

Sedimentology of the Swift Formation (Jurassic)
in the Little Rocky Mountains of Montana

A Thesis

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Department of Geological Sciences
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By

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ABSTRACT

The marine strata of the Swift Formation (Upper Callovian-Oxfordian) are widely distributed and well exposed in the Little Rocky Mountains of north-central Montana. The contact between the Swift and the underlying marine Rierdon Formation is sharp, whereas the upper contact with the non-marine Morrison Formation is gradational. The Swift Formation is about 30 m to 50 m thick and is divided into two members: a lower shale and an upper sandstone.

Detailed sedimentological analysis defined six facies; three in each member. The shale member contains a conglomerate facies (Facies A), a shale-siltstone facies (Facies B), and a bioclastic limestone facies (Facies C). The facies of the sandstone member comprise a sandstone-siltstone-shale facies (Facies D), a cross-bedded sandstone facies (Facies E), and a limestone facies (Facies F). The Swift Formation forms a coarsening-upward sequence from mud to sand-silt-mud intercalations to sand, which has been interpreted by other people as a progradational sequence across a shelf.

The Rierdon-Swift contact is a disconformity spanning three ammonite zones. The whole section of the Swift Formation is considered to be a shallow marine shelf deposit that formed in the course of a transgressive-regressive episode during Late Callovian-Oxfordian time. Facies A was

produced by the reworking of sediment by waves in a near-shore setting during the early stage of the transgressive sea. Facies B was deposited from suspension in relatively deep, open, marine waters during the maximum expansion of the Oxfordian sea. Facies C was formed by the winnowing effect of frequent storm-generated waves, reworking the muddy platform deposits of Facies B. Facies D and E form a continuous regressive sequence that was deposited in a storm-dominated, lower shoreface environment. Facies F was deposited in a shallow, relatively protected setting.

The depositional model proposed for the Swift Formation in the study area is one of a shifting pattern of sedimentation in a shallow marine setting, where inner shelf deposits passed transitionally into lower shoreface deposits; these, in turn, gave way to middle-to-upper shoreface sediments.

The sea-level changes during the deposition of the Swift Formation were as a result of mainly local and regional tectonism; eustatic factors, if any, were minor.

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1. INTRODUCTION

1.1 Location, physiography and drainage

The Little Rocky Mountains in north-central Montana are located between the Missouri River in the south and the Milk River in the north (Fig. 1). They lie about 100 km south of the USA-Canada International boundary, about 10 km west of Highway 191 (Fig. 1).

They form dome-shaped outcrops rising 600 to 900 metres above the surrounding plains and were produced by Tertiary uplift that exposed Archean, Paleozoic, and Mesozoic rocks. The mountains are dissected by numerous radiating, seasonal streams, which drain onto the surrounding plains (Fig. 3).

1.2 Geological setting

The Swift Formation in the Little Rocky Mountains consists of the uppermost marine Jurassic deposits in the area. It belongs to the Ellis Group which was defined by Cobban (1945) as including, in ascending order, the Sawtooth, Rierdon, and Swift Formations (Fig. 2). In the Little Rocky Mountains, the Jurassic rocks are represented by the Rierdon, Swift, and Morrison Formations. They crop out in the foothills of older Paleozoic rocks (Fig. 3). The Swift Formation is well exposed throughout most of the area. It is underlain by the Rierdon Formation and overlain by the Morrison Formation. The thickness of the Swift Formation ranges from 30 to 50 metres,

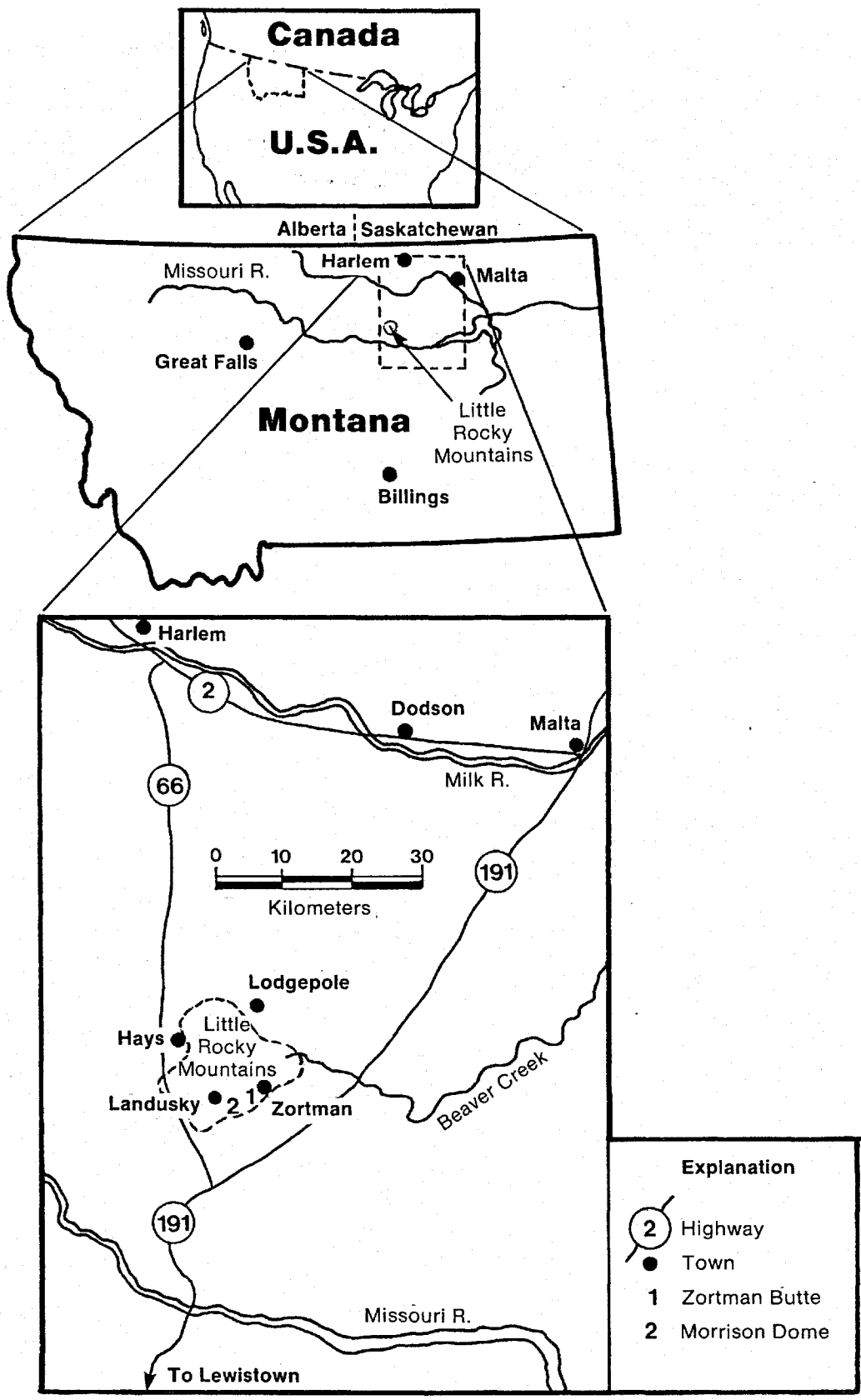


Fig. 1. Location map of the Little Rocky Mountains in north-central Montana.

Middle Jurassic		Upper Jurassic	Series
Bathonian	Callovian		Oxfordian
Ellis Group			Group
			Area
Sawtooth Formation	Rierdon Formation	Swift Formation	N.W. Montana
	Rierdon Formation	Swift Formation	Little Rocky Mountains
Piper Formation	Rierdon Formation	Swift Formation	N.E. Montana

Fig. 2. Marine Jurassic rocks (Ellis Group) in northern Montana (modified from Imlay, 1980; Peterson, 1957; Hayes, 1983; Poulton, 1984; Shurr et al., 1989).

with a maximum thickness of about 60 metres (Imlay, 1982, p. 3). The sediments of the Swift Formation were deposited during the maximum Jurassic transgression across the Western Interior of North America (Shurr et al., 1989, p. 75).

1.3 Previous work

Weed and Pirsson (1896) treated the Jurassic rocks of the Little Rocky Mountains area as one undifferentiated unit, but Imlay (1948, 1982) suggested that the Swift Formation was separated from the underlying Rierdon Formation by a disconformity which spanned at least three ammonite zones. Knechtel (1959) divided the Swift Formation informally into two members: a lower shale and an upper sandstone. Brooke and Braun (1972), after studying micropaleontology of the marine Jurassic rocks of the area, established two assemblage zones within the Swift Formation. Hayes (1983) studied the sedimentology of the Swift Formation in north-central Montana, including the Little Rocky Mountains, and southern Alberta.

1.4 Purpose of the study and methodology

The aims of this thesis are to study the sedimentological characteristics of the Swift Formation in the Little Rocky Mountains, and to reconstruct its depositional environment. The marine Oxfordian rocks of the Western Interior of North America are well constrained in many areas, both litho- and

biostratigraphically; their stratigraphic relationships are established over these areas, and their depositional environments are adequately understood (for references see Imlay, 1980, 1982; Poulton, 1984; section 2.2). Up to date, no detailed sedimentological analysis has been conducted in the study area in relation to ongoing and changing concepts of shelf-sedimentation, physical processes and depths. The present study presents a detailed facies analysis which contributes to the understanding of the marine Oxfordian rocks in the Little Rocky Mountains.

The fieldwork was conducted during two short trips in May and September of 1987. The outcrops of the Swift Formation provide good access for section measurements. Eight stratigraphic sections were measured in the Zortman and Morrison Dome areas (Fig. 3). The Jacob's staff method described by Compton (1985, p. 229) was used in measuring the sections. Each section was divided into units based on lithology, sedimentary structures and textures, and fossils. Additional sedimentological data were obtained from outcrops exposed in several other places across the area.

2. REGIONAL OVERVIEW

2.1 Paleostructural Setting

The Swift Formation and its equivalents in the Western Interior of North America were deposited in a partially enclosed, epeiric seaway (Oxfordian or Sundance sea) which extended southward from the Canadian Arctic as far as northern New Mexico (Fig. 4). These rocks record the events of a third and final major Jurassic transgressive-regressive cycle across the Western Interior of North America (Peterson, 1972, p. 177, 1986, p. 75; Shurr et al., 1989). The paleotectonic setting of the Oxfordian sea has been discussed in several publications (e.g., Peterson, 1954, 1957, Brooke & Braun, 1972; Brenner & Davies 1973, 1974; Hayes 1983). This sea covered a greater area than any other previous Jurassic sea in the region. It was characterized by shallow, shelf-like marine conditions which were comparable, in many ways, to conditions prevailing in modern shallow continental shelves (e.g., Brenner, 1980; Nelson, 1982; Swift, 1985).

Within the seaway, there were some structural elements that influenced the pattern of sedimentation. The most prominent were the Williston Basin, the Alberta Trough, the Twin Creek Trough, the Hardin Trough, the Sweetgrass Arch-Belt Island uplift, and the Sheridan Arch (Fig. 5). Pertinent to the sedimentation in north-central Montana during that interval were the Williston Basin and the Sweetgrass Arch-

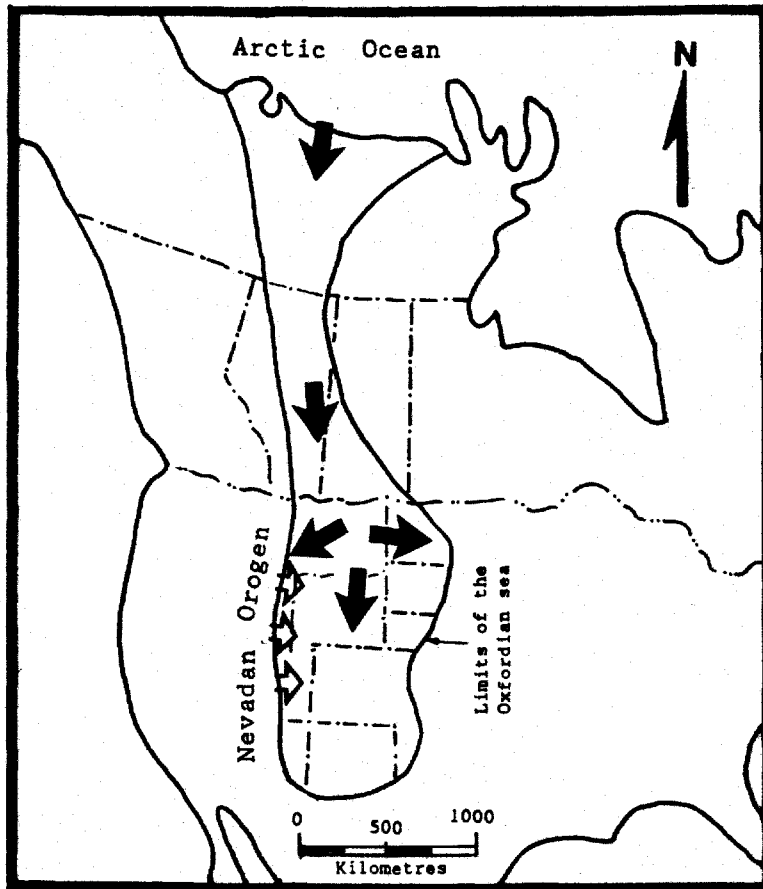


Fig. 4. Paleogeography of the Oxfordian sea in the Western Interior of North America. Modified from several sources, including Imlay (1957); Brenner (1983); Brenner and Davies (1974). Black arrows show direction of advancing sea, and blank arrows indicate direction of clastic sediments shed onto and dispersed across the shelf.

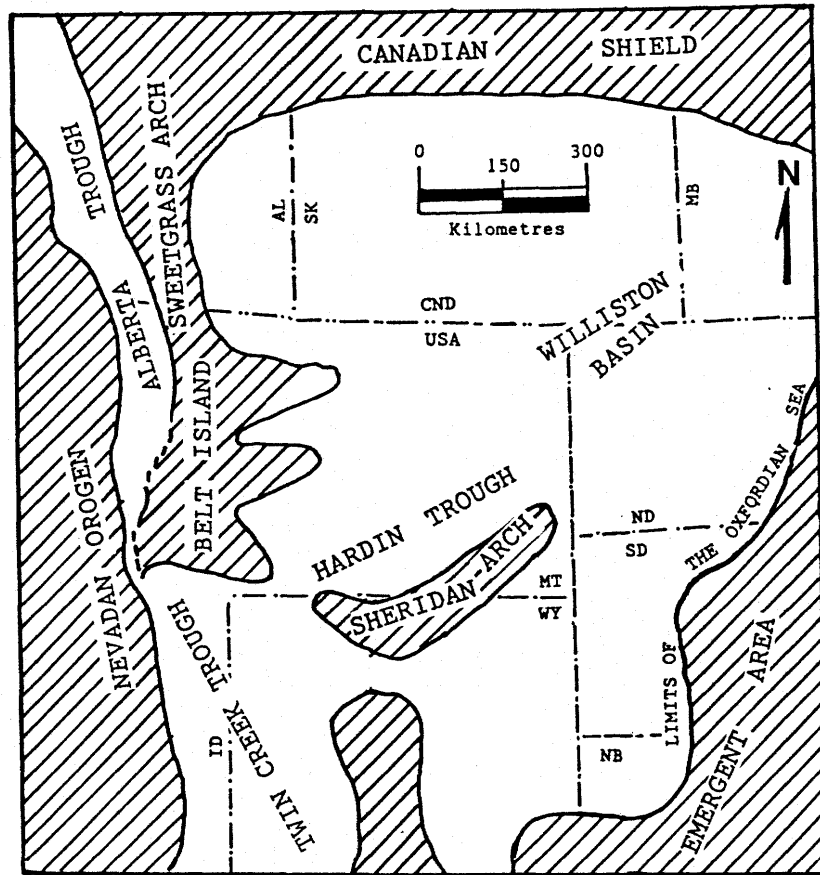


Fig. 5. Paleotectonic setting of the Western Interior of North America during Jurassic time (modified from Peterson, 1957; Brenner, 1983; Hayes, 1983).

