 25, 30, 28, 37, 33, 38, 19, 30, 20, 19, 32, 24, 17, 29 → $\bar{\theta}$=34, R=34.33, L=98.10%, P=23×10^{-15}
lineations	152/332 → $\bar{\theta}$=152/332, R=1, L=100%, P=3.7×10^{-1}

Hanson Member

nil

Chungo Member

lineations	5/185 → $\bar{\theta}$ = 5/185, R=1, L=100T, P=3.7×10^{-1}
rib and furrow	10.5 → $\bar{\theta}$=8°, R=1.99, L=99.90%, P=1.4×10^{-1}
flutes	45, 35 → $\bar{\theta}$=40°, R-1.99, L=99.62%, P=1.4×10^{-1}

Nomad Member

nil

Bighorn River

Muskiki Member

asymmetrical ripples 60, 75, 40 → $\bar{\theta}$=58, R=2.91, L=96.89%, P=5.9x10^{-2}

Marshybank Member

nil

Dowling Member

asymmetrical ripples 25, 20, 70, 55, 91, 42, 45, 33, 40, 40, 42, 30, 35, 40, 25, 31, 23, 42, 35, 38, 44, 43, 43 → $\bar{\theta}$ =40, R=22.24, L=96.68%, P=5x10^{-10}

Thistle Member

asymmetrical ripples 20, 25, 31, 37, 27, 36, 25, 41, 50, 34, 37, 30, 20, 30, 35, 40, 33, 32, 22, 39, 30, 32, 38, 44, 135, 117, 115, 116, 140, 120, 100, 109, 121, 100, 115, 114, 129, 140, 143, 105, 110, 115, 130, 80, 75, 81, 115, 110, 130, 120, 111, 106, 115, 118, 126, 116, 112, 119, 134, 125, 116, 122, 40, 55, 30, 60, 115, 100, 30, 60, 61, 45 56, 24, 35, 42

rib and furrow 41, 55 → $\bar{\theta}$=77, R=53.96, L=75.58%, P=4.4x10^{-20}

oscillation ripples 138/318, 95/275, 3/188, 108/289, 130/310, 150/330, 155/335, 105,285, 105/285, 101/281, 122/302, 109/284 → $\bar{\theta}$ =114/294, R=10.04, L=83.68%, P=2.2x10^{-4}

Hanson Member

asymmetrical ripples 52, 50, 41 → $\bar{\theta}$=43, R=2.99, L=99.65%, P=5.1x10^{-2}

Chungo Member

txb 140
rib and furrow 136, 149, 140, 145, 150 } $\bar{\theta}$=143, R=5.98, L=99.61%, P=2.5x10^{-3}
asymmetrical ripple 155 → $\bar{\theta}$=155, R=1.0, L=100%, P=3.7x10^{-1}

Nomad Member

nil

Blackstone River

Muskiki Member

nil

Marshybank Member

nil

Dowling Member

asymmetrical ripples — 48, $\bar{\theta}$=48, R=1, L=100%, P=3.7x10^{-1}

Thistle Member

asymmetrical ripples — 14, 44, 35, 20, 36, 35, 22, 12, 18, 10, 30, 28, 30 → $\bar{\theta}$=26, R=12.80, L=98.45%, P=3.4x10^{-6}

symmetrical ripples — 118/298, 105/285, 119/299, 50/320, 112/292, 123/303, 125/305, 129/309, 108/288, 142/322, 132/312, 112/292, 130/310, 135/315, 99/279, 138/318, 122/302, 130/310, 50/280, 50/320, 90/270, 125/305, 105/285, 170/350, 100/280, 130/310, 140/320, 80/260, 140/320, 170/350, 110/290 → $\bar{\theta}$=117/297, R=27.37, L=88.30%, P=2.3x10^{-11}

Hanson Member

asymmetrical ripples — 35, 39, 40 → $\bar{\theta}$=38, R=2.30, L=99.93%, P=5.0x10^{-2}

Chungo Member

rib and furrow — 55, 10, 63 → $\bar{\theta}$=43.11, R=2.76, L=91.88%, P=7.9x10^{-2}

symmetrical ripples — 100/280, 105/285, 125/305, 105/385, 105/283, 110/290, 100/280, 103/283, 130/310, 106/286, 114/294, 140/320, 132/312, 115/295, 110/290, 105/285, 130/310, 110/290, 115/295, 103/283, 123/303, 120/300, 123/303, 118/298, 115/295, 112/292 → $\bar{\theta}$=114/294, R=25.56, L-98.30%, P=1.2x10^{-11}

