AN ABSTRACT OF THE THESIS OF

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BEAVERHEAD COUNTY, MONTANA

Abstract approved: Signature redacted for privacy.
Keith F. Oles

The Middle and West Forks of Little Sheep Creek in the southern Tendoy Range have incised valleys across Cenozoic structural features exposing strata that range in age from the Mississippian to the Neogene. Paleozoic strata are 1,349 m thick and belong to the Mission Canyon Limestone, the Big Snowy, Amsden, Quadrant, and Phosphoria Formations. The Scott Peak Formation of the White Knob Group of Idaho is allochthonous and forms the upper plate of the Medicine Lodge thrust. Paleozoic and Triassic strata of the thesis area represent sedimentation across a transition zone between a stable craton to the east and the Cordilleran miogeosyncline to the west. Regional unconformities are recognized locally at the top of the Mission Canyon Limestone and the Phosphoria and Thaynes Formations, but not at the top of the Big Snowy, Amsden, or Quadrant Formations.

Mesozoic strata have a total thickness of 1,404 m and belong to the Dinwoody, Woodside, and Thaynes Formations, the newly recognized Gypsum Spring Tongue of the Twin Creek Formation, the Sawtooth and Rierdon Formations of the Ellis Group, the Morrison and Kootenai Formations, the Colorado Shale, and the Beaverhead Formation. Cenozoic strata are represented by the newly named Round Timber limestone (informal) of the

Medicine Lodge beds (Miocene) and the Edie School rhyolite (Pliocene).

Detailed stratigraphic and petrographic analyses were made of the Triassic Dinwoody, Woodside, and Thaynes Formations, the Jurassic Gypsum Spring Tongue, and the Miocene Round Timber limestone in order to determine environments of deposition.

The limestones, calcareous siltstones, and silty limestones of the Dinwoody and Thaynes Formations were deposited in a shallow marine shelf environment as a result of two transgressive pulses separated by an Early Traissic regression. The Triassic seas had transgressed eastward onto the craton from the miogeosyncline. The Dinwoody and Thaynes fauna indicate normal salinities and open marine conditions; the widespread regional distribution of the limestone-siltstone facies indicates broad equable conditions for sedimentation. Deposition was primarily a tractive process generated by storm-driven, tidal, and longshore currents within a maximum depth of approximately 50 m. The Early Triassic regressive phase is represented by the deposition of the variegated siltstone, sandstone, limestone, and dolomite of the Woodside Formation in a tidal flat environment.

The Gypsum Spring Tongue consists of interbedded variegated siltstone, sandstone, limestone, dolomite, and limestone conglomerate that were deposited in a tidal flat and restricted marginal marine environment extending east from southwestern Montana and eastern Idaho across Wyoming and southern Montana. The Middle Jurassic sea transgressed south across the North American continent.

The Miocene Round Timber limestone was deposited in a fresh-water lake in which calcite was being deposited as encrustations on the green algae Chara and as a precipitate directly from solution.

The folding and faulting within the thesis area are the result of cratonic and miogeosynclinal responses to Cretaceous-Early Tertiary orogenesis. A southwestward-plunging anticline has been refolded into northeastward-yielding, overturned, doubly-plunging folds oriented northwest-southeast. High angle reverse faults of minor displacement have occurred along the southeastern limb of Garfield anticline and within the axis of the Seybold syncline. Post-Laramide relaxation of compressional forces has caused north- and northwest-oriented normal faults that transect earlier structures. Sandstone and limestone of the Scott Peak Formation of the White Knob Group were implaced along the Medicine Lodge thrust.

Structure and Stratigraphy of the Little Sheep Creek Area, Beaverhead County, Montana

by

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THE STRUCTURE AND STRATIGRAPHY OF THE LITTLE SHEEP CREEK AREA, BEAVERHEAD COUNTY, MONTANA

INTRODUCTION

Location and Accessibility

The Little Sheep Creek area lies within the drainage systems of the West and Middle Forks of Little Sheep Creek in the southern part of the Tendoy Range, Beaverhead County, Montana and Clark County, Idaho. The Tendoy Range is a low-lying range of mountains that trends northwest for approximately 70 km from the northern margin of the Snake River Plain in Idaho to the Hap Hawkins Reservoir. The mountains are bounded on the east by the Red Rock basin and the western part of the Centennial basin. The Beaverhead Range parallels the Tendoys to the west and is separated from the Tendoys by the Horse Prairie basin and the North and South Medicine Lodge basins (Figure 1).

The thesis area is approximately 13 km southwest of Lima, Montana on the west flank of Garfield Mountain. The area is accessible by a county and Forest Service road that serves the East Creek campground on the East Fork of Little Sheep Creek and the White Pine Ridge area north of Gallagher Gulch. The West and Middle Forks are accessible by private roads that serve several homesteads. The southeastern part of the thesis area is barred to motorized vehicles and access is by foot or horseback. Access to the southern part of the area can be made by various jeep trails from Gallagher Spring, Four Eyes Canyon, or Medicine Lodge Pass.