Belt Island complex. The Williston Basin was a subsiding area, located eastward of the Little Rocky Mountains area, which received sediments throughout its geologic history. The Sweetgrass Arch, an uplift or structural high, was gently subsiding and then completely submerged during the time of the highest stand of the Oxfordian sea (Imlay, 1980; Hayes, 1983). The Belt Island trend of central Montana was envisaged by Peterson (1957, p. 403) as a broad submarine swell for most of the Upper Jurassic time. It rarely, if ever, formed an island that could have contributed much clastic sediment to the sea. Instead, it strongly affected the pattern of sedimentation by controlling regional conditions involving marine current systems, water salinity, and faunal distribution. Recently, Porter (1989) suggested that the Belt Island may have acted as a local source area during the regressive phase of the Oxfordian sea. In north-central Montana, the Sweetgrass Arch may have played a similar role at that time.

2.2 Stratigraphic correlation and literature review

The regional correlation of the Swift Formation and its equivalents in the Western Interior of North America is shown in Table 1. The Swift Formation is the youngest formation of the marine Ellis Group of Cobban (1945). It was named and defined at the Swift Reservoir, northwestern Montana. At its type locality the Swift Formation was divided into two

Callovian	Oxfordian	Stage / Area
Fernie Formation		S.W. Alberta- S.E. BC
Green Beds	Passage Beds	Gray Beds
Vanguard Group		S.E. Alberta- S.W. Manitoba
Masefield Formation	Roseray Formation	
Ellis Group		Montana
Rierdon Formation	Swift Formation	
Stump Sandstone		E. Idaho
Limestone	sandstone & siltstone	
Sundance Formation		W.C. Wyoming
L. Sundance Formation	U. Sundance Formation	
Sundance Formation		E. Wyoming & W. S. Dakota
L. Sundance Formation	Redwater Shale Member	
Stump Sandstone		N.E. Utah
Curtis Formation	Redwater Member	
Sundance Formation		S. Utah & N. Colorado
L. Sundance Formation	Redwater Shale Member	

Table 1. Nomenclature and stratigraphic correlation of the marine Upper Jurassic rocks in the western interior of North America (modified from Imlay, 1980; Hayes, 1983; Poulton, 1984).

gradational members: a lower shale and an upper sandstone. The lower shale member is composed of fissile shale that is highly glauconitic at the base. The upper sandstone member is composed predominantly of fine-grained sandstone, also highly glauconitic and with abundant micaceous shale partings. The term "Swift" has since been applied to similar strata across Montana and adjoining areas of the United States and Canada (Weir, 1949; Moritz, 1951; Imlay, 1956; Peterson, 1954; Francis, 1957; Mudge, 1972).

The regional stratigraphic correlation of the marine Oxfordian rocks in the Western Interior of the United States was established by Imlay (1952, 1957, 1980, 1982). Peterson (1957, 1972) also contributed significantly to the understanding of the Oxfordian sedimentation in this region. The Oxfordian rocks of Wyoming and adjoining area were studied by Imlay (1947), Peterson (1954), Pipringos (1968), and Wright (1973). In the Williston Basin area, the Swift Formation was studied by Francis (1957). Poulton (1988) correlated the major interregional events during the Jurassic period in the Western Interior of North America. a detailed account of paleotectonism and depositional environments in Williston Basin and vicinity during Late Jurassic time was given by Shur et al. (1989).

Carlson (1968) correlated the Upper Jurassic (Oxfordian) rocks of Montana and North Dakota with their Canadian equivalents. Recently, Poulton (1984) published results of a

detailed study on the Jurassic marine sedimentation in the Canadian western interior. Other information in the Oxfordian rocks of the western interior of Canada are included in studies by Hamblin and Walker (1979), Hall (1984), Marion (1984) and Stronach (1984).

Regional sedimentological analyses of the Upper Jurassic rocks in the Western Interior region were made by Brenner and Davies (1973, 1974), Specht and Brenner (1979), Brenner et al. (1985) and Uhlir et al. (1988). Brenner (1980) established a regional process-response depositional model for the Oxfordian rocks of the Western Interior. Review of the depositional environments of the Oxfordian rocks of Montana and Wyoming is given by Bouma et al. (1982). Porter (1989) interpreted the depositional environment of the Swift Formation of western Montana in terms of sea-level changes and local tectonics.

2.3 Geological history

The geological history of the Upper Jurassic (Oxfordian) rocks in the Western Interior region has been studied by several researchers (e.g., Cobban, 1945, p. 1287-1291; Imlay, 1947, 1956, 1980; Peterson, 1954, 1957; Hayes, 1983). Significant changes in Swift-age deposits mark a turning point in the sedimentary history of the Western Interior North America (Peterson, 1957). These changes include the end of carbonate deposition, the beginning of a rising area to

the west -- due to rejuvenation of the Nevadan orogen, and a change in faunal assemblages, characterized by a concomitant southward spread of belemnites and decrease in Gryphaea assemblages (Brenner, 1983; Peterson, 1986, p. 77; Shurr et al., 1989). The rising area released and dispersed clastic sediments into the shelf to the east throughout the remainder of the Mesozoic Era.

At the close of deposition of the Rierdon sediments, the sea withdrew northwestward owing to both eustatic and regional tectonic factors (Imlay, 1957, p. 475, 1980, Hayes, 1983; Peterson, 1954, p. 502; Hallam, 1978). Subsequently, erosion took place across most parts of the Western Interior region, causing parts of the pre-Swift strata to be partially or totally removed (Cobban, 1945; Imlay, 1947; Peterson, 1954).

Later, during early Late Callovian time, the sea entered north-central Montana and the Williston Basin area from the northwest, probably across the Alberta Trough (Figs 4 & 5). The exact route of entrance has been a matter of conjecture. Brooke and Braun (1972, p. 21) questioned the hypothesis of an Arctic or Alaskan source for the latest Callovian transgression across the craton; instead, they suggested that there may have been a direct incursion from the west. However, several lines of evidence drawn from more recent studies contradict this theory. For example, the occurrence of the oldest basal Swift beds in north-central Montana, and

the presence of younger beds of the Swift Formation in western and northwestern Montana, tend to diminish the likelihood of a relationship of the late Callovian incursion to the Pacific ocean (Imlay, 1982; Hayes, 1983). Also, paleontological data, especially ammonites, strongly suggest a boreal origin rather than a Pacific link (Poulton, 1984, p. 20; Callomon, 1984, p. 154). Furthermore, the Belt Island, then an emergent element, and the rising western source area would have made it difficult for such a link to occur.

No matter where the transgression originated, evidence suggests that the sea flooded the north-central Montana and Williston Basin area in early Late Callovian time, after having breached the low emergent Sweetgrass Arch. The exact position of this breach is also dubious. Whereas Imlay (1980) considered that the break in the arch was located in southern Alberta, Hayes (1983) argued that the sea breached the arch about 80 kilometres south of the USA-Canada border. Both faunal and lithologic evidence indicate that the Early Oxfordian sea did not spread further south, west, or east beyond northwestern Colorado, the eastern side of the Sweetgrass Arch, and South Dakota respectively (Imlay, 1948, 1980).

As the transgression continued through Early to early Middle Oxfordian, a thick section of mud with intercalations of silt and sand was deposited over vast areas of the Western Interior. In the central parts of the sea, sedimentation

continued under normal marine conditions --far away from any environmental stresses such as high salinity, shoaling, or clastic influx. This interpretation is supported by the association of open marine fauna within the lower parts of the Swift-age rocks, which occur over large areas both in the USA and in Canada (Arkell, 1956; Imlay, 1982; Hall, 1984, pp. 240, 246). During Middle Oxfordian times, the sea attained its maximum lateral extent (Fig. 5). At that time, near-shore siliciclastic deposits, mainly sands, were laid down on the western edge of the shelf, whereas mud and silt dominated the central part and calcareous mud accumulated farther east in the Williston Basin (Brenner and Davies, 1974). By the end of the Oxfordian time, the sea withdrew northwards leaving the region under non-marine sedimentation.

3. GENERAL CHARACTERISTICS AND AGE

3.1 Lithology and distribution

The Swift Formation in the study area consists of a sequence of alternating shales, siltstones, and sandstones, with variable thin beds of limestone, limestone concretions, and coquinas which are more common in the lower portion. Its base is commonly marked by thin beds of water-worn belemnite and quartz pebble conglomerate or brecciated limestone beds (Fig. 6). The Swift Formation can be divided into two gradational members: a lower shale and an upper sandstone.

The lower shale member is poorly exposed, in most places forming muddy lower-angle slopes littered with belemnite guards and selenite crystals (Fig. 7). Its thickness varies greatly from one place to another, ranging from 15 to 25 metres and averaging about 17 metres. Locally, the lower shale member has been partially or completely faulted out. The shale member consists predominantly of dark gray, non-calcareous shale and siltstone with thin interbeds of sandstone and limestone.

The upper sandstone member is also widely distributed across the area and mostly crops out as ledge-forming, resistant cliffs on top of the shale member. The thickness of the upper sandstone member is inversely proportional to that of the underlying shale member, being thin where the shale member is thick and vice versa. It consists of a series

Fig. 6. Belemnite/chert pebble conglomerate near the base of the Swift Formation, measured section no. 1, eastern side of Zortman Butte -- Dump area. This photo was taken at a locality in the lower centre of Fig. 7.

Fig. 7. Outcrop of the Swift Formation showing the shale member, eastern side of Zortman Butte. Looking north and upsection from a cliff in the Rierdon Formation. Notebook at the upper centre is 15x22 cm.



of fine- to medium-grained sandstones interbedded with variable thicknesses of siltstone, silty mudstone and shale.

3.2 Fossils

The Swift Formation is fossiliferous throughout its vertical and lateral extent, but the abundance of fossils decreases up-section. Belemnite fossils are extremely abundant, but decrease towards the top of the shale member. They are rarely found in the upper sandstone member. These fossils were identified by Santon (in Weed and Pirsson, 1896, p. 408) as Belemnites densus Meek and Hayden. Ammonite fossils occur within the lower ten to fifteen metres of the section and have been identified at various localities within the area (Imlay, 1948, 1982). Oyster and other bivalve fossils are also present, but are not so abundant as the belemnites. Marine microfossils are also common within the lower part of the Swift Formation; these include marine Ostracoda and Foraminifera (Peterson, 1954; Brooke and Braun, 1972; for a more detailed analysis see the following section).

3.3 Biostratigraphy and age of the Swift Formation

Imlay (1948) designated two ammonite zones within the shale member of the Swift Formation of the Little Rocky and Bearpaw Mountains (see section 4.2): the Quenstedtoceras collieri and the Cardioceras cordiforme zones.

The first zone is characterized by two species of Quenstedtoceras, one of Prosphinctes, and two of Cardioceras together with several pelecypod fauna. This zone has not been recognized anywhere in the Western Interior region other than at these two localities (Peterson, 1954; Imlay, 1956, 1982, p. 26). Imlay (1948) explains the sparse occurrence of the Q. collieri zone as follows:

"Failure to find the zone elsewhere may be explained in part by the sandy character of beds at that stratigraphic position in many parts of the Western Interior [especially, those on the western margin of the Oxfordian sea]. Also in places the zone may not be represented by sediments, as at Red Dome on the west side of the Pryor Mountains in south-central Montana, where sandstone containing Cardioceras rests directly on the Rierdon Formation."

The Q. collieri zone has been correlated with the upper part of the Q. lamberti and the lower part of the Q. mariae zones of Europe (i.e. latest Callovian-earliest Oxfordian).

The Q. lamberti zone may be identified over vast areas in the region (see Imlay, 1947, 1956). In the Little Rocky Mountains, the C. cordiforme zone occurs directly above the Q. collieri zone (Imlay, 1982). The C. cordiforme zone seems to correspond to the upper part of the Q. mariae and the lower part of the C. cordatum zones in Europe (i.e. Early Oxfordian).

Peterson (1954) established two ostracode subzones within the Upper Jurassic (Oxfordian) rocks of Montana and Wyoming and compared them to the ammonite zones of Imlay (1948), as

follows:

1. The Prognocythere subzone appears to correspond to the Q. collieri and C. cordiforme of Imlay (1948), and,
- 2) The Leptocythere imlayi-Cytherura lanceolata subzone, was found to occur within the 3-6 metres interval above the base of the Swift Formation in the Little Rocky Mountains. Stratigraphically, this subzone overlies the C. cordiforme ammonite zone.

Brooke and Braun (1972, pp. 13-14) established two microfaunal assemblage zones, identified in ascending order as (VI and VII), within the Swift Formation of the Little Rocky Mountains and in southern Saskatchewan (see Fig. 11 of section 4.2, p. 28).

The first assemblage zone, VI, occurs within the lower portion of the shale member. It is characterized by abundance of nodosariids, and to a lesser extent, agglutinated lituolids -- foraminiferal species with a few ostracode species.

The second assemblage zone, VII, was recognized within the upper part of the shale member in a sequence of shale, siltstone and sandstone. It is characterized by a predominance of agglutinated foraminifera (Haplophragmoides cf. linki) and a scarcity of nodosariid foraminifera. Unidentifiable ostracode and moulds and fragments of calcareous foraminifera are also present, but not in the same abundance as agglutinated species.

The upper sandstone member has not furnished any ammonites, which makes its age difficult to establish. A pelecypod fossil, Buchia concentrica (Sowerby), was recovered from the basal part of the Swift Formation in northwestern Montana (Mudge, 1972). This occurrence is of stratigraphic importance, because the B. concentrica did not appear anywhere in the world before late Oxfordian time (Imlay in Mudge, 1972, p. A48). Hall (1984, p. 246) also discovered B. concentrica in the Green Beds of the Fernie Formation in Alberta, stratigraphically equivalent to the shale member of the Swift Formation.

