Sedimentology of the early cretaceous lower sandstone member of the Thermopolis Shale, southwestern Montana
by Alan Dennis Stine

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Earth Sciences
Montana State University

Abstract:
The Early Cretaceous (Albian) lower sandstone member of the Thermopolis Shale in the Bozeman, Montana area has previously been interpreted as a transgressive marine shoreface sequence. Detailed observation and interpretation of lithologies and sedimentary structures in the lower sandstone member document deposition in a setting that more closely resembles modern shallow marine shelves rather than the marine shoreface. The proposed shelf environment is characterized by simultaneous deposition of sand and mud, with their distribution controlled by sediment availability, bathymetry, and distribution of marine currents and storm paths. First order geomorphic features on this shelf consist of sand ridges which exhibit a hierarchy of bedforms. Megaripples and ripples are the second and third order bedforms, respectively. The sand ridges, as well as isolated sand bedforms, migrated upon a substrate comprised of mud and complexly interbedded sand and mud (heterolithic deposits). Heterolithic intervals are interpreted as storm deposits while the mud is thought to be deposited in low energy environments where storm influence is minimal. The hydraulic regime controlling deposition on the shelf was characterized by genetically related uni-directional and oscillatory currents, that share an origin in storm events.

Documented transgressive processes recorded in the strata include ravinement, or shoreface erosion, of marginal marine deposits between the lower sandstone member and underlying non-marine Kootenai Formation. Additionally, rapid transgression subsequent to lower sandstone member deposition resulted in the sharp contact separating the lower sandstone member from the overlying offshore marine, middle shale member of the Thermopolis Shale. This transgression is also recorded by deposition of a thin, discontinuous, argillaceous micrite near the base of the middle shale member.

SEDIMENTOLOGY OF THE EARLY CRETACEOUS LOWER SANDSTONE MEMBER OF THE THERMOPOLIS SHALE SOUTHWESTERN, MONTANA

by

Alan Dennis Stine

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APPROVAL

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This thesis has been read by each member of the thesis committee and has been found to be satisfactory regarding content, English usage, format, citations, bibliographic style, and consistency, and is ready for submission to the College of Graduate Studies.

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ABSTRACT

The Early Cretaceous (Albian) lower sandstone member of the Thermopolis Shale in the Bozeman, Montana area has previously been interpreted as a transgressive marine shoreface sequence. Detailed observation and interpretation of lithologies and sedimentary structures in the lower sandstone member document deposition in a setting that more closely resembles modern shallow marine shelves rather than the marine shoreface. The proposed shelf environment is characterized by simultaneous deposition of sand and mud, with their distribution controlled by sediment availability, bathymetry, and distribution of marine currents and storm paths. First order geomorphic features on this shelf consist of sand ridges which exhibit a hierarchy of bedforms. Megaripples and ripples are the second and third order bedforms, respectively. The sand ridges, as well as isolated sand bedforms, migrated upon a substrate comprised of mud and complexly interbedded sand and mud (heterolithic deposits). Heterolithic intervals are interpreted as storm deposits while the mud is thought to be deposited in low energy environments where storm influence is minimal. The hydraulic regime controlling deposition on the shelf was characterized by genetically related uni-directional and oscillatory currents, that share an origin in storm events. Documented transgressive processes recorded in the strata include ravinement, or shoreface erosion, of marginal marine deposits between the lower sandstone member and underlying non-marine Kootenai Formation. Additionally, rapid transgression subsequent to lower sandstone member deposition resulted in the sharp contact separating the lower sandstone member from the overlying offshore marine, middle shale member of the Thermopolis Shale. This transgression is also recorded by deposition of a thin, discontinuous, argillaceous micrite near the base of the middle shale member.

INTRODUCTION

Purpose

Previous stratigraphic investigations of Early Cretaceous rocks in the Bozeman, Montana area interpret the lower sandstone member of the Thermopolis Shale as a nearshore marine unit deposited during initial transgression of the Cretaceous seaway (Fig. 1) into southwest Montana (Roberts, 1972; Schwartz, 1972, 1982; James, 1977; Vuke, 1982). However, these earlier investigations do not address in detail the sedimentology and hydraulic regime of the depositional environment of the lower sandstone member. Because transgressive marginal marine deposits are rarely preserved in the rock record (Ryer, 1977), this investigation was undertaken to examine the sedimentology of what has been interpreted as an ancient transgressive sandstone.

The specific goals of this investigation are to: 1) test the validity of earlier transgressive marine interpretations of the lower sandstone member; 2) define a more specific depositional environment for the lower sandstone member; 3) determine types and mechanisms of sediment transport in the lower sandstone member; 4) evaluate possible provenances of the lower sandstone member; and 5) contribute to knowledge of Early Cretaceous paleogeography in southwest Montana. This study is based on detailed comparison of sedimentological attributes of the lower sandstone member with characteristics of modern shallow marine sands.

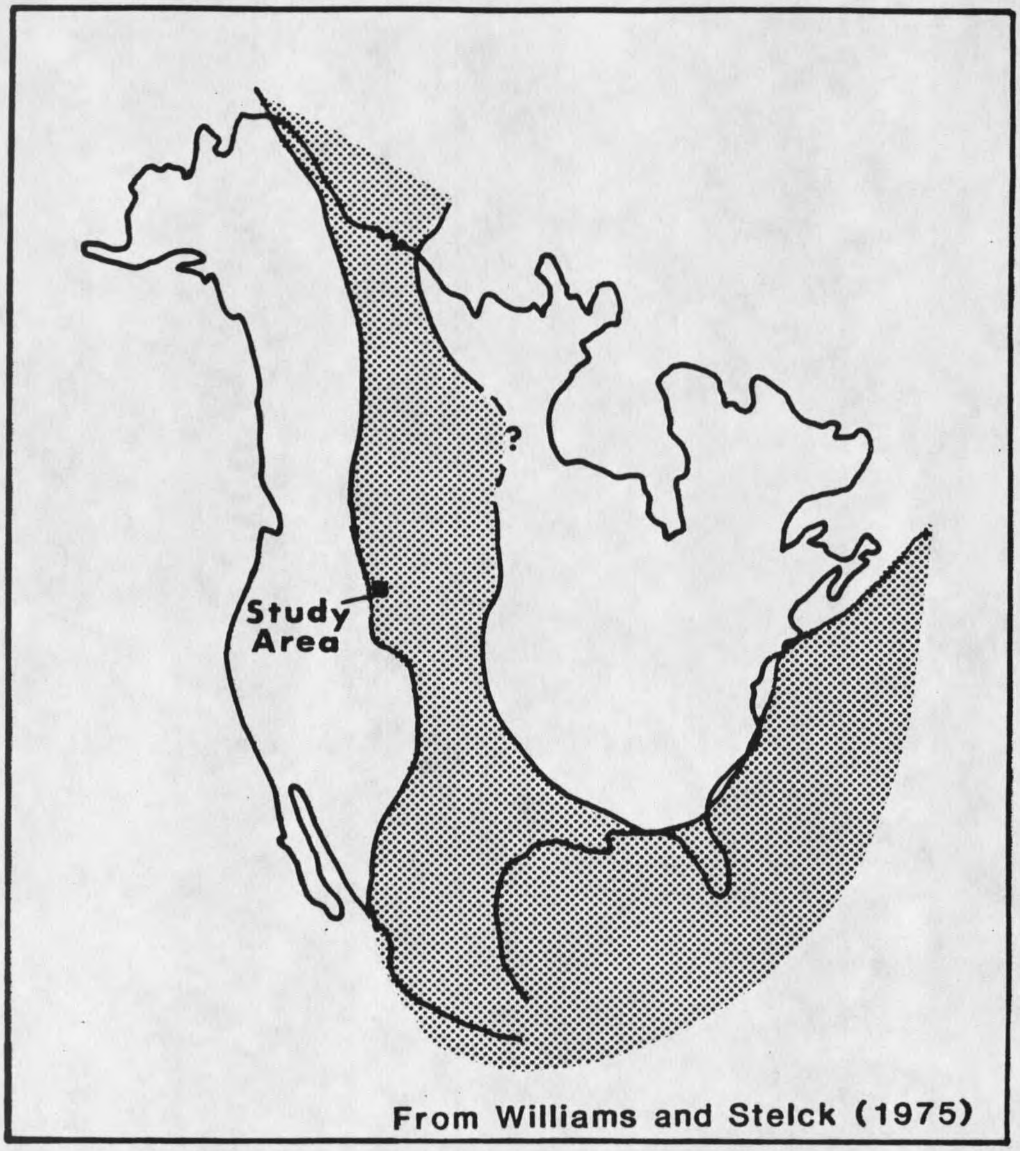

Figure 1. Map showing extent of early late Albian seas in North America.

Study Area

Exposures of the lower sandstone member occur in the north end of the Gallatin Range, Bridger Range, Sixteenmile area, Horseshoe Hills and Jefferson Canyon in Gallätin, Park, Jefferson and Broadwater Counties, southwest Montana (Fig. 2). Here, the lower sandstone member crops out along structures produced by Laramide and Sevier style deformation and subsequent Cenozoic Basin and Range extension.

The displacement of strata associated with these orogenies has a potential for disrupting facies relations within the lower sandstone member. Restoration of section locations to their approximate depositional positions (Fig. 3) was accomplished using estimates of displacement from Schmidt and others (1981), Woodward (1981), and D. Lageson, (personal communication, 1985). Accuracy of the estimates of tectonic displacement is uncertain due to limited knowledge of subsurface structures. The study area was divided into three structurally distinct geographical areas in order to assess the amount of displacement. Strata in the Elkhorn Mountains and Jefferson Canyon have been displaced 30-60 kilometers eastward by movement along the Lombard-Eldorado thrust zone (Schmidt and others, 1981; Woodward, 1981). The Bridger Range, Horseshoe Hills, and Sixteenmile area are part of the frontal fold and thrust belt where displacement of strata is to the east and on the order of 5 kilometers (Woodward, 1981; Lageson, D., personal communication, 1985). Strata in the northern Gallatin range have been displaced 1-2 kilometers to the southwest by Laramide style basement uplift (Lageson, D., personal communication, 1985).

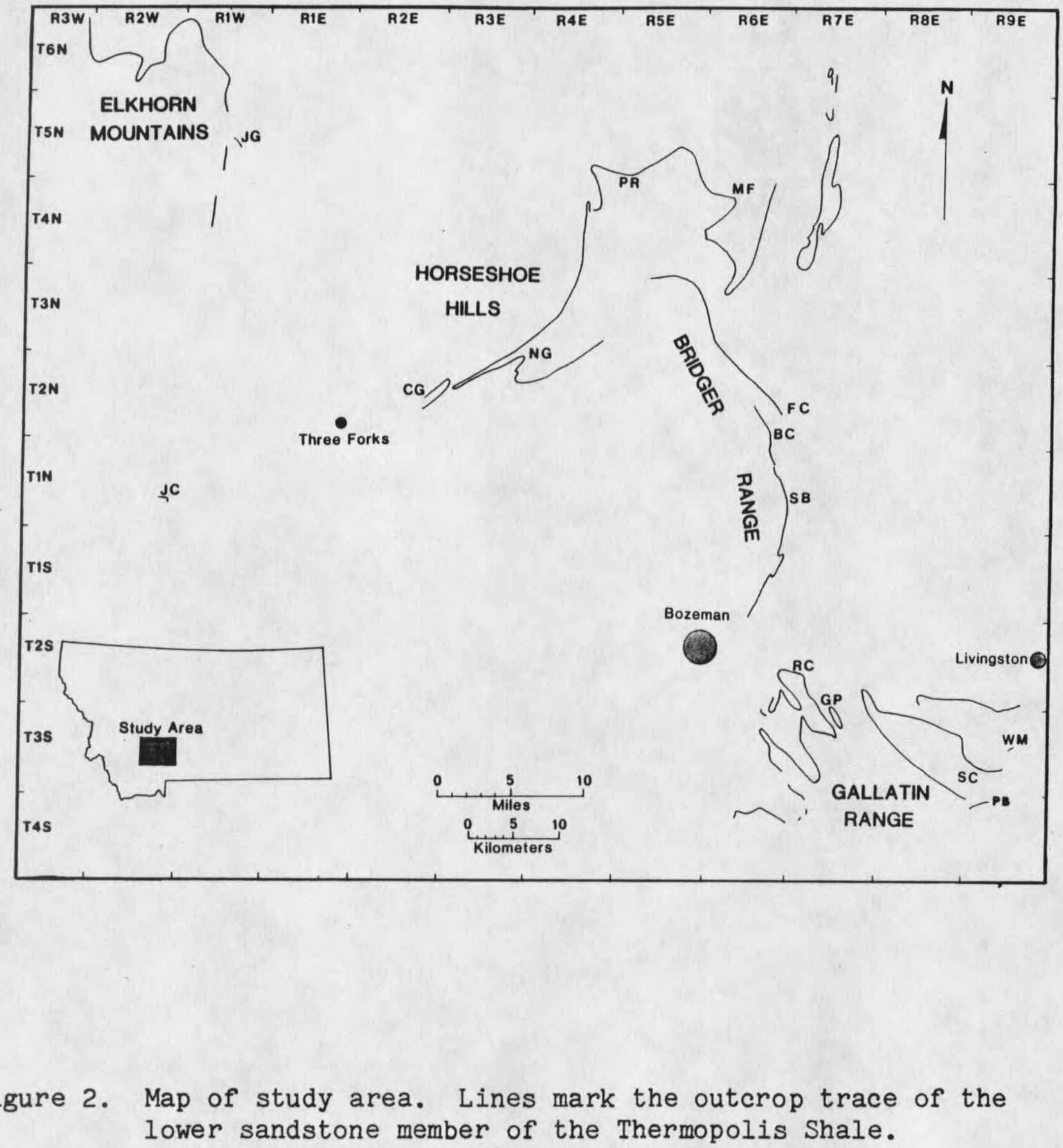

Figure 2. Map of study area. Lines mark the outcrop trace of the lower sandstone member of the Thermopolis Shale. Letters note the locations of measured sections.

