



Facies relationships and provenance of the Swift Formation (Jurassic) southwestern Montana
by Norman Albert Fox

A thesis submitted in partial fulfillment of the requirements for the degree of MASTER OF SCIENCE
in Earth Science

Montana State University

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Abstract:

A study of the Upper Jurassic Swift Formation indicates that the Belt island uplift of west-central Montana provided more of the sediment composing that formation than has previously been thought. The study area extends north and west of Yellowstone Park and includes the southern flank of the Big Belt Mountains.

The Swift is a moderately well-sorted, very-finegrained, calcareous, quartz sandstone. It is rich in glauconite and in black and brown chert similar to that found in the Madison and Phosphoria Formations respectively. Oysters and pelecypods are abundant throughout the Swift Formation and constitute the major faunal elements. Several generalizations may be made about the Swift Formation in area. Although it thins to the north and pinches out near the southern margin of Belt island, it forms a depositional blanket over most of the study area. Carbonate sedimentation increases in the southern part of the study area culminating in an oolite-rich facies. Swift sandstones in the northern part of the study area are coarser than those to the south and are characterized by an extensive basal conglomerate containing many angular pebbles and cobbles. Pebble angularity decreases to the south. These generalizations point to a shallow, marine shelf environment of deposition. The northern Swift rocks are interpreted as the deposit of a wide littoral zone on the rocky southern shore of Belt island. Angularity of pebbles in the Swift conglomerates plus the abundance of chert indicate that Belt island was a local source of clastic material. Lower-energy environments to the south were far enough away from that source to allow abundant carbonate sedimentation.

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Approved:


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ABSTRACT

A study of the Upper Jurassic Swift Formation indicates that the Belt island uplift of west-central Montana provided more of the sediment composing that formation than has previously been thought. The study area extends north and west of Yellowstone Park and includes the southern flank of the Big Belt Mountains.

The Swift is a moderately well-sorted, very-fine-grained, calcareous, quartz sandstone. It is rich in glauconite and in black and brown chert similar to that found in the Madison and Phosphoria Formations respectively. Oysters and pelecypods are abundant throughout the Swift Formation and constitute the major faunal elements. Several generalizations may be made about the Swift Formation in area. Although it thins to the north and pinches out near the southern margin of Belt island, it forms a depositional blanket over most of the study area. Carbonate sedimentation increases in the southern part of the study area culminating in an oolite-rich facies. Swift sandstones in the northern part of the study area are coarser than those to the south and are characterized by an extensive basal conglomerate containing many angular pebbles and cobbles. Pebble angularity decreases to the south. These generalizations point to a shallow, marine shelf environment of deposition. The northern Swift rocks are interpreted as the deposit of a wide littoral zone on the rocky southern shore of Belt island. Angularity of pebbles in the Swift conglomerates plus the abundance of chert indicate that Belt island was a local source of clastic material. Lower-energy environments to the south were far enough away from that source to allow abundant carbonate sedimentation.

INTRODUCTION

General Problem

The Middle and Late Jurassic was a time of fundamental changes in depositional character in North America. With the start of Late Jurassic time, significant carbonate and evaporite deposition effectively ceased on the craton (Dott and Batten, 1967, p. 363). Widespread, shallow epeiric seas remained the regional environment of deposition in the western interior of the United States, and a highly variable group of clastic, epeiric sea type sediments were deposited, the complexity of which "almost defies analysis" (Dott and Batten, 1967). This complexity not only applies to the facies variations within the Upper Jurassic sediments, but also to the source areas of those sediments. Before problems involving facies and source areas in the Upper Jurassic of Montana can be clearly defined and tackled, the regional stratigraphic framework of the Upper Jurassic of the western interior of the United States must be established.

Marine Jurassic Stratigraphy

The Jurassic System of the western interior of the United States is represented by four transgressions and

regressions, occurring in the Lower, Middle, and Upper Jurassic (Imlay, 1957; Peterson, 1957). The first three transgressive-regressive sequences are responsible for the Lower Jurassic (Hettangian-Toarcian) Fernie Group of British Columbia and Alberta, the Middle Jurassic (Bajocian-Bathonian) Sawtooth-Piper Formations of Montana, and the Upper Jurassic (Callovian) Rierdon Formation of Montana (Fig. 1). Each succeeding transgression of the Jurassic seas extended farther south than the last. The maximum extent of the Upper Jurassic seaway is shown in Figure 2. The western boundary of the seaway is obscure due to post-Jurassic uplift and thrust faulting in western Montana.

Upper Jurassic (Oxfordian) Sedimentation

The Upper Jurassic (Oxfordian) Swift Formation was deposited during the fourth transgression of the Jurassic sea following a lengthy (1-1.5 m.y.) period of erosion that occurred at the close of Rierdon time. Near the end of Oxfordian deposition, the sea withdrew from southwestern Montana as a result of Nevadan Orogenic uplifts to the west (Peterson, 1957). Imlay (1957) stated that

M. Y. B. P.	EUROPEAN AGES	SOUTHERN ALBERTA	EASTERN IDAHO	NORTH- EASTERN UTAH	MONTANA W. -E.	CENTRAL WYOMING	BLACK HILLS	
150	KIMMERIDGIAN	LOWER KOOTEN- AY	MORRISON	MORRISON	MORRISON	MORRISON	MORRISON	
	OXFORDIAN	GROUP	STUMP	CURTIS	SWIFT	UPPER SUNDANCE	SUNDANCE RED- WATER	
160	CALLOVIAN					RIERDON	LOWER	LOWER
	BATHONIAN							
170	BAJOCIAN			TWIN CREEK	CARMEL	SAWTOOTH - PIPER		
	TOARCIAN	FERNIE						
180	PLEINS- BACHIAN		NUGGET	NUGGET				
	SINEMURIAN			NAVAJO				
HETTANGIAN				KAYENTA WINGATE			NUGGET	NUGGET

Figure 1. Correlation Chart, Jurassic System.
Adapted from: Schmitt (1953), p. 358,
and Van Eysinga (1978).

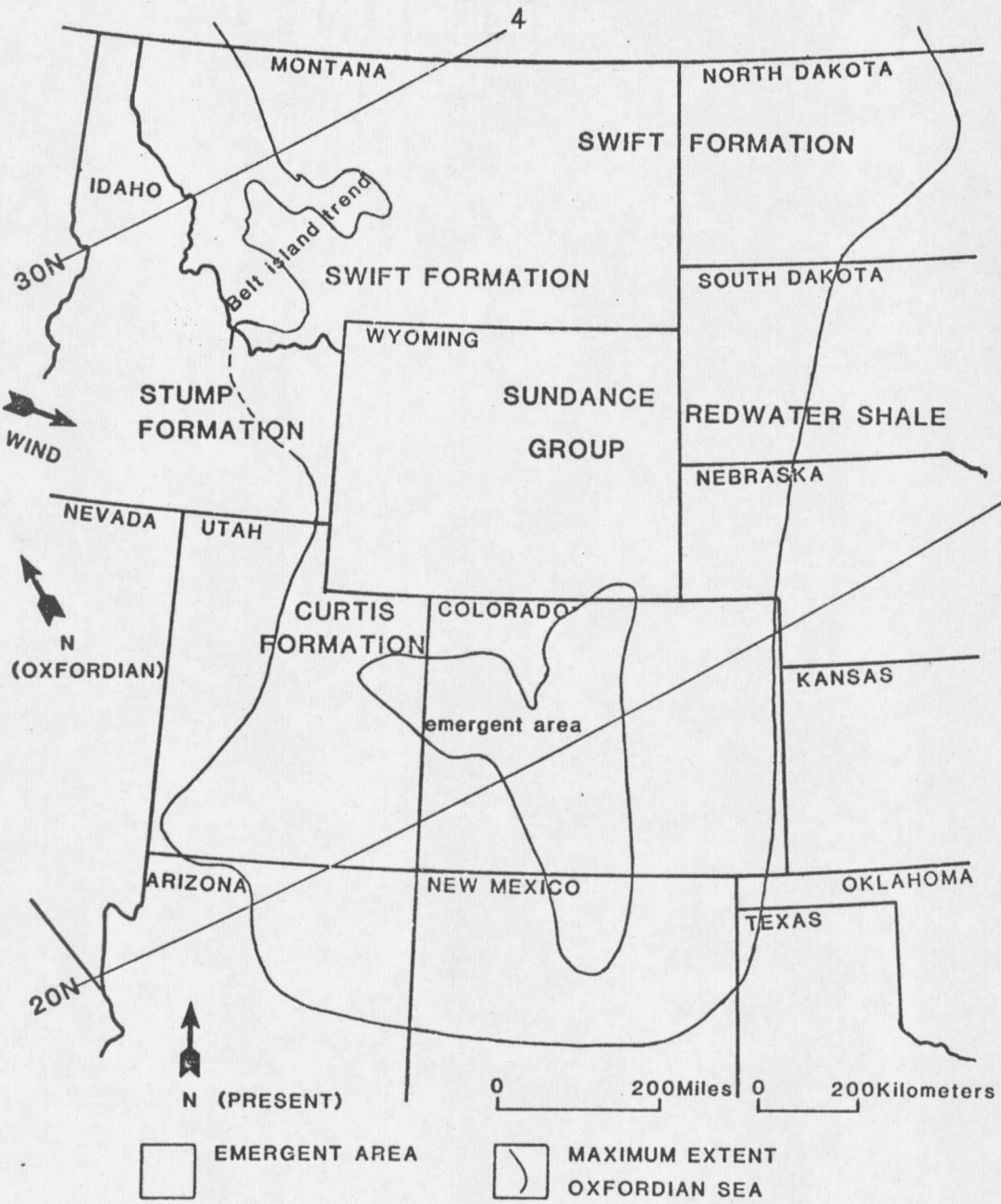


Figure 2. Maximum Extent, Upper Jurassic (Oxfordian) Seaway. After: Peterson (1972), p. 184.; Imlay (1947), p. 476.; and Dott and Batten (1967), p. 353.

there is no evidence for the existence of marine waters in the western interior after the close of the Oxfordian.

The Swift was deposited on a stable to unstable shelf in southwestern Montana (Schmitt, 1953), roughly similar to the present southeastern shelf of the United States. Water depth was probably [considerably] less than 300 feet and the sea was locally dotted with islands and shallowly submerged areas (Imlay, 1957). Some of these islands occurred on a tectonic positive area in west-central Montana that influenced sedimentation throughout the Middle and Upper Jurassic. This positive trend is known as Belt island or the "Belt Island Trend" (Imlay and others, 1948; Peterson, 1957).

At the time of Swift deposition, southwestern Montana was located between 35 and 40 degrees north latitude (Firstbrook and others), (Fig. 2). With Montana at this latitude, winds would have generally blown from the west (Dott and Batten, 1971). The shaded areas in Figure 2 representing Belt island would have been located down wind of highlands of the Sonoma Orogeny to the west.

Transgression of the Oxfordian seas was more widespread than the earlier Jurassic seas, resulting in widespread normal marine conditions within the western

interior (Imlay, 1957; Peterson, 1957). The Swift, Stump, and Curtis Formations of Montana, Idaho, and Utah are similar in general lithologic character (Imlay, 1957), consisting mostly of calcareous, glauconitic sandstones. To the east, the sandy character diminishes and finer grained deposits of the Sundance Formation of central Wyoming and the Redwater Shale of the Black Hills represent Oxfordian deposition (Fig. 1 and 2).

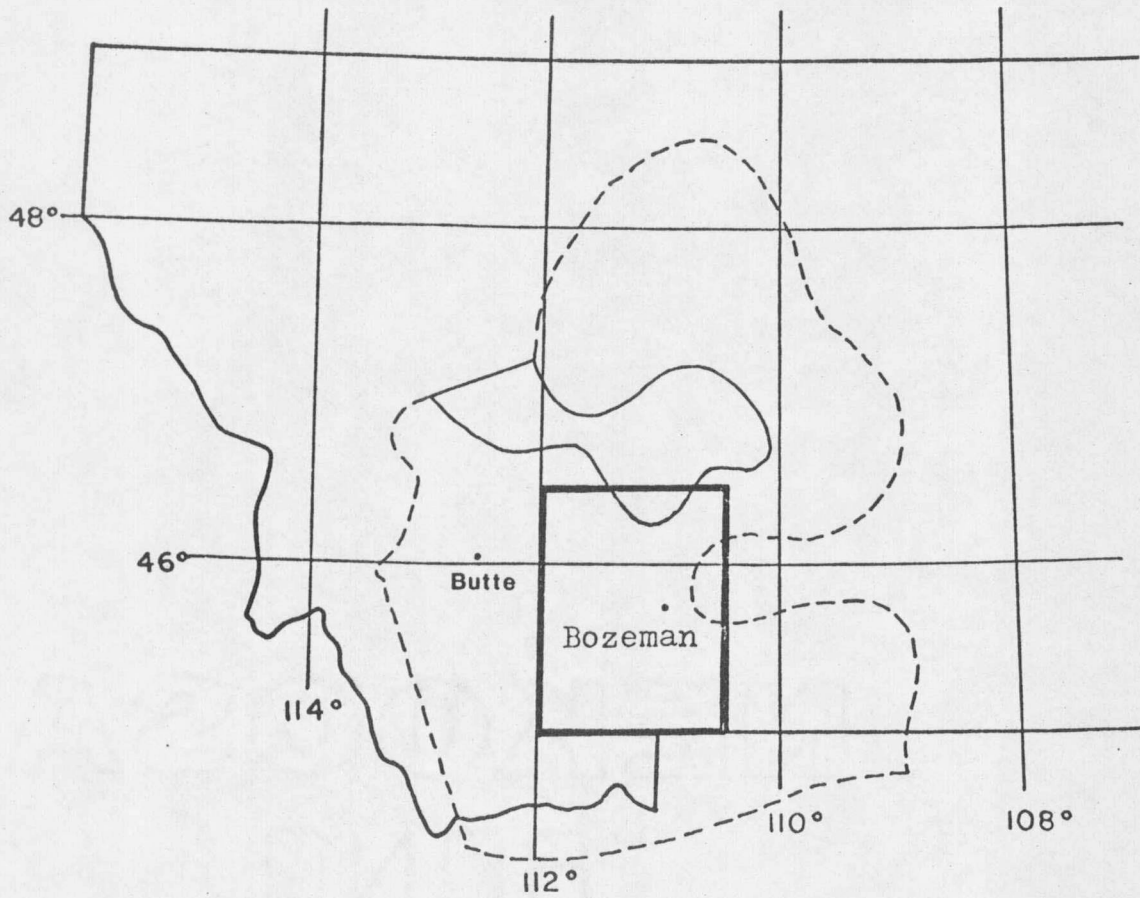
The duration of Oxfordian sedimentation in southwestern Montana is unclear. According to the Geological Society of London Time Scale (1964), the duration of the Oxfordian Age is 6 million years (Firstbrook and others). However, Brenner and Davies (1974) indicated that southwestern Montana was the last area in Montana to be inundated by the Oxfordian sea and the first area in Montana to be exposed during the regression of the sea near the close of the Oxfordian. Deposition of the Swift in southwestern Montana, therefore, probably represents only a portion of the Oxfordian Age. In any case, a comparison of the duration of the Oxfordian with the average thickness of Swift deposits in southwestern Montana indicates a low rate of sediment accumulation in southwestern Montana.

Purpose

The geometry and environment of deposition of the Swift Formation have been described in some detail in the literature on a regional scale. There is general agreement that the Swift is a shallow-marine shelf deposit, and that a tectonic positive area (Belt island) existed in west-central Montana. However, two basic unsolved problems exist concerning the paleogeography of Swift seas and the provenance of Swift sediments. These are: (1) Was Belt island an emergent land mass during the Oxfordian? and (2) if Belt island was emergent, did it serve as a significant sediment source for the Oxfordian deposits of southwestern Montana. Belt island is well established in the literature and is represented in west-central Montana by a depositional pinchout of the Swift Formation (Fig. 2). The zero edge of the Swift in this area may correspond to the shoreline of Belt island during the Oxfordian (Schmitt, 1953).

Fundamental differences of opinion exist regarding the emergent or submergent character of Belt island. According to Peterson (1957), (1) Belt island was not a major source of clastic sediment during the Jurassic and

(2) Belt island had little influence on Oxfordian sedimentation. He pointed out the extreme sandiness of the Swift rocks in western Montana as evidence of the major sediment source in western Montana and Idaho. Brenner and Davies (1974) acknowledged that Belt island was a local sediment source but indicated high relief areas in western Montana as the major sources of Swift sediment (Fig. 2). Schmitt (1953) found that both grain size of Swift rocks and prominence of disconformities within the Swift increased toward Belt island. He used these trends to interpret Belt island as an "active landmass during most of Swift time". Klemme (1947) proposed the existence of two emergent islands on the Belt island trend and described them in some detail on the basis of depositional thinning and facies relationships. In his view, these two islands were locally important as sediment sources during the Oxfordian. In an attempt to solve the two problems outlined above, and answer the questions arising from them, Swift rocks in southwestern Montana were studied with two goals in mind. These are: (1) The establishment of a group of lithofacies that describe the rocks in southwestern Montana, and (2) a reconstruction of the depositional history of the Swift Formation in south



--- > 75% Sandstone

Figure 3. Study Area Location. Heavy black line surrounds study area boundary.

western Montana. In order to achieve these goals, sedimentologic, stratigraphic, and petrographic evidence was collected from Swift rocks in an area of southwestern Montana. The study area for this thesis was chosen for two reasons: (1) Most of the Swift Formation outcrops between the south flank of Belt island and the Yellowstone Plateau are included within this area (Fig. 3), and (2) marked changes in lithologic character of Swift rocks occur from north to south within this area.

Text Organization

The text to follow is organized into a sequence of data, starting with descriptions of the study area and of Jurassic rocks in general in Chapters 2 and 3. Descriptions of Swift Formation rocks follow in Chapter 4. Chapter 5 deals with interpretations of Swift rocks based on models of similar deposits proposed by other geologists. Following the models, evidence based on descriptive data from Swift rocks is given in support of the interpretation. Stratigraphic cross-sections which provide this descriptive data are shown in Plate 1. Chapters 6 and 7 deal with conclusions concerning the provenance and depositional history of Swift sands.

Each lithofacies and subfacies proposed in this study is given a descriptive name in Chapter 4 and an interpretive name in Chapter 5. This may help clarify the separation between description and interpretation characteristic of these two chapters.