The area mapped covers approximately 60 km^2 in Sections 31-34,

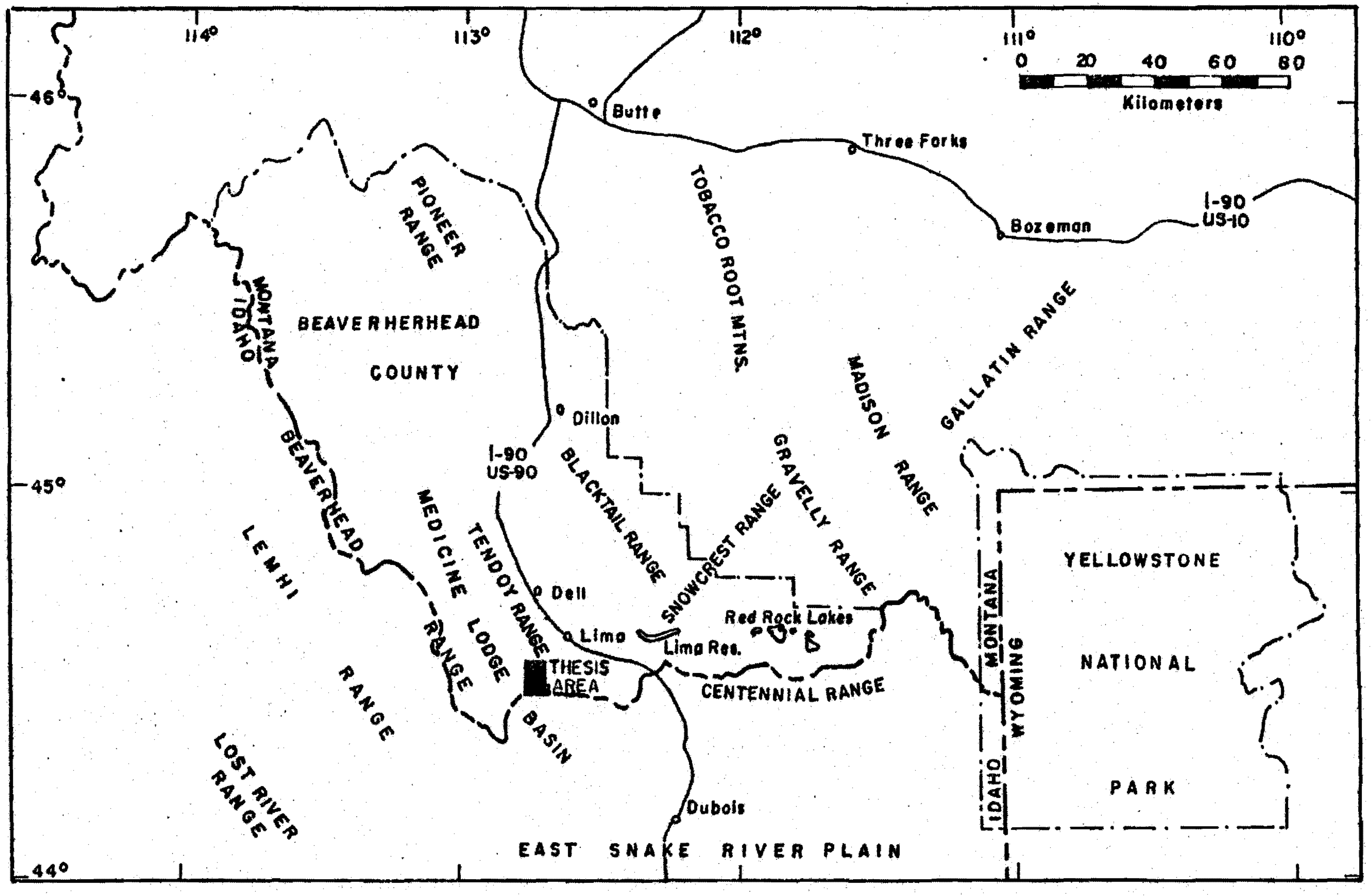


Figure I. Location map of the Little Sheep Creek thesis area.

T. 14 S., R. 9 W., and Sections 3-10, 7-10, 15-23, 26-30, T. 15 S., R. 9 W. in the Gallagher Gulch Quadrangle, Beaverhead County, Montana and Clark County Idaho, and the Caboose Canyon Quadrangle, Beaverhead County, Montana. Except for two homesteads each in the Middle and West Forks of Little Sheep Creeks, the thesis area lies within the confines of the Beaverhead and Targhee National Forests of Montana and Idaho, respectively.