The dating of the Oxfordian rocks in the region is based mainly on ammonite biostratigraphy established by Imlay (1948, 1982). The basal beds vary remarkably in age from one place to another. The oldest beds of the Swift Formation, dated as latest Callovian, have been encountered only in north-central Montana (Little Rocky Mountains and Bearpaw Mountains). In other places across the region, including the type locality in northwestern Montana (e.g., Cobban, 1945; Mudge, 1972), the oldest beds of the Swift have furnished ammonites of Early to early Middle Oxfordian time. This implies that the Swift section exposed in north-central Montana is more complete than any other Oxfordian section exposed elsewhere in the Western Interior; it also suggests that the area was flooded earlier than elsewhere.

4. RIERDON-SWIFT BOUNDARY

4.1 Description

The basal contact of the Swift Formation with the underlying Rierdon Formation in the study area is everywhere sharp, it can be recognized easily and traced laterally over considerable distances. The contact is characterized by three conspicuous features: a change in lithology from micritic limestone above the Rierdon Formation to thin, patchy conglomeratic beds at the base of the Swift Formation; a bored and/or brecciated surface above the Rierdon Formation; and a change in marine fauna.

Below the contact, the topmost beds of the Rierdon Formation are normally light gray and consist of thinly-bedded argillaceous limestone containing abundant Gryphaea. This limestone is widely exposed across the area and forms highly resistant outcrops. In contrast, the basal beds of the Swift Formation are dark and consist predominantly of water-worn belemnite conglomerate (Fig. 8a) with some chert or quartz pebbles, greenish yellow sandstone beds, or limestone conglomerate (8b). The belemnite conglomerate overlies bored surfaces, whereas the limestone conglomerate overlies brecciated surfaces on top of the micritic limestone of the Rierdon Formation (Figs. 9a,b respectively).

At some localities, e.g. Morrison Dome, the uppermost limestone of the Rierdon Formation is highly brecciated; in

8a



8b



Fig. 8: a) Belemnite/chert pebble conglomerate at the base of the Swift Formation (measured section no. 1, Fig. 12).

b) Sample from a limestone conglomerate at the base the Swift Formation (base of measured section no. 6, Fig. 12)

9a



9b



Fig. 9: a) Uppermost micritic limestone of the Rierdon Formation with borings.

b) Brecciated surface on top of the Rierdon Formation. Both photos were taken at Morrison Dome.

some places, the breccia is incorporated in the basal part of the overlying Swift Formation (Fig. 9, section 5.1.1). At other localities, the contact is characterized by presence of biogenic sedimentary structures. Borings (cf. Ekdale *et al.*, 1984, p.302) on top of the micritic limestone of the Rierdon Formation are most abundant structures and vary remarkably in shape and size; they may be pot-like or spherical, or cylindrical in shape (Fig. 10). Most of them are usually perpendicular or highly inclined to the bedding plane, ranging in diameter from a few millimetres to more than one centimetre, and their length may be more than three centimetres. Some borings are filled with very fine to silty sand of the Swift Formation; others contain bivalve fossils and dark brown silty sediments. Water-worn belemnite fragments, chert and quartz grains may be present in the borings. Some borings are iron-stained and have dark brown, rusty dendritic features on the walls. At certain localities (e.g., Morrison Dome), iron-staining may be seen on the surface on top of the Rierdon limestone. No karstification, such as that noted in the underlying Mission Canyon Limestone, was recognized in the surface of the Rierdon Formation.

4.2 Interpretation

The Rierdon-Swift boundary is interpreted as a disconformity. Imlay (1948, 1982), after studying the



Fig. 10: a) Sample of a bored, micritic limestone of the Rierdon Formation, b) cross section of the same sample through one of the boring (outlined) near the number 30. Sample taken at a locality near the base of measured section no. 6 of Fig. 12 (in pocket).

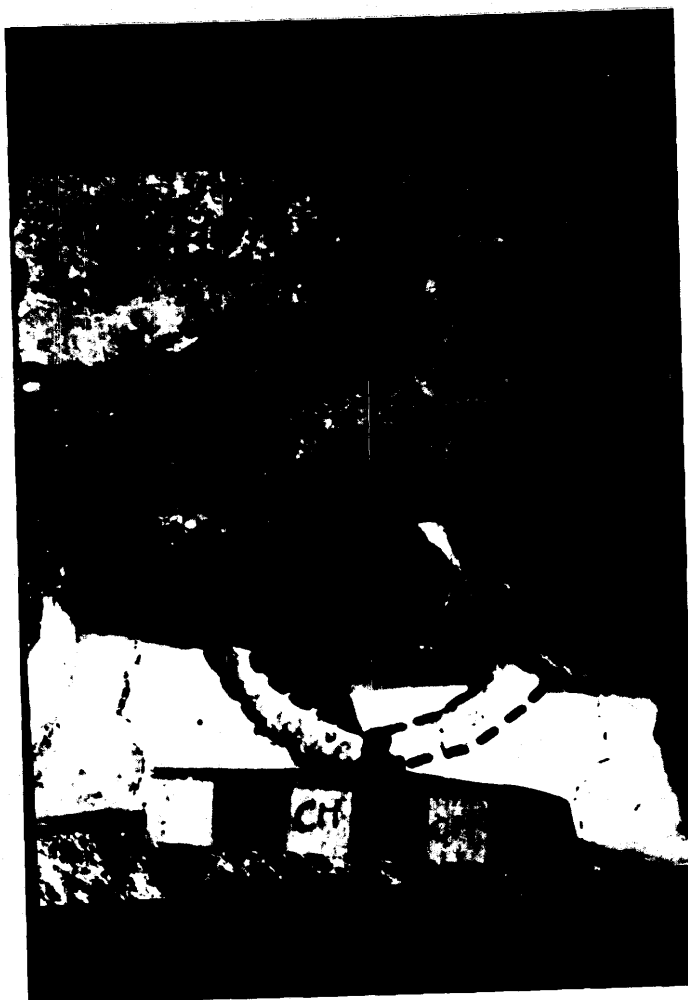


Fig. 10: a) Sample of a bored, micritic limestone of the Rierdon Formation, b) cross section of the same sample through one of the boring (outlined) near the number 30. Sample taken at a locality near the base of measured section no. 6 of Fig. 12 (in pocket).

biostratigraphy of the Little Rocky Mountains area, noted that three ammonite zones are missing between the Swift and the underlying Rierdon Formation (Fig. 11). After correlating the ammonite zones in the area to their standard counterparts in Europe, and after determining the age of upper part of the Rierdon Formation and the lower part of the Swift Formation, Imlay (1948, p. 17) concluded that

"[the] age determination means that in the Little Rocky Mountains and in the Bearpaw Mountains the Swift Formation is separated from the underlying Rierdon Formation by an unconformity representing most of the Middle and Upper Callovian and corresponding to the European zones of Cosmoceras [Kosmoceras] jason, Erymnoceras coronatum, and Peltoceras athleta."

The disconformity between the Swift and Rierdon Formations and their equivalents has also been reported in various other sites across the Western Interior of North America (Imlay, 1947, 1956; Peterson, 1957; Mudge, 1972; Pipringos, 1968; Poulton, 1984; Porter, 1988). The Rierdon-Swift boundary represents the "J-4" unconformity of Pipringos and O'Sullivan's (1978), who studied the Triassic and Jurassic unconformities in the Western Interior of the United States.

The sedimentological features discussed in the present study further support the previous paleontological evidence considered by Imlay (1948, 1982) to establish the disconformable character of the Rierdon-Swift boundary. The brecciated surface on top of the Rierdon Formation, and the overlying thin limestone conglomerate at the base of the

Age (M.Y.)	Series	Stages	Ammonite Zones		Assemblage Zones	Formation	Lithology
			European Zones	W. Interior Zones			
150	U. Jurassic	Oxfordian	<i>Quenstedtoceras mariae</i>	<i>Quenstedtoceras collier</i>	VII	Swift Form.	
	Jurassic	Callovian	<i>Quenstedtoceras lamberti</i>		VI		
<i>Peltoceras athleta</i>							
<i>Erynoceras coronatum</i>							
			<i>Kosmoceras jason</i>				
			<i>Sigaloceras calloviense</i>				
	Middle	Bathonian		<i>Kepplerites mclearnii</i>	V	Rierdon Form.	
					IV		

Fig. 11. Generalized figure to show Rierdon-Swift disconformity in terms of newest Interpretations of biostratigraphy. Ammonite zones are adopted from Imlay (1948, 1982) and Poulton (1984), and microfaunal assemblage zones after Brooke and Braun (1972). Age is based on Van Hinte (1976) and Haq et al. (1987).

Swift Formation, indicate that there was a break in sedimentation during which the Rierdon Formation was subject to erosion.

The occurrence of conglomerates at the base of marine deposits has been considered by several authors to be an acceptable criterion for recognizing unconformities (Krumbein and Sloss, 1963; Davies, 1983; Kidwell, 1984, p. 42). However, the absence of conspicuous topographic irregularities in the contact, the constant thickness of the Rierdon Formation across the area, and the thinness of the conglomerate above the disconformity imply that the uplift and erosion of the Rierdon Formation was insignificant. Moreover, this may indicate that the Little Rocky Mountains area was tectonically more stable during the time of development of the unconformity, as compared to the regions to the west, southwest, or south. In the latter areas, the uplift and erosion led to partial or total removal of the Rierdon Formation (cf. Cobban, 1945; Imlay, 1947). The absence of much incision at the disconformity surface implies that the emergent surface lay close to sea-level base for most of the unconformity period.

The bored surface underlying the belemnite/chert conglomerate is interpreted as a hardground (cf. Ekdale et al., 1984) formed before the deposition of the Swift sediments. The borings are similar to the Trypanites or Glossifungites ichnofacies (cf. Frey and Seilacher, 1980) and

were probably made by bivalves and worm-like organisms. The Trypanites and Glossifungites ichnofacies are commonly associated with disconformities or hardground (Frey and Pemberton, 1984; Ekdale et al., 1984). Pemberton et al. (1980) described similar Trypanites ichnofacies from a Silurian-Devonian disconformity in southern Ontario. They suggested that the Silurian-Devonian disconformity and its associated Trypanites trace fossils were developed during a regressive phase before the deposition of the overlying Devonian sediments. Other similarly bored surfaces in different places in the world have likewise been interpreted as disconformities (e.g., Purser, 1969; Kidwell, 1984).

The iron-staining on the walls of some borings may indicate that oxidation took place during the exposure of the Rierdon Formation. The staining was probably associated with an unconsolidated soil horizon, which might have been removed by waves and currents during deposition of the basal conglomerate of the Swift Formation. As Weller (1960) and Davies (1983) have pointed out, at some unconformities the soil profile is washed out during transgression and deposition of overlying sediments, an iron oxide-stained surface being the only preserved evidence of weathering criterion.

Based on the mechanism of interpolation used by Van Hinte (1976) and adopted by Kent and Gradstein (1985), the time represented by the Rierdon-Swift disconformity can be

estimated. The interpolation hypothesis assumes equal duration of biozones. According to the ammonite zonation suggested by Hallam (1975, 1978), the Callovian stage is represented by six ammonite zones. Each ammonite zone is given an age range of 1.0 Ma by Van Hinte (1976) and 1.04 by Kent and Gradstein (1985). Taking the average, each ammonite zone can be assigned to 1.02 Ma. The Callovian stage has been given a duration of 7.0 Ma. by Kent and Gradstein (1985).

Using these arguments, the duration of the Rierdon-Swift disconformity in the present study area is estimated as 3.06 Ma. This does not take into account local tectonic effect during that time.

5. SEDIMENTOLOGY

Based on gross lithology, textures, and primary and biogenic sedimentary structures, the Swift Formation in the present study area has been divided into six facies (Table 2):

- 1) Facies A: conglomerate facies;
- 2) Facies B: shale-siltstone facies;
- 3) Facies C: bioclastic limestone (coquina) facies;
- 4) Facies D: sandstone-siltstone-shale facies;
- 5) Facies E: cross-stratified sandstone facies; and,
- 6) Facies F: limestone facies.

The following is description and interpretation of each facies.

5.1 Facies A: conglomerate facies

5.1.1 Description

Facies A is represented by relatively thin (15 to 20 centimetres) conglomeratic beds which mark the base of the Swift Formation. Two types of conglomerate were recognized: belemnite/chert-pebble conglomerates and limestone conglomerates.

The belemnite/chert-pebble conglomerate has been identified in sections exposed in the Zortman Butte area (section 1 and 3 of Fig. 12 in pocket). In this area, the conglomerate forms thin (10 to 30, average 20 centimetres), discontinuous beds at the base of the Swift section or

Table 2: Characteristic features of the different facies.

Facies	Lithology	Sedimentary structures	Trace fossils	Body fossils
F	limestone, siltstone, and calcareous siltstone	small-scale ripple cross-lamination in siltstone and calcareous siltstone	horizontal small burrows are present in siltstones	absent or very rare
E	sandstone with some siltstone and shale partings	SCS, various types of wave- and current-generated cross-bedding	absent or very rare	very rare belemnite fossils in siltstone and shale
D	interbedded fine-grained sandstone, siltstone and shale	HCS, climbing ripple cross-lamination, horizontal and low-angle cross-lamination	common, dominated by horizontal burrows, mainly <u>Planolites</u> sp.	belemnite fossils common, some bivalve moulds
C	bioclastic limestone "coquinas"	absent	absent	<u>oyster shells</u> , mainly <u>Gryphaea</u>
B	shale and siltstone with a few sandstone and limestone layers or thin beds	absent in shale, but different types of cross-lamination occur in siltstone, sandstone, and limestone	siltstone is bioturbated locally, few U-shaped burrows may occur in sandstone	belemnite fossils occur in profusion, some pelecypods also occur
A	belemnite/chert and limestone conglomerates	absent or very rare	borings are common (mainly <u>Trypanites-Glossfungites</u>)	dominated by water-worn belemnites

slightly above it. The beds are poorly exposed at the surface and laterally discontinuous. In most places, these beds are overlain by thick, dark-gray shale. A single bed cannot be traced laterally for more than a few metres before it pinches out. Where well exposed, the conglomerate beds are seen to be commonly underlain by bored surfaces of micritic limestone of the Rierdon Formation. At some localities, the conglomerate is overlain or underlain by a thin bed of yellowish green, fine- to medium-grained, well-sorted sandstone. This sandstone closely resembles the basal glauconitic sandstone of the Swift Formation in northwestern Montana (Mudge, 1972, p. A48).

The belemnites/chert-pebble conglomerate is varicoloured; being limonitic, yellowish gray or green, or dark brown (Figs. 8a). The conglomerate is composed predominantly of belemnite fragments (up to 60%), chert or quartz pebbles (15%) and limestone clasts with subordinate clay clasts (10%), embedded in a yellowish brown matrix of highly calcareous clay and silt, sand-sized quartz grains, mica flakes and comminuted shell fragments.