STUDY AREA

The boundaries of this area were chosen to include most of the Swift outcrops in southwestern Montana and consist of the 45 and 46 30' latitude lines, and the 110 30' and 112 longitude lines (Fig. 3). Major structural features are the Three Forks Basin (Robinson, 1961), the Bridger Range (McMannis, 1955), and the Crazy Mountain Basin in the northern part of the area, and the Beartooth Uplift, the Gallatin Range, and the Madison Range in the southern part of the study area. The eastern boundary passes east of Livingston, and extends into the Beartooth and Absaroka Ranges to the south. The southern boundary is flanked by the Yellowstone Plateau. To the west, the Greenhorn Range, the Tobacco Root Mountains, and the west flank of the Elkhorn Mountains lie along the study area boundary. The northern boundary cuts through the southern part of the Big Belt Mountains north of Maudlow. Exposures of the Swift are found on the flanks or within the mountains mentioned above and in several other areas within the Three Forks basin.

REGIONAL STRATIGRAPHY OF THE ELLIS GROUP

Historical Review

Cobban and others (1945) designated a type section for the Ellis Formation at Rocky Creek Canyon, about eleven kilometers southeast of Bozeman. The type section consists of 88 meters of sedimentary rocks of various lithologies. Several months later, Cobban (1945) elevated the Ellis Formation to group rank and named the Sawtooth, Rierdon, and Swift Formations. The type sections for the Sawtooth and Rierdon Formations are located in Rierdon Gulch, roughly 16 kilometers west of Choteau, Montana and both are 41 meters thick at this location. The Swift Formation is 41 meters thick in its type section at Swift Reservoir, roughly 9 kilometers west of Dupuyer, Montana.

Subcrop and Supercrop Relationships

Southward regional tilting of rocks in western and central Montana, accompanied by a period of regional erosion and base leveling in the early Jurassic, resulted in beveling of Mesozoic and Paleozoic rocks in Montana (McMannis, 1965). Moritz (1951) estimated that more than 300 meters of sedimentary rock was eroded from southwestern Montana during this erosion period.

The oldest Paleozoic rocks exposed during pre-Ellis erosion are the Mississippian Madison Group Limestones in north-central Montana (Fig. 4). To the south, progressively younger Paleozoic rocks were exposed in roughly east-west trending bands across Montana. Lower Triassic rocks are the youngest rocks exposed and include the Thaynes, Woodside, and Dinwoody Formations in extreme southwestern Montana and the Chugwater (Spearfish) Formation of south-central Montana. The Ellis Group is conformably overlain by the Morrison Formation which commonly consists of variegated siltstone, shale and lenticular sandstone beds (Suttner, 1969). The contact between the Morrison and Swift Formations is normally easy to distinguish due to the presence of glauconite and chert in the Swift rocks (Suttner, 1969). There are, however, areas in which non-glauconitic, fine-grained sandstone of the Swift Formation is overlain by fine grained Morrison sandstone. In these areas, the Swift-Morrison contact is less easily defined, although the reddish color, and fine-grained, "tighter" fabric of the Morrison sandstones help to resolve this problem.

Ellis Group Formations

Sawtooth and Piper Formations

The Sawtooth Formation extends throughout west-central and southwestern Montana and grades eastward into the Piper Formation (Fig. 1). In west-central Montana, the Sawtooth consists of limestone, sandstone and shale ranging in thickness from a zero edge in west-central Montana to 60 meters in southwestern Montana (Peterson, 1957). Sawtooth sandstone, probably derived from northwestern and west-central Montana (Peterson, 1957), becomes finer-grained and grades into limestone to the south and east. Normal marine conditions dominated sedimentation during Sawtooth deposition.

The Piper Formation of central and eastern Montana extends into eastern North Dakota and consists of 120 meters of redbeds, anhydrite, and limestone deposited in a restricted marine environment. Belt island and the Sheridan arch (Fig. 5) were largely responsible for the restricted conditions in central and eastern Montana at this time (Peterson, 1957).

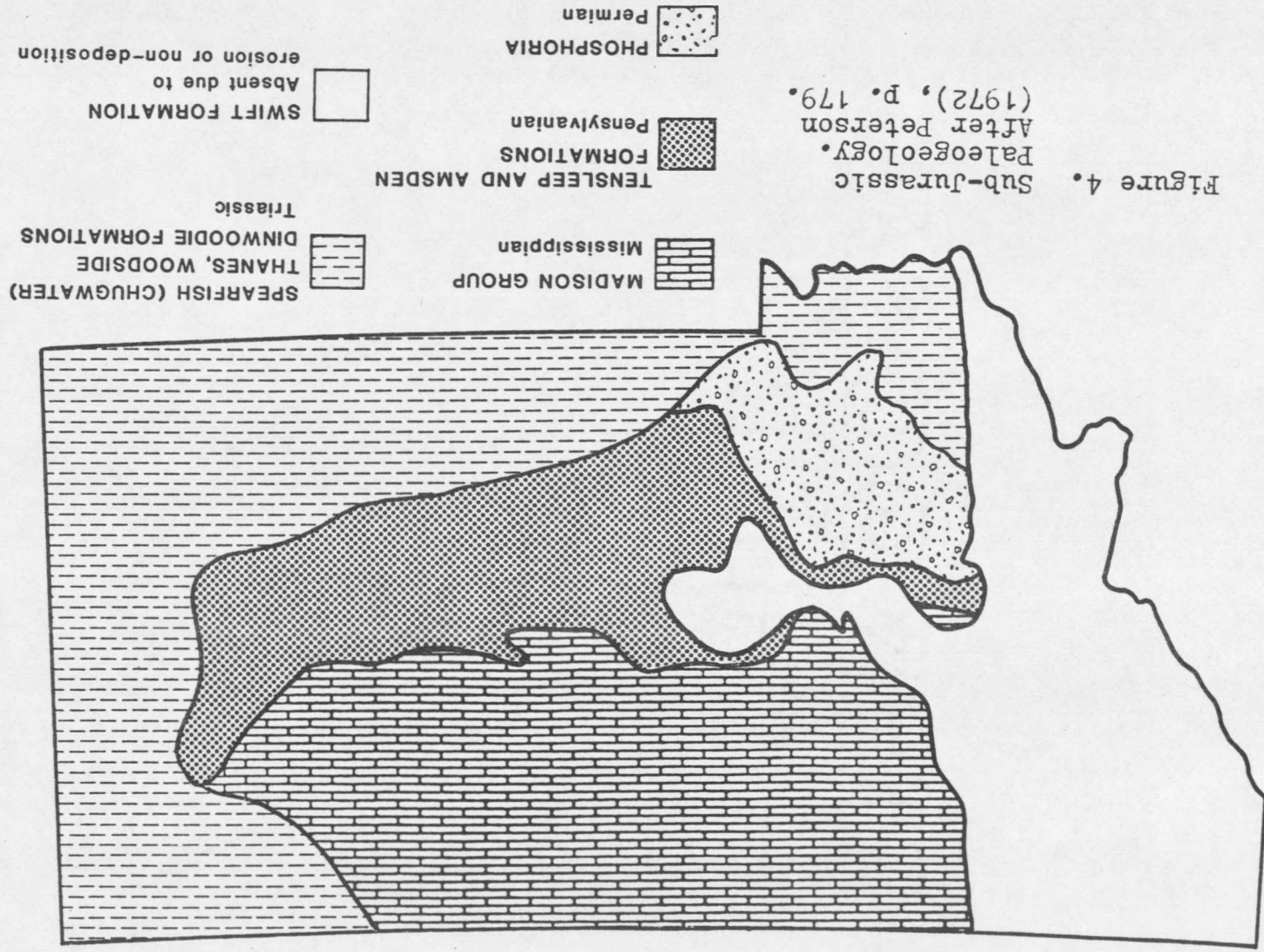


Figure 4. Sub-Jurassic
 After Peterson
 Paleogeology.
 (1972), p. 179.

Rierdon Formation

The Rierdon Formation occurs throughout Montana and North Dakota, conformably overlying the Sawtooth and Piper Formations in most areas. The Rierdon ranges in thickness from a zero edge in west-central Montana to 100 meters in northeastern Montana (Peterson, 1957). In west-central Montana, Rierdon deposits consist of dense to oolitic, locally sandy limestone. Sand grains in the Rierdon of southwestern Montana may have been derived from Sawtooth or Paleozoic rocks from Belt island (Klemme, 1947). Rierdon deposits of the Alberta trough in northwestern Montana (Fig. 5) consist of dark shale and evaporites similar to those of the Piper Formation (Peterson, 1972). In north-central and central Montana, the Rierdon grades into shaly limestone and calcareous shale due to the decrease in carbonate deposition in eastern Montana (Peterson, 1957). Toward the end of Rierdon deposition, Belt island was uplifted and the sea withdrew from Montana (Peterson, 1957). This resulted in a regional shallowing and "silting-in" of the sea. Rierdon deposition during this regression changed from fine-grained limestone and shale to coarser grained,

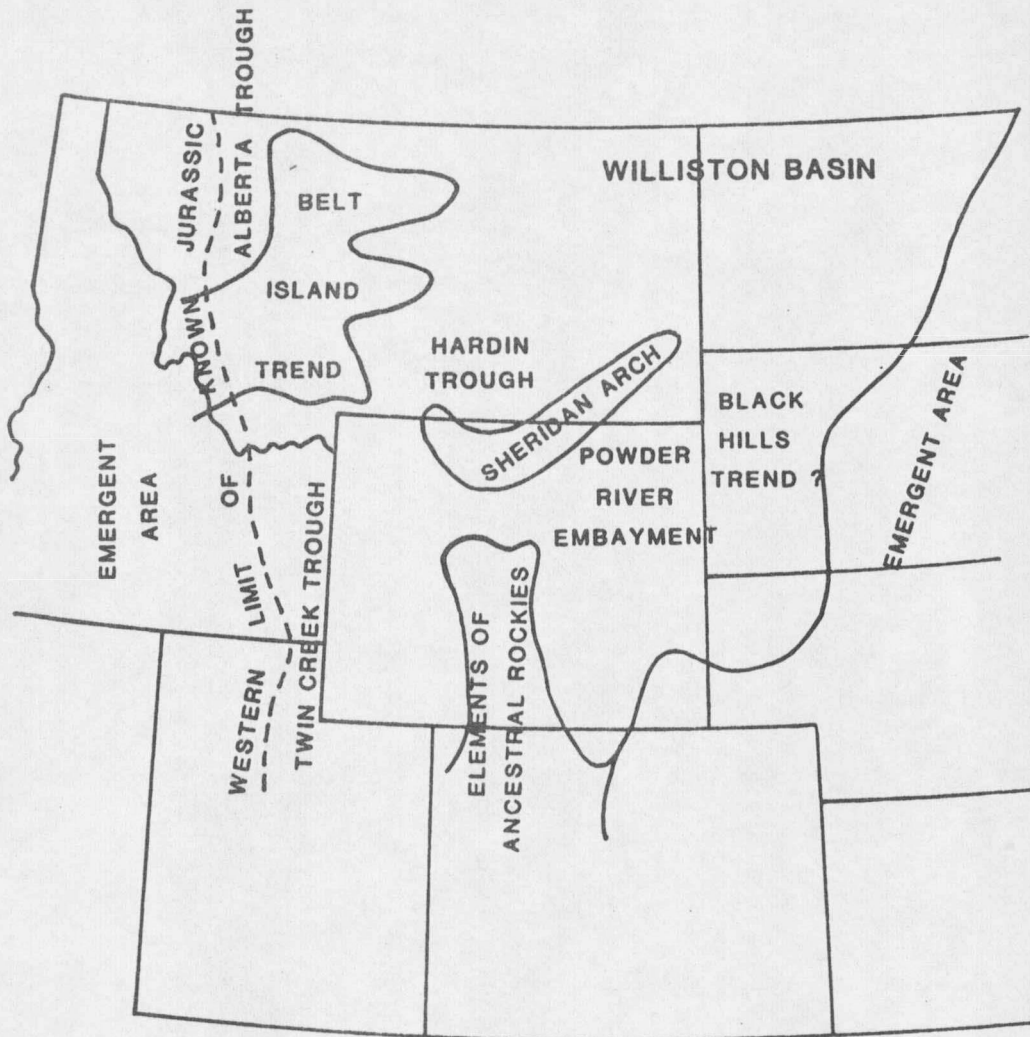


Figure 5. Jurassic Paleotectonic Elements. "Belt Island Trend" is shown here at maximum size during Jurassic time. After Peterson (1957), p. 403.

oolitic, "terminal clastic" sediments marking the close of Rierdon deposition (Peterson, 1957):

Swift Formation

The Swift Formation is the most widespread of the Ellis Group formations (Fig. 6). It is similar in extent to the Rierdon Formation but extends farther into eastern North Dakota. Correlative units are known as far east as central South Dakota and Nebraska (Peterson, 1972). The Swift is absent from the area over Belt island which was diminished in size during the Oxfordian due to the extent of the Oxfordian transgression. In the Williston Basin, the Swift reaches a maximum thickness of 120 meters. In northwestern Montana, it ranges in thickness from 3 to 7.5 meters (Peterson, 1957). In west-central and southwestern Montana, the Swift consists of fine-grained, cherty, glauconitic, calcareous sandstone deposited under normal marine conditions (Peterson, 1957). In northwestern Montana, the Swift assumes a two-member character with gray marine shale overlain by fine-grained, flaggy, glauconitic sandstone (Cobban, 1945). In eastern Montana, the two-member character persists. The shale member increases in thickness to the east, becoming increasingly

sandy in its lower part as the eastern depositional edge is approached.

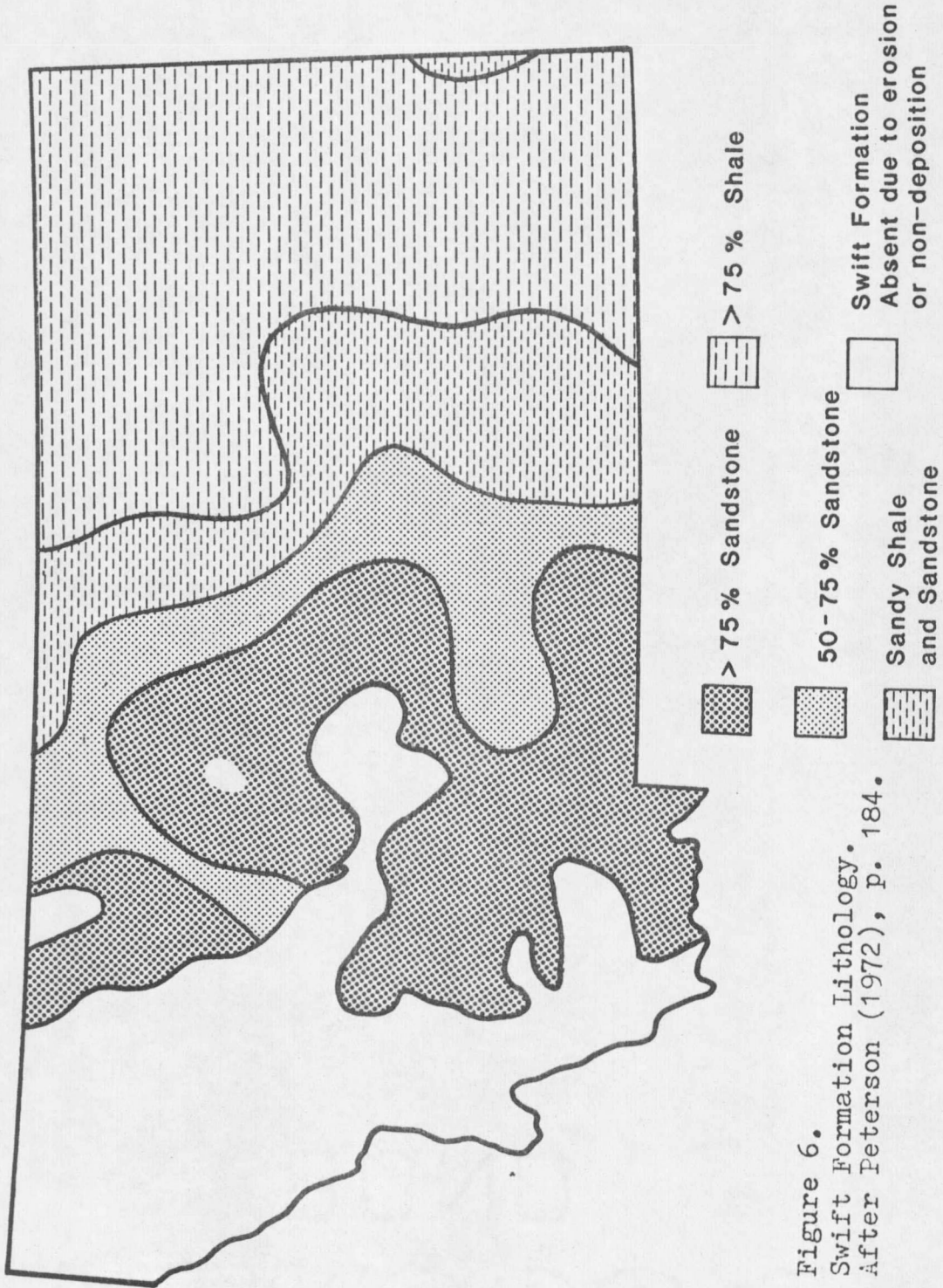


Figure 6.
Swift Formation Lithology.
After Peterson (1972), p. 184.

PETROGRAPHY

Generation and Analysis of Petrographic Data

Measured stratigraphic sections and hand specimens of Swift Formation outcrops form the foundation of all petrographic data used in this study. The locations of the nineteen stratigraphic sections which I measured are shown as solid circles in Figure 7 and are shown in detail in Plate 1. The stratigraphic interpretations discussed in Chapter 5 are largely based on data collected from these stratigraphic sections. Much supportive data was gained from twenty-nine stratigraphic sections measured by other geologists. These were found in the literature and in various other unpublished sources and are shown as hollow circles in Figure 7. Hand specimens were collected at changes in lithology from all of my stratigraphic sections and from several of the stratigraphic sections measured by others.

Petrographic studies of Swift Formation rocks include binocular microscope examinations of the hand specimens and petrographic microscope studies of thin sections. Hand specimens were described in terms of their lithology, grain size, and percentages of quartz, chert, allochemical grains, and carbonate cement (Appendix).

Forty thin sections were cut from hand specimens of selected stratigraphic sections. Data collected from the thin sections were used to supplement the binocular microscope studies.

From these field and microscope studies, a group of distinct sedimentary lithofacies emerged. Their names and general descriptions are given in Table 1. Due to the general homogeneity of Swift Formation deposits, the lithofacies differ in subtle ways. This is evident in overlap between the lithofacies in terms of sedimentologic, stratigraphic and petrographic characteristics such as cross-bedding, grain size, lithology, and component distribution. Detailed descriptions of the lithofacies and subfacies are given later in this chapter along with facies maps based on the stratigraphic data.