Nomad Member

asymmetrical ripple — 20 → $\bar{\theta}$=20, R=1, L=100%, P=3.7x10^{-1}

Thistle Creek

Muskiki Member

asymmetrical ripples 60 → $\bar{\theta}=60$, $R=1$, $L=100\%$, $P=3.7\times10^{-1}$

Marshybank Member

nil

Dowling Member

asymmetrical ripples 75, 120, 80, 85 → $\bar{\theta}=90$, $R=3.81$, $L=95.33\%$, $P=2.6\times10^{-2}$

Thistle Member

asymmetrical ripples 50, 40, 101, 29, 51, 109, 63, 25, 25, 111, 96, 135, 162, 18, 40, 41, 56, 57, 45, 45, 5, 20, 15, 30, 50, 55, 65, 26, 24, 26, 358, 355, 52, 35, 31, 20, 25, 30, 357, 31, 13, 55, 20, 27, 22 35, 51, 49, 25, 30, 35, 36, 11, 4, 30, 6, 15, 10, 25, 5, 11, 10, 10, 15, 15, 53, 20, 25, 30, 40, 25, 30, 40, 20, 30, 40, 85, 143, 113, 47

$\bar{\theta}=36$, $R=68.82$, $L=86.02\%$, $P=1.9\times10^{-26}$

lineation 10/190 → $\bar{\theta}=10/190$, $R=1$, $L=100\%$, $P=3.7\times10^{-1}$

Hanson Member

nil

Chungo Member

txb 180, 202, 200, 180, 55, 135, 140, 215
rib and furrow 20, 170, 163, 177, 171, 175, 100 } $\theta=163$, $R=10.19$, $L=67.90\%$ $P=9.9\times10^{-4}$

flutes 35, 40 → $\bar{\theta}=38°$, $R=1.99$, $L=99.9\%$, $P=1.4\times10^{-1}$

symmetrical ripples 60/240 → $\bar{\theta}=60/240$, $R=1$, $L=100\%$, $P=3.7\times10^{-1}$

Nomad Member

nil

McLeod River

Hanson Member

lineation	40/220 → $\bar{\theta}$=40/220, R=1, L=100%, P=37.10^{-1}	
asymmetrical ripple	10 → $\bar{\theta}$=10, R=1, L=100%, P=3.7x10^{-1}	

Chungo Member

asymmetrical ripples	130, 190	θ=94° R=3.68, L=61.30%, P=1.0x10^{-1}
flutes	95, 80, 73, 8	
rib and furrow	80, 40, 185, 160, 195, 150	$\bar{\theta}$=126°, R=5.09, L=63.60%, P=3.9x10^{-2}
trough cross-bedding	90, 100	
symmetrical ripples	94/274, 120/300, 90/270, 150/330, 120/300 $\bar{\theta}$=115/295, R=4.65, L=93.03%, P=1.3x10^{-2}	

Nomad Member

asymmetrical ripple	15 → $\bar{\theta}$=15°, R=1, L=100%, P=3.7x10^{-1}	

Little Berland River

Muskiki Formation

nil

Bad Heart Formation

asymmetrical ripples	125, 70	$\bar{\theta}$=99, R=4.15, L=83.08%,
flutes	70, 75, 153	P=3.2x10^{-2}
lineations	153/333, 149/329 → $\bar{\theta}$=151/331, R=200, L=99.94% P=1.4x10^{-1}	

Dowling Member

nil

Thistle Member

asymmetrical ripples	9, 19, 34, 356, 15	$\bar{\theta}$=18. R=5.83,
rib and furrow	35	L=97.18%, P=3.5x10^{3}

Hanson Member

asymmetrical ripples	21 → $\bar{\theta}$=21, R=1, L=100%, P=3.7x10^{-1}

Chungo Member

asymmetrical ripples	25, 20, 22, 5, 24, 20, 15, 11, 22, 13 → $\bar{\theta}$=18°, R=9.94, L=99.43%, P=5.0x10^{-5}
rib amd furrow	19 → $\bar{\theta}$=19, R=1, L=100%, P=3.7x10^{-1}

Nomad Member

nil

PETROGRAPHIC DATA

APPENDIX IV

The following Tables summarize the raw petrographic data collected during this study.