Purpose

The area encompassed by the Tendoy Range and the eastern flank of the Beaverhead Range north of the Snake River Plain acted as a transition zone between two distinct styles of sedimentation from the Late Precambrian to the Triassic. East, on the relatively stable craton, Cambrian and younger rocks rest on Precambrian crystalline basement. Precambrian crystalline rocks are not known to the west; instead, Cambrian and younger rocks lie on a thick sequence of Precambrian sedimentary rocks.

There are considerable changes in the thicknesses of the sedimentary strata across the transition zone; commonly sedimentation that was continuous in the miogeosyncline is marked by widespread unconformities on the craton. Facies in the two sedimentation provinces overlap or intertongue within the transition zone; depending on regional diastrophism or the rates of sediment supply, the unconformities may or may not be recognized within the transition zone.

The transition zone is coincident with Kay's (1951) Wasatch Line, or the hinge line as defined by Scholten (1957), that extends as a curvilinear belt from northern Mexico through the middle western states and into Canada. Besides marking a transitional zone between two styles of sedimentation, the Wasatch Line was also the area affected most intensely by the Sevier and Laramide orogenies during the Early Cretaceous to the Early Tertiary; thrusting and folding typical of the ancient miogeosynclinal sequence are superimposed on basement-controlled uplifts on the craton.

The Little Sheep Creek area was selected for study because it lies within the transition zone between two distinct sedimentological and tectonic provinces. A reconnaissance map prepared by Scholten et al. (1955) at a scale of 1:126,720 of the Tendoy and Beaverhead Ranges also indicated that a stratigraphic section typical of southwestern Montana from the Ordivician to the Neogene was represented in the proposed thesis area. As revealed in the following text the structure and the stratigraphy mapped during the.preparation of this thesis vary in some detail from the initial reconnaissance map.

The purposes of this thesis are three-fold. The first is to prepare a detailed geologic map of the bedrock geology; the second is to determine by outcrop, hand sample, and petrographic analysis the depositional environments of the Triassic Dinwoody, Woodside, and Thaynes Formations, and the newly discovered Jurassic Gypsum Spring Tongue of the Twin Creek Formation; the third is to reconstruct the geological history of the Little Sheep Creek area.

Investigative Methods

The field investigation was accomplished during a nine-week period from July 7 to September 5, 1979. Six hundred and thirty field stations were plotted on United States Department of Agriculture air photographs at a scale of 1:15,840. The data were later transferred to United States Geological Survey 7.5 minute topographic maps enlarged to a scale of 1:12,000.

Bedding thicknesses and stratification were measured in the field using a metal tape; for purposes of discussion the thicknesses have been converted to a standard terminology for stratification, cross stratification, and splitting properties established by McKee and Weir (1953). Paleocurrent directions were corrected for strike and dip after the method described by Briggs and Cline (1967). Bedding attitudes were measured by a Brunton compass. Stratigraphic thicknesses were measured directly by using a Jacob's staff and Abney level, a clinometer, or by pacing and then adjusting the thickness for bedding attitude.

Rock colors were determined using the Geological Society of America Rock-Color Chart (1963). Sandstones were classified according to Williams, Turner and Gilbert (1954). A reference sand grain gauge was used to determine sand size, degree of rounding, and sorting.

Carbonates have been classified according to Folk (1962). The degree of crystallinity of the carbonates was compared to the standard Wentworth sand size scale of clastic sediments; fine, medium, and coarsely crystalline carbonates can be compared to the fine, medium, and coarse sand size fractions. The presence of dolomite was determined by comparing the reaction rates of carbonates to 10% hydrochloric acid: dolomites react very slowly to the acid unless powdered. Field determinations for dolomite were later confirmed by using the dolomite staining method outlined in Appendix A.

Selected samples of Mississippian rocks, the Round Timber limestone, and all of the samples of Triassic rocks and the Gypsum Spring Tongue of the Twin Creek Formation were cut, polished, and examined using a binocular dissecting microscope. Eighty-nine samples were then selected for further petrographic analysis as thin sections.

I have tried to be consistent with the terminology outlined above. When using information derived from other authors I have followed the terminology used by the authors in order to avoid altering their initial interpretation through translation. Measurements in the field or in the laboratory were made using the SI system of measurements. I have converted FPS system measurements used by other authors to the nearest whole SI number and have provided both in the text.

Reference is often made to the "Lima region" in this text. This is the region defined by Scholten et al. (1955, p. 346) that encompasses the southern part of Beaverhead County, southwestern Montana and extends into Clark County, east-central Idaho.

Previous Investigations

The first geological observations of southwestern Montana were made by the Hayden Survey and published in 1872 (Scholten et al., 1955, p. 346). Douglass (1905) and Condit (1918) discussed the Paleozoic and Mesozoic stratigraphy of southwestern Montana on a reconnaissance basis. In 1927, Kirkham described 3,000 feet (915 m) of light colored Cretaceous sandstones and shales in the Centennial Range. He also described Tertiary lake deposits along Medicine Lodge Creek southwest of Garfield Mountain and the Medicine Lodge thrust where it extends into Idaho.