Literature Review of Swift Facies

Klemme (1947) defined a group of four facies to describe the Swift rocks in the Sixteen-mile Creek area near Maudlow in south-central Montana. Included in this group are the transgressive, marginal, inundative, and regressive facies. The transgressive facies is usually a conglomerate or conglomeratic sandstone with rounded

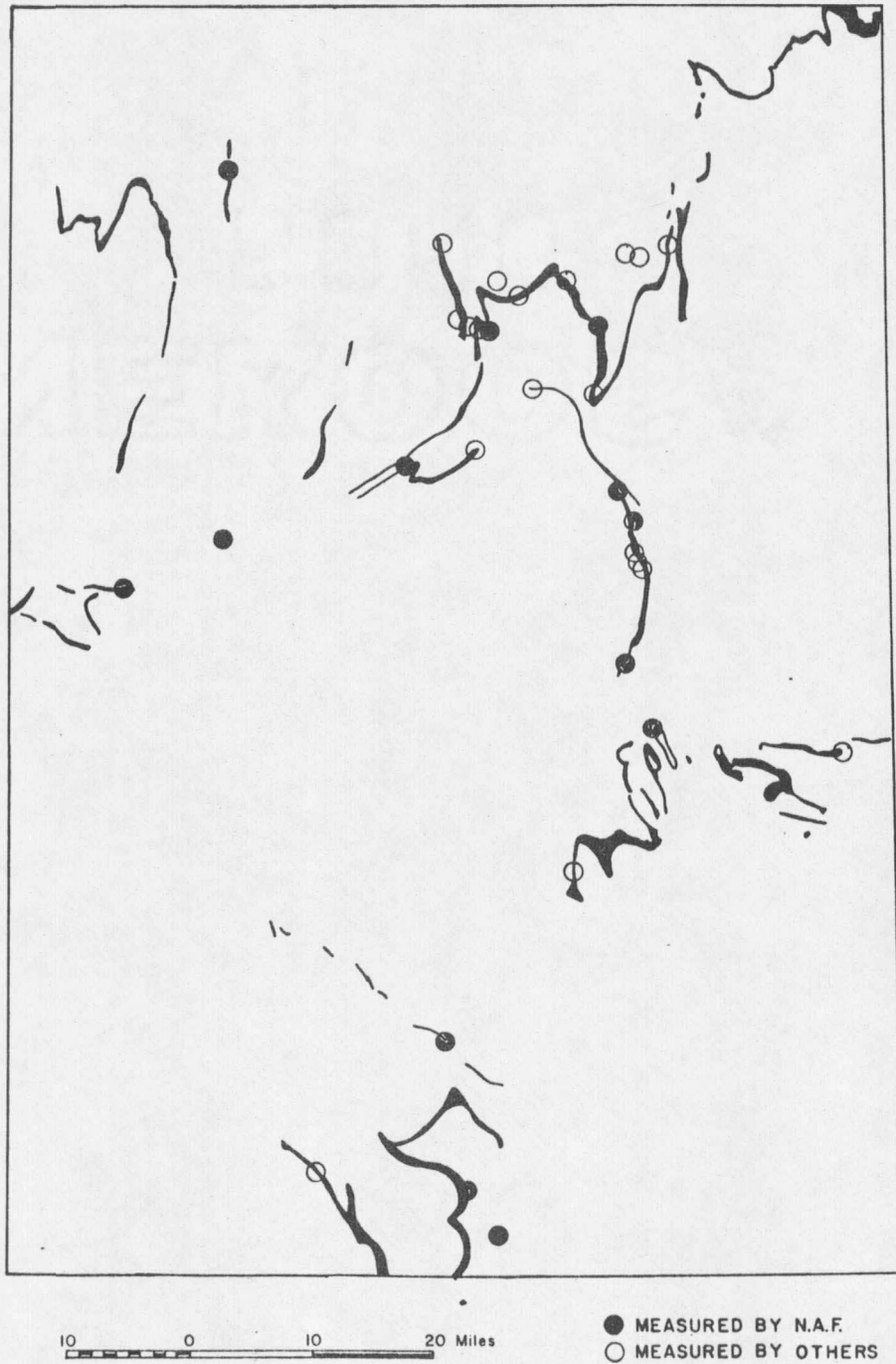


Figure 7. Outcrop Pattern Map of Ellis Group Rocks Within The Study Area. (N.A.F. is N.A. Fox).

DESCRIPTIVE NAME

LITHOLOGY

Calcareous basal
conglomerate facies

Coarse-grained chert-pebble
conglomerate, matrix to grain-
supported, non-fossiliferous
to locally fossiliferous,
horizontal to low-angle cross-
bedding.

Calcareous
sandstone facies

Medium-grained,
coquinoid sub-
facies

Medium to fine-grained sandstone,
chert-rich, non-fossiliferous to
fossil-rich, small to medium-
scale horizontal to cross-bedding,
and trough-bedding.

Fine-grained,
carbonate-rich
lithology

Very-fine to fine-grained sand-
stone, and sandy limestone, highly
calcareous, non-fossiliferous to
locally fossiliferous, small-
scale horizontal to cross-bedding
and ripple bedding.

Fine-grained,
well-sorted
subfacies

Fine-grained sandstone, chert-
poor, well-sorted, sparsely
fossiliferous, small to medium-
scale horizontal to cross-bedding.

Oolitic carbonate
facies

Fine to medium-grained sandy lime-
stone, oolitic, sparsely to highly
fossiliferous, small to medium-
scale trough to cross-bedding.

Table 1. Descriptive Names and General Descriptions of
Swift Formation Lithofacies and Subfacies.

pebbles in a fine-grained, calcareous, glauconitic sand matrix. It is generally similar to my calcareous, basal conglomerate facies in thickness and lithology. Klemme's marginal facies has no counterpart farther south in the thesis area. It is described as a very fine-grained sandstone rich in pyrite, and poor in glauconite and chert. The inundative facies is the most widespread facies in Klemme's study area and correlates closely with similar rocks in my thesis area. This facies is usually a fine-grained sandstone relatively rich in chert and glauconite grains. Klemme describes his regressive facies as a coarse to very fine-grained sandstone rich in pyrite and low in glauconite. He states that it is poorly developed and difficult to identify. No counterpart to this facies was found in my thesis area.

Descriptions of Facies and Subfacies

Calcareous Basal Conglomerate Facies

The basal conglomerate facies occurs throughout the northern half of the study area (Fig. 8) but exists only locally near the southern margin of Belt island (Klemme, 1947). To the south, distribution of the basal facies is

unclear due to lack of stratigraphic control, but it is locally present in the south-western part of the thesis area (Fig. 7).

The facies is best developed in Milligan Canyon near Three Forks (Fig. 8) where it is about 3 meters thick and composes the lower half of the stratigraphic section. In all directions from that area, the basal facies becomes thinner and makes up a relatively small percentage of the stratigraphic sections.

The base of the basal conglomerate facies lies with regional angular unconformity on progressively younger rocks from north to south. In stratigraphic section IC-A, B and C west of Townsend (Plate 1), the basal conglomerate overlies the Shedhorn Member of the Phosphoria Formation. In stratigraphic section IC-D west of Townsend, MC-A, B and C in Milligan Canyon and in two stratigraphic sections described by Klemme (1947) north of Maudlow, the conglomerate facies lies on bedded cherts of the Phosphoria Formation. In all other stratigraphic sections within the study area, the Swift overlies the Rierdon erosion surface, except in several areas near the southern margin of Belt island where the Swift rests directly on either the Sawtooth Formation or the Quadrant

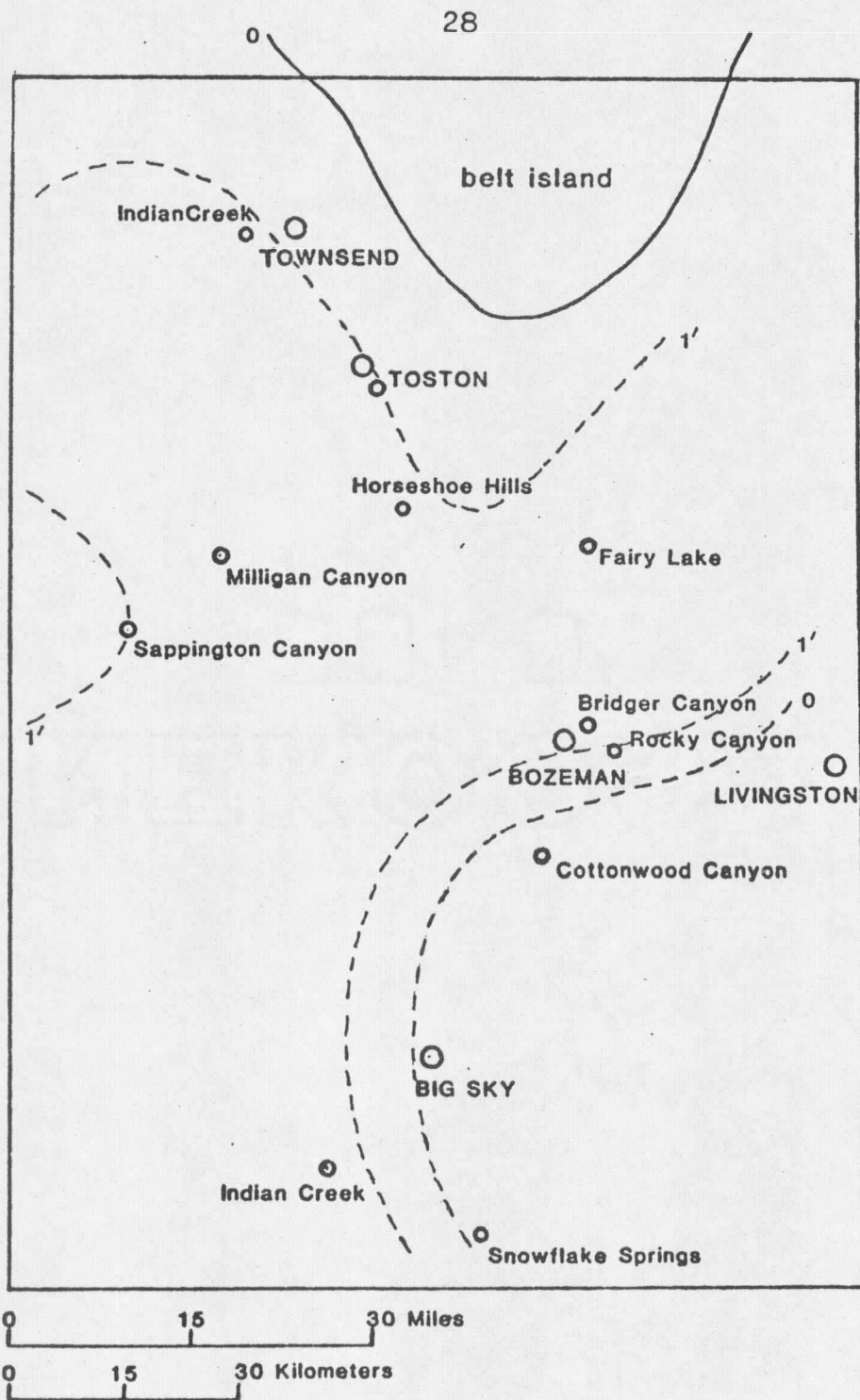


Figure 8. Isopach Map, Calcareous Basal Conglomerate Facies. Contour Interval = 1'.

Formation (Klemme, 1947). In all areas, the basal contact of the Swift is a sharply defined erosional surface. Cut-and-fill structures and crevice fillings are common in the Indian Creek, Milligan Canyon, and Rocky Canyon areas. In stratigraphic section IC-D west of Townsend, basal Swift rocks occur in drape structures over cobbles of Phosphoria bedded chert which lie on the pre-Oxfordian erosion surface.

The basal conglomerate facies is commonly overlain by rocks of the medium-grained/coquinoid subfacies. Exceptions to this are the areas near the southern margin of Belt island where the conglomerate facies is directly overlain by the Morrison Formation (Klemme, 1947).

The basal facies shows significant differences in lithology and stratigraphy from north to south. In the northern part of the study area, the facies consists of coarse- to medium-grained chert pebble conglomerate, often grain-supported at the base and becoming matrix-supported toward the top. In Milligan Canyon, grain-supported or pebble-rich, matrix-supported conglomerate beds alternate with more matrix-supported beds (Plate 1). The chert consists in large part of orange brown, angular to subrounded pebbles commonly found in the Phosphoria

Formation, with lesser amounts of black chert commonly found in the Madison Group Limestones. The black and brown chert pebbles generally show a similar degree of angularity and the black pebbles are slightly more rounded. At the base of section MC-C in Milligan Canyon, the conglomerate is characterized by flat chert pebbles strongly oriented parallel to bedding. South of Milligan Canyon, the conglomerate lithology is less well developed. The pebbles are more varied in their lithology, being not only black and brown chert, but also white quartzite, minor green chert, and minor fine-grained sandstone. The basal conglomerate facies south of Milligan Canyon is matrix-supported conglomerate and conglomeratic sandstone with pebbles composing a relatively small percentage of the rock. Average pebble size and angularity generally decrease from north to south in this facies.

The paleontologic character of the basal facies varies considerably. Conglomerate beds in the northern and western parts of the study area are generally more fossiliferous. Conglomerates in the central part of the area contain a variable assemblage of oysters and thick shelled bivalves. In the Horseshoe Hills and Fairy Lake

sections, horizons of abundant rynchonellid brachiopods occur in the conglomerate beds. In general, fossils in this facies are commonly broken and often lie parallel to bedding. Sedimentary structures in this facies include horizontal bedding and low-angle planar and trough-cross-bedding. Trough-cross-bedding is especially well developed in the Toston and Fairy Lake sections. Apparently no bioturbation occurs in this facies.

Calcareous Sandstone Facies

The calcareous sandstone facies, which includes all predominantly terrigenous clastic sediments above the basal conglomerate, is the most widespread of the three facies of the Swift Formation in the study area. The calcareous sandstone facies rests conformably on the basal conglomerate facies in the northern half of the study area, and rests on the Rierdon erosion surface in the southern half of the area. However, there are exceptions to this. In the northernmost stratigraphic section (IC-B) west of Townsend, the calcareous sandstone facies lies unconformably on the Shedhorn Member of the Phosphoria Formation. The facies is best developed in the southern Bridger Range where it reaches a thickness of 23

meters. In all areas, the calcareous sandstone facies is directly overlain by the Morrison Formation.

The calcareous sandstone facies is divided into two subfacies on the basis of lithology and stratigraphy (Table 1). Although the differences between the subfacies are subtle, they are observable in the field as well as under the microscope.

Medium-Grained/Coquinoid Subfacies

This subfacies is similar in extent to the basal conglomerate facies and is present throughout the study area except in the southeast corner (Fig. 9). However, it is only locally present near the southern margin of Belt island. The subfacies overlies the basal conglomerate facies wherever the latter is developed, but extends farther to the south (Fig. 9).

In the northern and western parts of the study area, this subfacies comprises most of the stratigraphic section. West of Townsend and in the Sappington Canyon and Milligan Canyon areas, this subfacies makes up 50 percent or more of the stratigraphic sections. From the Horseshoe Hills south, it makes up less and less of the stratigraphic sections, pinching out in the area south of Big Sky (Fig. 9).

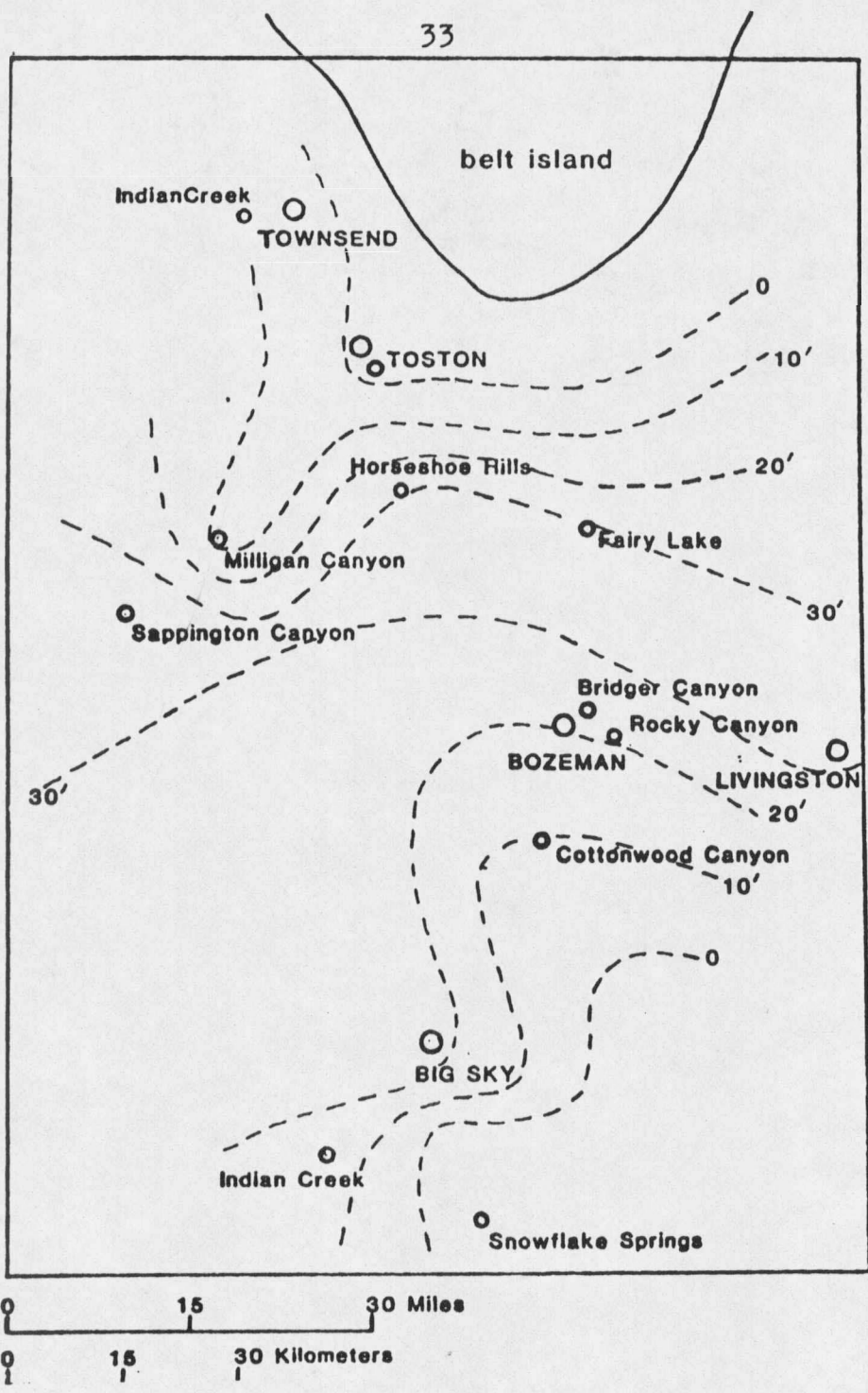


Figure 9. Isopach Map, Medium-Grained/Coquinoid Subfacies. Contour Interval = 10'.