JG-Johnny Gulch	BC-Brackett Creek
JC-Jefferson Canyon	SB-Sluicebox
CG-Cottonwood Gulch	RC-Rocky Canyon
NG-Nixon Gulch	GP-Goose Pond
PR-Paddys Run	SC-Strickland Creek
MF-Middle Fork Sixteenmile	WM-Wineglass Mountain
FC-Fairy Creek	PB-Park Branch

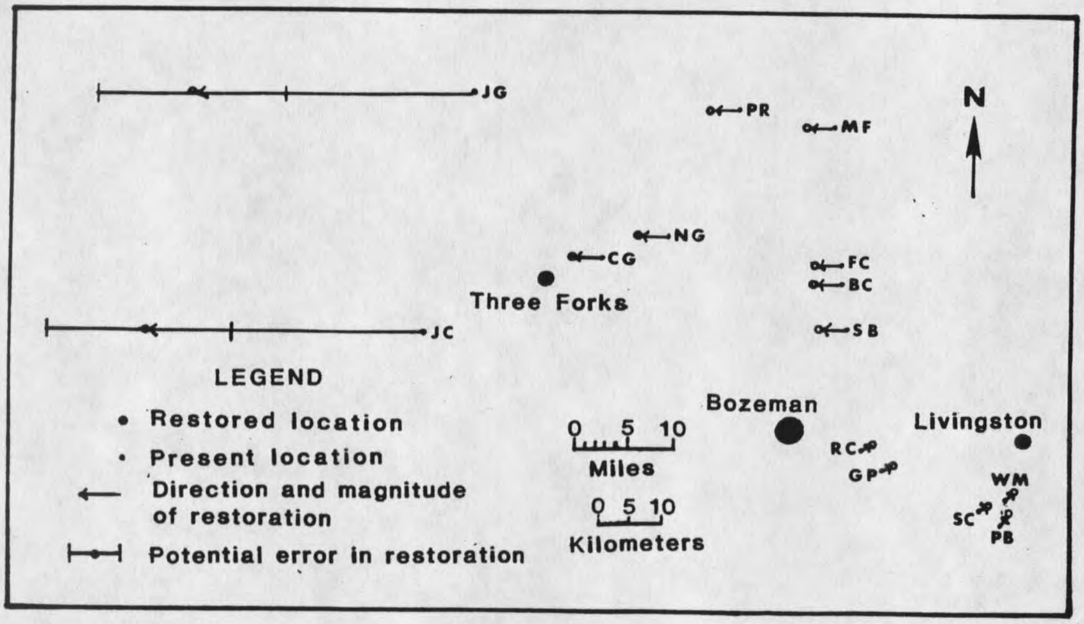

Figure 3. Map showing palinspastically restored locations of measured sections of the lower sandstone member, Thermopolis Shale.

Within each structural area, strata have undergone similar amounts of displacement. While the general trend of the paleoshoreline may be somewhat affected, major juxtaposition of depositional environments within each structural area is not thought to have occurred.

The amounts of differential movement vary between the structural areas. Differential movement between strata in the frontal fold and thrust belt and strata in the northern Gallatin Range is estimated to be only 6-7 kilometers. The depositional environment of the lower sandstone member in these two structural regimes is thought to be similar, as will be discussed later, and the displacement is not thought to juxtapose depositional environments. On the other hand, tectonic movement in the Elkhorn Mountains and at Jefferson Canyon produced approximately 25-60 kilometers of telescoping of strata relative to the frontal fold and thrust area. Strata of the two regions are much closer together than when they were deposited. Despite the eastward compression, the Elkhorn Mountains and Jefferson Canyon exposures are still located more than 15 kilometers west of the rest of the study area. Thus, although the absolute position is not the same, the relative east-west positions remain similar.

Previous Stratigraphic Investigations

Previous stratigraphic investigations of the lower sandstone member describe regional correlations and give general environmental interpretations (Roberts, 1965 and 1972; Schwartz, 1972 and 1982). A thorough review of the nomenclatural history of the lower sandstone member is given by Roberts (1972, p.C16). This terminology for the Lower Cretaceous stratigraphy in the Livingston, Montana area is used here. Regional rock stratigraphic equivalents (Fig. 4) of the lower sandstone member of the Thermopolis Shale include the basal sandstone of the Flood Member of the Blackleaf Formation in northwest Montana (Roberts, 1965); Interval A of the Blackleaf Formation in extreme southwest Montana (Schwartz, 1972); the First Cat Creek Sand in central Montana (Richards, 1957); the "Rusty Beds" and Greybull Sandstone Member of the Thermopolis Shale in central Wyoming (Roberts, 1965); and the Fall River Formation of the Black Hills in northeast Wyoming and western North Dakota (Roberts, 1965). Between central Montana and the Black Hills this interval changes to a siltstone or shale (Roberts, 1972).

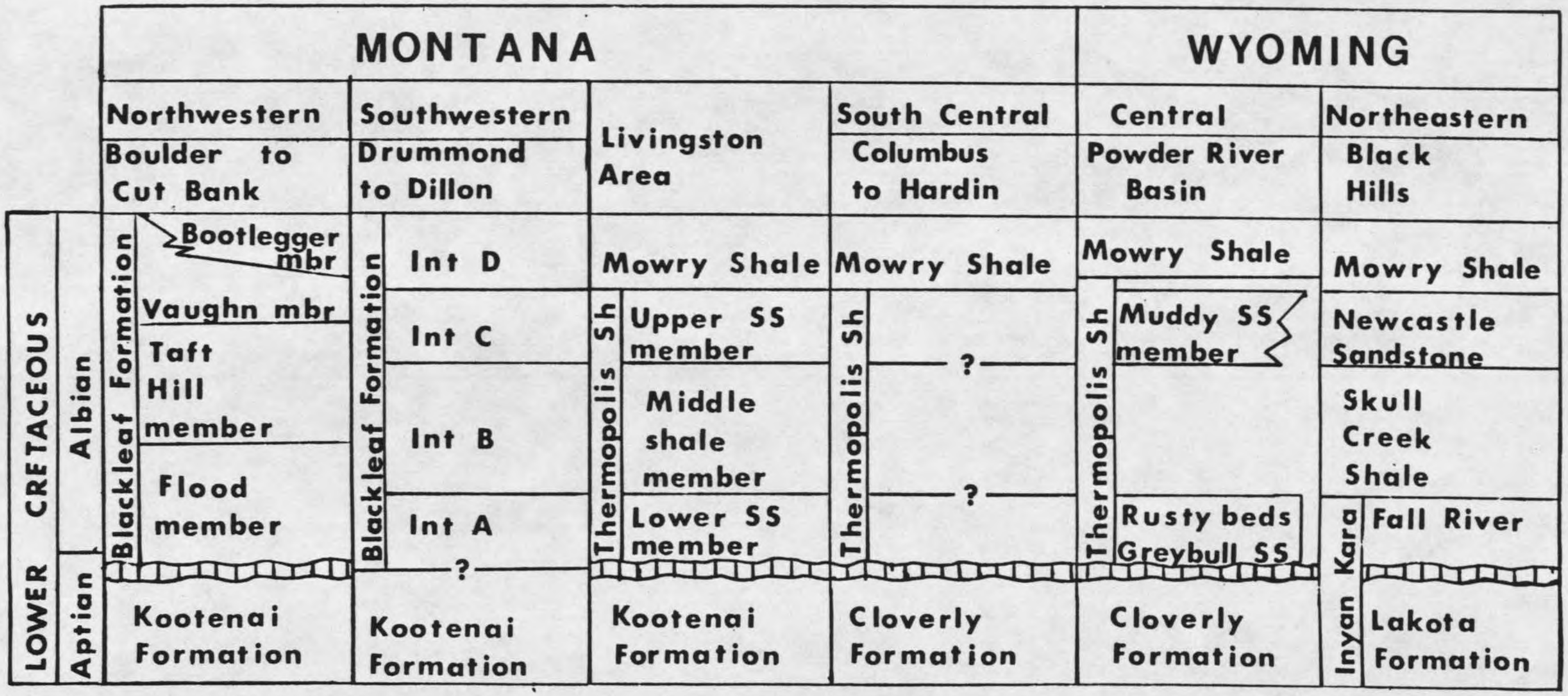

Figure 4. Stratigraphic correlation chart of Early Cretaceous rocks in Montana and Wyoming. Modified from Roberts (1965 and 1972) and Schwartz (1972).

The lower sandstone member of the Thermopolis Shale has been described as a 15 to 30 meter thick sequence of interbedded very fine to medium grained quartzarenite, siltstone and shale (Roberts, 1965 and 1972; Schwartz, 1972). The sandstones are thin to thick bedded and contain a variety of sedimentary structures, including crossbedding, ripple cross-lamination, and symmetric and asymmetric ripple forms. Burrows and trails within the lower sandstone member, including Ophiomorpha, are ascribed to marine bottom dwelling organisms. Shales are dark gray and appear similar to shale in the middle shale member of the Thermopolis Shale. Roberts (1972) and Schwartz (1972) both interpret the lower sandstone member as a nearshore marine unit deposited during initial transgression of the Cretaceous seaway into southwest Montana.

Throughout central and western Montana the lower sandstone member is underlain by the Lower Cretaceous (Aptian) Kootenai Formation (Suttner, 1969; Roberts, 1972). Roberts (1972) describes the upper 75 meters of the Kootenai Formation as a poorly exposed sequence of variegated red, purple, green and gray calcareous siltstone, claystone, mudstone, limestone and tuff that contains discontinuous beds of calcareous sandstone. This sequence was deposited in stream channels and upon associated flood plains of an extensive lowland within the embryonic western interior foreland basin. At the top of the interval is a biomicrite containing unidentifiable leached fossil molds and locally fossiliferous limestone containing small fresh water gastropods, pelecypods and ostracodes. The limestone was deposited in a fresh to possibly brackish water lake or series of lakes extending from northwest Wyoming into Canada and as far east as Harlowtown in central Montana (Suttner, 1969; Roberts, 1972; Holm and others, 1977).

The nature of the contact between the Kootenai Formation and the overlying Thermopolis Shale varies throughout Montana. Roberts (1972) states that the lower sandstone member commonly fills topographic depressions on an underlying erosional surface in the Livingston, Montana area. The Blackleaf-Kootenai contact is also disconformable on the Sweetgrass Arch in northwest Montana (Cobban and others, 1976). Schwartz (1972) observed the contact to be gradational in westernmost Montana and noted that it becomes more unconformable eastward. For the purposes of this study, the contact is recognized by the presence of non-calcareous quartzarenite or gray shale of the lower sandstone member, overlying variegated, calcareous mudstone or biomicrite of the Kootenai Formation.

The middle shale member of the Thermopolis Shale lies in sharp contact with the underlying lower sandstone member. The middle shale member is composed of dark gray shale, contains minor thin sandstone lenses and was deposited in an offshore marine environment subject to reducing conditions (Roberts, 1972; Vuke, 1982).

Methods

Stratigraphic and lithofacies data were gathered from outcrop exposures. Several of the exposures were cited in the literature (Roberts, 1972; Schwartz, 1972; James, 1977), most were located by reconnaissance. Fourteen sections (Fig. 2) were measured in detail using a Jacobs staff or tape. Many other more poorly exposed sections were briefly described. Lithology and physical and biogenic sedimentary structures were described in detail. Sandstones are named using the classification scheme of Folk (1980). Mudrocks are classified according to the method of Lundegard and Samuels (1980). Samples were collected of all lithologies at each measured section. Thirty thin sections were prepared, of which fifteen were stained to determine potassium feldspar content. In order to assess possible provenance of the lower sandstone member, all samples were studied with a binocular petrographic microscope and ten thin sections were point counted to determine relative abundances of detrital grains.

FACIES DESCRIPTIONS

Roberts (1972) and Schwartz (1972) have both described general characteristics of the lower sandstone member and have assigned it to a nearshore marine setting. This study expands upon their work by providing a more detailed interpretation of the sedimentology and hydraulic regime of the depositional environment of the lower sandstone member. In order to facilitate interpretation of depositional processes, several facies have been discriminated in the lower sandstone member. A primary way in which different depositional processes may be expressed is in the type of sediment deposited by each process. Initial discrimination of facies types is therefore based upon lithologies and lithologic associations present. Sandstone and shale are the primary lithologies in the lower sandstone member and facies are subdivided on the relative abundance of each. This method of facies discrimination is commonly used in siliciclastic facies analysis (Reading, 1978). Facies types defined for the lower sandstone member are: Sandstone (>90% sand), Heterolithic (50-90% sand and 10-50% mud) and Mud (>90% mud). These facies represent a near continuum of physical energy conditions with the exception of the absence of deposits with 10-50% sand and 50-90% mud. These lithologies may actually be present in the lower sandstone member but were not observed because of their low resistance to weathering coupled with the generally incomplete exposures of the lower sandstone member. As will be shown later, there is no consistent, regional stacking order to these facies in the lower sandstone member and the order in which they are discussed does not imply any vertical order to the lower sandstone member.