Post World War II interest in economically recoverable resources of uranium, phosphate, graphite, and petroleum resulted in extensive investigations of the stratigraphy and structure in southwestern Montana. Perry (1949) described the gypsum mine in the Big Snowy Formation on the East Fork of Little Sheep Creek. During the summers of 1947 and 1948 field investigations were made by three students from the University of Michigan pursuing Doctor of Philosophy degrees. The results of these field investigations were combined into a paper discussing the geology of the Lima region by Scholten, Keenman, and Kupsch in 1955. This paper has been a standard for subsequent investigations in the region of the Tendoy and Beaverhead Ranges. Klepper (1950) had included the Lima region in a reconnaissance of Beaverhead and Madison Counties, Montana. Sloss and Moritz (1951) provided a comprehensive description of the Paleozoic stratigraphy of southwestern Montana, and Moritz (1951) discussed the Triassic and Jurassic strata. In 1953 Lowell and Klepper described the Cretaceous-Paleocene Beaverhead Formation. Weaver (1955) described the petrography and nature of the Quadrant-Phosphoria contact in southwestern Montana. Cressman and Swanson (1964) later published a report on the stratigraphy and petrology of the Permian rocks of southwestern Montana and Idaho as part of a broad study of western phosphate fields.

A study of the Triassic System in western Wyoming, eastern Idaho, northeastern Utah, and southwestern Montana was undertaken by Kummel in the early 1950's; this culminated in papers on the distribution (1954) and the paleoecology (1957) of the Early Triassic strata in the region and in 1960 a brief discussion of the Triassic in southwestern Montana.

In 1945 Imlay published a paper on the occurrence of Jurassic rocks in the western interior United States. Further investigations ensued resulting in publications by Imlay in 1950, 1952, 1953 and 1957, and Imlay et al. in 1948 on the stratigraphy, paleoecology, and correlation of Jurassic strata in Montana, Wyoming, and Idaho. In 1969 Suttner discussed the distribution of the Morrison and Kootenai Formations in light of the initial effects of Late Mesozoic orogenesis and changes in the tectonic behavior of the western edge of the craton.

Huh (1967) described in detail the stratigraphy and correlation of the Mississippian strata from Montana to Idaho across the Wasatch Line. Maughan and Roberts (1967) published a paper that provided an insight into deposition of the Big Snowy and Amsden Formations and the nature of the Mississippian boundary in Montana. In 1976 Rose presented a model for sedimentation on Mississippian shelf margins in the central interior of the United States that included deposition of the Mission Canyon, Big Snowy, and White Knob Formations in southwestern Montana and Idaho. Rose's model was improved upon by Gutschick, Sandberg, and Sando (1980) who presented a history of the Kinderhookian to middle Meramecian carbonate shelf from Montana to Nevada that was synchronized by foraminiferan, conodont, and coral zonations.

Scholten (1957) discussed evidence for recurrent Paleozoic positive elements along the Wasatch Line. In 1960 Eardley described the phases of Laramide orogeny in the deformed belt of southwestern Montana and adjacent areas of Idaho and Wyoming. In 1960 Scholten described the depositional and tectonic events in southwestern Montana. Ryder (1968) published a comprehensive study of the Beaverhead Formation as a syntectonic conglomerate deposited in the Beaverhead and Tendoy Ranges. Ryder and Ames (1970) were able to date the Beaverhead Formation and Ryder and Scholten (1973) further clarified the stratigraphy, origin, and tectonic significance of the Beaverhead Formation.

Scholten and Ramspott (1968) discussed the mechanism for the folding and thrusting evident in the Beaverhead Range. Scholten (1973) provided a model linking the emplacement of the Idaho Batholith with the folding and the low-angle faulting found in the mountains to the east. He considered the folds to be the result of compressive forces on the flanks of the emerging batholith. These stresses were released by the development of low-angle thrust faults which propogated east, while local uplifts gave rise to epidermal glide sheets such as the Medicine Lodge thrust of the Tendoy Range. Skipp and Hait (1977) and Ruppel (1978) present interpretations of the thrusting that differ from Scholten's.

The most recent work available to me is a M. S. thesis by R. A. Klecker completed in 1980 on the Little Water Canyon area approximately 17 km to the north, and a M. S. thesis, still in progress, by G. D. Hildreth concerning the Mississippian biostratigraphy on the east flank of the Armstead anticline, both of Oregon State University.

Geomorphology

The maximum relief of the area is 1,758 feet (537 m), ranging from 6,892 feet (2,083 m) at the mouth of Gallagher Gulch to 8,650+ feet (2,638+ m) on the Continental Divide. The topography is both structurally and lithologically controlled; folded strata of resistant limestone and sandstone form cuestas and hogbacks, and less resistant mudstones underlie valleys; flat-lying strata weather and erode to form broad, flat divides.