The geometry of this subfacies can be described as a thin, east-west trending lens-shaped body. The thickness ranges from zero near the south margin of Belt island and in the area around Big Sky, to 7.5-9 meters at the Sappington Canyon and Horseshoe Hills sections, and in the central Bridger Range (Fig. 9).

Three lithologies of the medium-grained/ coquinoid subfacies. The course-grained subfacies is characterized by two lithologies (Fig. 12). In Milligan Canyon, the subfacies is composed of medium- to fine grained, pelecypod-rich, coquinoid sandstone, commonly containing brown and black chert sand grains and pebbles and abundant carbonate cement. In all directions from this area, the lithologic character of the subfacies changes. For example, all Swift rocks of the Sappington Canyon section belong to this subfacies and are composed of medium-to-fine-grained unfossiliferous sandstone containing abundant black and brown chert sand grains with a few pebbles and relatively little carbonate cement.

West of Townsend, the subfacies grades laterally from coquinoid sandstone to rocks similar to those at Sappington Canyon. From the Horseshoe Hills south, The

subfacies seems to be dominated by medium-to-fine-grained, sparsely fossiliferous sandstone with little carbonate cement. These rocks are interbedded with coquinooid sandstone similar to that at Milligan Canyon. Both lithologies have relatively poor sorting and are similar in grain angularity. The chert-pebble population ranges from angular to rounded but shows a relative abundance of angular chert pebbles compared with the other subfacies.

Fossil density in this subfacies is high in the coquinooid lithology and low in the cross-bedded sandstone lithology. Fossils consist of abundant pelecypods, fewer brachiopods, and a few gastropods. These fossils are commonly broken, although whole fossils are present, often lying convex-up in the rocks. Trace fossils in the form of burrows occur frequently in the facies in Bridger and Rocky Canyons. North of these canyons, trace fossils are relatively rare. Carbonized wood fragments occur locally in Milligan and Rocky Canyons.

The subfacies is characterized by abundant, low angle trough-cross-bedding, although horizontal bedding is common and is the dominant stratification in some beds. Bedding is often accentuated by fossils lying parallel to it.

An exception to the two basic lithologies discussed above is found in the area south of Toston along the Missouri River. Although lithologically different from the medium-grained/coquinoid subfacies, the rocks in the Toston area are included within this subfacies because they are geographically and stratigraphically surrounded by it. Because of lack of stratigraphic control, the actual extent of this lithology is obscure, but it was not found in any other stratigraphic section within the study area. The lithology may be developed only locally.

The facies reaches a maximum thickness of 4.5 meters. (Fig. 9). It overlies the basal conglomerate facies in two of the stratigraphic sections (Toston A and B), and possibly in the third, although the basal contact of Toston A is covered. In all three stratigraphic sections, the lithology is overlain by the Morrison Formation.

Rocks composing this carbonate rich lithology change rapidly both vertically and horizontally within the area south of Toston. The rocks consist, in general, of highly calcareous sandstone interbedded with sandy limestone. Subangular black and brown chert pebbles commonly occur

throughout these sections. These pebbles, along with limestone pebbles which occur less frequently, are "floating" in sandstone or carbonate matrix. (Plate 1).

Fossils in this lithology are sparse except in oolitic beds near the top of section Toston A (Plate 1). The fossils are thin-shelled bivalves that lie approximately parallel to bedding. No trace fossils or evidence of bioturbation were noted although the massive appearance of many of the beds indicates possible total destruction of bedding by bioturbation. Stratification, where it does exist, consists of thin to medium horizontal beds with minor localized small-scale ripple-cross-bedding and wavy lamination.

Fine-Grained, Well-Sorted Subfacies

This subfacies is developed in areas around the Horseshoe Hills, the Bridger Range, and the Gallatin Range (Fig. 10). It extends as far west as Three Forks, although this is uncertain due to lack of stratigraphic control. Throughout the east-central part of the study area, the medium-grained/coquinoid subfacies grades upward into this fine-grained subfacies. The contact is

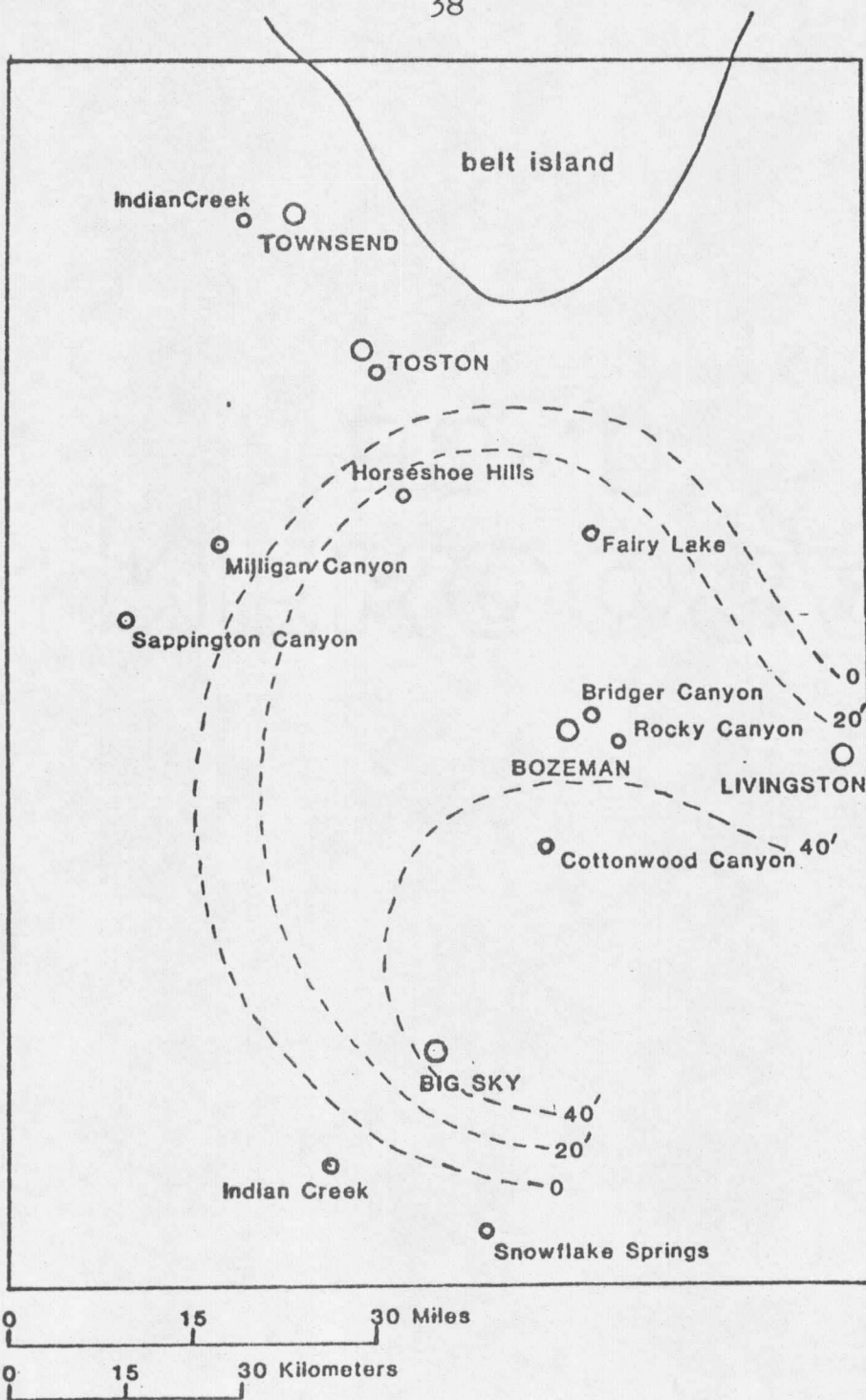


Figure 10. Isopach Map, Fine-Grained, Well-Sorted Subfacies. Contour Interval = 20'.

gradational and difficult to define. Throughout its area of extent, the fine-grained subfacies is conformably overlain by rocks of the Morrison Formation.

The fine-grained, well-sorted subfacies is the thickest of the three subfacies, reaching a maximum thickness of 15 meters in the central Bridger Range. Throughout the area in which it is developed, the subfacies is at least 10 meters thick and makes up greater than 50 percent of the stratigraphic sections (Plate 1).

The rocks of this subfacies are relatively homogeneous lithologically, consisting primarily of quartz sand with smaller amounts of chert sand grains and carbonate cement than the medium-grained/coquinoid subfacies (Fig. 12). The amount of angular sand grains in these rocks is generally similar to that of the fine-grained fraction of the medium-grained/coquinoid subfacies.

Fossils in this subfacies are sparse, and consist of bivalve fragments, burrows and feeding trails, and possible minor bioturbation in the form of mottling. These fossils are restricted to isolated horizons and beds within the subfacies.

The rocks in this subfacies are generally horizontally bedded. Bed sets range from laminae to medium-thick (McKee and Wier, 1953), and give the rocks a flaggy appearance. In several places in the Bridger Range, horizontal bedding alternates with thin to medium sets of low-angle cross-bedding. In other areas, horizontal bedding is poorly developed, giving the rocks a massive appearance.

Cross-Bedded, Oolitic Carbonate Facies

The oolitic carbonate facies is best developed at the southern end of the study area near Snowflake Springs. Although the facies occurs as far north as Big Sky, its actual extent is not known due to lack of stratigraphic control. However, the facies is absent from the Indian Creek section in the Madison Range and the Cinnabar Mountain section in the Gallatin Range to the east (Fig. 11). This indicates that the facies may be restricted to the south-central part of the study area.

The oolitic carbonate facies probably was deposited at about the same time as the basal conglomerate facies and the inundative facies to the north (Fig. 17). At Snowflake Springs, the facies rests conformably on the Rierdon Limestone and is overlain by rocks of the

Morrison Formation. Here it reaches a maximum thickness of about 16 meters.

The lithology of the oolitic carbonate facies is unique compared with other Swift rocks (Fig. 12). Rocks included in this facies are composed almost entirely of calcite in the form of ooids, calcite cement and bioclasts. The remainder of the rock is relatively large, well rounded quartz sand grains and angular to rounded chert sand grains and pebbles. The sand grains and pebbles are poorly sorted and are scattered throughout the rock. Sparry calcite cement composes nearly all of the interstitial material between the framework grains.

Fossils found in this facies consist of thin-shelled bivalves and oysters. In many horizons, beds rich in whole or broken fossils alternate with sparsely fossiliferous beds. Other beds range in fossil density from those having scattered shell fragments to those forming a coquina.

Sedimentary structures are well developed in this facies and are similar to those found in the medium-grained/coquinoid subfacies. Large-scale, low-angle cross-bed sets with tangential bases are common and are

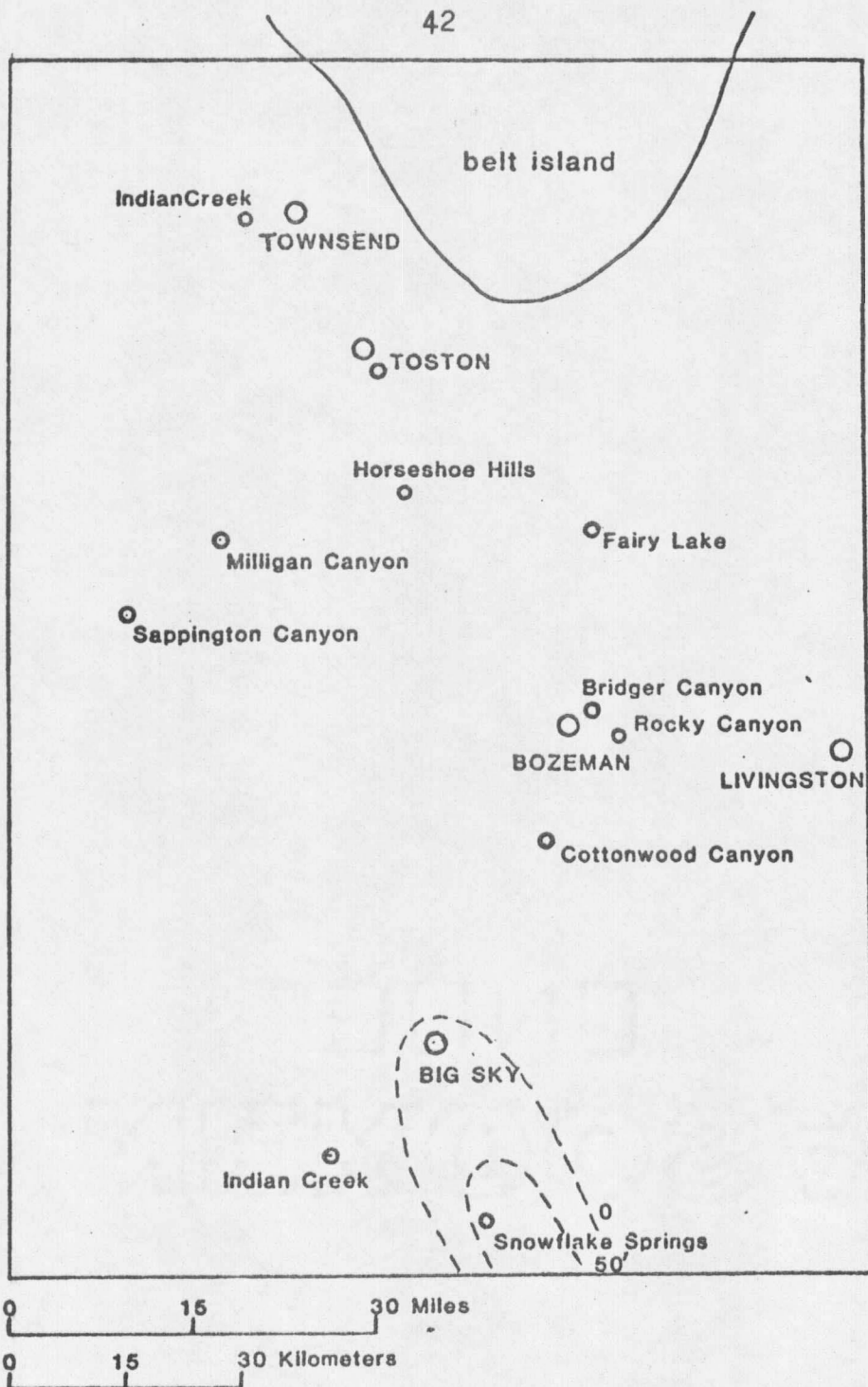


Figure 11. Isopach Map, Cross-Bedded, Oolitic Carbonate Facies. Contour Interval = 50'.

PHOTOMICROGRAPH A. Calcareous Basal Conglomerate Facies. From the base of the Rock Canyon A section (Plate 1). Note limestone pebble on right side of photo.

PHOTOMICROGRAPH B. Medium-Grained/Coquinoid Subfacies, Fine-Grained, Carbonate-rich Lithology. From the Toston B section (Plate 1). Note the abundant carbonate cement.

PHOTOMICROGRAPH C. Medium-Grained/Coquinoid Subfacies, Coquinoid Lithology. From the Sixteenmile Creek area. Note gastropod in lower left corner of photo.

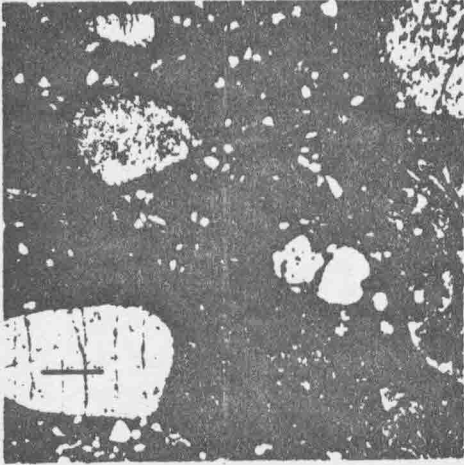
PHOTOMICROGRAPH D. Medium-Grained/Coquinoid Subfacies, Fine- to Medium-Grained Lithology. From the Horseshoe Hills section (Plate 1). Note large, angular chert grain in lower part of photo.

PHOTOMICROGRAPH E. Fine-Grained, Well-Sorted Subfacies. From Horseshoe Hills section (Plate 1).

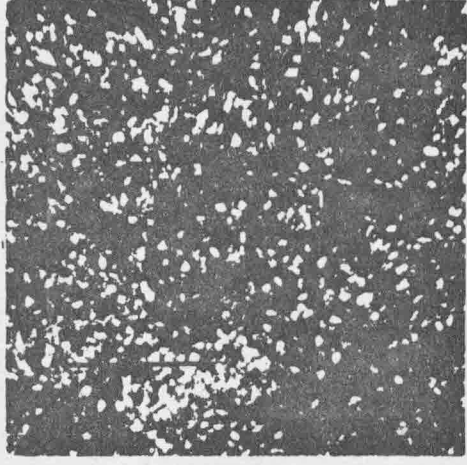
PHOTOMICROGRAPH F. Cross-Bedded, Oolitic Carbonate Facies. From the Snowflake Springs section (Plate 1). Note fossils and angular- to rounded clastic grains.

Figure 12. Photomicrographs of Swift Formation Lithofacies and Subfacies. The photographs were taken under plane light with a scale of one (1) mm indicated on the photos.

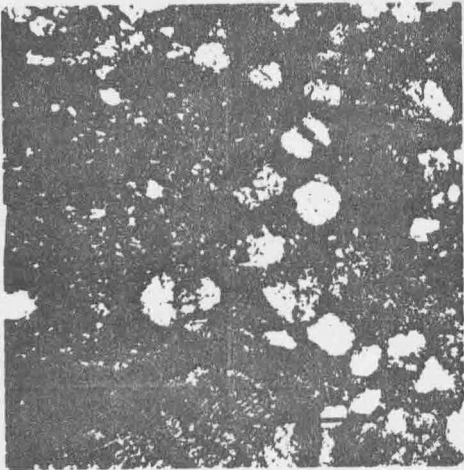
A



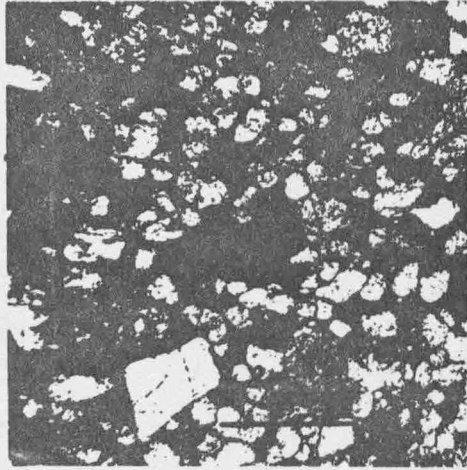
B



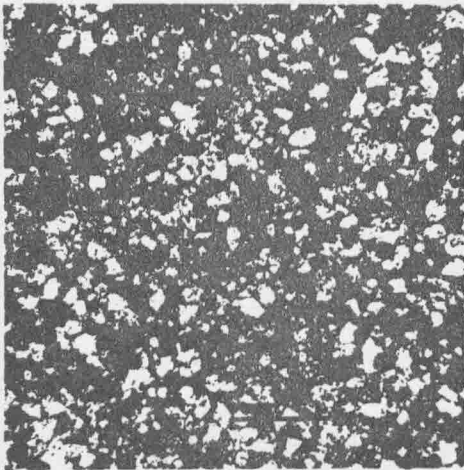
C



D



E



F



commonly interbedded with smaller scale trough-cross-bedding. No bioturbation was noted in this subfacies.