Sandstone Facies

The sandstone facies comprises 64% of the lower sandstone member and consists of >90% sand. Sandstones are composed of very fine to medium grained, chert-bearing, silica-cemented quartzarenite. Mudstone-clasts (to pebble size) are a minor, yet persistent, component of this facies and occur as isolated clasts or discontinuous lags within sandstone sequences. Shale interbeds comprise <10% of the sandstone facies and are usually present as very thin (<5 millimeters thick), discontinuous lenses or slightly thicker (approximately 1 centimeter thick), persistent beds.

Sandstones are medium to thick bedded (to 2 meters) and contain decimeter scale, high angle (15-30 degree) tabular to wedge planar (Fig. 5) and trough crossbeds; horizontal to low angle planar stratification; hummocky cross-stratification; and ripple cross-lamination. Several types of ripple cross-lamination are present in the sandstone facies. Foreset type cross-laminations with a tabular or trough geometry are present but the most abundant cross-lamination type is characterized by an internal structure of strongly tangential cross-laminations which may form offshoots that ascend onto adjacent ripples (Fig. 6). These cross-laminations have a variable angle of repose and the lower bounding surface of ripple sets is scooping and undulatory. Even, low angle and undulatory laminations are frequently associated with these cross-laminations. Symmetric and asymmetric,

Figure 5. Planar crossbeds in the sandstone facies. Lines on Jacob staff represent 6 inch intervals. From Wineglass Mountain exposure.

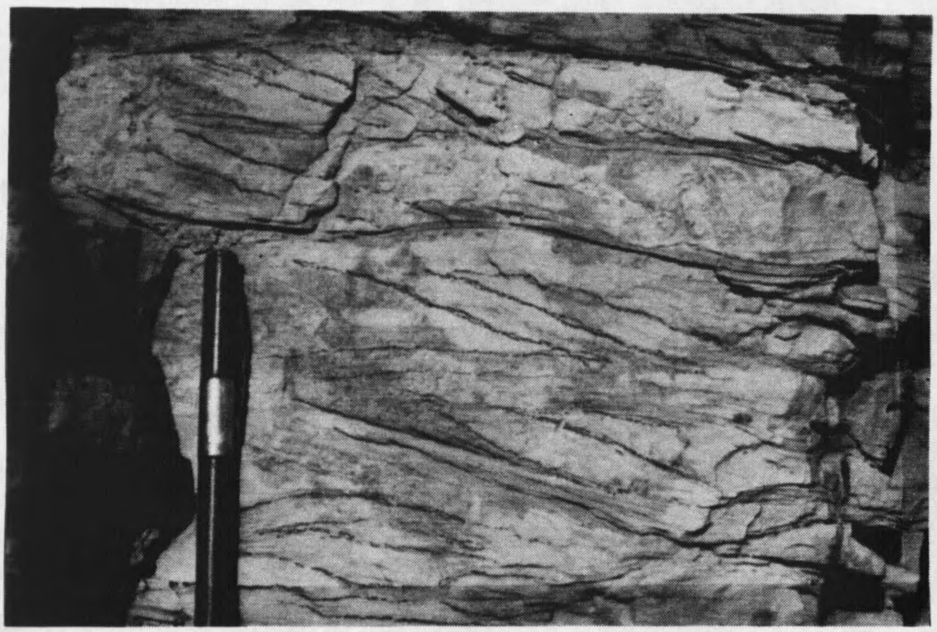

Figure 6. Wave ripple cross-lamination in crossbedded sandstone facies. From Wineglass Mountain exposure.

lingoid (Fig. 7) ripple forms are common in the sandstone facies with symmetric forms the more common type.

Figure 7. Asymmetric lingoid ripple marks in the sandstone facies. From Wineglass Mountain exposure.

Associations of sedimentary structures within the sandstone facies are highly variable. Many sandstone intervals are composed entirely of one type of sedimentary structure. These intervals may consist of either single bed or multiple bed sequences. In addition, some sandstone intervals consist of interbedded sedimentary structures, especially decimeter scale crossbeds, planar laminations and cross-laminations. No ordered spatial arrangement of these structures was observed. This may in part be due to the often poor exposure of sedimentary structures. Relative abundance of these structures is highly variable. In general, tabular to wedge-planar crossbeds are most common, while trough crossbeds, ripple cross-lamination, and horizontal to low angle stratification are less abundant. Hummocky cross-stratification is present only in the Elkhorn Mountains. The geometry of these sandstone intervals is mostly obscured because sandstone intervals usually persist beyond the limits of most exposures. Where observed (Wineglass Mountain exposure), the geometry is markedly planoconvex, with a flat base and convex-up, furrowed, upper-bounding surface (Fig. 8). The internal stratification of the sandstone bodies is form discordant and is comprised of crossbeds, cross-laminations, and planar laminations.

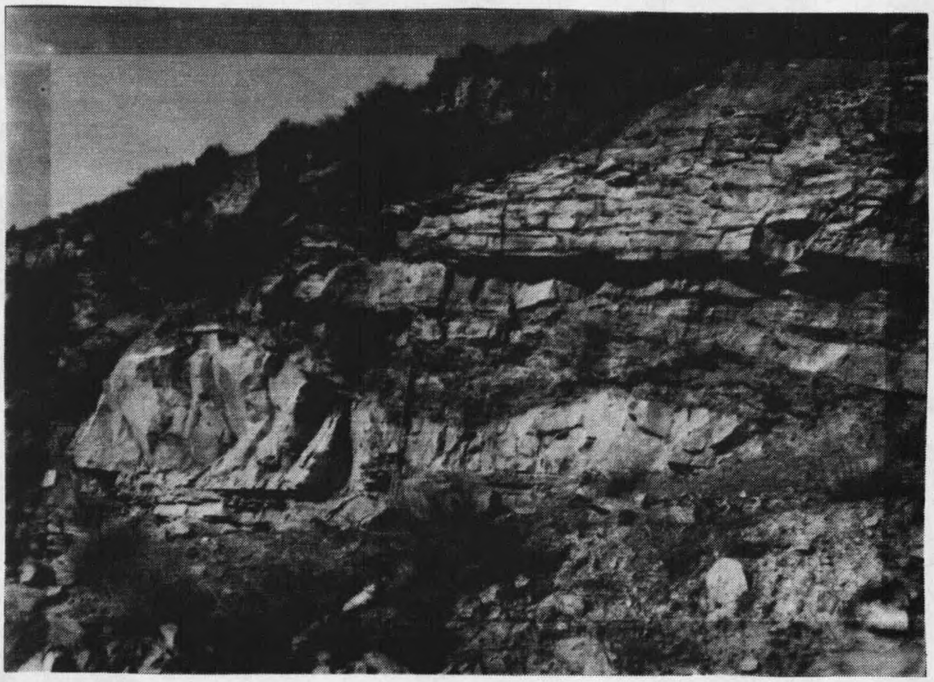

Figure 8. Photograph showing planoconvex geometry of sandstone body in the crossbedded sandstone facies at Wineglass Mountain.

Bioturbation in the sandstone facies takes two forms. In general, trace fossils are rare, but where observed are usually curvilinear, wrinkled, knobby cylinders, 1-1.5 centimeters in diameter and several centimeters long that emerge obliquely from bedding planes. These are probably a form of Ophiomorpha. A concentration of larger Ophiomorpha, 5 centimeters in diameter and 30 centimeters long is present at the top of a crossbedded sandstone interval at the Brackett Creek section. Horizontal, curvilinear, non-branching to slightly branching burrows, 1-2 millimeters in diameter and several centimeters long, occur in convex epi- and hypo-relief at many sandstone/shale contacts. These unidentified trails are common features throughout the lower sandstone member.

Sandstone intervals may also be thoroughly bioturbated with intensely disturbed laminations or a mottled texture of white sand blebs in a gray sandstone matrix. Vertical burrows, several millimeters in diameter, are present in several bioturbated beds though in general, individual burrows cannot be distinguished in these beds. Where bioturbated sandstone beds overlie and are in sharp contact with shale intervals the sandstone bed frequently contains numerous U-shaped burrows on its lower bounding surface.

Sandstone Facies Interpretation

The sandstone facies is interpreted to have been deposited under a variety of flow conditions. Tabular to wedge-planar decimeter scale crossbeds (Fig. 5) in the sandstone facies were produced by migrating straight crested megaripples. (Reineck and Singh, 1980, p. 38). Decimeter scale trough crossbeds are attributed to migrating dunes or lunate megaripples (Reading, 1978, p. 233; Reineck and Singh, 1980, p. 39).

Megaripples are lower flow regime bedforms, produced under uni-directional flow conditions, which have dimensions of 60 centimeters to 30 meters in length and 6 centimeters to 1.5 meters in height (Reineck and Singh, 1980, p. 41). Lower flow regime bedform development is mostly dependent on flow velocity and grain size, with a relatively minor influence by water depth in well sorted quartz sands like those of the lower sandstone member. For fine to medium sands, velocities of 40-70 cm/sec are necessary for straight-crested megaripple formation while lunate megaripples require slightly higher flow velocities, mostly more than 70 cm/sec in water depths of more than 2 meters (Reineck and Singh, 1980, p. 43).

Ripple cross-lamination in the sandstone facies is diagnostic of both oscillatory and uni-directional currents. Wave-generated ripple cross-laminations (Fig. 6) were recognized using criteria developed by Boersma (1970; cited in Reading, 1978, p. 250). These criteria include cross-laminations that are characteristically form discordant with internal structure that may: 1) contain bundled upbuilding where bundles are thin and oppositely dipping, 2) contain strongly tangential cross laminations which in extreme cases form offshoots that sweep up and ascend onto adjacent ripples, 3) have a variable angle of repose, and/or 4) have cross-lamination structure which differs in separate cross sections through a single ripple. In addition, the lower bounding surface of ripple sets is scooping and undulatory and the structures are frequently associated with even lamination of the straight, hummocky or low angle type. Symmetrical and less frequent asymmetrical ripple marks (Fig. 7) are associated with the ripple cross-laminations. This variety of ripple structure indicates that bedform development during deposition of the lower sandstone member was influenced by both uni-directional and oscillatory flows.

Some thin sequences of ripple cross-lamination, interbedded with megaripple crossbeds, exhibit sharp basal contacts and interwoven upper contacts with the ripple foresets dipping in a direction opposite to the overlying megaripple foresets. These small ripples were probably formed by backflow currents in the flow separation zone to the lee of migrating megaripples (Reineck and Singh, 1980, p. 28).

Horizontal to low-angle planar stratification may be attributed to either uni-directional or oscillatory flow. Associated symmetrical ripple forms suggest that these structures were probably produced under oscillatory flow conditions. However, if horizontal stratification was produced under uni-directional flow conditions, flow velocities of >100 cm/sec (Harms and others, 1982, p. 2-14) must have been common during deposition of the lower sandstone member.

Hummocky cross-stratification, which was observed only in the Elkhorn Mountains, is attributed to storm wave action with surges of greater displacement and velocity than those required to form wave ripples (Harms and others, 1982, p. 7-9). Hummocky cross-stratification is typical of a wave-dominated shoreface setting (Reineck and Singh, 1980, p. 398; Harms and others, 1982, p. 7-9).

Highly bioturbated sandstones indicate periods of slow deposition. The low sedimentation rates may be due to local factors which cause sediment-bearing currents to circumvent the area or low rates of subsidence that inhibit accumulation of sediment.

Multiple-bed sandstone intervals (e.g. at Wineglass Mountain exposure) are characterized by the presence of interbedded crossbeds and cross-laminations that are form-discordant to the sandstone bodies' planoconvex geometry. This reflects a hierarchy of bedforms in which the large, first order features are the sand bodies themselves. The internal structure of these sandstone bodies is form discordant, as the structures were produced by the smaller megaripples and ripples which represent the second and third order bedforms, respectively.

While most sedimentary structures in the sandstone facies are attributed to bedload transport, there is also evidence for cessation of flow. The presence of shale interbeds within some sandstone intervals documents quiet periods, as do horizontal fossil trails produced by deposit feeders typical of low energy conditions (Howard, 1978).

Direct evidence for marine influence in the sandstone facies is lacking due to the absence of body fossils in the lower sandstone member. The occurrence of Ophiomorpha is usually indicative of a high energy marine environment, though it has also been found in turbidites and fluvial deposits (Hoyt and Weimer, 1965; Frey and others, 1978). In the absence of body fossils, Ophiomorpha is the best faunal evidence for marine influence in the lower sandstone member.

Heterolithic Facies

The heterolithic facies comprises 14% of the lower sandstone member and consists of 50-90% very fine to fine grained, chert and mudstone-clast bearing quartzarenite interbedded with gray shale. Fine-grained carbonaceous debris is a common yet minor component of this facies and is usually concentrated on lamination or bedding surfaces. Trace fossils vary in abundance from non-existent to abundant on bedding surfaces. They include small horizontal trails and less frequent U-shaped burrows. Where U-shaped burrows are present, they are abundant and usually merge into vertical burrows in the overlying sandstone bed. Small, horizontal Thalassinoides burrows are present along sandstone-shale contacts at several sections. The heterolithic facies is further subdivided into the sand-dominated and mixed sub-facies on the basis of relative abundances of sand and mud.