TABLE AIV-1. Composition of sandstones from selected thin sections of the Chungo Member, Wapiabi Formation (this thesis) in the central Foothills, Alberta. MY = Mount Yamnuska, BTC = Burnt Timber Cr., CC = Cripple Cr., BHR = Bighorn River, BR = Blackstone River, TC = Thistle Cr., MR = McLeod River, and LBR = Little Berland River.

sample	quartz	chert	rock fragments	feldspar	detrital mica	detrital carbonate	unknown	other	matrix/ cement
MY-6	65	29.3	2.7	-	-	-	0.7	-	2.3
MY-8	54.7	41.7	1	-	-	-	0.3	-	2.3
MY-5	54.7	37.3	0.7	0.0	1	-	1.3	-	5
BTC-11	48	35.7	0.7	0.3	0.7	4.7	1.7	-	8.3
BTC-16	54.3	32	0.3	-	1.3	6.7	0.7	-	4.7
CC-16	35.7	53	3	-	-	-	0.3	-	8
BHR-8	53.3	33	-	-	0.7	3.7	-	-	9.3
BHR-6	73.7	17.7	1	0.3	-	1.7	-	-	5.7
BR-7	56.3	33	0.7	0.3	0.3	6	-	-	3.3
TC-16	62	28.7	3.7	-	0.3	-	-	0.3	5
MR-6	65	16	0.7	-	-	2.3	0.3	0.3	15.3
MR-11	64.3	26	0.7	-	0.7	1	-	0.7	6.7

TABLE AIV-1 (continued)

sample	quartz	chert	rock fragments	feldspar	detrital mica	detrital carbonate	unknown	other	matrix/ cement
LBR-9	54	23.7	0.7	-	0.7	4	-	-	17
LBR-11	62.7	30.7	-	-	0.3	0.3	-	0.3	5.7

TABLE AIV-2. Shows the facies from which each thin-section was derived, the average grain size of the slide, and the cements which were identified. The cements are listed in decreasing order of abundance. Grain sizes were estimated using a Canstrat grain size card.

sample	facies	ave. grain size	cements
MY-4	6	vf_u	siderite, minor calcite
MY-5	7	f_l	clay minerals, quartz, minor siderite and hematite
MY-6	8	$f_u - m_l$	quartz, minor hematite and siderite
MY-8	8	$f_u - m_l$	quartz
MY-10	8	$vf_u - f_l$	siderite
BTC-11		$vf_l - vf_u$	siderite, clay minerals, calcite
BTC-16		siltstone-vf_l	clay minerals, calcite, siderite
CC-14	5	siltstone-vf_l	clay minerals, calcite
CC-16	4	$f_u - m_l$	calcite, clay minerals
BHR-6	4	$f_l - f_u$	calcite, quartz
BHR-8	7	vf_u	siderite, minor clay minerals
BR-7	5	vf_l	clay minerals, calcite
TC-16	9	f_u	quartz, siderite
MR-4	4	siltstone-vf_l	clay minerals

TABLE AIV-2 (continued)

sample	facies	ave. grain size	cements
MR-6	4	vf_u	siderite
MR-9	4	vf_l	clay minerals
MR-10	4	siltstone-vf_l	clay minerals
MR-11	4	vf_l-vf_u	clay minerals, minor calcite
LBR-9	5	vf_l	clay minerals, minor calcite
LBR-11	6	vf_l-vf_u	clay minerals, minor siderite and quartz

TABLE AIV-3. Data from four pebble conglomerate samples used in the petrographic analysis.

sample	maximum pebble size	average pebble size	pebble rounding	dominant pebble lithology	matrix	cement
MY-9	6mm x 3mm	4mm x 2mm	well rounded	chert	sandstone	calcite first, then siderite
BHR-9	8mm x 4mm	4mm x 3mm	well rounded	chert	sandstone	siderite
BR-10	10mm x 5mm	9mm x 4mm	well rounded	chert	carbonate mud	clay minerals
MR-5	15mm x 6mm	5mm x 4mm	well rounded	chert	sandstone	calcite, first, then siderite