Petrographic Data

Glauconite

Glauconite is a common constituent of Swift Formation rocks in southwestern Montana, varying between 1% (or less) and 30% of the rock volume. Of the hand specimens studied, 37% contain no glauconite. Of the remaining 63%, all but a few contain 10% or less (Appendix). Rocks relatively rich in glauconite occur in stratigraphic sections south of Toston. In all other stratigraphic sections, the maximum amount was 7%. In the northern half of the study area, an apparent subtle increase in glauconite content occurs from north to south. No glauconite occurs in the Indian Creek section west of Townsend or in Milligan Canyon. Glauconite occurs in the Sappington Canyon and Horseshoe Hills sections. Klemme (1947) describes rocks in the Maudlow area as containing variable amounts of glauconite. Swift rocks of the Bridger Range

and Rocky Canyon sections contain up to 5% and 7% glauconite respectively. The glauconite content of Swift rocks south of Bozeman is obscure due to lack of stratigraphic control, although values of 2% and less are characteristic of the oolitic carbonate rocks at Snowflake Springs.

Glauconite grains in Swift rocks of southwestern Montana have shapes ranging from spherical to angular. No irregularly shaped glauconite grains such as mammillary forms were found. The generally finer-grained Swift rocks of the Bridger Range and Rocky Canyon areas seem to contain more spherical and oval glauconite grains and less angular glauconite grains than the coarser-grained Swift rocks to the north. Most of the glauconite grains are generally similar in size to the fine-grained quartz sand fraction of the Swift rocks of the study area (Appendix). Aside from the apparent increases in glauconite content and glauconite grain roundness from north to south, there are no significant differences in either amount or character of glauconite grains between the various lithofacies. The lithology south of Toston is an exception to this general trend and is enigmatic.

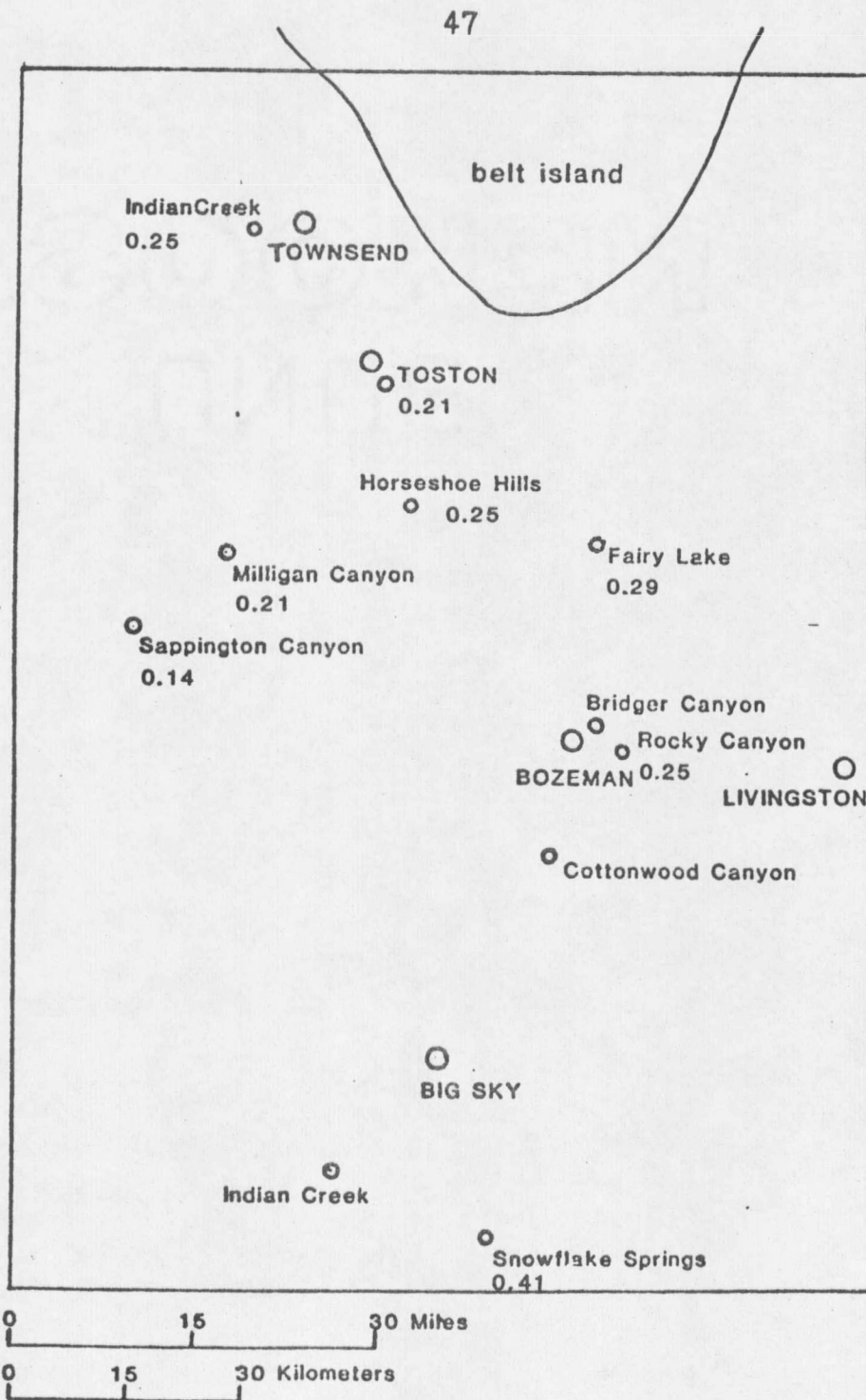


Figure 13. Average Grain Sizes (mm) of Calcareous Basal Conglomerate and Cross-Bedded, Oolitic Carbonate Facies.

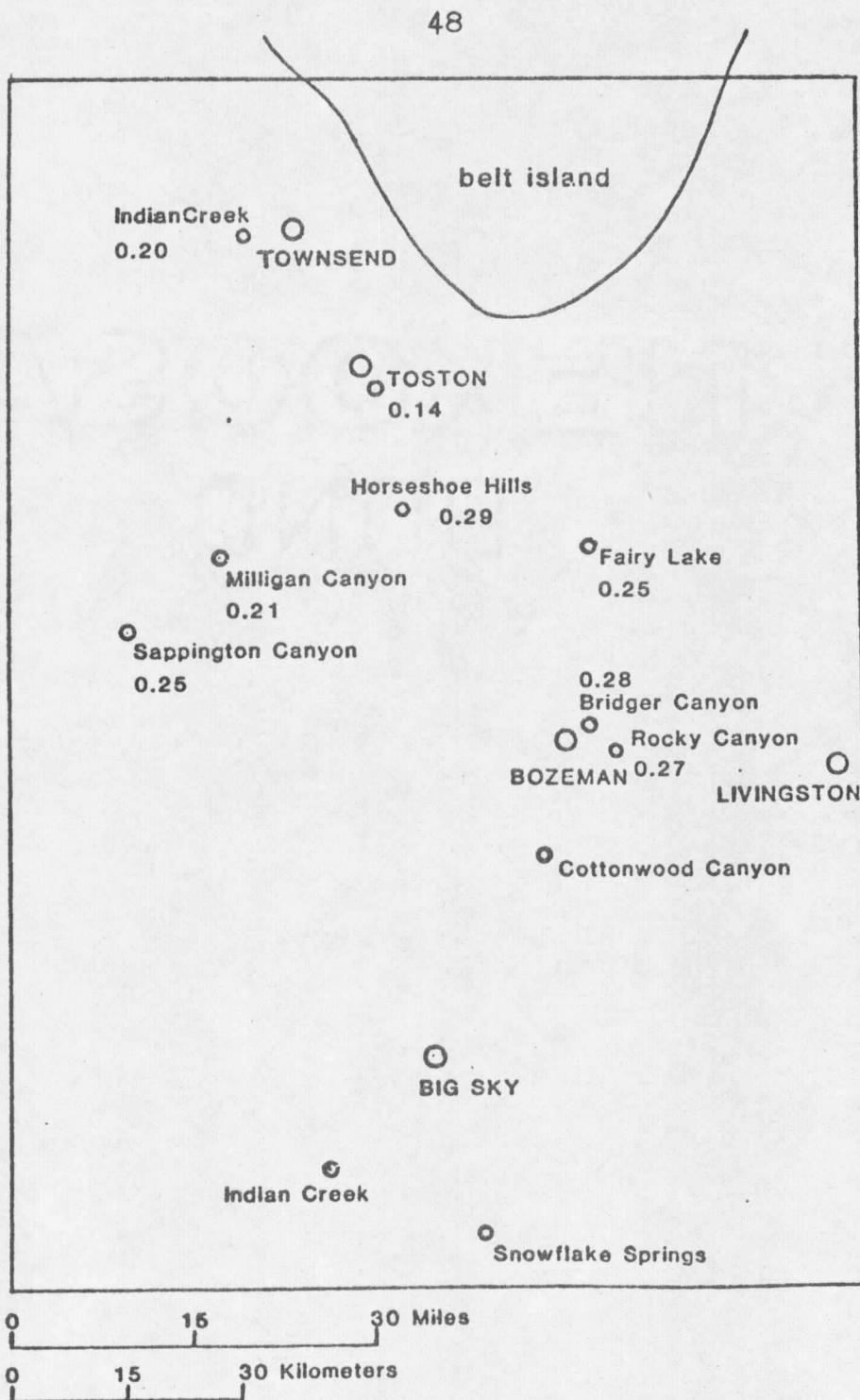


Figure 14. Average Grain Sizes (mm) of Medium-Grained/Coquinoid Subfacies.

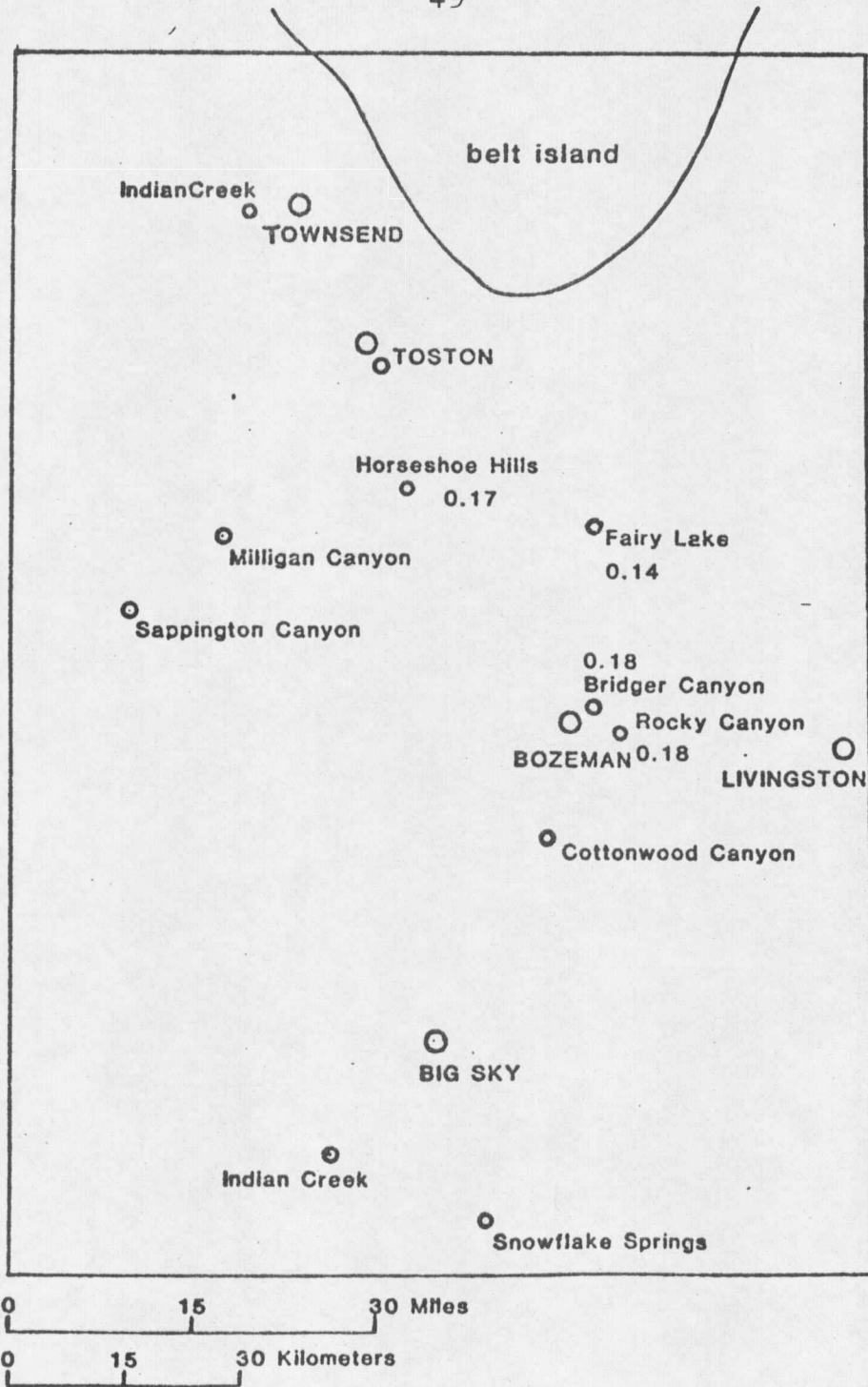


Figure 15. Average Grain Sizes (mm) of Fine-Grained, Well-Sorted Subfacies.

Average Grain-Size

Differences between facies and subfacies of the Swift are subtle, and overlap in characteristics is common. Data which clearly show other similarities and differences are measurements of grain size and component percentages.

Average grain sizes vary rapidly both vertically and horizontally in rocks of the Swift Formation, although trends exist which help to characterize each facies and subfacies. Grain size maps showing these trends are depicted in Figures 13, 14, and 15. The grain size analyses shown on these maps are the result of visual estimates made with a binocular microscope and are used here to define the facies and subfacies. Grain size values are not as important as the differences in average grain sizes between the facies and subfacies and the distribution of those average grain sizes. Visual estimates of average grain sizes include only that fraction of the rocks smaller than 2.00 mm. The basal conglomerate facies and the calcareous sandstone facies both have average grain sizes ranging from 0.15 to 0.30 mm. The average grain size of the detrital fraction of the oolitic carbonate facies is considerably coarser at 0.40 mm.

Although the basal conglomerate and calcareous sandstone facies are similar in average grain size, the basal conglomerate shows a decrease in average grain size away from Belt island. No such trend is evident in the calcareous sandstone facies.

Average grain size maps of the subfacies of the calcareous sandstone facies show that although there are no apparent trends in average grain size distribution either within or between the subfacies, there are definite differences in range of average grain size values. The medium-grained/coquinoid subfacies and the fine-grained, well-sorted subfacies have average grain size ranges of 0.20 to 0.30 mm and 0.15 to 0.20 mm respectively.

Component Percentages

Percentages of quartz, chert, allochemical grains, and carbonate cement also vary rapidly both vertically and horizontally in Swift rocks. They were measured with a binocular microscope, using the same method as that used for average grain size, and should be viewed as having the same limitations regarding the way in which they are used in this study.

Component percentage data for each facies and sub-facies are shown in ternary percentage diagrams in Figure 16. As depicted in these plots, the component percentages vary widely within and between facies and subfacies, making further analysis necessary. The scattering of the percentages in the diagrams is caused primarily by vertical and lateral changes in lithology. Because the amount of lithologic variability differs considerably between the facies and subfacies, the entropy, or degree of mixing of the components within the rocks is a useful criteria on which to differentiate them.

Entropy function behavior diagrams were first introduced by Pelto (1954) as a means of mapping multicomponent systems. While several examples of applications of entropy diagrams to facies interpretations exist in the literature, a good description of the use of entropy diagrams and modified entropy diagrams is given by Forgotson (1960).