The sand-dominated sub-facies consist of 75-90% sand with the remainder being mud. This sub-facies is characterized by tabular to slightly lenticular, persistent sandstone beds, 5-25 centimeters thick, separated by shale intervals approximately 5 millimeters thick (Fig. 9). Sandstone beds are typically in sharp contact with the adjacent shale and upper bedding surfaces are frequently symmetrical and slightly asymmetrical ripple-marked. Horizontal, parallel laminations are present in the lower 1/3 of sandstone beds. The middle and upper portions of each bed usually appear massive, though some exhibit discontinuous ripple cross-lamination. Small, horizontal trails are common on upper rippled bedding surfaces as are minor amounts (<1%) of fine-grained carbonaceous debris. This sub-facies

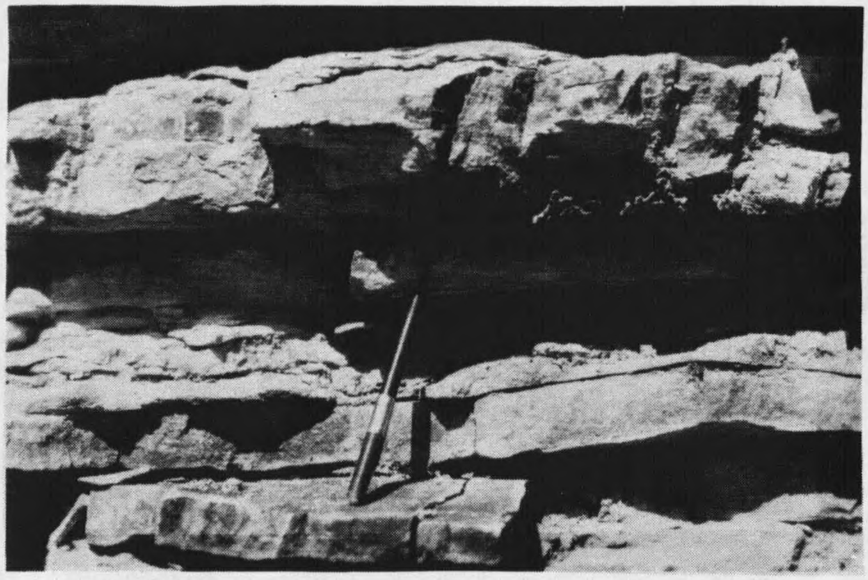

Figure 9. Sandstone beds separated by thin shale intervals in sand-dominated sub-facies. From outcrop along South Fork Sixteenmile Creek.

usually forms repetitive sequences of alternating sandstone and shale in bed sets up to 3 meters thick.

The mixed sub-facies contains 50-75% sand and 25-50% mud. This sub-facies is usually comprised of ripple cross-laminated sandstone with associated symmetrical and asymmetrical ripple forms (Fig. 10). Again, ripple cross-laminations with tangential cross-laminations and scooping to undulatory lower bounding surfaces are more abundant than foreset type cross-laminations characterized by a tabular or trough geometry. Horizontal, planar to wavy laminated and structureless thinly-bedded sandstones are associated with the ripple cross-laminated sandstone. Shale is present as drapes, thin lenses, or

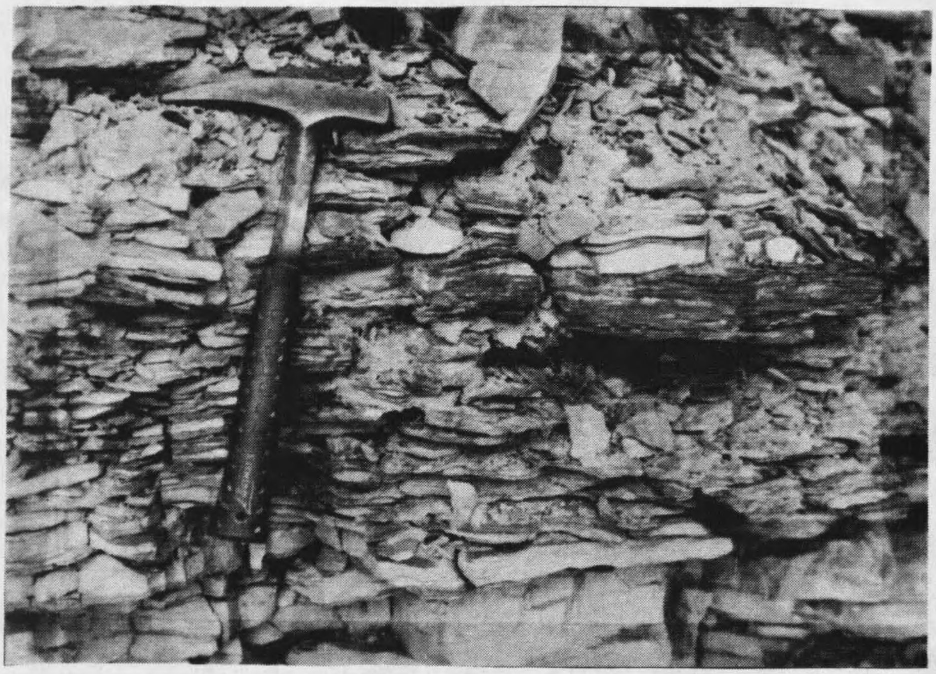

Figure 10. Flaser bedded sandstone and shale in the mixed sub-facies. From Wineglass Mountain section.

persistent beds generally <1 centimeter thick. This variety of complexly interbedded sandstone and shale is often referred to as flaser bedding (Reading, 1978, p. 233; Reineck and Singh, 1980, p. 113). Fining-upward sequences that are locally present within this sub-facies are characterized by a basal bed, 5-30 centimeters thick, which usually appears structureless but may contain horizontal, planar to wavy laminations. The basal bed is overlain by several centimeters of thinly lenticular-bedded, planar to wavy laminated and cross-laminated sandstone with minor interbedded shale. Shale content increases upward in these sequences.

Heterolithic Facies Interpretation

Deposition of complexly interbedded sandstone and shale requires high energy events, in which sand is introduced, alternating with quiet periods, during which clay is deposited from suspension. In shallow marine environments, these requirements may be met by tidal action (Allen, 1982, p. 511) or episodic storm events (Goldring and Bridges, 1973; De Raaf and others, 1977).

Goldring and Bridges (1973) describe a series of thin, sheet sandstones, which they refer to as sublittoral sheet sandstones, that are common to ancient sandy shelf sequences. Typically the main part of the sandstone bed is fine to very fine, well sorted, parallel laminated sandstone. Wave-generated ripple cross-laminations may be present near the top of sandstone beds and upper bounding surfaces may be modified by organic burrowing. Because only the tops of sandstone beds are burrowed, it is thought that these sandstone beds are single event deposits. Goldring and Bridges (1973) suggest that storm waves, possibly in combination with ebb tidal currents or storm surges, were important processes during deposition of sublittoral sheet sandstones. Shallow water turbidity currents, rip currents and tsunamis may have also played a role in their deposition. Goldring and Bridges (1973) note that sublittoral sheet sandstones occur either as amalgamations of individual sandstone beds separated by thin shale intervals or as amalgamated beds with little or no interbedded shale. They suggest that amalgamated beds that lack interbedded shale are associated with shorelines while sheet sandstones separated by shale interbeds are characteristic of a shallow marine open shelf. In the lower sandstone member, deposits of the sand dominated sub-facies are similar to Goldring and Bridges' (1973) shallow marine open shelf, sheet sandstones. The sand-dominated sub-facies is therefore interpreted as a storm- and wave-generated sequence deposited in a shallow marine open shelf setting.

Flaser bedded sandstone and shale, characteristic of the mixed sub-facies, is generally considered to be evidence of a tidally influenced depositional setting (Allen, 1982, p. 511). However, De Raaf and others (1977) have demonstrated that flaser bedding can also be produced in a wave dominated environment practically devoid of uni-directional currents. Their model is based on interpretation of sedimentary structures in flaser bedding in the Lower Carboniferous near Cork, Ireland. They suggest that the presence of wave-generated cross-lamination and near absence of uni-directional current generated cross-lamination in flaser bedding is diagnostic of a wave dominated environment.

Ripple cross-lamination structure in the mixed sub-facies of the lower sandstone member is interpreted as being dominantly wave-generated (wave-generated cross-lamination structure is discussed in sandstone facies interpretation section). Uni-directional current generated cross-lamination is present, but not nearly as abundant. Furthermore, symmetric ripple forms are much more common than asymmetric ripple forms in the mixed sub-facies. This evidence indicates that the mixed sub-facies was deposited in a wave dominated rather than tidally influenced setting.

Mud Facies

The mud facies comprises 22% of the lower sandstone member and is characterized by light to medium gray clay and mud shale with <10% interbedded very fine grained sandstone and siltstone. The sandstone usually occurs as symmetrical and rarer asymmetrical rippled lenses <1 centimeter thick and several centimeters long. These lenses contain horizontal to wavy laminations and ripple cross-laminations, or may be structureless. Rare, more persistent, structureless sandstone and siltstone beds up to several centimeters thick are also present. Small, horizontal trails in convex epi- and hypo-relief are fairly abundant at some sandstone/shale contacts. Minor amounts of fine carbonaceous debris are present on many of the lamination surfaces.

Mud Facies Interpretation

The presence of both shale and horizontal fossil trails indicates a quiet depositional setting that periodically experienced minor low energy events in which sand was introduced and worked into ripple forms. The uni-directional currents required for sand transport onto the mud substrate were probably storm induced as symmetrical, wave-generated ripple forms dominate. The gray shale in the lower sandstone member is almost identical to marine shale in the middle shale member and does not resemble the variegated, calcareous mudstone present in the non-marine Kootenai Formation. This supports interpretation of a marine environment for the lower sandstone member.

FACIES RELATIONS

Observation and interpretation of facies relations in the lower sandstone member is hampered by generally spotty and incomplete exposures. Complete, well exposed sections are present at only two locations (Rocky Canyon and Paddys Run) and even these are not laterally extensive. Despite the poor exposure, the facies relations observed provide valuable information in deciphering the depositional history of this unit.

Vertical Facies Relations

Vertical facies relations within the lower sandstone member are very similar throughout the study area with the exception of exposures in the Elkhorn Mountains and Jefferson Canyon. The following description covers all exposures except these anomalous sections.

All facies are thought to be present at each outcrop, though sometimes less resistant heterolithic, bioturbated sandstone, and shale intervals are covered at the poorer exposures. Vertical facies changes within the lower sandstone member typically occur on a decimeter to meter scale (see stratigraphic columns of measured sections in Appendix). No apparent stacking order was observed for the lower sandstone member as a whole, as each facies can occur at any vertical position within the section.

In spite of the absence of an overall, vertical stacking order to the lower sandstone member, potential exists for smaller scale facies trends. In order to determine the presence of these trends, Markov chain analysis was applied to facies transitions within the lower sandstone member (Table 1). In this method, observed transition probabilities are determined and random transition probabilities are calculated. Random transition probabilities depend only on the absolute abundance of the various facies and can be calculated using the equation $r_{ij} = n_j/(N-n_i)$ where r_{ij} is the random probability of transition from facies i to facies j, n_i and n_j are the number of occurrences of facies i and j respectively, and N is the total number of occurrences of all facies (Walker, 1979, p. 3). The difference between the numerical values of observed transition probabilities and random transition probabilities is used to determine whether transitions are more common than random. For facies transitions which occur more frequently than random, the observed minus random transition probabilities are greater than zero. Those transitions which have a positive numerical value for observed minus random transition probabilities in the lower sandstone member are those transitions from the heterolithic to the sandstone facies and from the mud to the sandstone facies (Table 1). These trends indicate the occurrence of coarsening-upward cycles that terminate in development of sandstone intervals.

Vertical facies relations at exposures in Jefferson Canyon differ from that pattern. Here crossbedded, planar laminated and cross-laminated sandstones are rare and instead the sequence is dominated by bioturbated sandstone. Additionally, the lower sandstone member is thinner than average, being only 10-15 meters thick.

		TO			
		S	H	M	
	S	---	.43	.57	Observed Transition Probabilities
	H	.83	---	.17	
	M	.95	.05	---	
FROM					
	S	---	.42	.58	Calculated Random Transition Probabilities
	H	.60	---	.40	
	M	.67	.33	---	
	S	---	.01	-.01	Observed Minus Random Transition Probabilities
	H	.23	---	-.23	
	M	.28	-.28	---	

Table 1. Markov chain analysis of facies transition within the lower sandstone member. S=Sandstone facies, H=Heterolithic facies, M=Mud facies.

Exposures in the Elkhorn Mountains differ in that a 5 meter thick interval of planar laminated sandstone caps the lower sandstone member. This is the only occurrence of a sandstone body in the lower sandstone member in which planar laminations are the dominant sedimentary structure. Symmetrical ripple forms and hummocky cross-stratification are intercalated with the planar laminated sandstone. The lower 12 meters of sections in the Elkhorn Mountains are comprised of interbedded shale, and crossbedded and ripple cross-laminated sandstone typical of other exposures of the lower sandstone member.