The belemnite fragments are highly calcified, pyritized, and mostly water-worn. They vary in size from less than two millimetres to more than 0.5 centimetres in diameter, and measure up to 7 centimetres in length (see Fig. 6). The chert and quartz pebbles are tan-coloured, rounded to sub-rounded, 0.1 to 0.7 centimetres in diameter, and are less

common than the belemnite fragments. The limestone clasts (up to 1.5 centimetre across) are light gray, angular to sub-rounded, and are less common than the chert or quartz pebbles. These clasts are similar in composition and texture to the micritic limestone of the Rierdon Formation. The claystone clasts are subordinate, angular, and measure up to one centimetre across.

The lower contacts of the belemnite/chert conglomerate beds are sharp everywhere, whereas the upper contacts are mostly gradational. Primary sedimentary structures are almost absent, sorting within the fragments is poor, and fabric is ill-defined.

The limestone conglomerate, a poorly indurated, clast-supported conglomerate, was identified only at Morrison Dome, where it forms thin (10 to 15 centimetres thick), patchy, discontinuous beds that overlies highly brecciated surfaces of argillaceous limestone of the Rierdon Formation (Fig. 8b)

The limestone conglomerate is composed predominantly of angular to sub-angular, fine-grained limestone clasts embedded in a sandy to silty calcareous matrix. The limestone clasts vary considerably in size, from a few millimetres to more than 5 centimetres across (Fig. 8b). They are similar in composition and texture to the underlying argillaceous limestone of the Rierdon Formation.

The lower contact of the limestone conglomerate is sharp and abrupt, and is always marked by a brecciated surface.

Both primary and biogenic sedimentary structures are lacking.

5.1.2 Interpretation

Both types of the conglomerates of Facies A are interpreted here as marine lags formed during the early stage of the Oxfordian transgression. They were deposited on a disconformable surface which developed at the top of the Rierdon Formation.

The belemnite/chert-pebble conglomerate is interpreted here as a condensed sequence that developed due to the reduced rate of sedimentation in the wake of the Early Oxfordian transgression (cf. Stronach, 1984, pp. 59-60; Poulton, 1984, p. 20). This interpretation is based on the following lines of evidence:

- 1) the conglomerate bed is a very thin, marine stratigraphic unit;
- 2) it overlies a disconformity;
- 3) the lack of terrigenous sediment; and,
- 4) the occurrence of calcite cement and pyrite, and the presence of a bored surface or hardgrounds.

Many researchers have considered the latter features as evidence for condensation (e.g., Vail et al., 1984, p. 134; Stronach, 1984, pp. 59-60; Haq et al., 1987, p. 1160; Kidwell, 1989). Haq et al. (1987) and Kidwell (1988, 1989, p. 2) pointed out that condensed sequences are composed predominantly of shell concentrations in terrigenous shelves

and record conditions of starved sedimentation and disconformities.

The Little Rocky Mountains area may have been gently sloping and tectonically stable, so that the sea spread quickly over the area and reduced the supply of terrigenous sediment to the shelf (cf. Brookfield, 1970; Rollins and Donahue, 1975, p. 258; Abbott, 1985, p. 158; Miller and Orr, 1988, p. 966). The absence of volcanic rocks and volcano-clastic intercalations attests to tectonic stability (Hallam, 1975).

The limestone conglomerate is interpreted as a transgressive lag derived locally from the underlying Rierdon Formation by strong waves and currents. The wide range in clast size in the limestone conglomerate may indicate either rapid accumulation and burial of fragments or a supply of a wide range of particle sizes with no subsequent reworking. MacCarthy (1987, p. 399) interpreted a similarly thin conglomerate layer as transgressive marine lag, produced by rip currents associated with storm conditions. Other examples of thin basal marine conglomerate facies have been described and interpreted as transgressive deposits (e.g., Bourgeois, 1980; Bose et al. 1988).

5.2 Facies B: shale-siltstone facies

5.2.1 Description

The shale-siltstone (Facies B) is widely distributed

across the area. Facies B, which makes up the lower half of the Swift Formation, is thick (15 to 25 metres thick, averaging 17.5 metres). In the section exposed on the eastern flank of Zortman Butte, Facies B is thick and well exposed, whereas at Morrison Dome it is relatively thin (10 to 15 metres thick) and more silty. Generally, the facies is eroded into lower-angle slopes that are interrupted locally by ledge-forming sandstones and limestones (see Fig. 7).

Facies B consists predominantly of intercalated shale and siltstone interbedded with thin beds or layers of sandstone, sandy limestone, and limestone concretions.

The shale and siltstone account for more than 80% of the total thickness of the facies. The shale is dark gray, fissile, non-calcareous, and fossiliferous. It forms relatively thick units (0.7 to 5 metres thick, averaging 2.5 metres) which form gentle slopes littered with abundant belemnites and selenite crystals. Both primary and biogenic sedimentary structures are absent within the shaly units. The siltstone is light gray, thinly laminated, ripple-cross-laminated (terminology of McKee and Weir, 1953) and bioturbated. It forms thin layers (1 to 10 centimetres thick) that increase in thickness and abundance toward the top of the facies.

The sandstones and limestones (<10% of the total thickness of the facies) commonly form thin, laterally persistent beds 5 to 30 centimetres thick (Fig. 13). They increase in



Fig. 13. Limestone bed within Facies B (unit 10, measured section no. 1, Fig 12). Note the sharp lower contact below hammer handle.

thickness and abundance toward the top of the facies. A single bed can be traced laterally throughout the extent of the outcrop, which commonly extends for several tens of metres, but outcrop-to-outcrop correlation is difficult.

The sandstones and limestones vary in colour from light gray to brown. They are fine- to medium-grained and composed predominantly of quartz and calcite respectively. Other minor constituents include mica and chlorite. Internally, the beds are thinly- and ripple-cross-bedded. Their lower contacts are sharp and slightly erosional. Locally, the contacts are marked with small claystone and siltstone pebbles, giving the bed a graded appearance. The upper contacts are transitional to the overlying siltstone and shale units.

Limestone concretions are of limited occurrence. They are composed mostly of septarian limestones, which are commonly covered by shaly units.

Facies B is generally fossiliferous, though the abundance of fossils decreases upwards through the section. Both macro- and microfossils are abundant. Belemnites are dominant within the shale and siltstone units, but occur rarely within the sandstone and limestone beds. Ammonites are relatively rare, though they have been reported from the lower 6 to 9 metres of the facies (Knechtel, 1959, p. 736; Imlay, 1982). Bivalves Gryphaea nebrascensis (Meek and Hayden) and others are also present and in places form thin coquina beds (see section 5.3). Brooke and Braun (1972) identified abundant

microfossils (foraminifera and ostracoda) within the shale and siltstone units.

5.2.2 Interpretation

The suite of rock types, sedimentary structures, and faunal assemblages indicates that Facies B was deposited in a shallow-water, open-marine inner shelf during the maximum transgression across the area (cf. Brenner and Davies, 1974, their mud facies). The relatively thick shaly units indicate that most deposition was from suspension -the environment was under quiet water conditions for long periods of time. The laminated to ripple-cross-laminated siltstones and thinly planar- to cross-bedded, fine-grained sandstone and limestone stringers represent a relative increase in current and wave strengths in an otherwise quiet environment (Cant, 1980, p. 123; Craft and Bridge, 1987, p. 353).

The sharp-based, relatively thick, calcareous sandstone beds are interpreted as storm-generated deposits, accumulated under foul-weather conditions (terminology of Dott, 1983). The sharp lower contacts of some beds, the local presence of coarse materials at the base of these beds, and the small-scale fining-upward pattern of many beds imply that the sandstones were deposited under waning current conditions (cf. Reineck and Singh, 1972; Banks, 1973). The calcareous sandstone and sandy limestone beds are comparable, in texture, thickness and lateral continuity, to the sandstone

beds described by Brenchley et al. (1979) from the Ordovician epicontinental platform of Norway; the Ordovician sandstone beds also were interpreted as storm-produced deposits.

Facies B resembles the clay-dominated facies described by MacCarthy (1987), who interpreted it as having been deposited from suspension in a shallow-water, marine environment under a reduced rate of sedimentation. Facies B also resembles the shale lithofacies of the Mesaverde Group, interpreted by Swift et al. (1987) as having been deposited in a shallow marine environment. Swift et al. (1987) envisaged an open, middle to inner shelf environment to account for the deposition of the shale lithofacies.

5.3 Facies C: bioclastic limestone (coquina) facies

5.3.1 Description

Facies C, the bioclastic limestone (coquina) facies, is of limited lateral and vertical distribution and does not account for more than one or two per cent of the total thickness of any measured section. This facies occurs exclusively within the lower shaly part (Facies B) of the Swift Formation, where there are coquina beds at various stratigraphic levels. This lower stratigraphic occurrence of the facies is in sharp contrast to the regional position of its equivalents (the coquinoid sandstones) that are exposed elsewhere in the southern part of the region (Brenner and Davies, 1974; Specht and Brenner, 1979).

Facies C is represented by thin coquina beds that range in thickness from a few to more than 30 centimetres, averaging 20 centimetres (Figs. 14a,b). Most of these coquina beds are sheet-like, laterally discontinuous; in some places they are concealed below a thick cover of weathered shale and siltstone units and can hardly be seen on the surface. Coquina beds vary in thickness and abundance from one place to another; individual beds cannot be traced laterally for more than a few metres before they die out or are covered by talus. Generally these bioclastic beds are thicker, more abundant, and more continuous laterally at Morrison Dome than in the Zortman area.

Lithologically, the coquinas range from packstone to grainstone (Dunham, 1962) and are composed largely of whole and fragmented molluscan shells (dominantly Gryphaea sp. and Ostrea sp.), together with a few belemnite fragments, fine-grained limestone or sandstone clasts, and silt or clay chips (Fig. 14b). The constituents are embedded in a matrix of varying proportions; this is composed of medium- to fine-grained, sand-size terrigenous material, small shell fragments, and calcareous silt and clay. The shells and shell fragments vary greatly in size and shape, from a few millimetres to more than 4 centimetres long and up to 2.5 centimetres across (Figs. 14b,c). These fragments are thick to medium walled and randomly oriented, with no sign of transportation.

14a



14b



Fig. 14: a) Coquina bed of Facies C exposed at Morrison Dome.
b) slab of a coquina bed at Morrison Dome to show the composition of the bioclastic limestone of facies C.

14c



Fig. 14c. Sample from a coquina bed (upper part of the photo), plan view, and across-section (lower part of the photo) through the same sample to show the composition of the bioclastic limestone and the range in shell size.

Where they are well exposed, the coquina beds are always sharp-based with erosional or undulatory contacts with the underlying shales and siltstones (Fig. 14a). Low-relief erosional surfaces are present locally, but they are difficult to identify. The upper contacts of the coquinas are more or less gradational to the overlying sediments. Sole marks, such as scour-and-fill structures (cf. Aigner, 1982), have not been recognized.

Internal primary sedimentary structures are generally lacking or poorly developed within these shelly limestone beds. Some beds appear to show weakly developed, graded bedding. Large scale, cross-stratified structures (cf. Brenner and Davies 1973; Uhler et al. 1988) are absent. Biogenic sedimentary structures have not been identified in the area.

5.3.2 Interpretation

The bioclastic limestone facies is interpreted here to be a product of high-magnitude storms that occasionally stirred the muddy platform. It was deposited as a result of the winnowing effect of storm-generated waves and currents on the muddy platform (cf. Specht and Brenner, 1979). The high textural contrast between the coquinas and the surrounding argillaceous rocks support the hypothesis of storm-origin of the former. Similarly the sharp and erosional nature of the lower contacts of these bioclastic beds, and the local

occurrence of lithic intraclasts, attest to the high intensity and the erosive nature of the storm-induced waves and currents (cf. Kreisa, 1981; Bloos, 1982). The absence of large-scale cross-bedding negates any tidal influence.

The role of storms in forming shelly bioclastic limestones has been acknowledged by several researchers over the last few decades (e.g., Powers and Kinsman, 1953; Gernant, 1971; Einsele and Seilacher, 1982; Hobday and Morton, 1984). The passage of periodic storms over muddy, shelf environments is known to introduce strong waves and currents in deeper waters (Kreisa and Bambach, 1982; Jeffery and Aigner, 1982; Dott, 1983). These waves and currents were capable of entraining shells from the sea floor, exposing previously buried ones, ripping up weakly consolidated rocks and resuspending finer sediment. After the storm began to wane, shells and intraclasts would have settled first to form lag deposits, whereas the fines would have been winnowed out by later wave processes.

In many ways, Facies C resembles the coquinoid sandstones described and interpreted as storm deposits by Brenner and Davies (1973, 1974) and Specht and Brenner (1979) from the Oxfordian rocks of Wyoming and southern Montana. Brenner and Davies (1973) divided the coquinoid sandstone beds into three types: channel, storm, and swell lags. The coquina beds of this study share many characteristics of the storm and swell lags described by these authors. The storm lags were

interpreted as having been deposited as the result of the flushing of surge channels during stormy periods and the spread of storm-transported debris on the leeward flanks of marine sand bars and inter-bar muddy areas. The swell lags were considered to be the product of the passage of storm swells over the muddy shelf areas that developed between marine sand bars. The absence of channel lags in the study area suggests that the coquina beds of Facies C must have been formed in a deeper setting than their counterparts to the south.

The coquina beds of this study also compare closely to the coquinoid sandstones that were reported from Alberta at the base of the Rock Creek "A" and "B" cycles by Marion (1984, p. 331), who interpreted these coquinoid sandstones as storm-generated deposits formed under the highest energy conditions in a muddy platform. Craft and Bridge (1987) described similar coquina beds from Devonian shallow-marine deposits in New York State; these beds were interpreted as swell lags that were deposited as a result of wave-induced winnowing.

Many workers have described similar coquina beds from both the geologic record (e.g., Kidwell and Jablonski, 1983; Einsele and Seilacher, 1982; Eyles and Lagoe, 1989) and from modern shelf environments (Kumar and Sanders, 1976). All of these examples were interpreted as the product of storm-induced waves which reworked muddy shelves, winnowed and concentrated bioclastic fragments, and removed the fines.

5.4 Facies D: sandstone-siltstone-shale facies

5.4.1 Description

Facies D consists predominantly of a series of sharp-based sandstone beds interbedded with siltstone and shale units of variable thickness (e.g., Fig. 12 -- units 21, 22, 23, and 24 of measured section No. 1, in pocket). Generally this facies comprises the middle part of an overall coarsening-upwards sequence that varies in thickness from one locality to another.

The siltstone-shale interbeds are relatively thick, but poorly exposed; usually they occur as distinct, lower-angle slopes below the sandstone beds. The shale is gray to dark gray, fissile and fossiliferous. Fossils are mainly belemnites with some bivalves. Both primary and biogenic sedimentary structures are lacking in the shale. The siltstone layers (a few millimetres to more than 10 centimetres thick) are more abundant than in Facies B. These layers are thinly laminated, cross-laminated, or ripple-cross-laminated (Fig. 12 -- unit 21 of measured section No. 1). Locally, the siltstone is moderately bioturbated.