In order to show relationships between component percentage distribution and entropy level of the Swift facies and subfacies, diagrams are depicted in Figure 16 which show the entropy function behavior diagram of

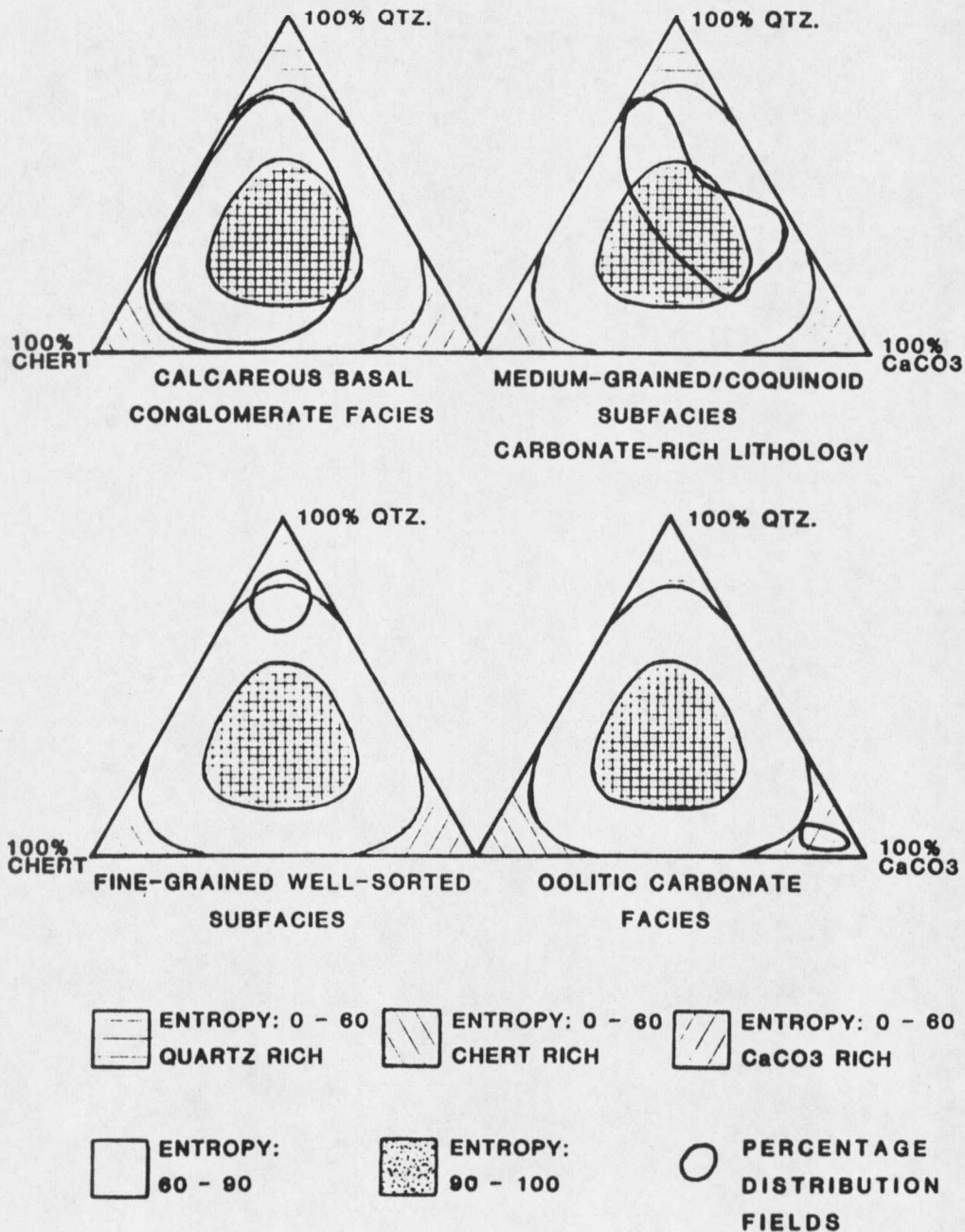


Figure 16. Component Percentage Diagrams Superimposed on Entropy Behavior Diagrams. Entropy Behavior Diagram From Forgotson (1960).

Forgotson (1960) superimposed on the ternary percent diagram. Component percentages are plotted on the ternary percent diagrams and translate directly into entropy values. Because entropy changes vertically as well as horizontally within the rocks, ranges of entropy values are shown on stratigraphic cross-sections A-A' and B-B' in Plate 1.

The component percentage diagrams in Figure 16 show differences in component percentages, ranging from subtle to obvious between the facies and subfacies. Figure 16 and Plate 1 show that differences in entropy between the facies and subfacies are subtle and show considerable overlap. However, it is apparent from Figure 16 that the medium-grained/coquinoid subfacies is dominated by entropy values greater than 90, and the fine-grained, well-sorted subfacies is dominated by entropy values less than 60, as is the oolitic carbonate facies. The entropy of the basal conglomerate facies is greater than 60 and occupies more of the entropy: 60-90 field than the medium-grained/coquinoid subfacies. These trends suggest a general decrease in entropy from north to south in the study area (Plate 1). This decrease in entropy is directly

related to the southward change in depositional environments discussed in Chapter 5.

STRATIGRAPHIC AND PALEOGEOGRAPHIC INTERPRETATIONS

Facies Reconstruction Diagrams

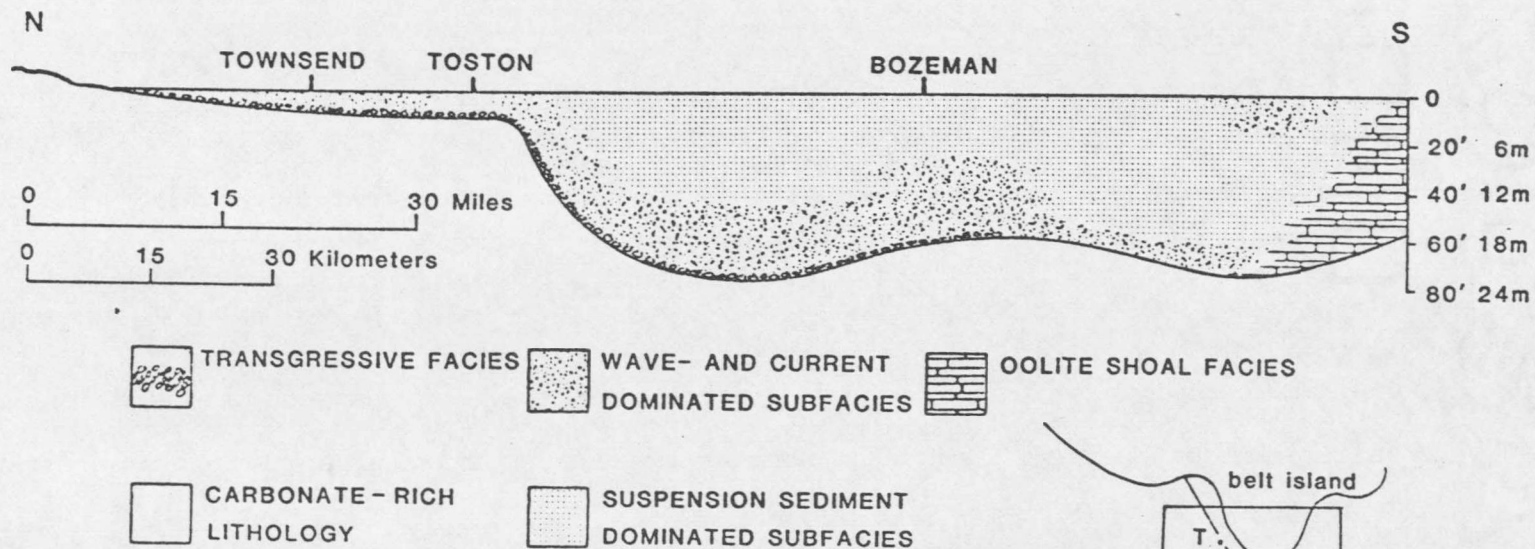
On the basis of information given in Chapter 4, interpretations of the depositional environment for each facies and subfacies described in Chapter 4 were worked out. Table 2 gives interpretive names to these facies and subfacies. Two facies reconstruction cross-sections based on stratigraphic cross-sections A-A' and B-B' (Plate 1) are shown in figures 17 and 18. The diagrams show the vertical and horizontal relationships between the facies and subfacies.

Factors Controlling Swift Deposition

Probably the most important single factor controlling the depositional character of the Swift is sediment starvation. Evidence of sediment starvation during Oxfordian deposition is of two types. First, the low rate of sediment accumulation characteristic of the Swift indicates either a slow overall sedimentation rate or periods of faster sedimentation followed by longer periods of very slow or no sediment accumulation. In either case, sediment starvation is a likely cause for the low rate of sedimentation which seems to be characteristic of all

DESCRIPTIVE NAME	INTERPRETIVE NAME
Calcareous basal conglomerate facies	Transgressive basal conglomerate facies
Calcareous sandstone facies	Inundative facies
Medium-grained, coquinoid sub-facies	Wave- and current-dominated subfacies
Fine-grained, carbonate-rich lithology	Shallow-water, low-energy lithology
Fine-grained, well-sorted subfacies	Suspension-sediment-dominated subfacies
Oolitic carbonate facies	Oolite shoal facies

Table 2. Descriptive and Interpretive Names of Swift Formation Lithofacies and Subfacies.



58

Figure 17. North-South Stratigraphic Cross-Section Showing Lithofacies Reconstruction. Datum for Swift Formation is Swift-Morrison Contact.

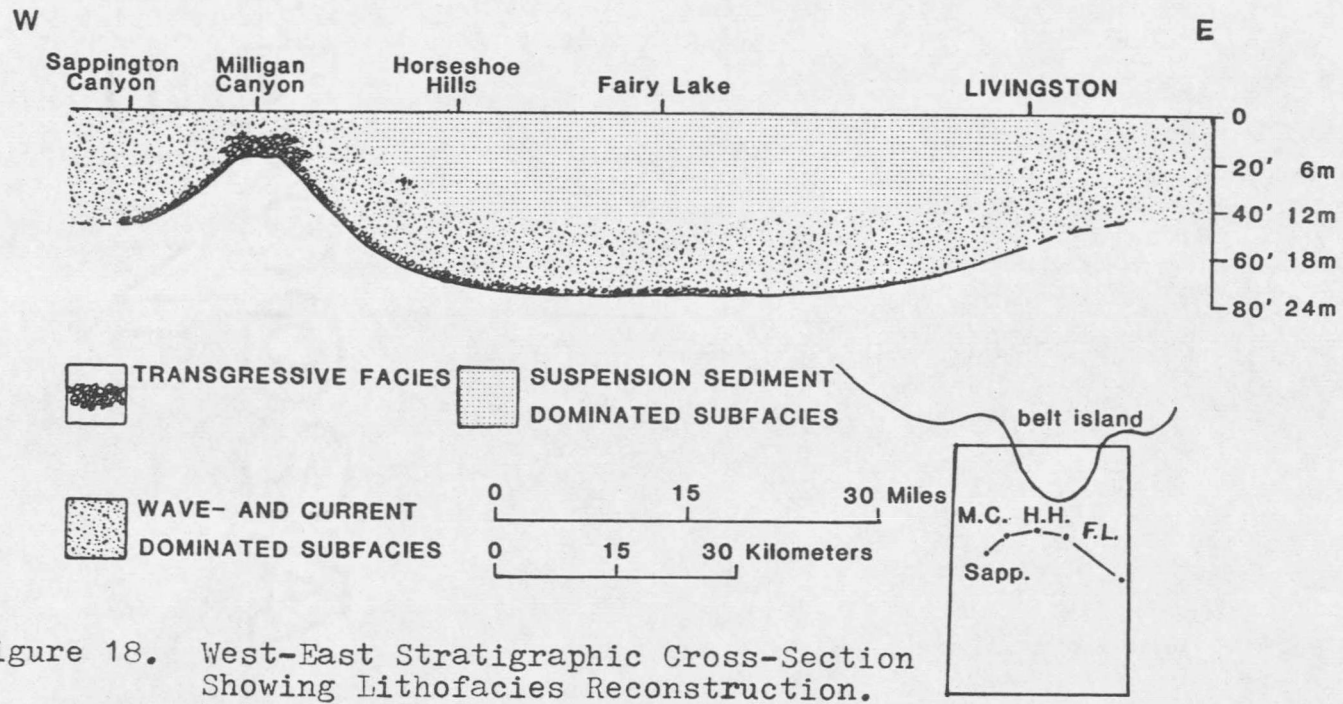


Figure 18. West-East Stratigraphic Cross-Section Showing Lithofacies Reconstruction. Datum for Swift Formation is Swift-Morrison Contact.

Swift rocks above the transgressive basal conglomerate in southwestern Montana. Second, probable evidence for a low rate of sedimentation is found in Swift rocks. Although the presence of glauconite in varying amounts throughout the Swift Formation is, perhaps, the strongest evidence for a low sedimentation rate (McRae, 1972), a general scarcity of biogenetic sedimentary structures and the general maturity of Swift sandstones indicates constant physical reworking of the available sediment. A low rate of sedimentation would allow the sediments accumulating above the basal conglomerate to be reworked many times before final burial. Deposition of the transgressive basal conglomerate probably occurred under conditions of rapid burial and will be discussed later.

The stratigraphic cross-sections in Figures 17 and 18 show that the Oxfordian depositional basin in southwestern Montana had a low relief bathymetric profile. Any changes in sea level and, especially sudden changes in energy level and wave agitation within the depositional environment (such as storms), would have had far-reaching effects on the character of the sediments.

Apparently, storm activity played a major role in sediment deposition and reworking in southwestern Montana.

It is evident from the duration of Oxfordian time (Firstbrook and others) that deposition of the Swift Formation occurred over at least several million years. Given a time period of that scale, "rare", less frequent, more violent events such as storms would have controlled shore and epeiric sea sedimentation to a greater extent than "normal" wave and current action (Gretenor, 1967; Dott and Batten, 1981). Many examples of profound effects on modern and ancient coastal sediments caused by storm activity are found in the literature (Brown, 1939; Hayes, 1967; Ball, and others, 1967; and Kumar and Sanders, 1976).

Source rocks for the Swift sediments were probably restricted to limestone, sandstone, chert and minor shale of the Phosphoria, Quadrant, and Amsden Formations, the Big Snowy and Madison Groups, and the Rierdon and Sawtooth Formations (Fig. 4). Quartz sand and chert, therefore, compose the bulk of the clastic sediment supplied to the Oxfordian sea.

Transgressive Facies

Although the Oxfordian sea transgressed generally from the north into southwestern Montana (Imlay, 1947),

the strand line in this area probably migrated from southeast to northwest onto the south flank of Belt island. The period of time necessary for the strand line to migrate across southwestern Montana may have been relatively short compared to the duration of Swift deposition. Lines of evidence that support these theories are the low relief, southward regional slope of the southwestern Montana shelf (Figs. 17 and 18) which may have allowed a rapid rate of transgression to the northwest, the presence of limestone pebbles and cobbles within the basal conglomerate facies which indicate little transport and rapid burial, and the angular to very angular chert pebbles within the conglomerate facies in some areas which indicate little transport and rapid burial, possibly under high energy conditions. Klemme (1947) found limestone pebbles as far north as the south margin of Belt island. Other evidence of rapid Oxfordian transgression occurs in the Black Hills where Oxfordian shale overlies Callovian redbeds with sharp unconformity (Imlay, 1947).

The basal conglomerate of the Swift is apparently the only deposit preserved of the Oxfordian transgressive phase. Littoral and sublittoral transgressive deposits

such as barrier and lagoon sediments do not occur. The literature of transgressive-regressive systems indicates that such deposits are commonly not preserved or are only partially preserved in the rock record (Fischer, 1961; Kraft, 1971; Swift, 1968).

Shoreface erosion of littoral sediments during transgression produces a disconformity within the littoral sediments (Swift, 1968) called a revinement by Stamp (1922). The revinement process acts to destroy marginal deposits unless it is counteracted by other factors within the littoral environment. The revinement process seems to work best in environments where "wave energy dominates over sediment supply" and not so well where "sediment supply is in excess of energy available to remove it" (Swift, 1968). If the shoreface becomes armored with a coarse-grained lag deposit, the revinement process will fail to operate (Swift, 1968).

The basal conglomerate facies is probably a near-shore lag deposit formed at or near the shoreline of the Oxfordian transgression. The high energy of this environment was enhanced by the rapidity of the Oxfordian transgression. These factors worked together to strengthen the revinement process and destroy the barrier and lagoonal

sediments in this area. The thin, widespread geometry of the conglomerate facies indicates deposition in a migrating littoral environment.

The basal conglomerate is a complex deposit in which beach-forming processes and storm activity both played an important part. Modern transgressive environments show a highly complex interplay of transporting agents and subenvironments (Kraft, 1971).

In many areas of southwestern Montana, the transgressive facies consists of a poorly sorted, chaotic deposit of sand and pebbles, often without fossils or well developed bedding. These deposits are interpreted as storm lag deposits, reflecting instantaneous depositional processes (i.e. storms) as opposed to normal marginal marine depositional processes. Kumar and Sanders (1976) presented a model of shoreface storm deposition in which basal lag deposits of course-grained gravel, overlain by laminated, fine-grained sand, constitute the storm deposit. In their model, this deposit is overlain by burrow-mottled, coarser sand deposited under fair weather conditions. It is likely that the basal conglomerate facies of the Swift fits this shoreface storm deposit model. Lack of burrow-mottled sediments in the transgressive facies

indicates that the reworking process was probably very active at this time. Laminated, fine-grained sandstone similar to the sand directly overlying the coarse gravel in Kumar and Sanders (1976) model is locally found overlying the basal conglomerate of the Swift in southwestern Montana.

Some conglomerate beds in this facies, however, are characterized by abundant fossil fragments, flat or elongated pebbles oriented parallel to bedding, and abundant, well sorted, sandy matrix in which the pebbles commonly float. Subtle horizontal and low-angle cross-bedding and local not-so-subtle trough bedding are also common. These characteristics suggest deposition in the littoral zone by normal, non-catastrophic beach-forming processes (Swift, 1976; Reading, 1978). Because of the complex interactions of storm activity with beach-forming processes, no attempt was made to resolve the transgressive facies into beach deposits or shoreface storm deposits.

The chert pebbles which compose most of the transgressive facies were largely derived from small islands located in the west-central part of the study area. These islands, composed of *Phosphoria* bedded chert, represent

topographic highs on the pre-Oxfordian erosion surface. The resistant chert beds were left as remnants as the Rierdon Limestones were eroded. These islands seem to have occurred as a localized group or groups, their position possibly coinciding with a facies change to bedded chert within the Phosphoria of this area.

The high energy of the Oxfordian transgression quickly eroded the jointed chert beds, producing a short-lived local source of abundant angular chert pebbles. Erosion of resistant rocks in a similar setting is presently occurring on the coast of California (Shepard and Grant, 1947).

Evidence for the existence of small islands of bedded chert is indirect because they were probably completely destroyed by wave action during the Oxfordian transgression. No rocks in the study area could be interpreted as remnants of islands with the possible exception of an area about 30 miles south of Belt island called Negro Hollow (Fig. 6). Here, fine-grained sandstone of the Morrison Formation directly overlies a chert-pebble and cobble bed which is probably part of the Phosphoria Formation.

The Milligan Canyon section, about 9 kilometers southeast of Negro Hollow, contains the strongest indirect evidence of small, short-lived chert sources. Here, the greater-than-average thickness and development of the basal conglomerate facies show that much chert was concentrated in local areas. This, coupled with the greater-than-average size and angularity of chert pebbles and cobbles indicates that Milligan Canyon is close to a chert source. During the erosion process, chert from the islands was broken up by wave activity and scattered across the shelf. Most of the chert was probably not moved very far, although storm activity may have moved pebbles considerable distances from the source (Barrell, 1925).

The sharp, undulatory contact between the basal conglomerate and the bedded cherts may represent a wave-cut surface (Shepard and Grant, 1947). This indicates that the chert islands may have been beveled off by wave activity, producing a local thickening of the transgressive conglomerate.

Inundative Facies

Introduction

During the rise in sea level and retreat of the Oxfordian shoreline, marine sediments were deposited in two distinct environments corresponding to the two sub-facies of the inundative facies given in Table 2. These environments differed in terms of water depth, energy level, and style of sediment accumulation. Similarities in their general lithology indicate deposition in a shallow marine shelf environment in which terrigenous clastic deposition dominated.

Wave- and Current-Dominated Subfacies

The wave- and current-dominated subfacies was deposited in an open marine shelf which extended throughout the northern part of the study area. Water depth in this part of the shelf was above fair-weather wave base, resulting in an environment dominated by relatively high energy processes. Deposition of bed forms in this sub-facies was controlled by wave and current activity.