Lateral Facies Relations

Lateral facies changes are rarely observed on outcrop scale. Most thick (>3 meter) sandstone bodies can be traced for at least several hundred meters, and at several locales, such as the central Bridger Range (see stratigraphic columns for Brackett Creek and Fairy Creek in Appendix) and the northeastern Gallatin Range, some sandstone intervals are believed to correlate across several outcrops for distances of at least 5 kilometers. This correlation is based on visual correlation of stratigraphic profiles and walking out sandstone intervals between the measured outcrops where possible. The sandstone intervals may, and probably do, persist beyond the correlated sections, but the lack of exposure prevents further correlation. The sandstone body at Wineglass Mountain, which is characterized by a planoconvex geometry, is typical of these thick, persistent sandstone intervals. Heterolithic intervals which bound these correlated sandstones probably have an equivalent lateral extent. Thinner facies intervals often do not correlate from one outcrop to the next and the precise extent of these thinner sequences cannot be ascertained. In general, thick sandstone sequences are sheet-like, with a planoconvex geometry observed at the edge of sandstone bodies as suggested by outcrops at Wineglass Mountain (Fig. 8). These sandstones occasionally have a lateral extent of at least 5 kilometers while thinner sequences exhibit more lateral variability.

Facies Relations Interpretation

The sharp contact at the base of the lower sandstone member is probably a result of shoreface erosion which occurred during transgression of the sea across non-marine coastal plain deposits of the Kootenai Formation. This type of disconformity has been termed a ravinement (Swift, 1968) where observed on the U.S. Atlantic coast. Marginal marine deposits such as tidal inlet fills, dune sands, beach sands, and shoreface sands are not present and were presumably removed by shoreface erosion due to wave action.

The sharp contact at the top of the lower sandstone member marks a change from a sand-dominated shallow marine environment in the lower sandstone member to a mud-dominated offshore marine environment (Roberts, 1972; Schwartz, 1972). Schwartz (1972) states that this results from further westward transgression of the sea. The thin, discontinuous, micrite near the base of the middle shale member supports this interpretation as limestone deposition is a common response to a rapid transgression, during which clastic sediment supply to the offshore environment is greatly reduced (Ryer, 1977).

The depositional environment for the lower sandstone member was probably a shallow marine setting similar to modern marine shelves in depositional process and style. Mud and sand were deposited simultaneously, with their distribution controlled by sediment availability, bathymetry, and distribution of marine currents and storm paths. Sandstone was deposited by migration of individual bedforms, including ripples and megaripples, and by migration of bedforms associated with sand bodies which may have had a lateral extent of at least 5 kilometers as indicated by correlation of sandstone intervals. However, the extent of sandstone bodies in the lower sandstone member may not merely represent the dimensions of the sand bodies as they appeared during deposition. The observed dimensions may in part be due to lateral accretion of the sand body as it migrated on the bottom. These sand bodies contain crossbeds, planar laminations and cross-laminations that are form discordant to sand body geometry.

Field (1980) describes a sand ridge topography which is a dominant feature of the entire U.S. Atlantic inner continental shelf. These sand ridges typically exhibit 6-9 meters of relief, and vary from 9-15 kilometers in length and 1-3 kilometers in width. Swift and Field (1981) note asymmetric and symmetric ripple forms from these ridge crests and symmetric ripples, megaripples and sandwaves in ridge troughs. Much of the ridge surface has been bulldozed smooth by echinoids (sand dollars). Field and others (1981) describe similar sand ridges in the Northern Bering Sea. They note a hierarchy of three sizes of superimposed bedforms including large sand waves (height 2 meters, spacing 200 meters), smaller sand waves (height 1 meter, spacing 200 meters) and oscillatory and uni-directional current generated ripples. Sand ridges in both the North Atlantic and Northern Bering Sea exhibit documented historical migration (Field, 1980; Field and others, 1981). Field and others (1981) speculate that sand ridges in the ancient rock record would consist of lenses of sand, 10-15 meters thick, encased in a sandy shale. Sedimentary structures in the sand bodies should consist of ripple cross-laminations and tabular- and wedge-shaped crossbeds. Bedding planes should show evidence of lingoid and possible oscillatory ripples. Their description is remarkably similar to the sedimentary structures and geometry of sand bodies observed in the lower sandstone member. Thus, the multiple bed sandstone bodies in the lower sandstone member are interpreted as sand ridges.

Migration or constructional upbuilding of sand ridges and individual sand bedforms upon associated muds and storm deposits resulted in the coarsening-up sequences suggested by Markov chain analysis. The absence of an overall vertical trend or stacking of facies in the lower sandstone member indicates that there was no directional change in depositional environment, such as a change in sea level, during deposition of the lower sandstone member.

Similar sand ridges exist where sufficient uni-directional currents are generated by storm (Swift and others, 1971; Stahl and others, 1974; Stubblefield and others, 1975; Hunt and others, 1977; Swift and others, 1979; Field, 1980; Field and others, 1981; Swift and Field, 1981) or tidal processes (Swift, 1975; Swift, 1976). The presence of interbedded storm sequences and wave-generated ripple structures with sand ridge deposits in the lower sandstone member suggests that sand ridge development and maintenance was due to storm and wave-dominated processes. Additional support for interpretation of the hydraulic regime can be found in related nearshore deposits and paleodispersal patterns.

Because they are farther west, sections of the lower sandstone member in the Elkhorn Mountains must have been deposited nearer the inferred paleoshoreline during lower sandstone member deposition (Vuke, 1984). These sections are capped by a sequence of horizontal to low angle, planar laminated, and symmetrical ripple marked sandstone. Hummocky cross-stratified sandstone is present near the base of this capping sequence. This variety of sedimentary structures is typical of wave dominated upper shoreface sequences (Reineck and Singh, 1980). Thus the presence of wave generated, rather than tidal, nearshore deposits supports a model in which storms dominated the hydraulic regime in this part of the seaway.

Paleodispersal patterns in the lower sandstone member were obtained from decimeter scale planar crossbeds. The limited sample population (n=14) is due to the general lack of three-dimensional exposures of sedimentary structures. Qualitative observation of two-dimensional exposures of crossbeds suggests that the observed paleodispersal patterns are probably similar to trends that would be obtained from a larger population. Paleodispersal directions range from 329 to 75 degrees azimuth with no modes to the south (Fig. 11). A bimodal dispersal pattern would be expected from tidal deposits though a unimodal pattern does not preclude the possibility of tidal influence (Reading, 1978, p. 237). However, trends of symmetrical ripple crests are strongly east-west (Fig. 11) and markedly perpendicular to dominant planar crossbed dips. This suggests a genetic relationship between the uni-directional currents which formed the megaripples and the oscillatory flows which produced the symmetrical ripples. Such a relationship has also been noted in late Pre-Cambrian shelf deposits in Norway. There, a model was developed in which uni-directional currents were genetically related to wave activity (Levell, 1980). Such a model may also apply to lower sandstone member deposition where the hydraulic regime was probably

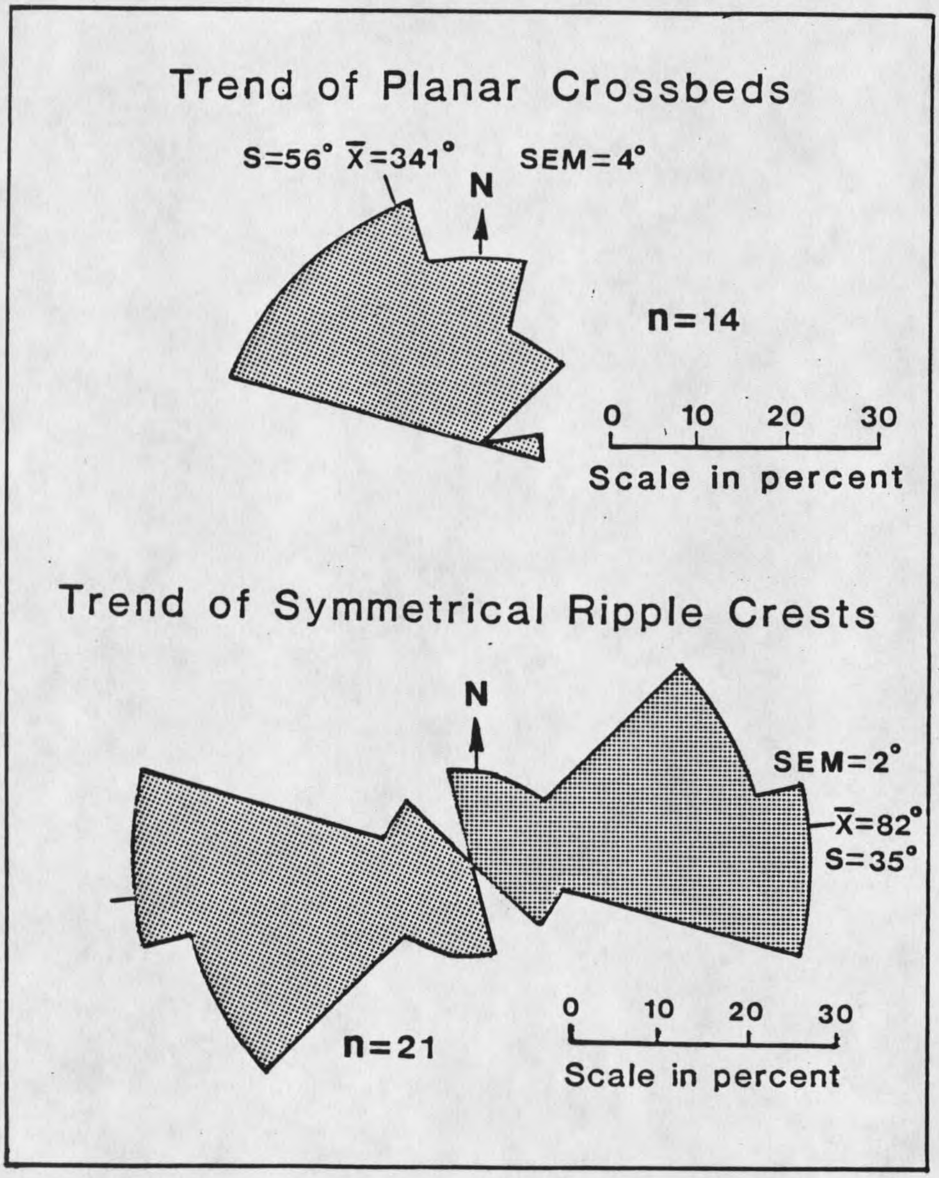

Figure 11. Rose diagrams showing both the trend of planar crossbeds and the trend of symmetrical ripple crests in the lower sandstone member of the Thermopolis Shale. X represents the vector mean, S the standard deviation and S.E.M. the standard error of the mean. Data from exposures at Fairy Creek, Rocky Canyon, Strickland Creek, Nixon Gulch, Cottonwood Gulch, Paddys Run, Sluicebox, and Wineglass Mountain.

characterized by genetically related uni-directional and oscillatory flow conditions which shared an origin in episodic storm events. The episodicity of these events is reflected in the presence of shale interbedded with the sandstone.

Interpretation of facies relations for the lower sandstone member in the Bozeman area also carries regional sedimentologic implications. A regional depositional model for the lower sandstone member, Interval A of the Blackleaf Formation (Schwartz, 1972), in southwest Montana has been presented by Schwartz (1972 and 1982). In this model, two distinct Early Cretaceous depositional provinces in southwest Montana have been defined (Fig. 12). They are: 1) a western province, located west and southwest of the Boulder batholith; and 2) an eastern province consisting of the area east and southeast of the Boulder batholith. The study area for this investigation is located in the eastern province.

Schwartz (1972 and 1982) notes marked differences in depositional environment for the lower sandstone member between the eastern and western provinces. In the western province, the lower sandstone member is comprised of interbedded very fine to fine grained litharenite, siltstone and shale. It is interpreted as a mixed distal-fluvial and lacustrine clastic sequence deposited in a fresh- to brackish-water setting. In contrast, the lower sandstone member in the eastern province consists of fine grained quartzarenite and subordinate sub-litharenite interbedded with minor amounts of siltstone and shale. Schwartz (1972 and 1982) and Vuke (1982) interpret the lower sandstone member in the eastern province as being