The Little Sheep Creek area exhibits a well-developed drainage system consisting of deep valleys with steep sides and rounded, wellvegetated divides. Two tributaries of Little Sheep Creek, the West and Middle Forks, form the dominant drainage systems of the area (Figure 2). The West Fork of Little Sheep Creek, and Little Sheep Creek below the confluence of its tributaries, are antecedent streams developed during the uplift of the present Tendoy Range that began in the Pliocene. The rate of uplift was slow enough for the streams to incise channels across the dominant Laramide structures. The Middle Fork demonstrates structural adjustment of its flow direction and is subsequent upon the softer Permian and Triassic strata through which it flows; the westward flow of the stream in the southeastern part of the map area roughly parallels the strike of the formations. The northward bend in Section 16, T. 15 S., R. 9 W. may have been controlled by the Middle Fork Fault. The Middle Fork becomes antecedent as it flows across the strike of the formations in the northern part of the area. Gallagher Gulch and Two Spring Gulch are also subsequent intermittent streams; northward-flowing tributaries of the Middle Fork in the southeast part of the map are antecedent to the structural trend of the underlying strata (Figure 3).

The topography is not consistent with that which is ordinarily found in semi-arid regions. Lacking are the deep canyons and gullies providing continuous vertical and lateral exposures of the strata. The Continental Divide and other prominent ridges and divides are capped, for the most part, by gravel. The gravel and the varying permeability of the underlying sedimentary rocks reduce the amount of precipitation runoff; water percolates through the strata to reappear as springs at the exposed interfaces of rock types of differing permeabilities. As a result, the area exhibits numerous springs whose numbers and flow rates vary according to the changes in the annual rate of precipitation.

Figure 2. View looking north northeast down the West Fork of Little Sheep Creek from the NW¼, Section 20, T. 15 S., R. 9 W. Red Rock basin distant center; White Pine Ridge left center; Middle Fork at base of tree-covered slope distant right center.

Figure 3. View looking east up the Middle Fork of Little Sheep Creek from the NW¼, SE¼, Section 21, T. 15 S., R. 9 W. Garfield Mountain (Elevation 3,343 m) in center left.

The movement of groundwater and the ubiquitous springs have been a factor in the type of mass wasting found in the area. Along with soil creep, there are noticeable examples of slumps and debris slides, some of which coincide with the springs. The locations are provided in a separate table in Appendix A. The mass wasting is not confined to any particular rock unit. The examples cited in Appendix A show that slumping has involved formations whose lithologies include mudstone, limestone, calcareous siltstone and conglomerate (Plate I). A slump found on the northern flank of Dixon Mountain above the Dry Creek drainage west of Dell, Montana involves sandstone of the Quadrant Formation. The degree of revegetation on the slump surfaces, and the modification of the surface features, indicate that the mass wasting is of Recent origin and the occurrence is periodic. The Dixon Mountain slump is reported by local residents to have occurred in 1914.

STRATIGRAPHY

The stratigraphic section in the Little Sheep Creek area is one of the most complete in the region and is representative of the standard stratigraphic section of southwestern Montana. The strata range in age from the Mississippian to the Pliocene and include 16 formations, two tongues of the Park City Formation mapped with the Phospharia Formation, and two informal Tertiary units. The approximate total thickness of the strata is 2,875 m. The distribution of the formations and informal beds is presented in the geologic map in Plate I and is tabulated in Figure 4. Regional correlations are provided in Plates III and IV.

THE CARBONIFEROUS SYSTEM

Mississippian strata of southwestern Montana and central Idaho reflect facies changes from those of a stable cratonic sequence in Montana to a miogeosynclinal sequence in central Idaho (Scholten et al., 1955, p. 365; Huh, 1967). A narrow linear zone, the Wasatch Line (Kay, 1951), is a sedimentological and tectonic boundary between the Cordilleran miogeosyncline and the craton. The Tendoy Range and the eastern flank of the Beaverhead Range fall within the boundaries established for the transition zone (Huh, 1967, p. 32). The cratonic facies of the Mississippian System are represented by the Lodgepole and Mission Canyon Formations of the Madison Group and the Big Snowy Formation. The miogeosynclinal facies are represented by the McGowan Creek Formation and by discrete formations of the White Knob Group. They are, in ascending order, the Middle Canyon, Scott Peak, South Creek, and Surrett Canyon Formations (Huh, 1967, p. 341; Sandberg, 1975).

SYSTEM	ROCK UNIT		THICKNESS
Tertiary	Edie School Rhyolite		49m
	Round Timber Limestone		73m
Cretaceous	Beaverhead		803m
	Colorado Shale		218m
	Kootenai		275m
Jurassic	Morrison		49m
	Ellis Group	Rierdon	31m
		Sawtooth	40m
	Gypsum Spring Tongue		73m
Triassic	Thaynes		207m
	Woodside		40m
	Dinwoody		226m
Permian	Phosphoria		199m
Pennsylvanian	Quadrant		730m
Mississippian	Amsden		180m
	Big Snowy		110m
	Mission Canyon		upper 130m
	Scott Peak		Allochthonous

Figure 4. Generalized stratigraphic column of the Little Sheep Creek area.