The relative scarcity of fine-grained sand in this subfacies indicates that frequent reworking of the

sediment selectively winnowed out the fine-grained detritus transported into this area. The amount of that fine-grained detritus may have been large considering the generally fine-grained character of the Paleozoic source rocks. The fine-grained sand was winnowed out and transported south by wave and current activity, leaving behind the medium and coarse-grained fraction of the sediment delivered into the environment. These medium and coarse quartz and chert sand grains and fossils formed a lag deposit which is the coquinoid and medium-grained, non-coquinoid lithologies of this subfacies.

Brenner and Davies (1973) described coquinoid sandstones in southeastern Montana and northeastern Wyoming. In their study area, "storm-lag" deposits occur in "coarse-grained, thin, laterally extensive beds interbedded with shale and fine-grained sandstone". They suggested that these deposits are the result of sudden, short-term changes in energy level caused by storms. A similar interpretation is found in a paper written by Spect and Brenner (1979) on bioclastic carbonate deposits in east-central Wyoming. In my study area, the wave- and current-dominated subfacies consists of thicker deposits of sediments similar in character to the storm-lag

deposits of Brenner and Davies (1973). Because of these similarities, the coquinoid sandstones in this subfacies are likely the result of storm activity. The relatively thick beds of this lithology are probably the result of reworking of the sediment by frequent storm activity, producing a thick storm deposit. Another possibility is "stacking" of thinner storm deposits one on top of the other by successive storms, although no evidence of erosional contacts within the subfacies was noted which would support this theory. Because of the weathering of the rock, such erosional contacts would be difficult to identify.

The non-coquinoid lithology of this subfacies may represent a slightly different depositional environment than the coquinoid sandstone. Deposits of this type are generally thicker than those of the coquinoid lithology, although the average grain size and sedimentary structures are similar. This may indicate deposition in slightly deeper water (Fig. 18) where sedimentation was controlled less by storm activity and more by wave activity and currents which are common in shallow water environments (Clifton and others, 1971). Moreover, this lithology may represent sediment moved from shallow areas

during storms. In the Bridger Range and Rocky Canyon areas, coquinoid sandstone and sandy limestone beds alternate with beds of the non-coquinoid lithology. Hummocky cross-stratification similar to beds described by Hamblin and Walker (1979) are developed to a limited extent in these non-coquinoid beds. Hummocky cross-bedded rocks in this area may represent deposition by ebb-flow and wave-surge currents as storm activity waned (Goldring and Bridges, 1973; Hamblin and Walker, 1979).

The fine-grained, carbonate-rich lithology of the wave- and current-dominated subfacies south of Toston is anomalous when compared to the other two lithologies of this subfacies. While the coquinoid and non-coquinoid lithologies represent relatively coarse-grained, high energy deposits, the carbonate-rich lithology seems to represent a lower energy environment within an area dominated by higher energy processes. Rocks of this lithology reflect a predominance of quiet, clear water sedimentation although the effects of frequent storm activity are commonly preserved in the form of pebbly sand beds which interrupt sandy carbonate deposition. The abundance of glauconite and carbonate cement in the rocks and the local occurrence of very very sandy sandy, oolitic

limestone beds indicate little sediment influx for at least part of the time (Drobnick, 1965).

The depositional environment of this lithology is difficult to visualize. Stratigraphic sections just south of Toston are relatively thin compared to the Swift sections in adjacent areas farther to the south (Fig. 19). This may reflect deposition in relatively shallow water on an existing pre-Oxfordian erosion surface or one created during the Oxfordian transgression. Explanation of the relatively low energy, sediment-starved character of this environment remains problematical.

Suspension Sediment Dominated Subfacies

The winnowed sediment, discussed above, probably consisted of very fine grained quartz sand with minor amounts of sand-sized chert grains. It was carried from the wave- and current dominated environment into deeper water in suspension clouds (Reineck and Singh, 1972). This occurred during periods of high wave activity (storms) which occurred repeatedly in this part of the shelf (Brenner and Davies, 1973).

The deeper water environment, therefore, was dominated by suspension deposition occurring as the wave

activity subsided and the suspension clouds settled. The homogeneous, well-sorted character of this subfacies, the thin, horizontal bedding, and the small-scale cross-bedding support this suspension deposition theory (Reineck and Singh, 1972; Masters, 1967). The cross-bedding was probably caused by sand waves or dunes migrating in response to storm-generated currents (Reading, 1978).

The water depth of this subfacies was below fair-weather wave base resulting in an environment where quiet water conditions prevailed and where sediment was only agitated during storm activity. However, two lines of evidence indicate frequent reworking by storm activity. Although some benthonic organisms might normally live in an environment such as this during relatively quiet periods, there is a general scarcity of body fossils, animal trails, and burrows in these rocks. Also, as indicated in Chapter 4, horizontal bedding ranging in thickness from fine laminations to medium beds is the predominant sedimentary structure in this subfacies.

Reineck and Singh (1975) proposed a model of sediment accumulation (Fig. 19) which is applicable to this suspension sediment dominated subfacies. The model shows

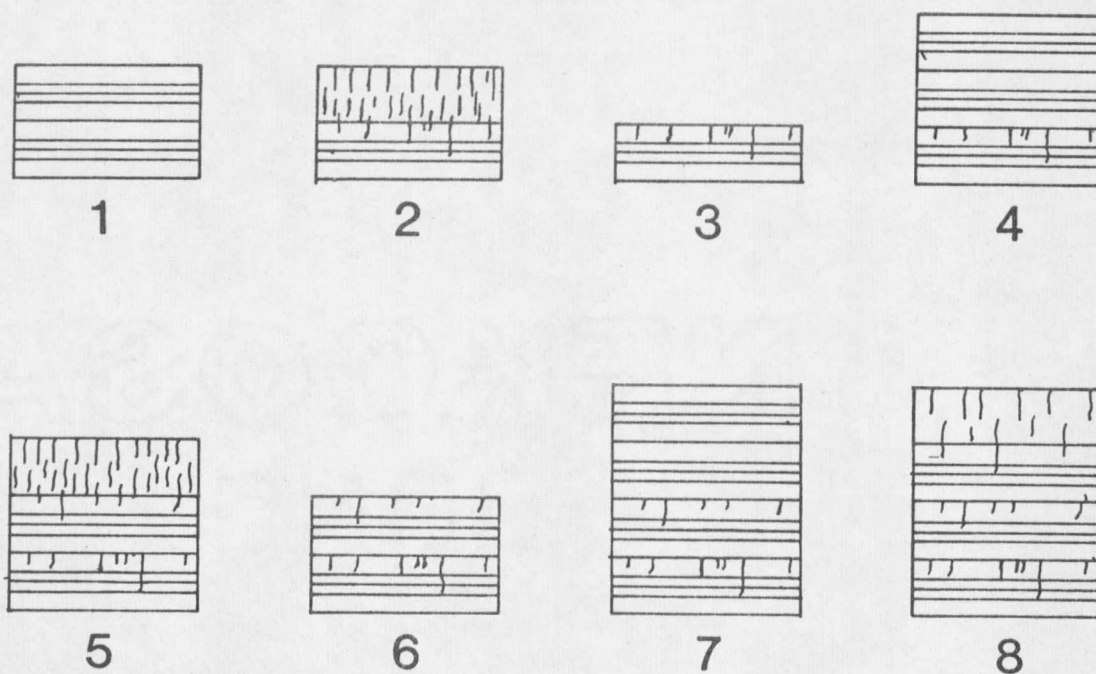


Figure 19. Destruction of Bioturbated Zones by Periodic Storm Erosion. 1: Deposition of Sediment Under Normal Marine Conditions. 2,5,8: Bioturbation of Upper Beds. 3,6: Storm Erosion of Upper Beds. 4,7: Redeposition of Upper Storm Sediment and Renewal of Normal Conditions. After Reineck and Singh (1973).

a sequence of sediments interrupted by periodic erosion. Indications of bioturbation were consistently destroyed by erosion and reworking by storms, resulting in accumulation of sediment in which only isolated bioturbation structures were left undisturbed. Storms that brought new sediment into this environment were also responsible for reworking the top layers of sediment already deposited there.

Oolitic Shoal Facies

During the period of transgression and inundation within which detrital sediment was deposited throughout much of southwestern Montana, a high energy, oolitic carbonate shoal environment occupied the southern third of the study area. Unlike the clastic-dominated, shallow areas to the north, calcium carbonate sedimentation dominated the deposits in this area. Several characteristics of oolitic carbonate deposits support a high energy, shoal environment interpretation for the oolitic carbonate facies in southwestern Montana. First, because extensive "spreads" of ooids are presently forming in agitated water less than 5 meters deep (Ginsburg and James, 1974), the initial oolitic carbonates of the Swift Formation in

southwestern Montana probably formed under similar conditions. Second, the presence of medium-scale planar and trough-cross-bedding indicates deposition by migrating ooid sand dunes (Reading, 1978). Ball (1967) indicated that large- to medium-scale cross-bedding in modern oolitic carbonate sand bodies may be preserved deposits of intermittent, high energy events (i.e. storms). Third, angular to rounded chert pebbles and quartz and chert sand grains present in the oolitic carbonate rocks may indicate high energy conditions. Both cross-bedding and clastic pebbles and grains are common in the oolitic carbonate facies in southwestern Montana.

The abundance of calcite and the presence of quartz and chert clasts in the rock also indicate that major sources of detrital sediment for the Swift Formation were far enough away from the area to allow calcium carbonate to dominate sedimentation, yet were close enough to provide a constant, low rate of detrital influx into the environment. Many of the rounded quartz sand grains could have been derived as residuum from the erosion of Rierdon Limestone. Other rounded grains and pebbles could have been derived from fluvial material transported into the area during the pre-Oxfordian erosion period. The angular

chert clasts, however, are more easily explained as coming from a source which was active during the deposition of the oolite shoals. The location of this source is not known, although the island mass in southwestern Montana proposed by Moritz (1951) is a probable source of detrital material during this time. As the sea transgressed, oolitic carbonate deposition apparently kept up with the rise in sea level, producing a deposit which correlates with both the transgressive and inundative facies to the north.

A depositional environment similar to the oolitic carbonate shoal model described here exists in an upper Cambrian limestone sequence of western Utah (Rees and others, 1976). In that paper, they described cross-bedded, oolitic sandy limestone beds which they interpreted as a "migrating dune complex" deposited on a "gently sloping bank margin or over a slight positive feature on a broad shelf". Ball (1967) described carbonate sand bodies in the Bahama Islands which are probably modern analogs to these deposits although there are no clastic grains in the carbonate sands of the Bahamas.

Depositional Relationships

Each of the lithofacies described in Chapter 4 and discussed as depositional environments above have characteristics which relate them to the other lithofacies or environments of deposition. These relationships may be described in terms of a regional depositional setting in which overall transport of sediment was from north to south. Sediment transport into the wave- and current-dominated environment enabled relatively thick deposits of storm-lag type material to be deposited. Thickness of the suspension sediment subfacies in turn indicates that much of the sediment transported through the wave- and current-dominated environment was carried south and deposited in the deeper water environment there. Finally, the oolitic carbonate facies in the southern part of the study area indicates deposition at some distance from the sediment source or sources, yet close enough to accumulate quartz and chert clasts. Any detrital clay introduced into this depositional setting was apparently carried farther offshore onto the distal parts of the shelf where the Swift is primarily shale (Peterson, 1975; Moritz, 1951).

There are many studies of transgressive depositional environments published in the literature which are similar in some respects to the Swift Formation in southwestern Montana. Examples include the Middle Devonian of New York State (McCave, 1973), Lower Silurian deposits of southwestern England and Wales (Bridges, 1975), Modern transgressive deposits on the north Atlantic coast of the United States (Fisher, 1961; Kraft, 1971) and the eastern, southeastern, and southern Recent continental shelves of the United States (Swift, 1974; Ginsburg and James, 1974).

PROVENANCE OF SWIFT FORMATION DEPOSITS

Several lines of evidence lead to the conclusion that the Belt island trend was more important as a sediment source during the Oxfordian than has previously been thought. First, the transgressive basal conglomerate facies is developed only in the northern half of the study area near the south flank of Belt island. It is thickest in the Jefferson Canyon area and generally thins to the south, pinching out in the area south of Bozeman (Fig. 8). Second, chert pebbles in the conglomerate generally decrease in size and angularity from north to south away from Belt island. Third, black chert pebbles and grains are generally more abundant in the northern part of the study area and were probably derived from the Madison Group limestones over Belt island (Fig. 5). Chert grains and pebbles from the Phosphoria Formation were derived from small islands in the west-central part of the study area (Fig. 20).

Marked changes in depositional environment from north to south occur in the study area. The northern third of the study area is dominated by shallow-water, high-energy conditions with sufficient sediment influx to

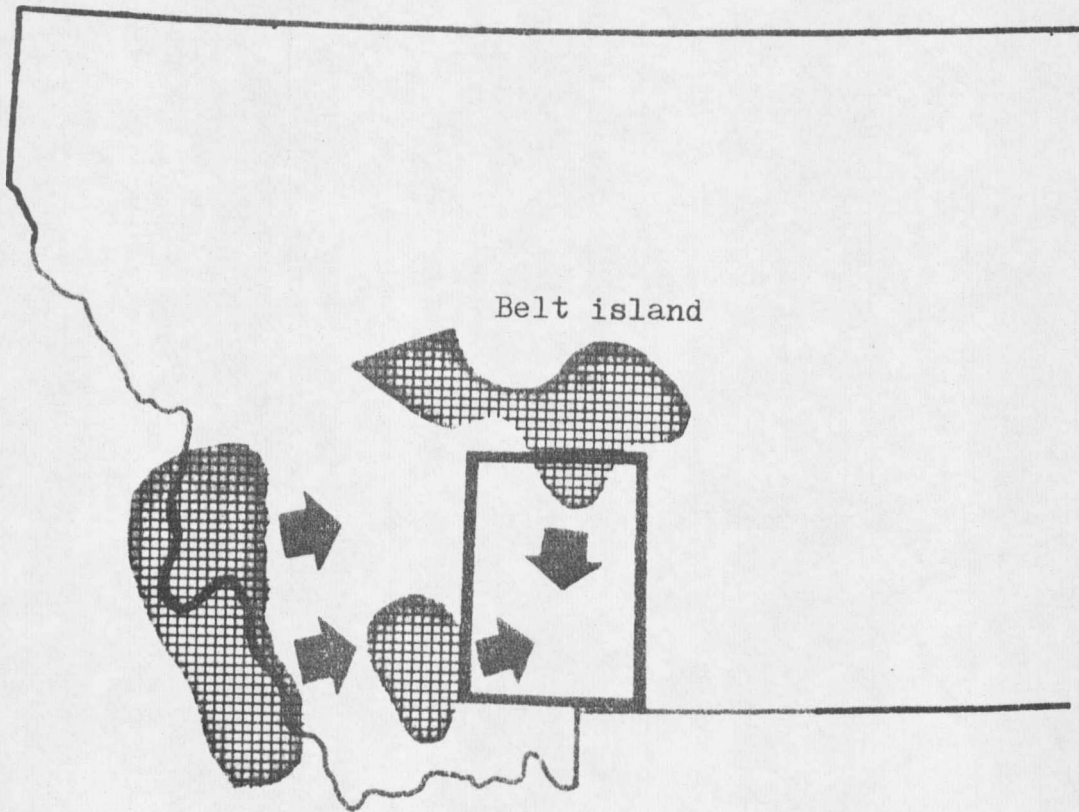


Figure 20. Sources and Direction of Transport of Sediments Composing the Swift Formation of Southwestern Montana. Heavy black line surrounds study area boundary.

produce relatively thick storm-lag and wave- and current-dominated deposits. In the central third of the area in the Bridger Range, quieter conditions in a deeper water environment controlled sedimentation of fine-grained sand winnowed from shallower areas to the north. In the area around Bozeman, higher energy conditions produced relatively thick deposits of storm-lag and wave- and current-dominated deposits which are overlain by quieter water deposits. In the southern third of the area, oolitic carbonates dominated sedimentation, reflecting the increased distance from the source of detrital sediments. In other areas of extreme southwestern Montana, the Swift consists of alternating beds of glauconitic sandstone and shale, reflecting quieter water conditions (Moritz, 1951).

The change in depositional environment from north to south indicates a pattern of overall southward transport of sediments into southwestern Montana (Fig. 20). This is an alternative theory to eastward transport of sediments from uplands to the west which has been postulated by other workers (Chapter 1).

However, evidence in southwestern Montana suggests that uplifts in western Montana and Idaho influenced by

the Nevadan Orogeny were of lower relief during the Oxfordian than has previously been thought. Characteristics which would indicate the presence of such uplifts are not apparent in Swift rocks. Major uplifts in close proximity to southwestern Montana would have made possible a higher sedimentation rate than is characteristic of Swift sediments. Along with a higher sedimentation rate, a general increase both in grain size and chert percentage toward the west would be expected. No such increases were noted in any of the Swift rocks of the study area. Also, the Morrison Formation is, possibly, too fine grained to be derived from such uplifts. Suttner (1969) found no positive evidence in his study of the Morrison in southwestern Montana to support higher relief sediment source areas to the west.

Absence of these characteristics, along with interpretations of Swift deposits indicate the existence of important low-lying sediment sources in the Belt island trend of west-central Montana (Fig. 20). Sediment from this source was shed into the surrounding seas and transported southward into southwestern Montana. Local sediment sources also existed in extreme southwestern Montana (Fig. 20) as indicated by angular chert pebbles in the

Swift rocks of Snowflake Springs. This source or sources also contributed sediment to the area.

OXFORDIAN DEPOSITIONAL HISTORY

The depositional history of the Swift Formation in southwestern Montana is summarized below with the aid of a series of diagrams shown in Figure 21. These diagrams evolved from the information presented in Chapters 4 and 5.