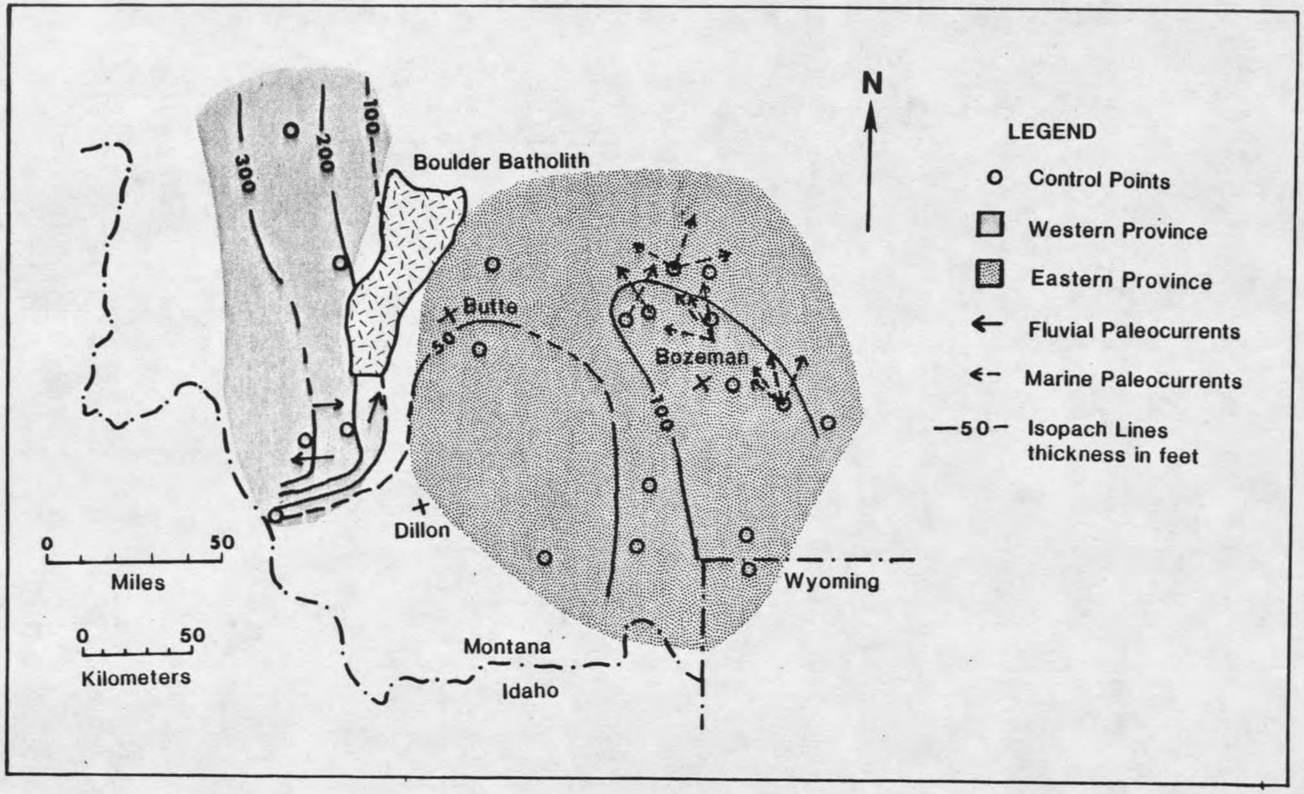

Figure 12. Isopach map of the lower sandstone member of the Thermopolis Shale in southwest Montana. All sections are restored to their approximate depositional locations. Also depicted are locations of western province (non-marine) and eastern province (marine). Data outside of study area from Schwartz (1982) and Vuke (1982).

deposited in a transgressive marine shoreface setting.

Observations and interpretations presented in this paper indicate that the lower sandstone member in the Bozeman, Montana area was deposited in a shallow marine shelf rather than the expected transgressive shoreface setting. This fact does not necessarily invalidate the model of Schwartz (1982). Because of shoreface erosion during transgression, marginal and nearshore marine intervals such as lagoonal and shoreface deposits were reworked and left no depositional record. Deposition of the lower sandstone member in the Bozeman area is therefore attributed to an influx of sediment into a shallow marine setting where deposition was controlled by shallow marine processes, rather than transgressive reworking of sediment in a nearshore environment. Since equivalent shoreline deposits were not recognized, it is difficult to ascertain actual shoreline fluctuations and possible influence of transgressive processes during lower sandstone member deposition. Because descriptions of the lower sandstone member in the rest of the eastern province (Schwartz, 1972; Vuke, 1982) closely resemble the lower sandstone member in the Bozeman area, it is likely that much of the lower sandstone member in the remainder of the eastern province was also deposited in a shallow marine rather than transgressive shoreface setting.

Schwartz (1982) also notes the existence of Early Cretaceous foreland structures in southwest Montana that influenced depositional patterns. The presence of an active positive element in southwesternmost Montana is indicated by an isopach map of the lower sandstone member (Fig. 12). This positive element is coincident with the northwesternmost expression of Laramide, basement involved uplifts in Montana. Though there is no evidence that this element contributed sediment during lower sandstone member deposition, it does appear to have been an area of slower subsidence as reflected by the thinning of the lower sandstone member. Exposures of the lower sandstone member in Jefferson Canyon occur along the northern edge of this trend. The lower subsidence rates there may have been instrumental in producing the anomalous section of bioturbated sandstone, which is usually indicative of slow deposition.

PETROGRAPHY AND PROVENANCE

The provenance for the lower sandstone member in the Bozeman area has not been well defined. Schwartz (1982) suggests that sands in the lower sandstone member of the eastern province were derived dominantly from the direction of transgression. It is unclear whether this means that sediment was derived by transgressive reworking of underlying sediment or if the southward transgressing sea was actively transporting sediment from the north. The following discussion reviews the petrography and paleodispersal patterns of the lower sandstone member in order to assess possible provenance for the lower sandstone member.

Petrographic Description

Petrographic analysis of the lower sandstone member includes point counts, to determine relative abundances of detrital mineral grains, of 10 samples collected from various exposures in the study area (Table 2). Cursory examination of 20 more samples indicated no differences that warranted further detailed examination. Coarser grained samples were chosen for petrographic analysis in order to provide consistency in sampling and because these samples generally contain a greater variety of framework grains and so should more accurately reflect composition of the source area. Most samples average fine sand size though they range from very fine to medium sand.

Sample Location	Mean Sand Size	Mono Quartz	Poly Quartz	Chert	Feldspar	Mud Clasts	Mud Matrix	FeO	Biotite	Muscovite	Zircon	Tourmaline	Porosity
Wineglass Mountain	fine	76	2	2	–	1	–	2	–	1	tr	–	15
Strickland Creek	fine	70	4	1	–	5	1	–	–	tr	–	–	18
Johnny Creek	very fine	81	3	4	–	9	3	1	–	tr	tr	–	–
Cottonwood Gulch	fine	86	2	1	–	8	2	–	tr	tr	–	–	–
Nixon Gulch	fine	72	5	2	tr	6	12	2	–	1	–	–	–
N. Fork Brackett	fine	72	6	1	–	4	5	1	–	–	–	–	10
Paddys Run	fine	90	4	2	–	tr	2	tr	–	tr	tr	–	1
Jefferson Canyon	very fine	54	1	3	tr	–	38	tr	tr	–	tr	tr	–
Rocky Canyon	fine	86	5	2	tr	1	1	tr	–	–	–	tr	–
N. Fork 16-mile	medium	71	8	7	tr	12	–	–	1	tr	–	tr	–

Table 2. Data set from point counts of thin sections of the lower sandstone member.

The fabric of the sandstone is characterized by long grain contacts, squashed mud clasts and generally low porosity. Quartz overgrowths are abundant in all but one sample, which was clay cemented. Quartz is the most common cement type, though clay, which was probably derived from squashed detrital mud clasts, acts as a patchy matrix or cement. Porosity is highly variable, ranging from 0-18 percent and is best developed in thicker sandstone bodies in the northeastern Gallatin Range. Qualitative observation of thin sections indicates that sandstones are mostly well sorted and detrital grains are subrounded indicating a high degree of textural maturity (Pettijohn and others, 1972, p. 588).

Quartz types present include monocrystalline and polycrystalline quartz. Authigenic quartz is common as quartz overgrowths, but due to the difficulty in distinguishing it from detrital grains, it is included in the monocrystalline quartz category. For relative abundances of detrital grains, see Table 2.

Mudstone-clasts are the second most common clast-type observed. They typically contain silt-sized quartz grains. Individual mudstone-clasts may range up to pebble size. Because of the compositional and textural maturity of this sandstone, mudstone clasts are thought to be intraformational and have no direct bearing on provenance determination for the lower sandstone member. Chert is present in all samples, but is a minor component and exceeds 5% in only one sample. Muscovite and biotite are present in trace amounts in most samples. Mica plates are usually bent around competent grains. Microcline and potassium feldspars are present in trace amounts in several of the samples. Zircon is also present in trace amounts as is tourmaline, which is always well rounded. Clay minerals are usually confined to mud clasts, though in one sample the clay matrix amounted to 37% of the sample. This clay did not appear to be well crystallized and was probably introduced into the sandstone bed by burrowing.

Provenance Interpretation

Provenance of the lower sandstone member is difficult to ascertain due to the compositional maturity of the sandstone. Possible sources include exposed land areas to the west of the Western Interior Cretaceous Seaway as well as transgressive reworking of the underlying Kootenai Formation. Since the Kootenai Formation had a source in the Sevier orogenic belt (Roberts, 1972; Suttner and others, 1981), the original source of sediment for the lower sandstone member should be west of the Western Interior Cretaceous Seaway. The following discussion evaluates this hypothesis.

MacKenzie and Poole (1962) delineate two petrologic suites of detrital framework grains in Dakota Group sandstones of the Western Interior, including Wyoming, South Dakota, Nebraska, Colorado and Kansas. The lower sandstone member of the Thermopolis Shale is lithostratigraphic equivalent to Dakota Group sandstones they studied in Wyoming. They found an "eastern" suite which was derived from the Continental Block Province east of the Cretaceous seaway and a "western" suite which was derived from the Sevier orogenic belt west of the seaway. Both suites are characterized by very high quartz abundance. In addition, the "eastern" suite contains, on the average, 2% feldspar and only trace amounts of chert. The heavy mineral fraction is dominated by zircon, tourmaline, rutile and chloritoid. The "western" suite contains, on the average, about 15% chert and only trace amounts of feldspar. The heavy mineral fraction is comprised mostly of zircon and tourmaline. Tourmaline grains found in the "western" suite are generally more rounded than tourmalines in the "eastern" suite.

Within the lower sandstone member, quartz grains are the principle component and chert ranks second in abundance of detrital grains. Feldspar is present in only trace amounts in a few of the samples. Zircon and tourmaline are the only heavy minerals observed and the tourmaline is well rounded. In comparison with MacKenzie and Poole's (1962) study, the composition and therefore provenance for the lower sandstone member most closely resembles that of the "western" suite whose source was the Sevier orogenic belt to the west. Support for this interpretation can be found in a plot of mean framework modes of the lower sandstone member and their relationship to different tectonic provenances (Fig. 13) as determined by Dickinson and Suczek (1979). Though most samples plot between defined provenances, some occur in the region occupied by the recycled orogen provenance while none occur in the continental block provenance. This indicates that the original source for the lower sandstone member was likely the Sevier orogenic belt rather than the eastern continental block provenance.

The Sevier orogenic belt may have provided sediment to the lower sandstone member by two different mechanisms. First, because the source for the underlying Kootenai Formation was the Sevier orogenic

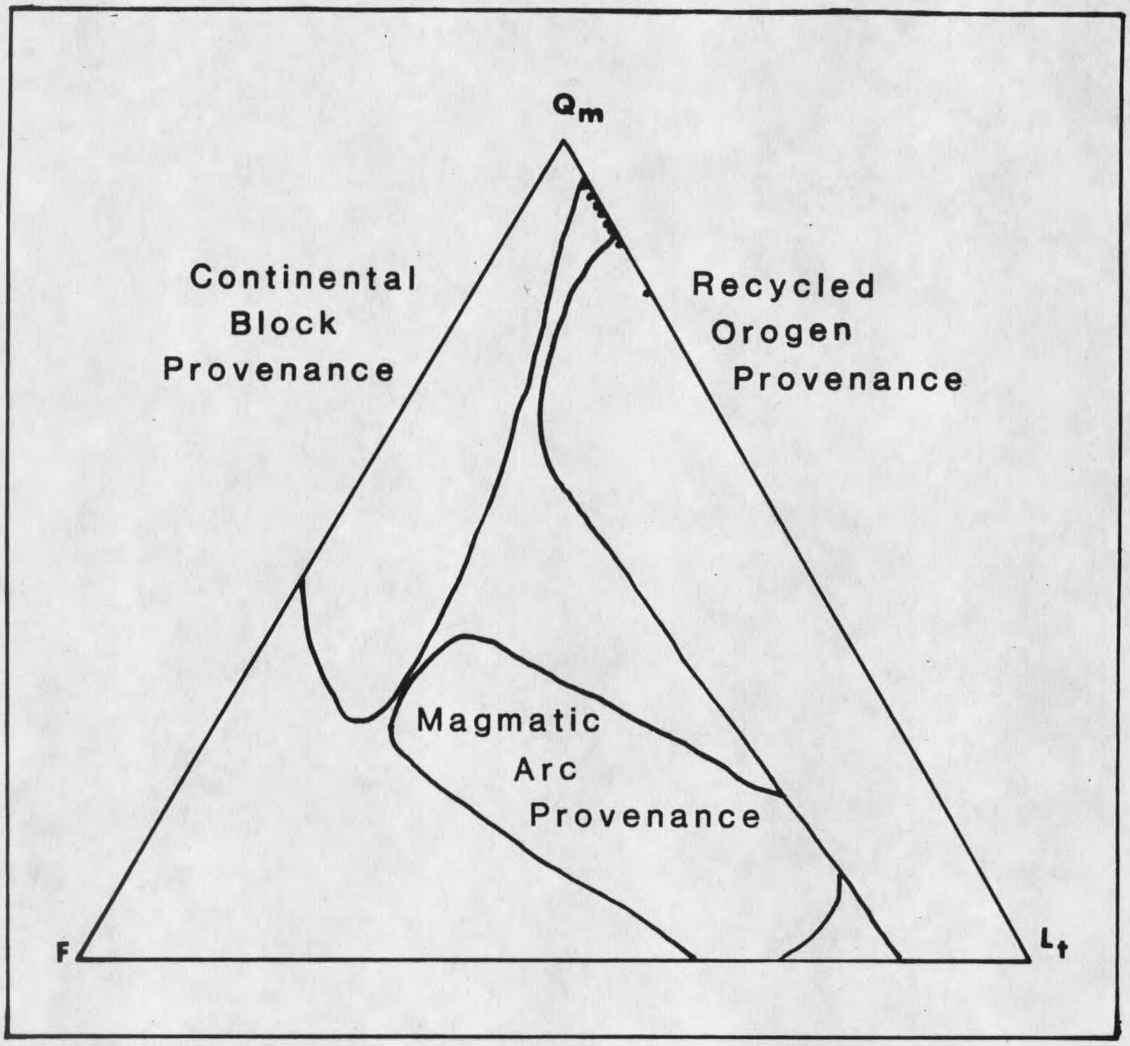

Figure 13. Ternary diagram showing mean framework modes of the lower sandstone member and the relationship to different tectonic provenances as determined by Dickinson and Suczek (1979): Qm is monocrystalline quartz grains, F is total feldspar grains, Lt is total lithic grains, including polycrystalline quartz grains.

belt (Roberts, 1972; Suttner and others, 1981), transgressive reworking of the Kootenai Formation may have supplied sediment to the lower sandstone member. Alternatively, fluvial processes may have transported sediment from the Sevier orogenic belt to the coast. Once there, the sediment may have been redistributed by coastal processes and incorporated into the lower sandstone member.