The sandstone beds are also more common and thicker than in Facies B. These beds vary in thickness from a few to more than 60 centimetres and are up to one metre thick in places; they are more prominent in the Zortman Butte area than elsewhere. Individual sandstone beds have considerable lateral continuity and can be traced for tens of metres along

the depositional strike, but in cross-section some of them appear wedge- or lens- shaped. Their width across the strike is difficult to estimate because they are poorly exposed, but judging from the high length-to-thickness ratio, most of them are sheet-like.

The sandstones are variegated; from light gray, yellowish gray, brown or salt and pepper to cream yellow. Their texture usually ranges from very fine- to fine-grained; rarely, they are medium-grained. They are composed predominantly of moderate to well-sorted quartz and calcite, in varying proportions.

Sedimentary structures are very common within the sandstone beds. The sandstone beds are thinly-bedded (1 to 2.5 centimeters; Fig. 15) with different types of cross-bedding. The lower contacts of beds are always sharp or erosional, whereas the upper contacts are commonly gradational. The most common primary sedimentary structures include horizontal or low-angle cross-lamination, hummocky cross-stratification (hereafter referred to as HCS), and various types of ripple cross-bedding. Biogenic sedimentary structures are common but are restricted mostly to the upper parts of the beds. The following paragraphs will be devoted mainly to a discussion of the sedimentary structures associated with the sandstone beds.

Each sandstone bed within Facies D is characterized by a fining-upward sequence, ranging from 15 to more than 60

centimetres in thickness, and has a specific suite of primary and biogenic sedimentary structures (Fig. 16). Most sequences display characteristics of the hummocky cross-stratified sequence proposed by Dott and Bourgeois (1979, 1982) or the modified sequence of Walker et al. (1983).

The suite of sedimentary structures is arranged vertically as follows: sharp basal contact --> horizontal parallel lamination --> hummocky cross-bedding --> horizontal parallel lamination --> ripple cross-bedding --> bioturbation (Fig. 16).

The basal contact of each sequence (cf. first-order contact of Dott and Bourgeois, 1982) is always sharp or slightly erosional, but without sole marks (Fig. 17a). No basal lag (cf. division B of Walker et al., 1983) was recognized. Instead, the basal contact is directly overlain by parallel or slightly undulating laminae up to 10 centimetres thick (cf. division P of Walker et al., 1983) (Fig. 17a). The parallel lamination is succeeded by low-angle (less than 10 degree) sets of truncated laminae, 5 to 15 centimetres thick. The latter sets closely resemble the hummocky cross-stratification structure variously described from the rock record (Dott and Bourgeois, 1982; Bourgeois, 1980; Brenchley, 1985; Atkinson et al., 1986; and many others). Internal erosional surfaces (second order contact of Dott and Bourgeois, 1982) are common within these sets of truncated laminae (Figs. 17a,b). Wave length and amplitude

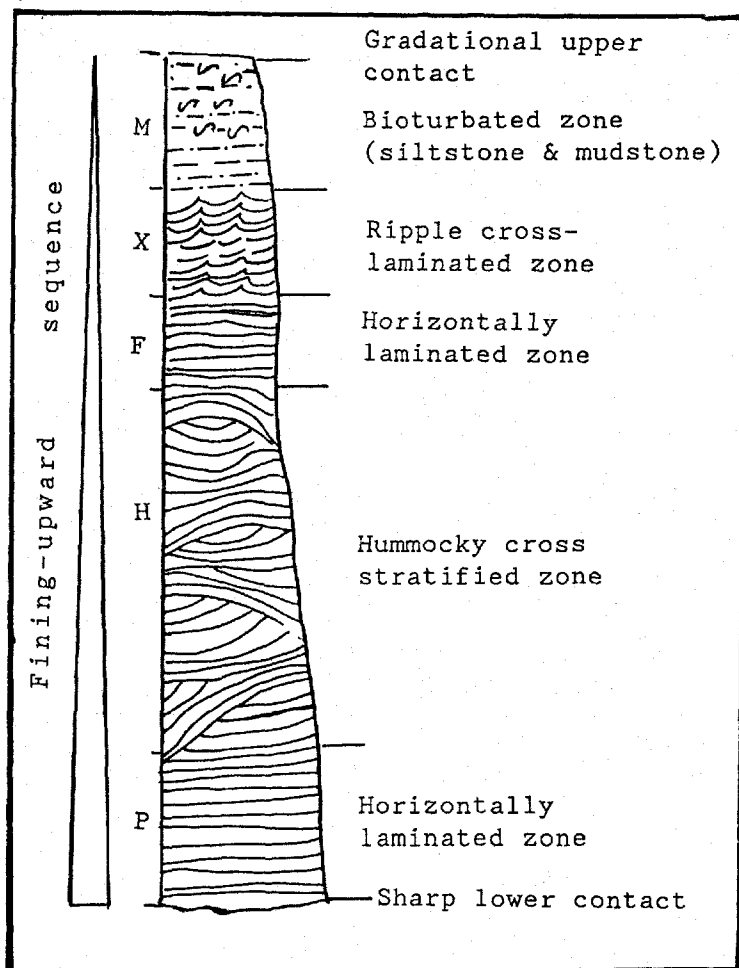


Fig. 16. An idealized suite of sedimentary structures associated with sandstone beds of Facies D. This suite is not always complete (thickness of the sequence range from 20 cm up to 60 cm).

17a



17b



Fig. 17. Hummocky cross-bedded sandstones of Facies D:
 a) sandstone bed exposed at Lodgepole area (note sharp lower contact, zones P, H, F, and X; zone M is covered).
 b) sandstone bed showing elements of zones P, H, and F of Fig. 16 (sandstone outcrop at Morrison Dome).

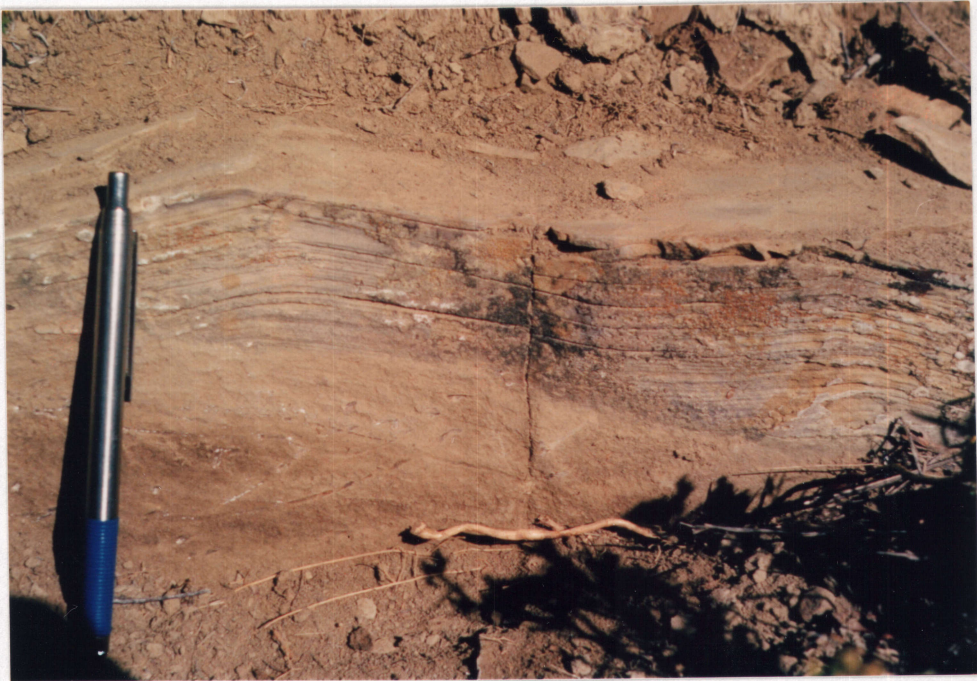
of the hummocks range from 30 and 5 to more than 90 and 15 centimetres respectively. These dimensions are less than those described by Harms et al. (1975), Hamblin and Walker (1979) and Surlyk and Neo-gaard (1986), but are similar to many others described from both recent deposits (Greenwood and Sherman, 1986) and the rock record (e.g., Campbell, 1966; Mount, 1982, Wu, 1982; Nottvedt and Kreisa, 1987). Locally, the hummocky beds may be replaced by, or grade upward into, a zone of horizontal or gently inclined laminae (cf. division F of Walker et al., 1983).

The zone of planar laminae (zone F of Fig. 16) grades upward into a variably rippled, cross-laminated zone of a few centimetres thick. The most common type of ripple cross-lamination in this zone is climbing ripple cross-lamination (Fig. 18). The climbing ripples are small-scale (cf. ripple laminae-in-phase of McKee, 1965). These ripples are 10 to 20 centimetres in wave length and up to 8 centimetres high. Bounding planes forming the top and bottom of sets are nearly parallel, with complete preservation of both lee and stoss sides.

The ripple cross-laminated zone commonly passes upwards into very thin layers of bioturbated, very fine sand or coarse silts (cf. laminated-to-burrowed sequence of Howard and Reineck, 1981; or division M of Walker et al., 1983). The top surfaces of these layers are highly bioturbated and contain abundant trace fossils (Fig. 19), mainly horizontal

Fig. 18. Sandstone bed of Facies D with climbing ripple cross-lamination (zone X). Note that the ripple zone grades upwards in bioturbated muddy zone (M). Sandstone bed exposed at Lodgepole (see location 2, Fig. 3).

Fig. 19. Planolites sp. trace fossils preserved on sandstone beds of Facies D, plan view: a) sample from unit 26 of section 1, Fig. 12; b) sample from a sandstone bed exposed at Morrison Dome.



to gently-inclined burrows assignable to the Planolites sp. of the Cruziana ichnofacies (Frey and Seilacher, 1980; Pemberton and Frey 1984). Individual examples are relatively large (up to 10 centimetres long). They are cylindrical, smooth-sided, and non-branched. They range in density from isolated to moderately crowded, with various types of crossovers or interpenetrations. The Planolites-like ichnospecies described here bears most of the characteristics of P. beverleyensis described by Pemberton and Frey (1982).

6.4.2 Interpretation

Facies D is interpreted as having been deposited in a storm-dominated, prograding shelf (lower shoreface) shallower than that of Facies B. This interpretation is based mainly on:

- 1) a high percentage of sand and coarse silt compared to that of Facies B;
- 2) common recurrence and up-section thickening of the sandstone beds;
- 3) abundance of storm-related sedimentary structures within the sandstone beds; and,
- 4) decrease in marine fauna.

The above factors are considered to be the consequence of a shorewards shift from the open marine conditions which had prevailed for a long time. This shift resulted in a high sedimentation rate as a result of continuous supply of

terrigenous sediments brought to the shelf during the regressive phase of the Oxfordian sea. The upward increase in the proportion of the sandstone and siltstone, and the concomitant increase in sedimentary structures, reflect a progressive upward shallowing and an increase in wave and current strengths.

In terms of hydrodynamic conditions, Facies D is interpreted as having been formed as a result of the interplay of fair-weather and storm conditions. The shale units were deposited from suspension during calm weather conditions, whereas the sandstone beds were formed by storm waves and currents during rough weather conditions.

The sandstone beds of Facies D are interpreted as storm deposits that were formed in a regressive shelf dominated by recurring storms. Each sandstone bed ("an event" deposit in the sense of Dott and Byers, 1989) and its associated sedimentary structures are considered to be the product of a single storm event. Initially the storm was powerful, but it waned gradually. The vertical sequence of sedimentary structures and the small-scale fining-up trend within each sandstone bed attest to the waning nature of the storm (Mount, 1982; DeCelles, 1987, p. 260; Craft and Bridge, 1987, p. 351; Dott and Byers, 1989). A similar interpretation was proposed by Hunter and Clifton (1982) for hummocky cross-stratified sequences within the Upper Cretaceous Cape Sebastian Sandstone. DeCelles (1987) described similar

sandstone sequences from the Middle Tertiary deposits of southern California.

The horizontal parallel laminae (zone P of Fig. 16) overlying the sharp basal contact could have been deposited from a higher flow regime than the subsequent one (zone H of Fig. 16), which deposited the hummocky cross-bedding. According to Walker et al. (1983) parallel laminae below HCS can be formed either by energetic unidirectional or by oscillatory flows. The parallel laminae encountered in the study area were probably formed under oscillatory flows, since they are not associated with the sole marks that characterize beds formed under unidirectional flows.

The hummocky cross-stratified sets are considered to have been deposited as a result of storm-dominated waves and currents scouring the underlying beds, suspending fine sand- and silt-sized sediments and draping them later, as the storm power abated, on the scoured surfaces. The HCS was first named, described and interpreted as storm structure by Harms et al. (1975). Many authors have discussed the morphology and genesis of HCS (e.g., Dott and Bourgeois, 1982; Brenchley, 1985). Several researchers have agreed with the above-mentioned authors that hummocky cross-bedded deposits were formed by storm-generated waves and currents (Hamblin and Walker, 1979; Bourgeois, 1980; Wright and Walker, 1981; Atkinson et al., 1986; Bose et al., 1988). However, the exact nature and mode of formation of the HCS is still a matter of

investigation (Swift et al., 1983; Duke, 1985, 1987; Klein and Marsaglia, 1987; Swift and Nummedal, 1987; Southard et al., 1990).

The parallel and low-angle cross-lamination, with various kinds of ripple cross-lamination above the hummocky cross-stratified zone, record a decrease in wave and current strengths that was caused by rapid attenuation of the storm strength (Cant, 1980, p. 129; Mount, 1982; Handford, 1986). The ripple cross-laminated zone toward the top of each sandstone bed records the latest stages of the storm and the decreasing hydrodynamic conditions (cf. Rice, 1984, p. 157). The climbing ripples suggest a rapid deposition from suspension and a high rate of sediment supply (Harms, 1979). They were preserved owing to a high rate of burial under the influence of wave-generated currents. Development of climbing ripple cross-lamination implies that abundant sediment is supplied continuously to the currents, so that ripples are built upward in overlapping series rather than migrating in a forward direction (McKee, 1965). The climbing ripples of the present study seem to be analogous to the sinusoidal ripples of Jopling and Walker (1968), who interpreted them as the product of a unidirectional current system, but Campbell (1966) interpreted similiary undulating laminations as wave ripples. The present ripples also resemble those of the Ordovician deposits in Virginia, which have been described by Kreisa (1981), who interpreted the climbing ripples as being

the product of rapid deposition under the influence of randomly-oriented wave currents. Hunter and Clifton (1982) described a similar kind of ripples as "symmetrical ripples" that were developed within the planar- to cross-bedded sandstone of the Upper Cretaceous Cape Sebastian Sandstone. These writers interpreted the symmetrical ripples as having been formed by wave-generated currents due to rapid deposition and a high rate of burial.