After deposition of the Sawtooth and Rierdon Formations, and the period of erosion following Rierdon deposition, northward transgression of the Oxfordian sea began (Fig. 21a). The strand line migrated rapidly toward the north into the study area and basal Swift sediment was deposited on progressively older rocks from south to north. The generally scoured character of the basal contacts indicates that deposition along the strand line was rapid and under high-energy conditions. Large limestone pebbles and cobbles within the basal zone support this interpretation. Few clues exist at the contacts or within the basal Swift rocks which indicate the nature of the pre-Oxfordian erosion surface. Cut-and-fill and drape structures and crevice fillings mentioned in Chapter 4 occur locally. Scattered, carbonized wood fragments are found along with rounded chert pebbles, some of which may

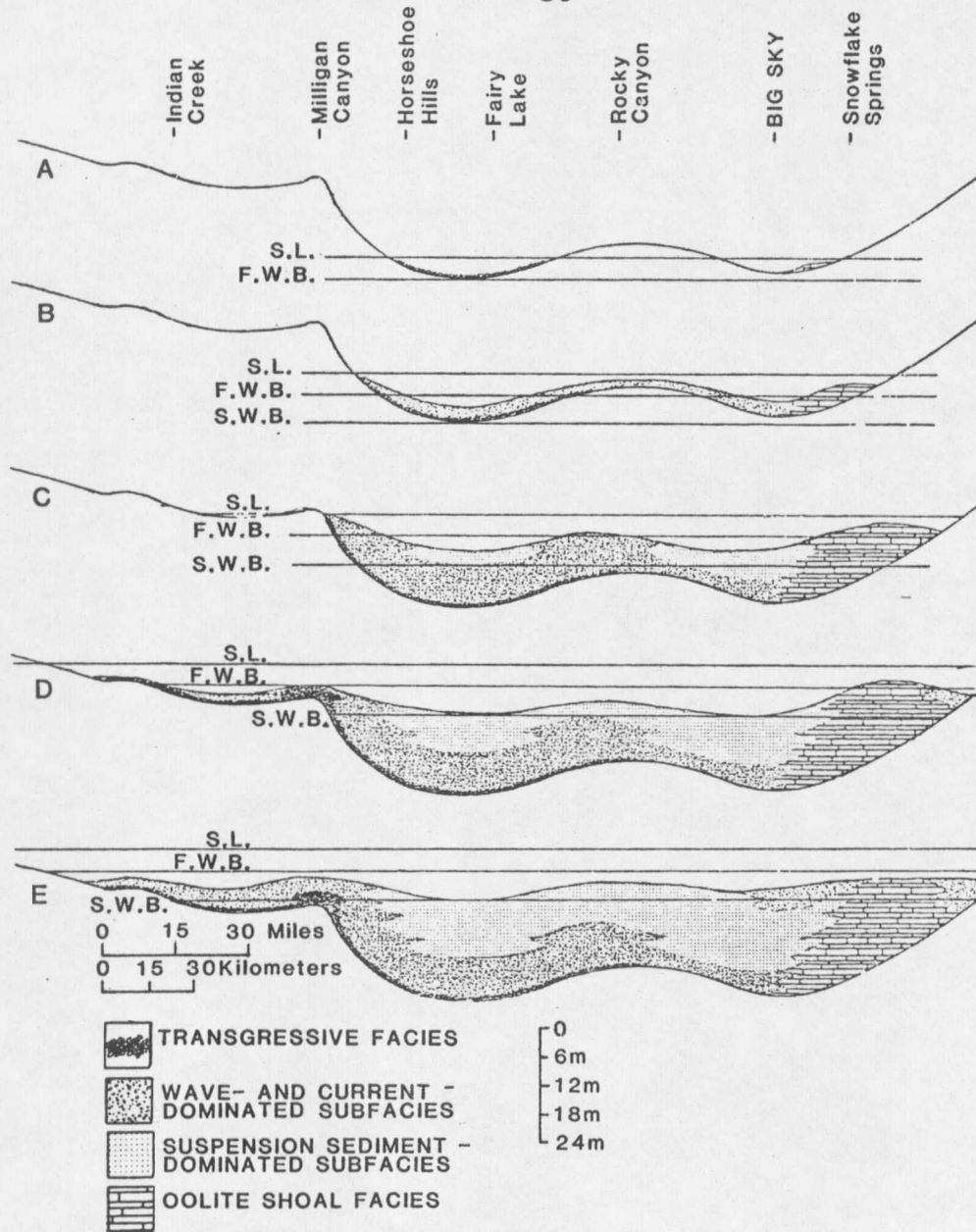


Figure 21. Depositional History of the Swift Formation Within the Study Area. S.L. - sea level, F.W.B. - fair-weather wave base, S.W.B. - storm wave base. "A" - Early Oxfordian time, "E" - Late Oxfordian time.

have been reworked from fluvial sediment on the erosion surface.

In the northern part of the study area, erosion of small islands produced a local source of angular chert pebbles. These pebbles were transported to the south as far as the Bozeman area during the transgressive phase (Fig. 21b). Water depth in the Rocky Canyon area was apparently shallower than to the north, resulting in little transport of pebbles from the north into this area, although high-energy conditions dominated sedimentation.

Deposition of wave- and current-dominated sediments, which was occurring in the area south and east of Bozeman during the transgressive phase, migrated to the north during the inundative phase (Fig. 21c). As deposition of wave- and current-dominated sediment occurred in the Jefferson Canyon and Townsend areas, finer-grained suspension deposits were forming in the deeper waters of the Bridger Range area (Fig. 21d). Continued deposition of wave- and current-dominated sediments in shallow water and suspension sediments in deeper water occurred through the remainder of Swift deposition (Fig. 21e). In the southern part of the study area, sandy carbonates were deposited

throughout the period of Swift deposition and interfinger with the more clastic deposits to the north.

Near the end of the Oxfordian, the sea withdrew from the shelf, leaving a broad, flat plane on which the Morrison Formation was deposited. Sedimentation of the Morrison progressed from west to east as low-gradient, eastward flowing streams brought sediments from sources to the west.

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APPENDIX

APPENDIX

PETROGRAPHIC DATA

ABBREVIATIONS

STRATIGRAPHIC SECTIONS

Indian Creek	I.C.
Toston	TOS.
Horseshoe Hills	H.H.
Fairy Lake	F.L.
Sappington Canyon	S.C.
Milligan Canyon	M.C.
Bridger Canyon	B.C.
Rocky Canyon	R.C.
Big Sky	B.S.

LITHOFACIES

Snowflake Springs	S.F.S.
Calcareous Basal Conglomerate Facies	CBC.
Medium-Grained/Coquinoid Subfacies	MGC.
Fine-Grained/Coquinoid Subfacies	FGS.
Oolitic Carbonate Facies	OC.

Stratigraphic section measured in meters.	Average grain-size in mm	% Quartz	% Chert	% CaCO ₃ Cement	% Bioclasts	% Glauconite	% Pebbles	Lithofacies
INDIAN CREEK								
I.C.-A-3.8	0.23	75.0	22.0	3.0	0	0	0	MGC.
I.C.-A-2.1	0.19	73.0	20.0	7.0	0	0	0	MGC.
I.C.-A-1.2	0.33	48.0	50.0	2.0	0	0	0	CBC.
I.C.-B-2.4	0.20	73.0	20.0	7.0	0	0	0	MGC.
I.C.-B-0	0.20	80.0	15.0	5.0	0	0	0	MGC.
I.C.-C-1.4	0.20	76.0	17.0	7.0	0	0	0	MGC.
I.C.-C-0.3	0.27	70.0	20.0	5.0	5.0	0	0	MGC.
I.C.-C-0.15	0.29	46.0	17.0	7.0	30.0	0	2.0	CBC.
I.C.-C-0	0.23	70.0	20.0	5.0	5.0	0	5.0	CBC.
I.C.-D-0.9	0.19	38.0	15.0	7.0	40.0	0	0	MGC.
I.C.-D-0	0.19	73.0	20.0	7.0	0	0	30.0	CBC.
TOSTON								
TOS.-A-4.2	0.17	15.0	11.0	49.0	25.0	0	0	MGC.
TOS.-A-2.4	0.17	38.0	12.0	50.0	0	0	0	MGC.
TOS.-A-2.0	0.36	40.0	20.0	40.0	0	0	0	MGC.
TOS.-A-1.8	0.23	30.0	10.0	60.0	0	0	0	MGC.
TOS.-A-1.2	0.19	35.0	20.0	45.0	0	0	0	MGC.
TOS.-A-0.5	0.20	77.0	13.0	10.0	0	0	0	MGC.
TOS.-B-2.4	0.17	45.0	8.0	40.0	0	7.0	0	MGC.
TOS.-B-1.5	0.17	38.0	12.0	40.0	0	10.0	50.0	MGC.
TOS.-B-1.2	0.17	43.0	20.0	7.0	0	30.0	0	MGC.
TOS.-B-0.9	0.10	82.0	8.0	10.0	0	10.0	25.0	MGC.
TOS.-B-0.6	0.10	45.0	3.0	45.0	0	7.0	0	MGC.
TOS.-B-0.15	0.23	35.0	20.0	45.0	0	5.0	50.0	CBC.

Stratigraphic section measured in meters.	Average grain-size in mm	% Quartz	% Chert	% CaCO ₃ Cement	% Bioclasts	% Glauconite	% Pebbles	Lithofacies
TOS.-C-2.1	0.12	45.0	8.0	40.0	0	7.0	0	MGC.
TOS.-C-1.5	0.14	72.0	8.0	10.0	0	10.0	0	MGC.
TOS.-C-1.2	0.10	48.0	20.0	7.0	0	25.0	0	MGC.
TOS.-C-0.6	0.07	5.0	0	94.0	0	1.0	0	MGC.
TOS.-C-0	0.20	15.0	20.0	60.0	0	5.0	60.0	CBC.
SAPPINGTON CANYON								
S.C.-12.1 A	0.20	60.0	30.0	7.0	0	3.0	0	MGC.
S.C.-12.1 B	0.23	58.0	30.0	10.0	0	2.0	0	MGC.
S.C.- 8.2	0.23	64.0	24.0	10.0	0	2.0	0	MGC.
S.C.- 4.2	0.12	63.0	30.0	5.0	0	2.0	0	MGC.
S.C.- 0.3	0.48	24.0	40.0	35.0	0	1.0	0	MGC.
S.C.- 0	0.14	70.0	25.0	5.0	0	F	0	CBC.
MILLIGAN CANYON								
M.C.-A-3.9	0.20	73.0	20.0	7.0	0	0	0	MGC.
M.C.-A-2.4	0.27	37.0	13.0	40.0	10.0	0	5.0	MGC.
M.C.-A-1.8	0.23	65.0	30.0	5.0	0	0	7.0	CBC.
M.C.-A-1.5	0.20	20.0	10.0	70.0	0	0	25.0	CBC.
M.C.-A-1.2	0.20	85.0	10.0	5.0	0	0	70.0	CBC.
M.C.-B-5.2	0.29	43.0	20.0	7.0	30.0	0	2.0	MGC.
M.C.-B-3.6	0.19	32.0	13.0	40.0	15.0	0	0	MGC.
M.C.-B-1.5	0.07	15.0	2.0	83.0	0	0	0	MGC.
M.C.-B-0.9	0.19	45.0	20.0	35.0	0	0	35.0	CBC.
M.C.-B-0.8	0.13	45.0	15.0	40.0	0	0	3.0	CBC.
M.C.-C-4.5	0.19	45.0	20.0	15.0	20.0	0	0	MGC.
M.C.-C-3.9	0.19	75.0	3.0	22.0	0	0	0	MGC.
M.C.-C-2.7	0.20	75.0	18.0	6.0	1.0	0	0	CBC.
M.C.-C-0	0.29	75.0	20.0	5.0	0	0	25.0	CBC.

Stratigraphic section measured in meters.	Average grain-size in mm	% Quartz	% Chert	% CaCO ₃ Cement	% Bioclasts	% Glauconite	% Pebbles	Lithofacies
HORSESHOE HILL								
H.H.-19.7	0.14	76.0	8.0	15.0	0	1.0	0	FGS.
H.H.-17.0	0.14	75.0	17.0	7.0	0	1.0	0	FGS.
H.H.-13.6	0.14	76.0	17.0	5.0	0	2.0	0	FGS.
H.H.-9.4	0.20	73.0	17.0	5.0	0	5.0	0	FGS.
H.H.-9.1	0.20	73.0	17.0	7.0	0	3.0	0	FGS.
H.H.-4.5	0.33	59.0	15.0	15.0	10.0	1.0	0	MGC.
H.H.-3.0	0.33	40.0	20.0	30.0	10.0	0	0	MGC.
H.H.-1.2	0.19	65.0	17.0	15.0	0	3.0	0	MGC.
H.H.-0.3	0.23	53.0	17.0	15.0	15.0	0	2.0	CBC.
H.H.-0.0	0.19	63.0	15.0	7.0	15.0	0	2.0	CBC.
FAIRY LAKE								
F.L.-A-22.1	0.13	82.0	8.0	10.0	0	0	0	FGS.
F.L.-A-10.6	0.14	82.0	8.0	10.0	0	0	0	FGS.
F.L.-A-5.5	0.20	64.0	20.0	15.0	0	1.0	0	MGC.
F.L.-A-3.6	0.27	50.0	17.0	7.0	25.0	1.0	10.0	MGC.
F.L.-A-3.3	0.23	76.0	14.0	5.0	3.0	2.0	0	MGC.
F.L.-A-1.5	0.38	30.0	20.0	30.0	20.0	0	0	MGC.
F.L.-A-0.9	0.29	68.0	15.0	12.0	5.0	0	13.0	CBC.
F.L.-A-0.6	0.29	57.0	18.0	10.0	15.0	0	5.0	CBC.
F.L.-A-0.3	0.29	65.0	15.0	10.0	10.0	0	25.0	CBC.
F.L.-B-10.6	0.14	73.0	14.0	10.0	0	3.0	0	FGS.
F.L.-B-9.7	0.29	43.0	20.0	7.0	25.0	5.0	0	MGC.
F.L.-B-8.2	0.19	73.0	15.0	7.0	0	5.0	0	MGC.
F.L.-B-2.7	0.27	60.0	20.0	15.0	5.0	0	0	MGC.
F.L.-B-2.1	0.20	77.0	13.0	10.0	0	0	0	MGC.
F.L.-B-1.5	0.20	40.0	15.0	30.0	15.0	0	0	MGC.
F.L.-B-0.6	0.29	40.0	20.0	30.0	10.0	0	10.0	CBC.

Stratigraphic section measured in meters.	Average grain-size in mm	% Quartz	% Chert	% CaCO ₃ Cement	% Bioclasts	% Glauconite	% Pebbles	Lithofacies
ROCKY CANYON								
R.C.-A-18.2	0.14	80.0	10.0	10.0	0	0	0	FGS.
R.C.-A-16.7	0.20	78.0	10.0	12.0	0	0	0	FGS.
R.C.-A-15.2	0.20	78.0	10.0	12.0	0	0	0	FGS.
R.C.-A-13.6	0.20	78.0	7.0	15.0	0	0	0	FGS.
R.C.-A-12.1	0.20	78.0	10.0	12.0	0	0	0	FGS.
R.C.-A-10.6	0.20	78.0	7.0	15.0	0	0	0	FGS.
R.C.-A- 2.4	0.20	57.0	25.0	18.0	0	0	0	MGC.
R.C.-A- 1.5	0.27	57.0	25.0	15.0	0	3.0	0	MGC.
R.C.-A- 1.2	0.23	38.0	20.0	30.0	5.0	7.0	20.0	MGC.
R.C.-A- 0.9	0.20	53.0	20.0	15.0	5.0	7.0	0	MGC.
R.C.-A- 0.3	0.27	52.0	20.0	15.0	10.0	3.0	2.0	CBC.
R.C.-A- 0.0	0.27	42.0	25.0	15.0	15.0	3.0	5.0	CBC.
R.C.-B-18.2	0.20	78.0	15.0	7.0	0	0	0	FGS.
R.C.-B-13.6	0.20	42.0	15.0	35.0	3.0	4.0	0	FGS.
R.C.-B-10.6	0.20	65.0	15.0	15.0	3.0	2.0	0	FGS.
R.C.-B- 7.0	0.20	76.0	15.0	5.0	3.0	1.0	0	FGS.
R.C.-B- 3.9	0.29	27.0	17.0	15.0	30.0	1.0	0	MGC.
R.C.-B- 0.6	0.38	67.0	15.0	15.0	3.0	0	0	MGC.
R.C.-B- 0.0	0.29	59.0	25.0	15.0	0	1.0	0	MGC.
BRIDGER CANYON								
B.C.-12.4	0.20	77.0	15.0	7.0	0	1.0	0	MGC.
B.C.-10.9	0.20	77.0	15.0	7.0	0	1.0	0	MGC.
B.C.-10.3	0.17	82.0	10.0	7.0	0	1.0	0	MGC.
B.C.- 7.6	0.19	74.0	10.0	15.0	0	1.0	0	MGC.
B.C.- 6.7	0.23	71.0	20.0	5.0	3.0	1.0	0	MGC.
B.C.- 3.3	0.14	48.0	30.0	20.0	0	2.0	0	MGC.
B.C.- 3.0	0.29	58.0	30.0	5.0	5.0	2.0	25.0	MGC.
B.C.- 2.1	0.29	53.0	30.0	15.0	1.0	1.0	5.0	MGC.

Stratigraphic section measured in meters.	Average grain-size in mm	% Quartz	% Chert	% Bioclasts + % CaCO ₃ Cement	% Oolite	% Glauconite	% Pebbles	Lithofacies
BIG SKY								
B.S.- 23.0	0.20	65.0	20.0	13.0	0	2.0	0	MGC.
B.S.- 21.2	0.20	60.0	25.0	13.0	0	2.0	0	MGC.
B.S.- 20.6	0.20	40.0	15.0	45.0	5.0	T	0	MGC.
B.S.- 18.2	0.20	60.0	25.0	13.0	0	2.0	0	MGC.
B.S.- 4.8	0.20	50.0	30.0	16.0	0	1.0	3.0	MGC.
B.S.- 2.1	0.31	17.0	30.0	42.0	5.0	1.0	10.0	OC.
B.S.- 1.5	0.40	7.0	25.0	65.0	17.0	0	3.0	OC.
B.S.- 0.0	0.31	5.0	15.0	80.0	30.0	T	0	OC.
SNOWFLAKE SPRINGS								
S.F.S.-15.2	0.49	1.0	5.0	78.0	15.0	1.0	5.0	OC.
S.F.S.-13.3	0.52	5.0	4.0	70.0	20.0	1.0	0	OC.
S.F.S.-11.5	0.33	5.0	7.0	18.0	70.0	T	0	OC.
S.F.S.-10.9	0.38	5.0	12.0	13.0	70.0	T	0	OC.
S.F.S.- 9.1	0.49	3.0	7.0	58.0	30.0	2.0	1.0	OC.
S.F.S.- 7.6	0.40	3.0	6.0	16.0	75.0	T	0	OC.
S.F.S.- 7.3	0.52	7.0	10.0	43.0	40.0	T	0	OC.
S.F.S.- 5.5	0.40	7.0	10.0	33.0	50.0	T	0	OC.
S.F.S.- 3.3	0.33	7.0	10.0	42.0	40.0	1.0	0	OC.
S.F.S.- 1.5	0.33	7.0	9.0	53.0	30.0	1.0	0	OC.
S.F.S.- 0.6	0.33	7.0	12.0	41.0	40.0	0	0	OC.

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