The cratonic and miogeosynclinal facies represent end members of nearly continuous deposition in two sedimentary provinces which were connected across the transitional zone. Facies of certain formations can be extended into the transition zone and have lateral correlatives in the adjacent province, but formational names do not extend across the transition zone from one area to another. An exception to this rule is the Big Snowy Formation, the black shales of which can be traced westward from Montana, to the eastern flank of the Beaverhad Range (Huh, 1967, p. 46; Maughan and Roberts, 1967, p. B11). Equivalent carbonates are found in the upper White Knob Group near Challis, Idaho. The South Creek and Surrett Canyon Formations are restriced to the miogeosyncline and are absent in the transition zone. The Scott Peak and Middle Canyon Formations have facies which extend from the miogeosynclinal area eastward to the eastern margin of the transition zone. East of the transition zone the Mississippian strata belong to the Lodgepole and Mission Canyon Limestones and the Big Snowy Formation.

The Mississippian strata in the Little Sheep Creek area are both allochthonous and autochthonous. Most of the Mississippian strata are involved with the Medicine Lodge thrust. Post-Laramide normal faulting, with creation of the Medicine Lodge basins, has made it impossible to follow the thrust trace to its origins. Origin of the Medicine Lodge thrust is discussed in a subsequent section. Suffice it to say that the Mississippian strata in the Medicine Lodge thrust are allochthonous and are more representative of the miogeosyncline-transition zone facies than those of the cratonic zone. Close lithologic similarities between the Mississippian rocks of the thrust sheet in the Little Sheep Creek area and strata described by Huh (1967) on the east flanks of the

Beaverhead Range permit a tentative assignment of the thesis area strata to the Scott Peak Formation. This assignment is made for convenience, albeit with trepidation; the thrust plane was not selective of lithologies through which it cleaved so that a variety of facies occur randomly along the thrust contact and within the thrust sheet itself. The rocks also are very fractured and locally folded, making correlation difficult but not impossible. The convenience comes from being able to apply a name to the strata for purposes of discussion, rather than repeated references to an alternative name, the Undifferentiated Mississippian Strata.

The second source is autochthonous. The Mississippian strata in the northeast corner of the thesis area occur on the upthrown side of the Tendoy fault (Scholten et al., 1955, p. 384). The Tendoy fault is a high-angle reverse fault which forms a major structural feature that trends in a northwest-southeast direction along much of the eastern flank of the Tendoy Range. The trace of the fault occurs about 1,000 m north of the thesis area. Limitations imposed by myself on the area mapped reduced the areal extend of the autochthonous Mississippian strata. However, differentiation into formal units is possible. The Big Snowy Formation can be defined, as well as an interval of thin-bedded limestone of the Amsden Formation. The strata beneath the Big Snowy Formation are distinct but not lithologically dissimilar to the Scott Peak Formation. Following the nomenclature established by Huh (1967) the oldest autochthonous strata exposed in the Little Sheep Creek area are assigned to the Mission Canyon Limestone.

THE WHITE KNOB GROUP

The name White Knob was proposed by Ross (1962, p. 385) for all of the Carboniferous limestone in south-central Idaho. The name was adopted from exposures in the White Knob Mountains, Idaho. These strata previously had been part of the Brazer Formation, an umbrella name which had at one time included all of the Paleozoic rocks of central Idaho and those which could be extended into Montana (Ross, 1934, p. 977-985). Huh (1967, p. 34) raised the White Knob to group status and described four included formations. They are, in ascending order, the Middle Canyon, Scott Peak, South Canyon, and Surrett Canyon Formations.

The Scott Peak Formation

The Scott Peak Formation is named for exposures in East Canyon near Scott Peak, Beaverhead Range, Idaho (Huh, 1967, p. 39-42). At the type locality the formation is 2,250 feet (686 m) thick and is made up of three members: lower and upper cyclic members, and a middle massive member.

The lower member consists of a cyclically interbedded sequence of crinoid-bryozoan calcarenite beds and thin beds of dark, fine-grained, chert-bearing limestone 650 feet (198 m) thick. In the middle of the lower cyclic member is an interval of calcareous quartz siltstone interbedded with the calcarenites. The siltstone is lithologically similar to calcareous quartz siltstone found in the underlying Middle Canyon Formation. The massive middle member is 850 feet (259 m) thick and is composed of predominantly microcrystalline limestone containing many visible calcite crystals and no chert. Transitional contacts exist with the overlying and underlying members. The upper cyclic member is a 750 foot (229 m) thick sequence of intercalated chert-bearing, fine-grained limestone, mixed crystalline calcarenite, and crinoidal calcarenite.