The upper bioturbated muddy sand layer is interpreted as a post-storm deposit formed under fair-weather conditions. The Planolites burrows on top of this layer suggest a return to fair-weather conditions and the recolonization of the sea-floor by opportunistic feeding organisms (cf. Mount, 1982, p. 836; Frey and Pemberton, 1984). Generally the Planolites ichnofossils characterize subtidal, poorly sorted and unconsolidated substrates (Frey and Pemberton, 1984; Ekdale et al., 1984). Hydrodynamic conditions of such substrates vary from moderate energy levels in shallow waters, between the daily fair-weather and the storm wave bases, to low energy levels in quiet waters.

The sequence of events that led to the deposition of the sandstone beds within Facies D was comparable to that proposed by Mount (1982) to account for the deposition of the hummocky cross-bedded units of the Andrews Mountain Member (Lower Cambrian), eastern California which were considered to be formed by storm-surge ebb currents that generated high-

energy conditions. Mount (1982) interpreted each sandstone unit in terms of two contrasting energy conditions. The lower part of the units, which are characterized by hummocky cross-bedding, was deposited under stormy conditions. The upper bioturbated part was interpreted as being the product of fair-weather conditions. Many other workers have now described similar beds to those of Facies D and all consider them as products of storm-generated processes (e.g., Bose et al., 1988; Brenchley et al., 1979; Moore and Hocking, 1983; Brenchley et al., 1986; Craft and Bridge, 1987; Davis and Byers, 1989).

5.5 Facies E: cross-bedded sandstone facies

5.5.1 Description

This facies transitionally overlies Facies D (the sandstone-siltstone-shale facies). It consists of sandstone beds with minor interbedded siltstone and shale partings. The sandstone beds are thick (Fig. 20a, 20b), amalgamated and more laterally continuous than the sandstone beds of Facies D. They range in thickness from a few decimetres to more than 1.5 metres; the amalgamated beds may be up to 2 metres thick. There is a vertical and lateral variation in the abundance and thickness of these sandstone beds. For example, the sandstone beds are either very thin (about 20 centimetres thick) or entirely absent on the eastern and southeastern side of Zortman Butte. On the western side of



20a



20b

Fig. 20: a) Thick, amalgamated sandstone bed of Facies E exposed on the western side of Zortman Butte (length of the tape is 55 centimetres.

b) Thick, hummocky cross-stratified sandstone bed of Facies E exposed at Stage Route Butte (location 1, Fig. 3).

Zortman Butte and the eastern side of Saddle Butte, at least two beds were recognized (Fig. 12 -- units 19 and 6 of measured sections 3 and 5 respectively). At Morrison Dome, the sandstone beds are exceptionally thick (one to three metres), amalgamated, abundant, and laterally persistent (Fig. 12 -- units 15 and 17 of measured section 6).

Individual sandstone beds can be traced several hundreds of metres along the extent of outcrop and sometimes down to the surrounding plains. Commonly they crop out as resistant, cliff-forming ledges above the lower-angle, slope-forming shale of Facies B (Figs. 20b, 24a of section 5.6.1).

The sandstones are variegated, medium-grained, and are well sorted. They range in colour from buff-gray, pale gray or greenish gray, to brown or reddish brown. These sandstones weather light gray and commonly split into flaggy slabs (1 to 2.5 centimetres thick). They are composed predominantly of quartz grains which account for more than 60% of the total composition (Fig. 20c). The quartz grains are angular to sub-angular, and rounded in parts of the section. Calcite is also common, mainly as a cement. Locally, clay clasts are present. Interlayers of siltstone and shales (up to 10 centimetres) are also common.

The sandstone beds of Facies E are characterized by a wide spectrum of primary sedimentary structures, but generally lack biogenic structures. The most common primary sedimentary structures include horizontal lamination and low-angle cross-

20c

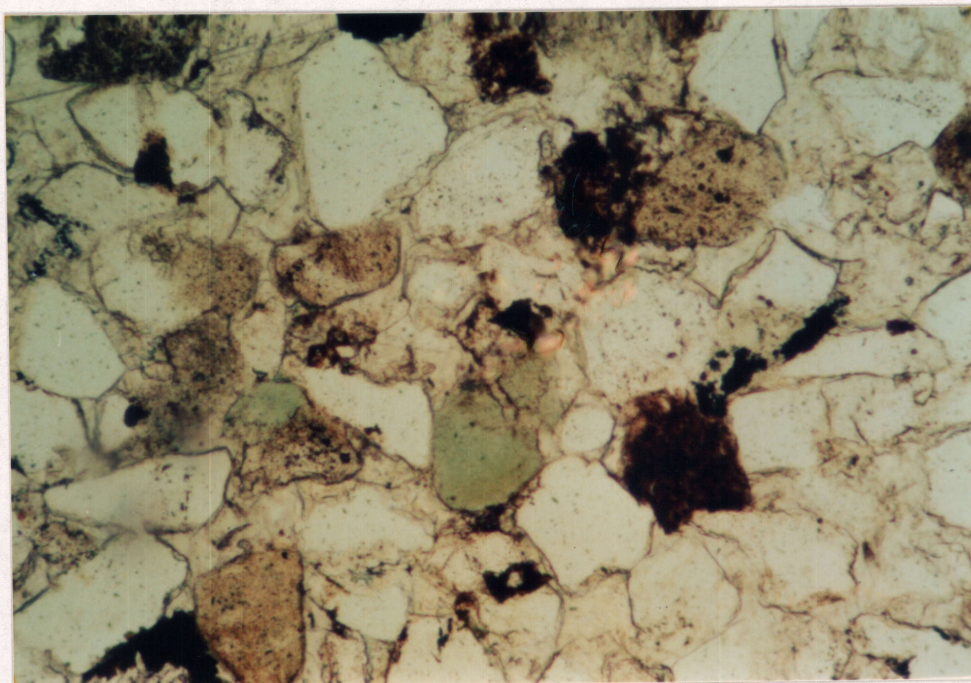


Fig. 20c. Thin section from a sample taken from sandstone of Fig. 20a. Note predominance of quartz (light gray) in the section. Plane polarized light, x150.

bedding, gently concave, low-angle cross-bedding, hummocky cross-bedding, and various types of wave-generated ripple cross-bedding. Trough cross-bedding and relatively high-angle cross-bedding structures occur in places.

The horizontal and low-angle cross-bedding (Harms, 1979) structures are very common in Facies E, especially in the sandstone beds exposed at Morrison Dome (Fig. 21). The cross-bedding structures occur in sets (up to 15 centimetres thick) that taper towards lower erosional surfaces and are commonly arranged in cosets (McKee and Weir, 1953) up to 50 centimetres thick.

The gently concave, low-angle cross-bedding structure is prominent and widely distributed within Facies E. It is common within thick sandstone beds and consists of slightly concave sets of laminae that drape mutually intersecting, shallow depressions or scours (Fig. 22). The sets of laminae range in thickness from a few cm to more than 15 centimetres; the scours are up to 80 centimetres wide. The laminae in one set commonly accrete upward from an erosional surface, drape over each other conformably and gradually become flat toward the top of the set to form horizontal bedding. Sets of laminae always have a sharp lower boundary.

There are some sporadic occurrences of HCS associated with the concave cross-stratification (Fig. 20b). These hummocky structures are composed of curving-upward sets of laminae up to ten centimetres thick; they are similar to the

Fig. 21. Horizontal and low-angle cross-bedded sandstone bed of Facies E exposed at Morrison Dome (length of the tape is 53 centimetres).

Fig. 22. Swaley cross-bedded sandstone of Facies E exposed at Morrison Dome.



accretionary hummocks described by Brenchley (1985). Other sedimentary structures include some wave-generated ripple cross-bedding (De Raaf et al., 1977) and rare trough cross-bedding (McKee and Weir, 1953; Harms, 1979). Some of the sandstone beds display straight-crested, symmetrical wave ripples on top, with wave length in the order of 2.5 to 10 centimetres (Fig. 23).

Neither tidal features, such as herringbone cross-bedding or reactivation surfaces, nor any signs of emergence such as plant roots, mud cracks or rain drops, have been identified within the sandstone of Facies E. The cross-bedded sandstone beds contain neither body nor trace fossils.

The lower contacts of the sandstone beds are always sharp or erosional into the underlying rocks, which are usually siltstone and shale. Upper contacts are also sharp.

5.5.2 Interpretation

In trying to interpret Facies E, one should keep in mind the abundance of the primary sedimentary structures and the stratigraphic position of the facies above previous Facies D. The predominance and thickness of the sandstone beds is also significant. The high percentage of sand in relation to the silt and mud and the absence of biogenic sedimentary structures must also be taken into account in making an interpretation.

The sedimentological attributes of Facies E suggest that



Fig. 23. Sample of sandstone of Facies E with symmetrical wave ripples, plan view. Wavelength of the ripples here is smaller than what was recorded in the field.

it was deposited in a storm-dominated, lower to middle shoreface (Harms et al., 1982; Walker, 1984). The absence of any sign of emergence suggests that Facies E was deposited in a sub-tidal environment, subject to both fair-weather and storm waves and currents (cf. Johnson, 1978, p. 253; Bouma et al., 1982; Krause and Nelson, 1984; McaCarthy, 1987).

In general, Facies E represents the top of a coarsening-upward sequence in which mudstone coarsens up into siltstone and sandstone intervals, which in turn grade up into medium-grained sandstone. In addition, the HCS and related structures of Facies D were replaced by other types of wave-generated sedimentary structures in Facies E. consequently, the shoaling pattern observed in Facies D must still have continued during the deposition of Facies E. Consequently Facies E is interpreted as having been deposited further shorewards, in an environment shallower than that of Facies D and subject to both storm and fair-weather processes.

The relatively thick, amalgamated sandstone beds of this facies suggest a continuous reworking of sediment by waves and currents in a shallowing shoreface setting. The waves and currents may have removed all mud- and silt-sized sediment before deposition of a subsequent bed took place. Bose and Das (1986) pointed out that shoaling-up leads to a gradual increase in the capability of waves and currents to rework sediments to greater depths below the water-sediment interface. As a result, any record of quiescent weather

conditions may have been removed.

The horizontal and cross-bedding were formed by migrating sand waves and dunes (Banks, 1973; Johnson, 1977). Such structures commonly indicate strong wave and current activities, but in terms of environmental reconstruction they are of little value since they form in a wide spectrum of environments (Harms et al., 1975, 1982). Vertical variation in the type of sedimentary structures indicate fluctuating hydraulic conditions.

The gently concave cross-bedding is considered to be swaley cross-stratification (cf. Leckie and Walker, 1982). Before interpreting this cross-bedding structure in the study area, it is worthwhile to provide a brief historical review of swaley cross-stratification (hereafter referred to as SCS) in general.

The SCS was first defined and interpreted as a storm-generated sedimentary structure by Leckie and Walker (1982) from Cretaceous deposits of the Moosebar and Gate Formation. Faulkner (1988) described a similar structure as "dish-like" from the Shipway Limestone of Gowerwaled and interpreted it as storm-generated structure. Since then, an overwhelming consensus of opinion holds that the SCS structure was formed and preserved by storm-generated processes in shallow marine environments above fair-weather wave base (Handford, 1986; Rosenthal and Walker, 1987; Bose et al., 1988). The SCS is always associated with thick amalgamated deposits that occur

in shallower parts of prograding stratigraphic sequences (Walker, 1982; Walker et al., 1983; Sun, 1989). Swaley cross-stratified deposits were commonly reported as occurring in a stratigraphic position above that of HCS, but below beach deposits (McCroory and Walker, 1986, p. 57).

Occurrences of SCS in the present study area suggest that the Facies E sandstones were deposited by storms and were later reworked by fair-weather waves and currents. These conditions led to the amalgamation of the sandstone beds.

The amalgamation of the sandstone beds may be explained in different ways. Some writers suggest it may have been due to the high recurrence of storm events, which removed, totally or partially, fair-weather deposits (mud and silt or burrowed zone) of a sandstone bed (Bourgeois, 1980; Kreisa, 1981; Mount, 1982; Dott and Bourgeois, 1982). In prograding sequences, the amalgamation of sand beds could have resulted due to a high influx rate from source areas, coupled with frequent storm recurrence (Brenchley, 1985). In distal areas, amalgamation of storm-dominated deposits may have been due to the low rate of mud accumulation in relation to sand influx (Brenchley, 1985, p. 381).

Sporadic occurrence of HCS associated with the SCS structures further supports the role of storms in the deposition of the sandstone beds of Facies E. The association of SCS with hummocky cross-bedding has been well documented in the rock record (Plint and Walker, 1987; Bose and Das,

1986; Faulkner, 1988). The ripples on top of cross-bedded sandstones are interpreted as wave-generated structures (cf. De Raaf et al., 1977), which reflect strong energy regimes during the deposition of the sandstone beds.

The absence of trace fossils attests to the strength of waves and currents, which could have made the organisms vulnerable to such conditions or reworked bioturbated layers. In current and wave-agitated environments, primary structures predominate, because biological structures either were not developed or were destroyed (Elliott, 1978, p. 148). Conversely, in quiet settings, biological structures dominate to the extent that the sediments become homogeneous or virtually structureless (see Krause and Nelson, 1984, p. 500).

The absence of herringbone cross-bedding, reactivation surfaces, and mud drapes precludes the possibility that any significant role was played by tidal current in the area at the time these rocks were deposited (McCrorry and Walker, 1986). The absence of plant roots or swash sedimentary structures indicates that these rocks were sub-littoral deposits.

5.6 Facies F: limestone facies

5.6.1 Description

Facies F (the limestone facies), which is 0.5 to more than two metres thick, forms the uppermost part of the Swift

Formation in the Little Rocky Mountains area; it occurs either as low relief beds in flat lying areas or as ledge-forming cliffs on top of the Swift sandstones (Figs. 24a, 24b). Locally (e.g., eastern side of Zortman Butte), Facies F is underlain by a one metre thick sandstone bed. This sandstone consists of medium to coarse-grained, clean and well sorted quartz grains (>85%). However, no sedimentary structures were recorded from this, mainly because of its poor exposure. Facies F consists of impure limestone beds interbedded with silty limestone or calcareous siltstone layers. In the Zortman area, the limestone facies is capped by a dark bed of concretionary manganiferous siderite (Knechtel, 1959, p. 737). At Morrison Dome, the manganiferous bed is dark brown, fine-grained, rippled and up to 30 centimetres thick. Knechtel (1959) considered this bed, with others at different stratigraphic levels, to be part of the Morrison Formation.

Where it has been observed closely, the limestone of Facies F is light to milky or buff gray and fine-grained; it weathers to pale yellow or yellowish gray, and fractures conchoidally or in blocky masses. Internal sedimentary structures are absent. Both trace and body fossils are scarce.

The lower 15- to 25-centimetres of the limestone facies consists predominantly of silty limestone that is thinly laminated, rippled, and bioturbated. This layer is characterized by presence of small halite impressions and



24a



24b

Fig. 24. Limestones of Facies F: a) low-lying limestone bed exposed on the western side of Zortman Butte.
b) an outcrop at Morrison Dome showing a ledge-forming sandstone bed of Facies E that overlies a slope-forming shale and is, in turn, overlain by a limestone bed of Facies F.

small-scale horizontal trace fossils (Fig. 25).