The Kootenai Formation varies from a chert-bearing quartzarenite to a chert-rich litharenite with upper Kootenai sandstones becoming progressively more quartz rich upwards (Suttner and others, 1981). Reworking of these sands in a shoreface and shallow marine setting could conceivably result in chert-bearing quartzarenites similar to those in the lower sandstone member. Kootenai sandstones in the Bozeman area occur approximately 30 meters below the contact with the lower sandstone member and are separated from the lower sandstone member by an interval of variegated, calcareous mudstone and biomicrite. No Kootenai sandstones were observed between the lower sandstone member and the mudstone and limestone at the top of the Kootenai Formation. It is possible that another interval of sandstone was present at the top of the Kootenai Formations but was totally reworked by the transgressing sea. However, no evidence was observed which supported this hypothesis and no mention of another sandstone interval above the limestone at the top of the Kootenai Formation was found in the literature (Suttner, 1969; Roberts, 1972; James, 1977; Suttner and others, 1981).

The Sevier orogenic belt may also have supplied sediment to the lower sandstone member in a more direct manner. Sediment may have been fluvially transported to the coast and distributed by coastal processes to the shelf where it was incorporated into the lower sandstone member. Evidence for this hypothesis can be found in fluvial equivalents of the lower sandstone member.

Distal-fluvial and lacustrine rock stratigraphic equivalents of the lower sandstone member in the western province include very fine to fine grained, litharenites and sub-litharenites of Interval A of the Blackleaf Formation (Schwartz, 1972, 1982). Interval A sandstones are composed dominantly of quartz and lithic fragments of "detrital carbonate clasts with minor presence of chert, quartzite, quartzarenite and quartz wacke fragments" (Schwartz, 1972). Winn and others (1984) have shown that significant compositional differences can exist between fluvial and marine sandstones derived from the same source. They found that marine sands of the Late Cretaceous Frontier Formation in southwestern Wyoming, contained 18% more monocrystalline quartz, 8% less chert and polycrystalline quartz and 9% less other rock fragments than related fluvial deposits. Causes of this are: 1) trapping of the coarser, more rock fragment rich sediment in fluvial environments and 2) destruction of softer rock fragments by wave action in either a beach or offshore setting. Thus further reworking of Interval A sands in a high energy nearshore setting could result in chert bearing quartzarenites similar to those observed in the lower sandstone member.

Though lower sandstone member equivalents in the western province may have supplied some sediment to the lower sandstone member in the eastern province, Schwartz (1982) suggests that positive elements in southwestern Montana may have acted as barriers to paleodispersal from the western to the eastern province. Thus, there may have been other locales in the Sevier orogenic belt that contributed sediment to the lower sandstone member in the eastern province. Figure 12 demonstrates a relatively thicker region of the lower sandstone member that extends from the Bozeman area south southeast into Yellowstone National Park. Paleocurrent data in the Bozeman area (Fig. 11 and 12) suggests that sand derivation may have been from the south-southeast. These paleocurrent data are not definitive since they rely upon such a small sample population (n=14). A larger sample population was not obtained due to the lack of well exposed, three-dimensional exposures of sedimentary structures necessary for paleocurrent measurement.

SUMMARY

Initial transgression of the Cretaceous seaway into southwest Montana was accompanied by shoreface erosion which produced the sharp contact or ravinement that separates non-marine deposits of the Kootenai Formation from the shallow marine lower sandstone member of the Thermopolis Shale. Subsequent to transgression, deposition of the lower sandstone member occurred in a shallow marine environment of low relief. Sediment was probably supplied to this setting by two processes. First, reworking of upper Kootenai Formation sands in the shoreface environment and ultimate stranding of the detritus on the shelf by transgression may have provided some of the sediment in the lower sandstone member. Second, sediment may have been introduced by fluvial systems to the coast from the Sevier orogenic belt located west of the seaway. Coastal storm processes then transported sediment offshore to the shelf where it was incorporated into the lower sandstone member.

First order geomorphic features on this shelf are large scale sand ridges, with dimensions of up to 5 kilometers. The sand ridges are characterized by an inherent heirarchy of bedforms. Megaripples and ripples are second and third order features respectively. The internal structure of sand ridges is form discordant and consists of a potpourri of interbedded decimeter scale crossbeds, ripple cross-laminations and horizontal to low angle planar laminations. Decimeter scale planar crossbeds are the most abundant structure in sand ridge

deposits. Development and maintenance of sand ridges is attributed to storm-induced uni-directional currents with a dominant northwestward (341 ±4 degrees) flow direction. Some sandstone intervals occur as isolated beds bounded by other facies intervals. The crossbedded, cross-laminated and planar-laminated sedimentary structures in the sandstone facies thus represent both sand ridges and isolated sand bedforms.

Storm deposits, bioturbated sandstone and shale occur in areas not occupied by sand ridges or isolated sand bedforms. The storm deposits are represented by intervals of the heterolithic facies.

Bioturbated sandstones represent deposition in areas of low sedimentation rates, allowing the fauna to totally disrupt internal structures. Low sedimentation rates may be accounted for by regional low subsidence rates, as in the Jefferson Canyon area, or local factors such as absence of sediment laden currents or temporary barriers which inhibited sediment accumulation. Mud was locally deposited in low energy or sheltered environments where storm influence was minimal.

Migration or repeated development and abandonment of sand ridges upon related deposits resulted in a complex vertical sequence of sediments characterized by interbedded sandstone, heterolithic and mud facies intervals. These local temporal changes may reflect changes in sediment supply, storm paths, distribution of marine currents, sea level, or a combination of these factors. Because of reinterpretation of the lower sandstone member as a shallow marine rather than shoreface sequence, it is not possible to ascertain whether the lower sandstone member is an actual transgressive deposit. This would require recognition of equivalent shoreline deposits and interpretation of shoreline fluctuations during lower sandstone member deposition.

Subsequent to lower sandstone member deposition, transgression reduced sediment supply to the marine environment. This produced the sharp contact separating the lower sandstone member from the offshore marine, middle shale member of the Thermopolis Shale. The low clastic sedimentation rate also allowed carbonate deposition to locally prevail, resulting in the thin, discontinuous argillaceous micrite bed near the base of the middle shale member.

Though this study describes the hydraulic regime and depositional characteristics of the lower sandstone member in the Bozeman area, little is known of the regional sedimentologic patterns of this shelf environment. Future work in the Crazy Mountains Basin and along the north flank of the Beartooth uplift would help in defining the lateral longshore as well as offshore extent of the paleo-sand ridge topography. Such work should prove useful in determining mechanisms of sediment transport as well as predictive models for ancient sand ridge fields.

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APPENDIX

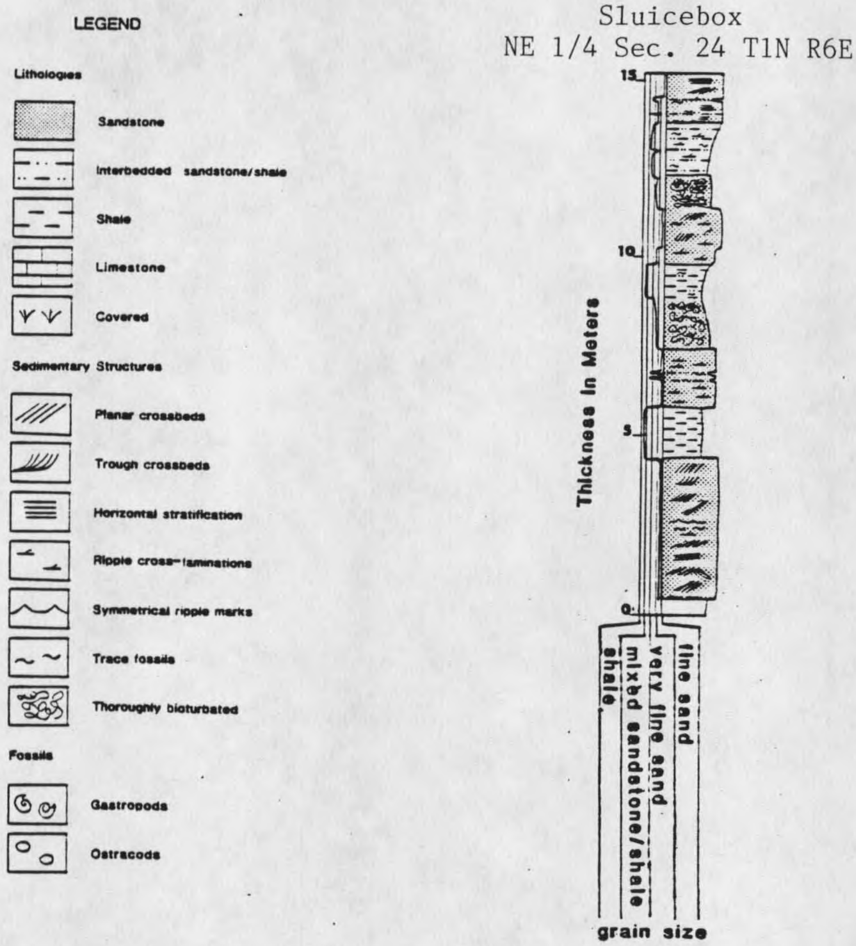

Figure 14. Location and stratigraphic column of Sluicebox outcrop.

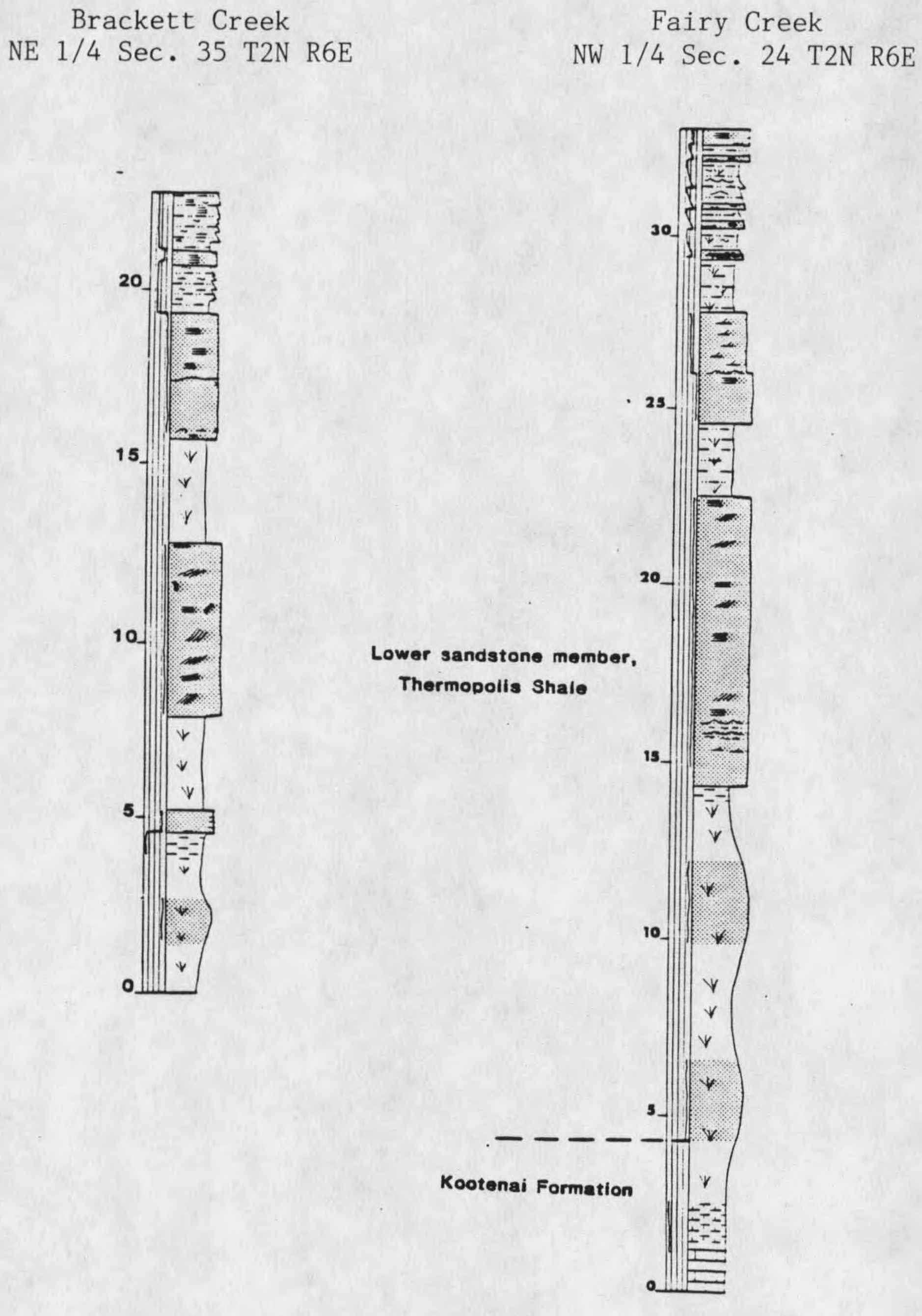

Figure 15. Locations and stratigraphic columns of Brackett Creek and Fairy Creek outcrops.