Only the lower cyclic member and parts of the middle massive member exist on the east flank of the Beaverhead Range and extend into the transitional zone. Huh (1967, p. 41) states that all miogeosynclinal strata above the lower cyclic member of the Scott Peak Formation are thinned or missing from the zone of transition to the craton.

Lithology and Distribution

Three lithologies are recognized in the limestones of the Medicine Lodge thrust which permit their assignment to the Scott Peak Formation in the thesis area. They are a very fine-grained sandstone, a thinly bedded limestone, and a massive limestone, arranged in presumed ascending order.

The vary fine-grained sandstone occurs at the base of the Medicine Lodge thrust near Seybold Spring on the north side of the canyon in Section 5, T. 15 S., R. 9 W. It also occurs in klippen overlying the Thaynes Formation in Section 18, T. 15 S., R. 9 W. The sandstone is a quartz arenite; the coloration varies from being pale yellowish brown (10 YR 8/6), very pale orange (10 YR 8/2), or white (N 7) on fresh surfaces, to dark yellowish orange (10 YR 6/6), pale yellowish brown (10 YR 8/6), and very light gray (N 8) on weathered surfaces. The rock is generally thin-bedded, well indurated and weathers to slabby talus. Outcrops near Seybold Spring are not as well indurated and weather to form a yellowish sandy soil. The rocks are crisscrossed with minute calcite veins which obscure any sedimentary structures that may exist in the sandstone. Samples of the sandstone are variably calcareous to siliceous.

Microscopically the siliceous sandstone consists of poorly sorted, silt-sized, very fine-grained, angular particles of quartz in grain-to-grain contact. There is an equal ratio of strained and unstrained quartz. Cementation is by silica overgrowth. Isolated occurrences of calcite and hematite cement also are present. The quartz arenites are very similar to sandstones of the Middle Canyon Formation described by Klecker (1980, p. 16) in Little Water Canyon.

The intimate relationship of the very fine-grained sandstone in the thesis area with the overlying thin-bedded limestone indicates that the sandstone may be part of the lower cyclic member of the Scott Peak Formation.

The thin-bedded limestone occurs at the base of the thrust sheet in the NE¼, Section 31, T. 14 S., R. 9 W. The beds are 10 to 25 cm thick and contain nodular and bedded cherts which are as much as 8 cm thick. The limestone is a finely crystalline, fossiliferous sparite that is medium light gray (N 6) on fresh surfaces, but light gray (N 7) where weathered.

The structureless limestone is the predominant lithologic type found in the thrust sheet. The limestone is generally very thick-bedded with beds ranging in thickness from about 2 m to at least 7 m. One outcrop, whose sheer size and height prevented detailed examination, was estimated to be 18 m thick; there are no obvious bedding planes. The outcrops are predominantly medium light gray (N 6), medium gray (N 5), or medium dark gray (N 4) on fresh surfaces. Minor variation in the coloration occurs in which the limestone is brownish gray (5 YR 4/1) and olive gray (5 Y 4/1) on fresh surfaces, and moderate olive gray (5 Y 5/1) on weathered surfaces.

Microscopic examination of thin sections of selected samples, or examination of cut and polished surfaces under a dissecting microscope, reveal that the very thick-bedded limestone consists of three basic lithologies. The first is a micrite with some echinoid plates; the second is a wackestone which alternately contains crinoid columnals, echinoid plates, molluscan fragments, or combinations of the three; the third is a crinoidal grainstone. The predominance of relative distribution of one lithotype over another was not determined.

The strata in the thrust sheet have been deformed to varying degrees. The thin-bedded limestone in Section 31, T. 14 S., R. 9 W., shows little deformation; however, disharmonic folds can be observed in a similar limestone on the south-facing canyon wall in the gully above Seybold Spring. An apparent extreme of deformation is the the sheared limestone found near the confluence of two drainages in the SW¼, SE¼, NW¼, Section 30, T. 15 S., R. 9 W. In the easternmost drainage are exposed approximately 45 m of thick-bedded limestone. Several of the limestone beds appear to be laminated and interbedded with the thick-bedded units. These laminated units are one to six meters thick and are very fissile, breaking into platy shards that have shiny and polished surfaces similar to phyllitic texture in metamorphic rocks. The fissility is not primary but is related to tectonic shearing of the limestone during thrusting. Examination of a thin section of the fissile units reveals flow lines of micrite outlined by streaks of black organic material enclosing calcite crystals (Figure 5).