Interestingly, no gypsum layers or moulds were observed within the silty limestone layer.

The silty limestone bed grades both laterally and vertically into highly calcareous siltstone which, in turn, is overlain by limestone beds. Locally, for example on the western side of Zortman Butte, the silty limestone is overlain directly by the limestone beds without intervening calcareous siltstone.

5.6.2 Interpretation

Facies F is considered to be the topmost stratum of the Swift Formation in the Little Rocky Mountains and is interpreted here as recording the shallowest stage of the Late Callovian-Oxfordian sea in the area. The moulds at the base of the facies suggest that deposition occurred in a very shallow, slightly restricted environment, most likely localized pools in tidal flats or small lagoons. They are similar to the "hopper-shaped calcite pseudomorph after halite" described by Handford and Moore (1976), who interpreted that depositional environment as a prograding tidal flat, frequently affected by mixed meteoric-marine waters. Schreiber et al. (1976), and Gornitz and Schreiber (1981) also reported similar structures (displacive hopper-shaped halite cubes) from the Dead Sea and interpreted their presence as indicative of very shallow environments.



Fig. 25. Halite impressions at the base of a limestone bed (unit 33, measured section no. 1, Fig 12) of Facies F exposed on the eastern side of Zortman Butte.

The sparse distribution of the pseudomorph moulds at the base of Facies F suggests that marine water at that time was not sufficiently saturated with respect to halite to allow its extensive precipitation (Arthurton, 1973).

The ripple cross-lamination implies that the sediments were subject to reworking by waves and currents during fair-weather conditions. The absence of mudcracks, rain drops, or plant roots (indicators of emergence) suggests that deposition of the facies took place in a subtidal to lower intertidal zone (Schreiber et al., 1976, p. 747).

However, the absence of gypsum, which is commonly associated with halite in evaporative marginal marine settings, is problematic. There are two possible explanations. Firstly, the marine waters at the time of deposition were deficient in calcium, because Ca^2 was consumed to form limestone. Alternatively the halite may have been precipitated by evaporative concentration from sulphate-poor waters, perhaps from continental sources. Neither explanation is satisfactory and the halite's origin requires further investigation.

6. DEPOSITIONAL ENVIRONMENT

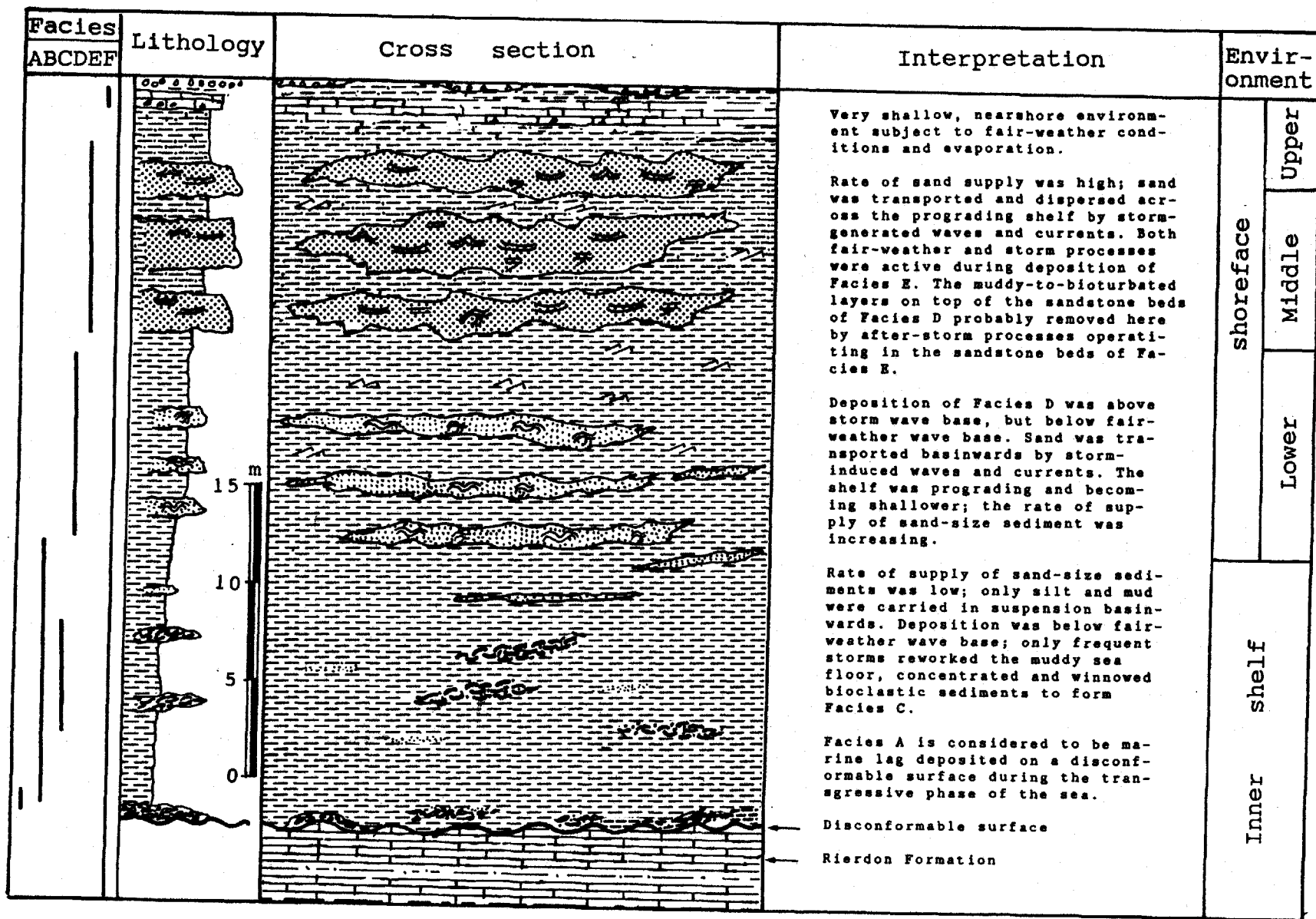
6.1 Facies relationships

The preceding analysis has shown that the Swift Formation in the Little Rocky Mountains area has six distinct facies, which are labeled A-F. Relationships among the facies are shown in Fig. 26. Facies A represents the base of the section and rests disconformably on the underlying Rierdon Formation; it is of limited lateral and vertical distribution, and commonly occurs as patchy lenses with sharp contacts.

Facies B is widely distributed laterally, with almost constant thickness. The sandstone and limestone beds of this facies are relatively thin; they are lens- or wedge-shaped, with sharp lower contacts and gradational upper boundaries. The bioclastic limestone beds of Facies C are randomly distributed -- both laterally and vertically -- within Facies B. They form thin sheets that taper laterally over short distances. As a rule, contacts with the enclosing shale are sharp.

Facies B becomes sandier toward the top, where it grades imperceptibly into Facies D; that facies, in turn, varies considerably in thickness from one place to another, thickening southwards and westwards of the Zortman area. The upward transition of Facies D into Facies E is gradational. Facies F is of restricted distribution, and occurs as discontinuous patches at the top of the Swift Formation.

Fig. 26. Facies relationship and depositional environment of the Swift Formation in the Little Rocky Mountains. For explanation see Fig. 12 in back pocket.



Facies B, D, and E are most prominent by virtue of their thickness and lateral continuity. They account for more than 90% of the bulk volume of the Swift Formation, and they are arranged vertically in a coarsening-up sequence.

6.2 Depositional model

The Swift section of the Little Rocky Mountains is considered to have been deposited in a shallow-marine, shelf environment, ranging from middle- through inner-shelf to lower shoreface (cf. Harms et al., 1982; Reinson, 1984; Walker, 1984); the depositional model is portrayed in Fig. 26. At the base of the section, Facies A rests disconformably on the underlying Rierdon Formation. It represents a transgressive lag that was deposited during the early stage of the Oxfordian transgression across the area. Facies A was deposited under a reduced rate of sedimentation, with a low influx of terrigenous sediments. Before the deposition of Facies A, there was no sedimentation at the site for a long period that spanned about three ammonite zones and perhaps as much as 3.06 Ma.

As the sea advanced from the north, the supply of coarse sediment to the shelf was highly reduced and sedimentation, mainly from suspension, continued for a long period of time -- resulting in the deposition of Facies B, which represents deposits of the inner shelf environment. Frequent storms in this environment, caused the deposition of thin layers of

coarse siltstone or very fine sandstone (cf. the distal tempestites of Aigner and Reineck, 1982). More intense storms operated periodically in the muddy shelf environment. These disturbances reworked bioclastic material in situ and resulted in the deposition of the bioclastic beds of Facies C.

Facies D was deposited on a shallow marine, progradational, muddy shelf to lower shoreface that was periodically affected by storms and waves that transported, deposited and moulded sand to form the sandstone beds. This facies was deposited under fair-weather, storm, and post-storm conditions.

Continuous shoaling and proximity to the source area are more evident in Facies E, which shows a marked increase in the sand proportion and bed thickness, grain size, and bedload traction sedimentary structures. The sedimentological attributes, and the stratigraphic position of Facies E above Facies D, indicate a shallower environment (middle to upper shoreface) that was subject to storm and fair-weather conditions (cf. Leckie and Walker, 1982; McCrory and Walker, 1986).

Facies E passes upward into thin beds of calcareous siltstone and the impure limestone of Facies F (the limestone facies) that forms the topmost part of the Swift Formation in the area. Facies F represents the product of a quiet-water environment that was subjected only to weak waves and

currents. Such an environment may indicate the presence of localized, small lagoons in which halite was precipitated periodically by evaporation.

6.3 Comparison with other exposed Oxfordian sequences in the Western Interior

The Swift section exposed at the Little Rocky Mountains is similar in many aspects to other described Oxfordian sections elsewhere in the Western Interior. However, despite the overall similarities, local variations exist from one place to another. For example, the present section differs from the western Montana exposure (Cobban, 1945; Mudge, 1972) in that it is older, thicker and more shaly. These differences suggest that the Swift Formation of the Little Rocky Mountains was deposited more distally than the Swift Formation of western Montana. These characteristics also provide evidence that north-central Montana was flooded earlier, and that eventually the sea spread westward.

The Swift Formation of the present study closely resembles the Upper Sundance Formation and the Redwater Shale Member of central and eastern Wyoming respectively (see Imlay, 1947, 1956; Pipringos, 1968; Wright, 1973). These deposits are dominated by shale and fine-grained sandstone, and suggest deposition far from the contemporary shoreline. They represent the mud and marine bar-sand facies of Brenner and Davies (1974). Nonetheless, no tidal sedimentary structures

have been recognized in the Swift deposits of the Little Rocky Mountains. In contrast, such structures have been described variously in Oxfordian rocks, at least from the regressive sequences, in Wyoming and Utah (cf. Brenner et al., 1985; Kreisa and Moiola, 1986, Uhlir et al., 1988).

6.4 Comparison with other shallow marine deposits outside the Western Interior

The Swift Formation of the present study also resembles many other shallow marine siliciclastic sequences from different parts of the world and of different geological ages (e.g., Levell, 1980; Brenchley and Newall, 1982; Atkinson et al., 1986; Craft and Bridge, 1987; Bose et al., 1988; Nealon and Williams, 1988). In all cases, the shallow marine deposits were considered to have been formed through an interplay of fair-weather and storm conditions. For example, most of the sedimentary structures described from the Precambrian Skaergadnes Formation of Norway (Levell, 1980) were attributed to the action of storm-generated waves and currents. The sedimentological attributes of the sandstone beds of the present section bear strong similarities to those described by Brenchley and Newall (1982) from sandstone beds of the Ordovician Caradoc rocks and show some similarity to the observation of Bose et al. (1988) from the Proterozoic Kaimur Formation, which passes from offshore to beach deposit across a storm-dominated shelf.

8. SEA-LEVEL CHANGE

After a period of non-deposition that spanned about three ammonite zones, the Sundance sea flooded the Little Rocky Mountains from the northwest, during early Late Callovian time. Evidence for this interpretation is the occurrence of the Q. collieri zone (Imlay, 1948) near the base of the Swift section. However, the incursion must have started earlier than the eustatic sea-level rise, which was inferred to occur no earlier than Early to Middle Oxfordian time (Brookfield, 1970, p. 350; Hallam, 1978; Vail et. al., 1984, p. 130; Brenner, 1983). Accordingly, the inundation across the present study area was more likely caused by local or regional tectonic factors than by eustasy. These factors were conceivably the product of the Nevadan orogeny to the west and differential subsidence across the whole shelf. The Little Rocky Mountains area lay on the western margin of the Williston Basin, which was subject to a considerable degree of subsidence (Shurr et al., 1989). The Sweetgrass Arch to the west was also experiencing some degree of subsidence (Brooke and Braun, 1972). That being so, subsidence in the Little Rocky Mountains during the same time was inevitable.

Initially, the sea spread quickly over the area, to the extent that only a thin, discontinuous part of the sedimentary record (Facies A) was preserved, which may be accounted for by the gently sloping shelf or because the

shore was bordered by low-lying lands. In either instance, the rapidly spreading waters led to the development of a thin, sediment-starved section (the belemnite/chert conglomerate of Facies A), as evidenced by the concentration of water-worn belemnite fragments and chert pebbles.

During early Oxfordian time, the sea deepened and spread rapidly westward, eastward, and southward. The deeper waters kept sand-size sediments from reaching the site of deposition; only silt- and mud-size sediments were transported basinwards and settled from suspension. As a result, laterally continuous, thick sheets of shale were deposited. The thickness of these sheets suggests that deposition continued under such conditions for quite a long period before regression began (cf. Hallam, 1978, his model D). The exact time when the sea-level changed from a rising to a falling state is unknown; this is mainly due to the transitional nature of the Swift sequence from the lower shaly portion to the mostly sandy upper part.

By Middle Oxfordian, slow regression across the shelf started; this is evident in the increase in the sand percentage over that of mud and silt in the upper portion of the Swift section. The regression was probably intermittent and punctuated by short periods of transgression. This is suggested by the occurrences of interbedded shale and sandstone beds within the upper part of the section.

The last problem to be addressed here is the cause of the

Oxfordian regression from the Little Rocky Mountains. Was this eustatically controlled or was it due to local tectonic parameters and an increased rate of sediment influx? As mentioned earlier, the sea-level rise over the study area does not correspond to the global eustasy; rather, the rise was due to local tectonic factors coupled with the rate of sedimentary influx. One must refer again to these local factors, since the Middle Oxfordian was a time of eustatic sea-level rise (Brookfield, 1970; Hallam, 1978; Brenner, 1983). Several authors agree that there was a rising area bordering the western side of the Sundance sea (e.g., Peterson, 1972; Wright *et al.*, 1978; Poulton, 1988). It was this rising area, corresponding to the Nevadan orogeny, that kept shedding and dispersing siliciclastic sediment eastward across the shelf. The rising nature of that source area may have been felt distally in the Little Rocky Mountains, as reflected by the thick pile of sandstone beds of the Swift section. Another possibility is that subsidence of the area ceased and may even have been reversed. However it is unlikely that the area was uplifted, since no part of the Swift section in the area is known to have been removed by post-Swift erosion. Moreover, the gradational nature of the contact and the absence of an unconformity between the Swift and the overlying Morrison Formations combine to negate the possibility that such uplift occurred.