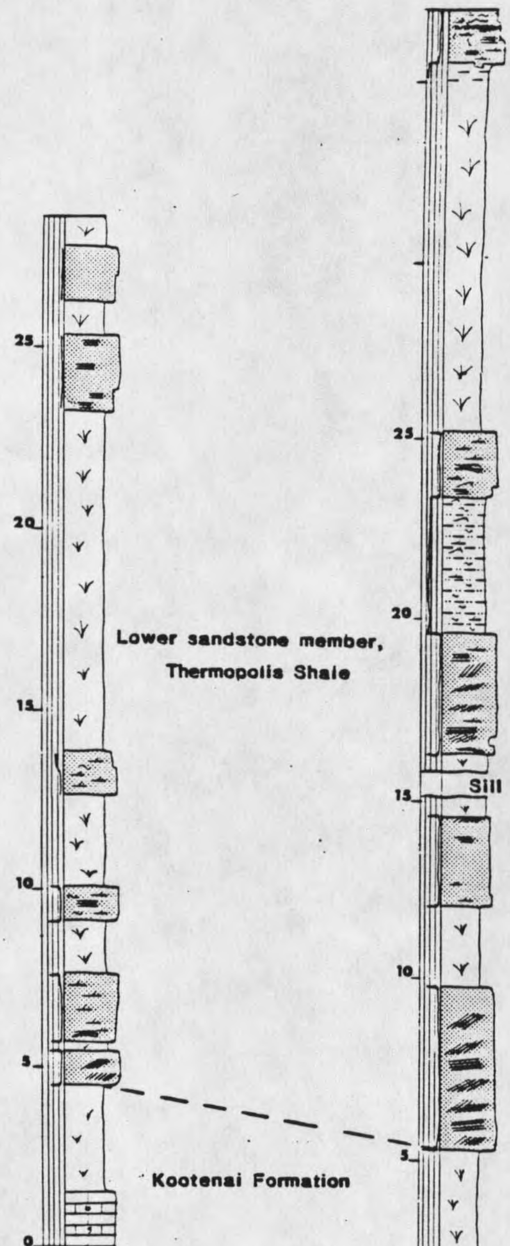

Figure 16. Locations and stratigraphic columns of Nixon Gulch and Cottonwood Gulch outcrops.

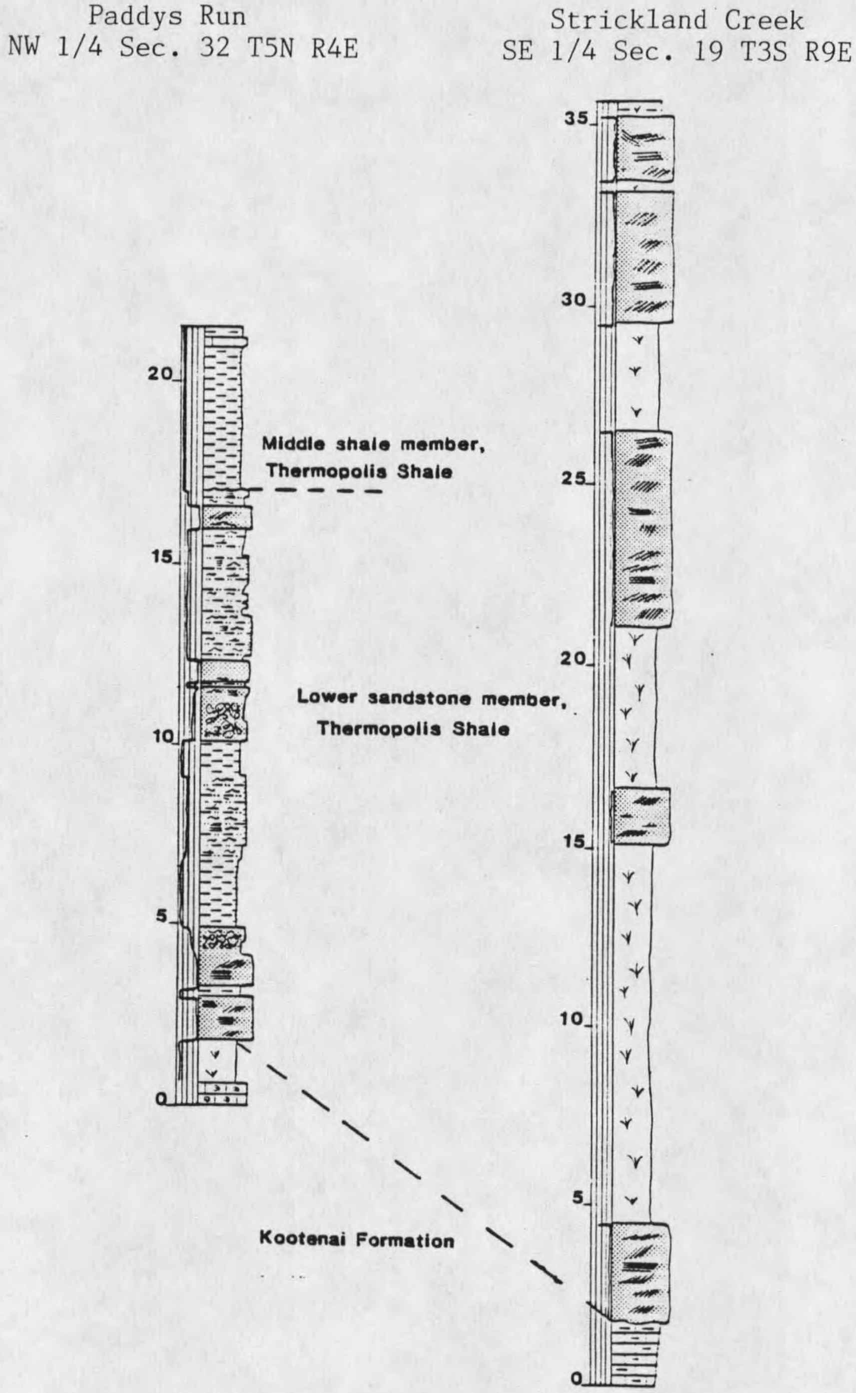

Figure 17. Locations and stratigraphic columns of Paddys Run and Strickland Creek outcrops.

Park Branch
NE 1/4 Sec. 28 T3S R9E

Wineglass Mountain
SE 1/4 Sec. 15 T3S R9E

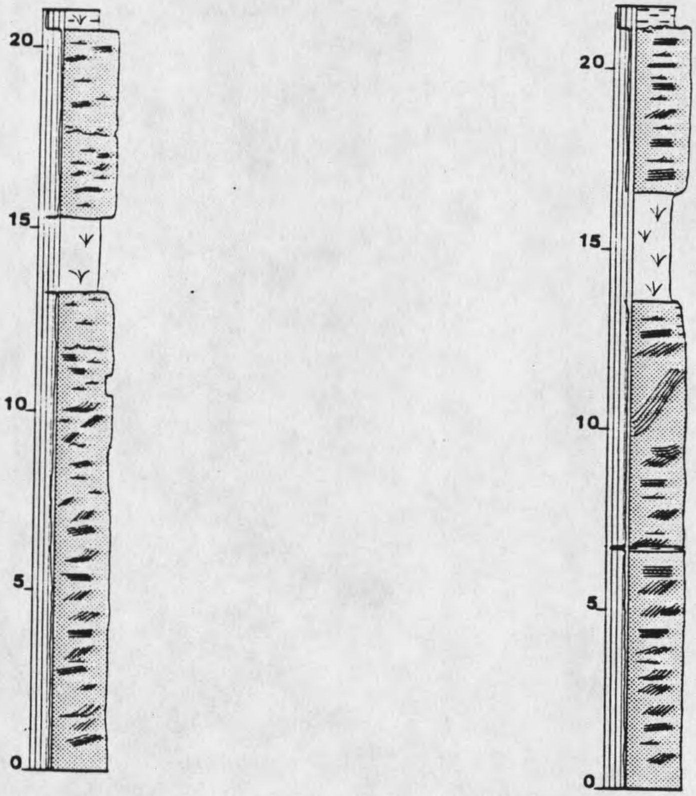

Figure 18. Locations and stratigraphic columns of Park Branch and Wineglass Mountain outcrops.

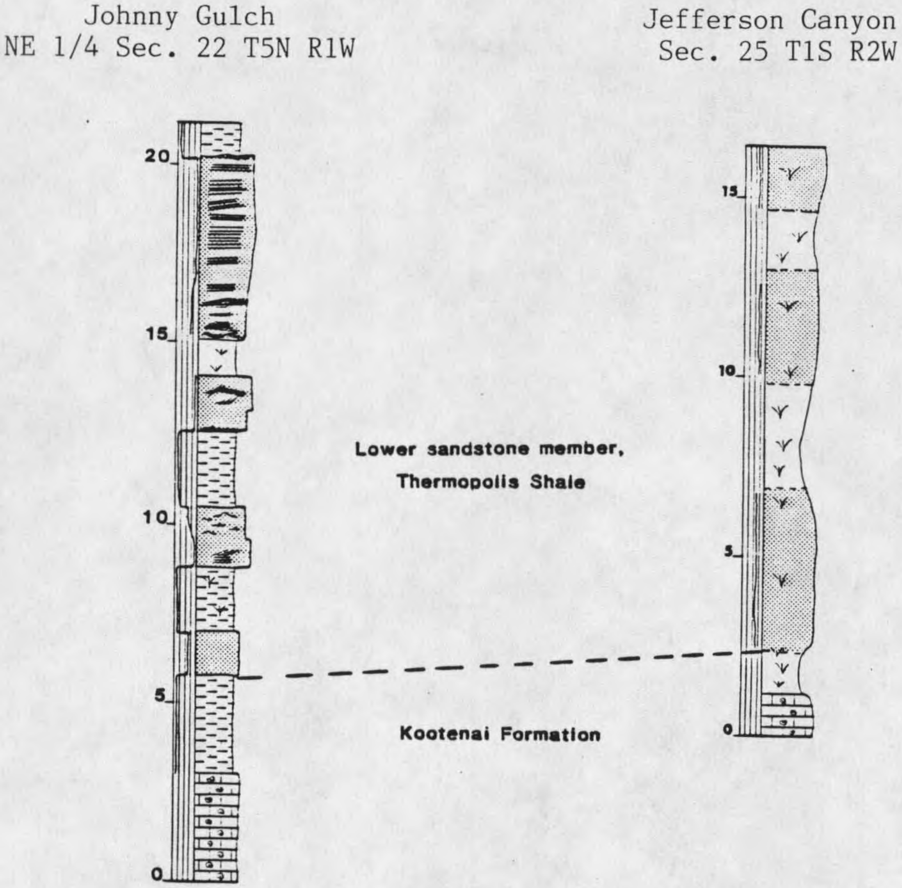

Figure 19. Locations and stratigraphic columns of Johnny Gulch and Jefferson Canyon sections.

Rocky Canyon
SW 1/4 Sec. 20 T2S R7E

Goose Pond
NE 1/4 Sec. 3 T3S R7E

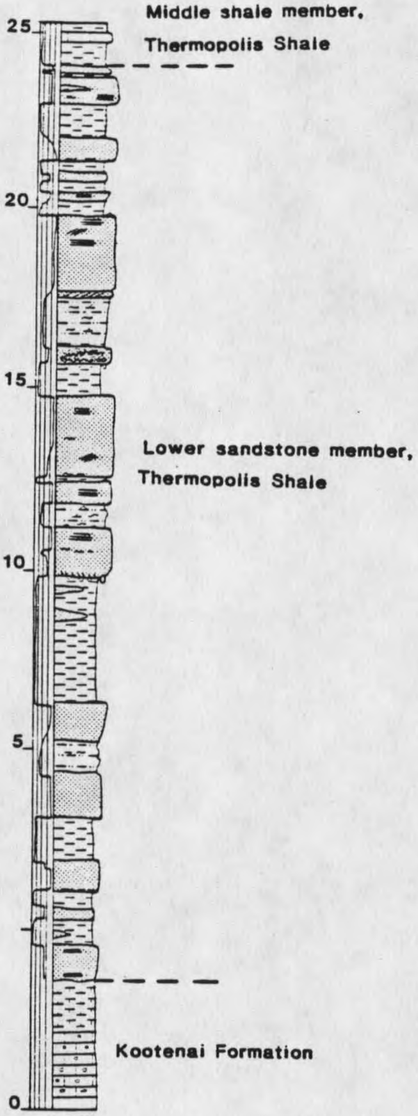

Figure 20. Locations and stratigraphic columns of Rocky Canyon and Goose Pond sections.

Middle Fork Sixteenmile
NW 1/4 Sec. 11 T4N R6E

Figure 21. Location and stratigraphic column of Middle Fork Sixteenmile outcrop