APPENDIX V

METHOD FOR PALINSPASTIC RECONSTRUCTION OF MEASURED SECTIONS

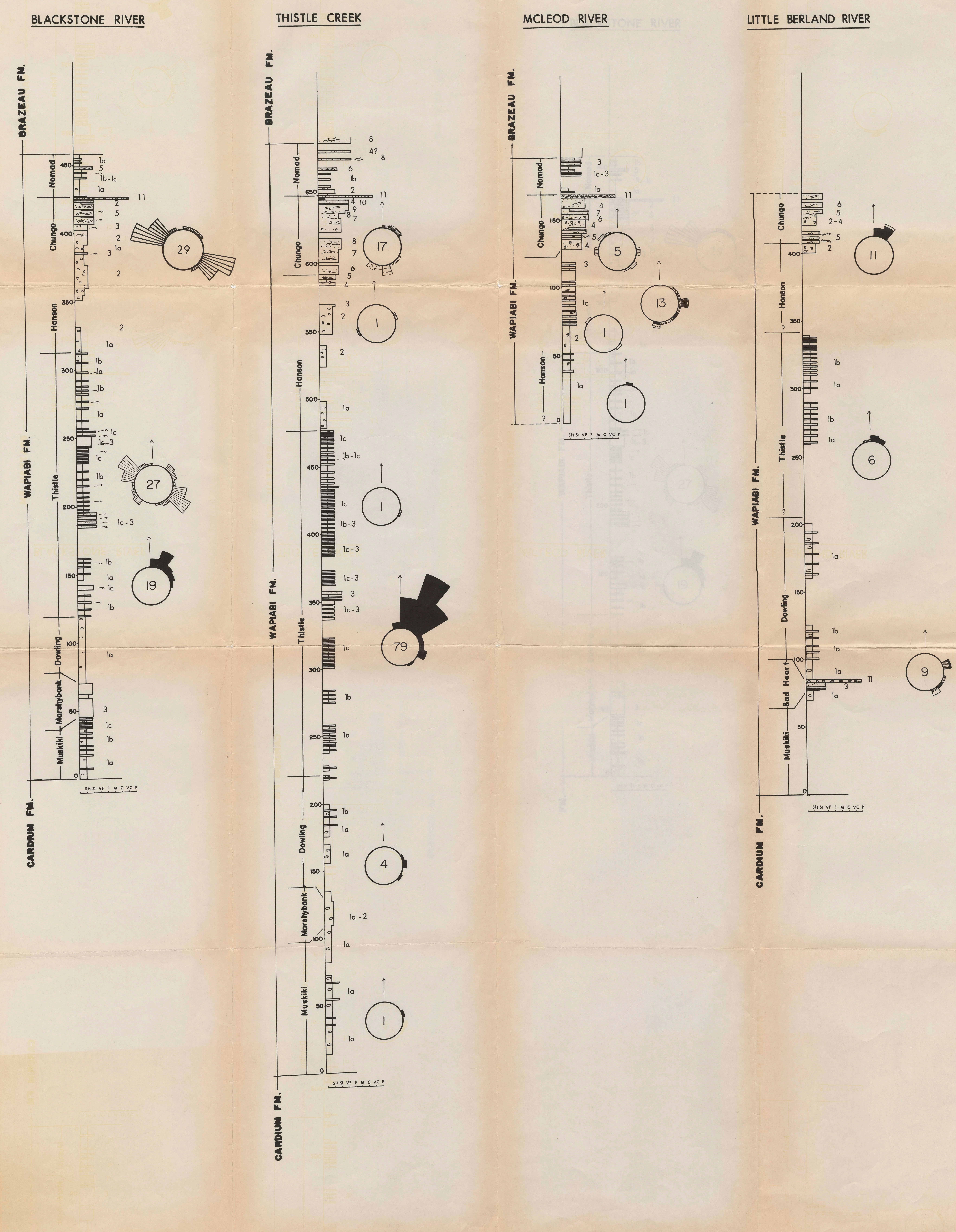
In order to understand the pre-thrusting relationships of the sections measured in this study, an attempt was made to palinspastically reconstruct the original depositional position of the sections. This was done by first plotting the present position of sections on the closest available structural cross-section of the Foothills. Cross-sections through the study area were obtained from Bally (1966) and Wheeler (1972). The particular thrust slice on which a measured section layed was recorded. These are listed under the "Structural Setting" portion of the thesis. Cross-sections were then "unthrusted", using the eastern edge of the disturbed belt as the stationary reference point. The datum along which the amount of thrusting was measured was either the Cardium Formation or the Wapiabi Formation - Belly River Formation boundary, depending on the detail of the cross-section. The difference between the present distance of the measured section from the reference point and the original distance of the measured section from the reference point was taken as the amount of shortening due to thrusting. The reconstructed positions of the measured sections are shown in Figure 2-3.