9. CONCLUDING REMARKS

Strata of the Swift Formation exposed in the Little Rocky Mountains have been studied in terms of six facies: a conglomerate facies at or slightly above the base; a shale-siltstone facies; a bioclastic limestone facies; a sandstone-siltstone-shale facies; a cross-bedded sandstone facies; and a limestone facies representing the uppermost part of the section. Detailed sedimentological study of these facies has shown that:

- 1) The Rierdon-Swift boundary is sharp and disconformable.
- 2) The Swift Formation represents a classical, shallow-marine, shelf deposit accumulated in the course of a transgressive-regressive episode spanning Late Callovian to Oxfordian time.
- 3) The Swift Formation represents an overall coarsening-upward sequence in which mud and silt pass upward into coarse silt, which in turn grades into fine- to medium-grained sand.
- 4) Storms played a significant role during deposition of the Swift strata, whereas tidal action was of minor importance.
- 5) The depositional model proposed for the Swift Formation in the Little Rocky Mountains is one of a shifting pattern of sedimentation in a shallow, prograding marine shelf. In this model, inner shelf deposits pass

transitionally into lower shoreface sediments, which in turn give place to upper shoreface or foreshore strata.

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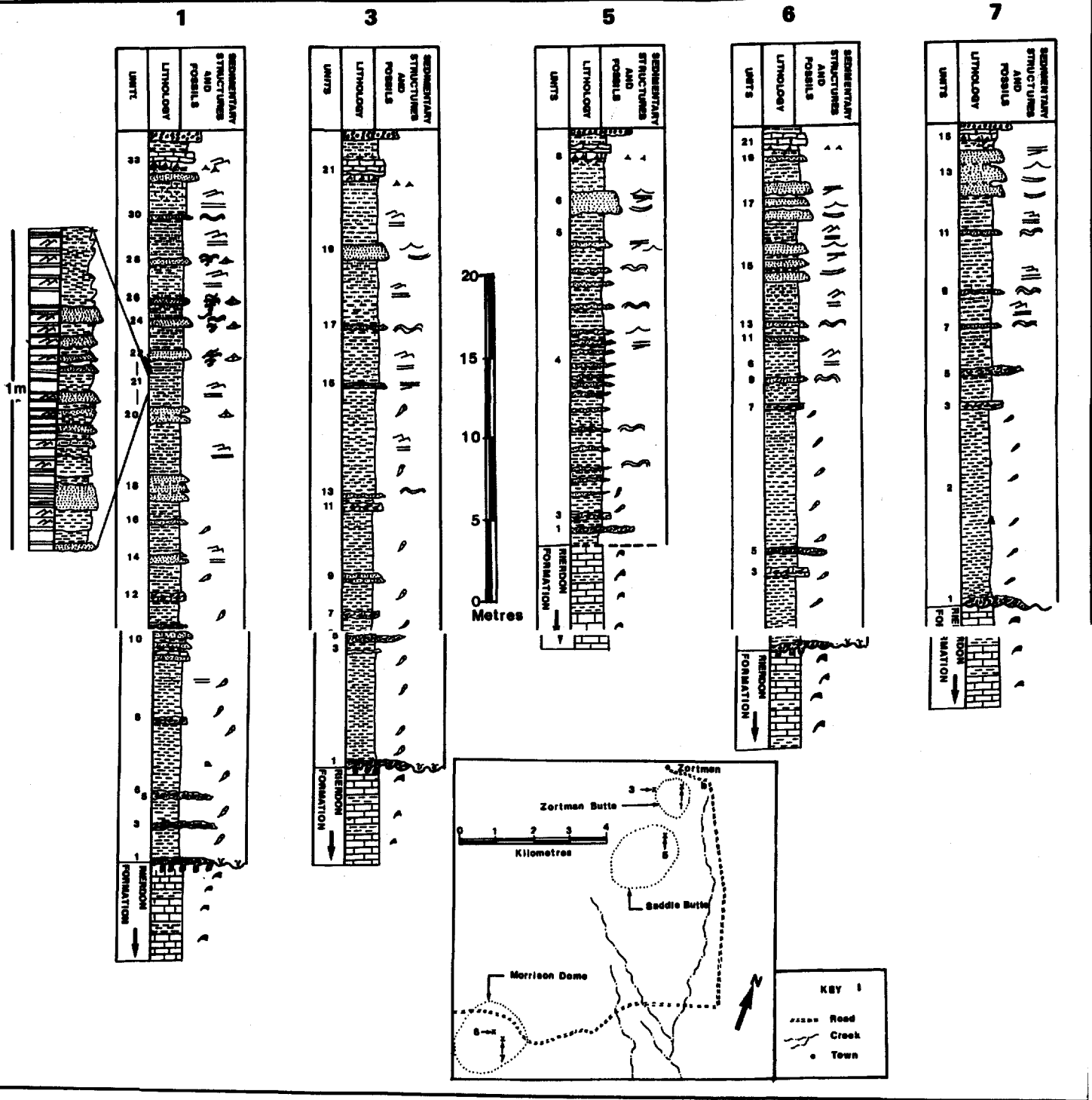
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Fig. 12. Stratigraphic sections - Swift Formation - measured at Zortman Butte (1 & 3), Saddle Butte (5), and Morrison Dome (6 & 7).
For location see attached map.



LITHOLOGY

	Sandstone		Biosteeic limestone (coquina)
	Siltstone		Belemnite/chert and limestone conglomerates
	Shale		Concretionary limestone
	Limestone		Calcareous siltstone
	Sandy limestone		Manganiferous siderite nodules of the Morrison Formation

EXPLANATION

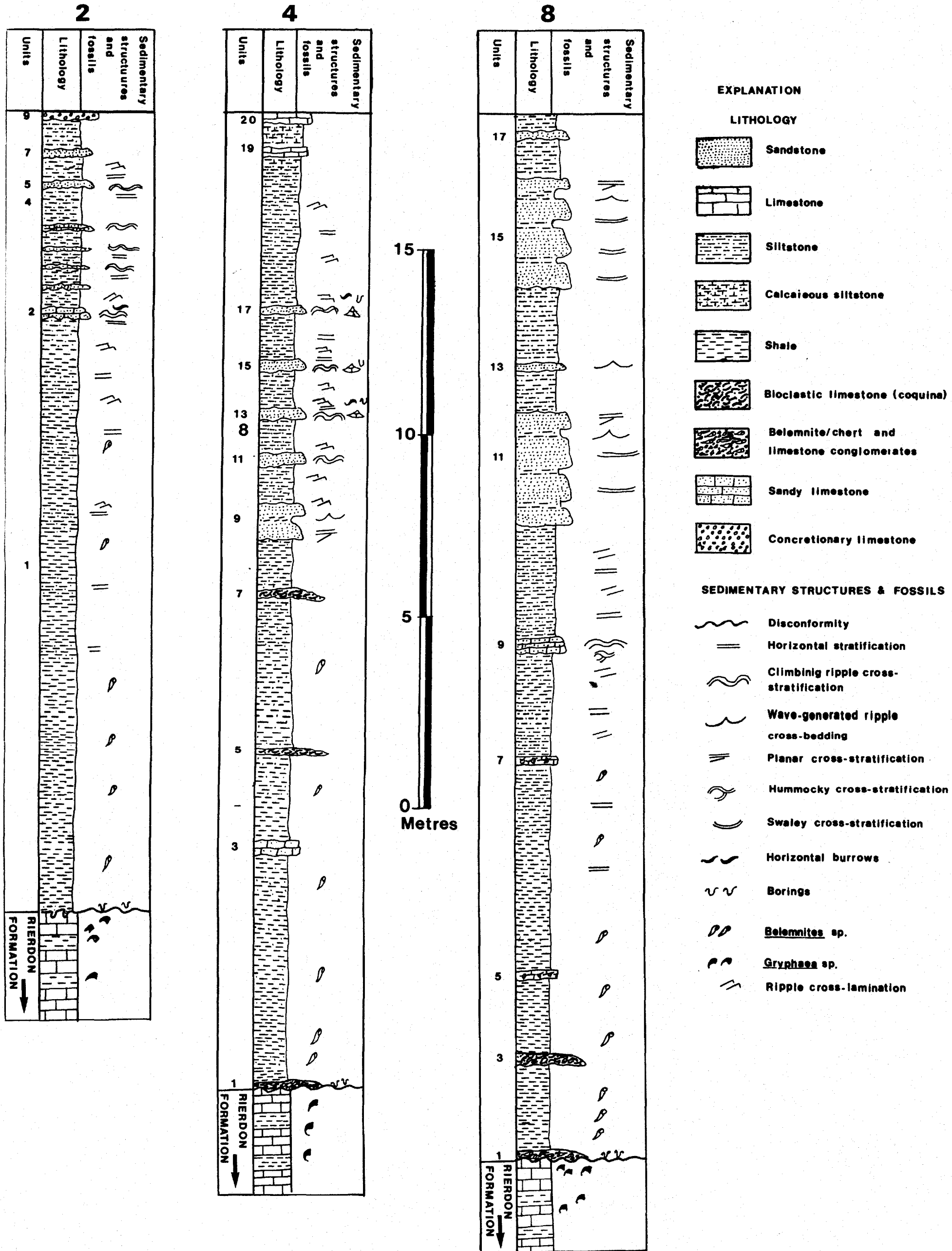
	Horizontal lamination
	Planar cross-bedding
	Ripple cross-lamination
	Hummocky cross-stratification
	Swaley cross-stratification
	Climbing ripple cross-lamination
	Wave-ripple cross-bedding
	Fault

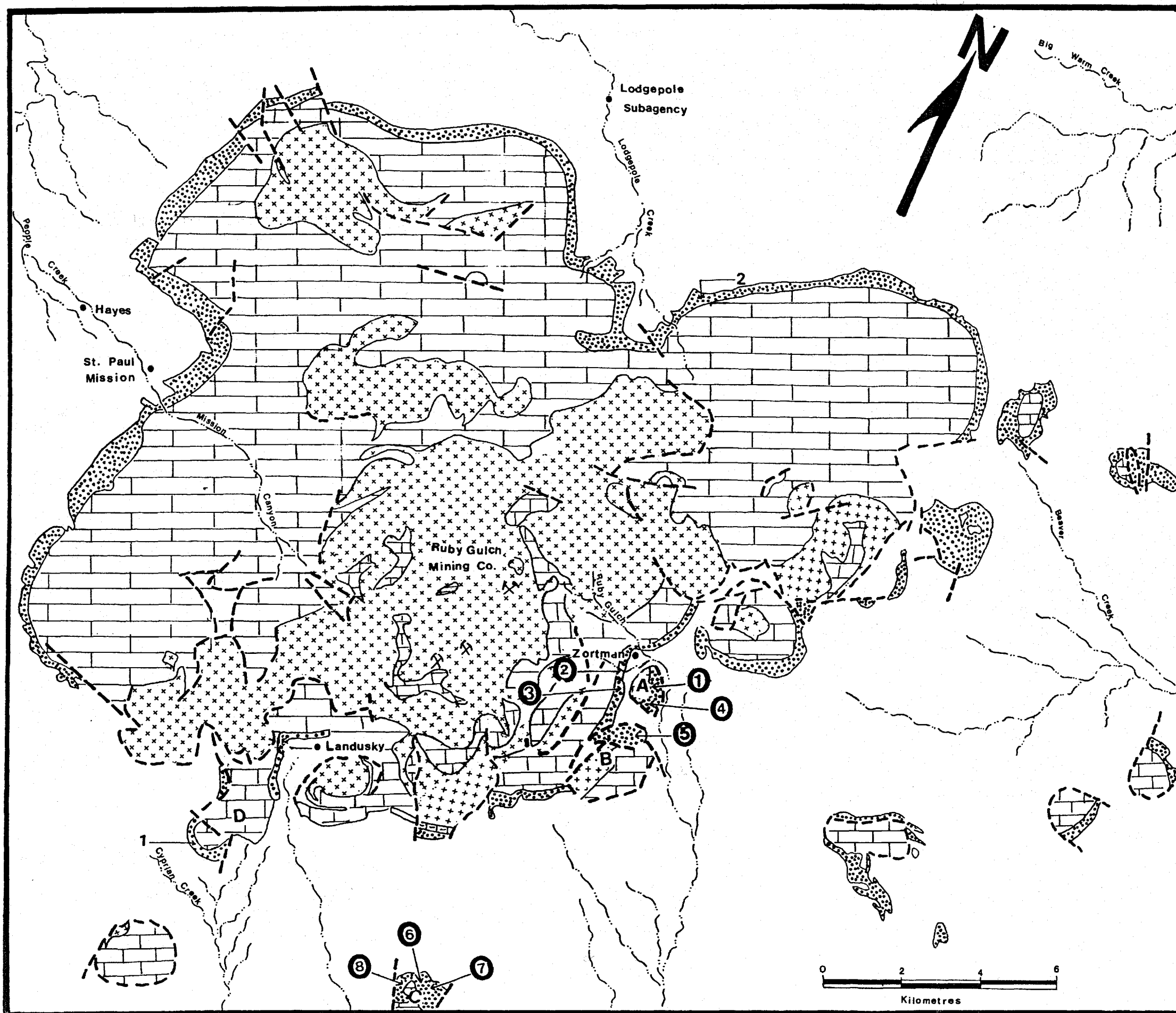
SEDIMENTARY STRUCTURES AND FOSSILS

	Disconformity
	Horizontal burrow
	Berings
	Bivalve moulds
	Nette impressions
	Belemnites sp.
	Oryzaea sp.

Appendix 1. Stratigraphic sections of the Swift Formation exposed at the Zortman Butte area

(2 & 4) and Morrison Dome (8). For location see Fig. 3.





Explanation



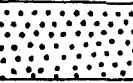
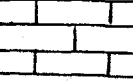


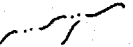
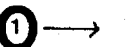



-  Syenite Porphyry (Tertiary)
-  Post-Jurassic rocks (Predominantly shales & sandstones)
-  Jurassic rocks (include three formations—two marine and one, youngest, non-marine). These are, in ascending order, Rierdon, Swift, & Morrison Forms.
-  Pre-Jurassic rocks (Predominantly limestones and dolomites with minor shales and sandstones)
-  Fault
-  Geological boundary
-  Creek "Khor"
-  Location of measured sections
-  Location of checked outcrops
-  Town
-  Zortman Butte, Saddle Butte, Morrison Dome, and Stage Route Butte respectively

Fig. 3. Simplified geological map of the Little Rocky Mountains (modified from Knechtel, 1959).