KEY TO ROSE DIAGRAMS

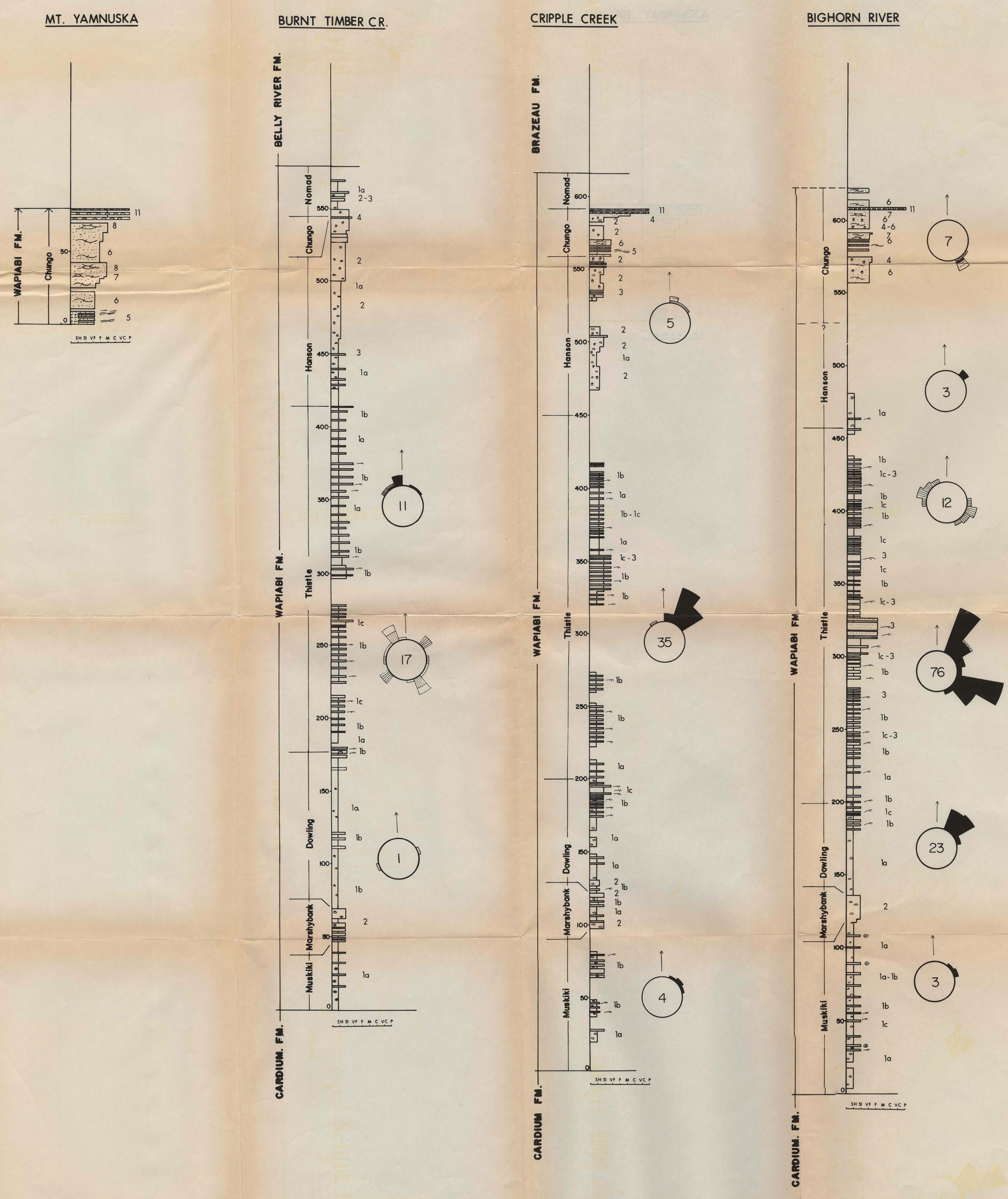
- ASYMMETRICAL RIPPLES
- SYMMETRICAL RIPPLE CREST TRENDS
- FLUTES
- TROUGH CROSS-BEDDING
- LINEATIONS
- RIB & FURROW

KEY TO MEASURED SECTIONS

- COAL
- BIOTURBATION
- SANDSTONE
- PEBBLE CONG.
- HUMMOCKY X-STRATIFICATION
- SWALEY X-STRATIFICATION
- TROUGH X-BEDDING
- RIPPLE MARKS
- SIDERITE CONCRETIONS
- AMMONITES



BLACKSTONE RIVER
THISTLE CREEK
MCLEOD RIVER
LITTLE BERLAND RIVER
BRAZEAU FM.
WAPIABI FM.
CARDIUM FM.
Nomad
Chungo
Hanson
Thistle
Dowling
Marshybank
Muskiki
Bad Heart
SH SI VF F M C VC P
29
27
19
17
1
79
4
5
13
11
6
9


MT. YAMNUSKA
BURNT TIMBER CR.
CRIPPLE CREEK
BIGHORN RIVER
WAPIABI FM.
BELLY RIVER FM.
BRAZEAU FM.
CARDIUM FM.
Nomad
Chungo
Hanson
Thistle
Dowling
Marshybank
Muskiki
SH SI VF F M C VC P
11
17
5
35
4
7
3
12
76
23