Fossils

Crinoid columnals, echinoid spines, ramose and fenestral bryozoans, molluscs, ostracods, and horn corals are found in the Scott Peak Forma-

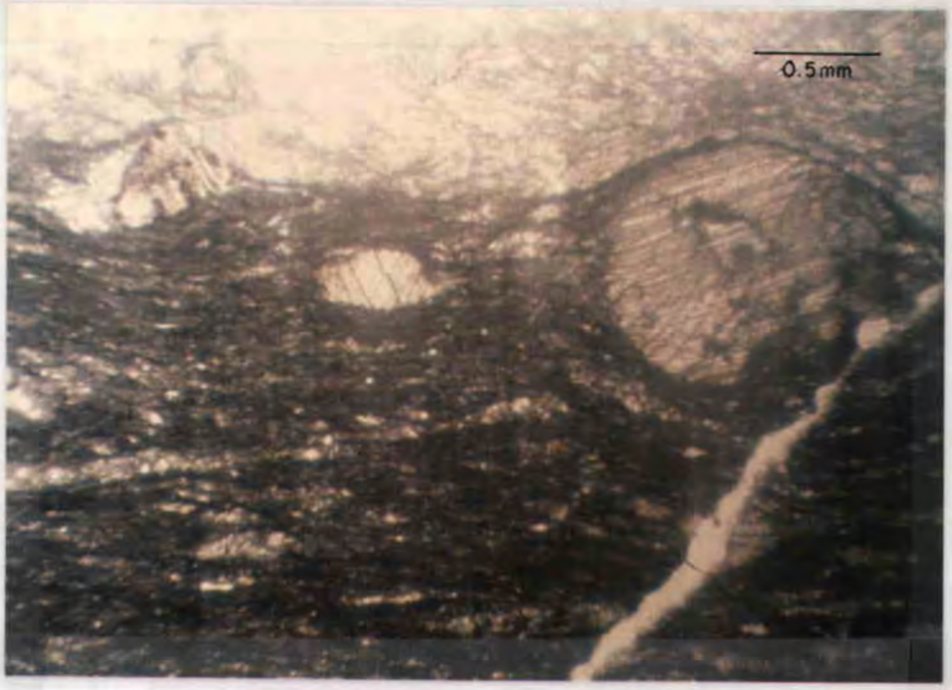


Figure 5. Photomicrograph, sheared limestone of the Scott Peak Formation from the Medicine Lodge thrust. Note flow lines of micrite (M) around calcite crystals (C). Circular crinoid columnal right center; calcite vein at filling top. Sample Kmb 79-269, SW¼, Section 30, T. 15 S., R. 9 W. Plane light.

tion. Generally the fossils are too fragmentary to warrant identification. One sample of thin-bedded limestone contains corals identified as Rotiphyllum by W. J. Sando of the United States Geological Survey, Washington, D. C.

THE MADISON GROUP

Peale (1893) applied the name Madison to a sequence of Lower Carboniferous limestones exposed near Three Forks, Montana. Collier and Carthcart (1922, p. 173) were the first to refer to the Madison limestones as a group by dividing the sequence of limestone into two formations, the Lodgepole and Mission Canyon Limestones. The names were derived from canyons in the northern part of the Little Rocky Mountains, north of Billings, Montana. Sloss and Hamblin (1942, p. 313-315) proposed that the nomenclature used by Collier and Carthcart be applied to the Madison Group throughout Montana. They also formally recognized two subdivisions of the Lodgepole Limestone, the Paine and Woodhurst members, units which had originally been described by Weed in 1899. Sloss and Hamblin (1942) used a generalized type section located near Logan, Montana for the Madison Group. They designated the type section for the Lodgepole Limestone to be in the Little Rocky Mountains, but locations were not precisely stated. No type section was described for the Mission Canyon. Sandberg and Klapper (1967) described a third, basal member, the Cottonwood Canyon member, of the Lodgepole Limestone in Wyoming. This member was acknowledged but not described by Sando and Dutro (1974) in their type section descriptions of the Madison Group limestones in Montana.

Sando and Dutro (1974) provided field descriptions, paleontological documentation, and precise locations of type sections of the Madison Group in Montana. The type section for the Madison Group is on the the north bank of the Gallatin River across from Logan, Montana. The type section for the Mission Canyon Limestone is immediately north of Monarch in the Little Belt Mountains. The section for the Paine and Woodhurst members of the Lodgepole Limestone is on Belt Creek in the Little Belt Mountains.

Madison Group limestones are distributed over most of Montana, parts of the Dakotas, Idaho, Wyoming, and Utah. Equivalent age limestones have been described in California, Nevada, Arizona, Utah, Wyoming and Colorado (Sando and Dutro, 1974), and Canada and Alaska (Oles, 1980, personal communication).

The Mission Canyon Limestone

The Mission Canyon Limestone is 1,050 feet (320 m) thick at its type section and 860 feet (262 m) thick at Logan, Montana. The lower half consists of thin- to very thick-bedded medium- to coarse-crystalline limestone with lenticular chert, interbedded with cherty dolomitic limestone and fine-crystalline limestone. The upper half is characterized by thin- to very thick-bedded fine-grained limestone, or dolomitic limestone, with lenticular and nodular chert. The upper part of the Mission Canyon is marked by fine-grained limestone with interbeds of quartz siltstone (Sando and Dutro, 1974).

The best exposures of the Madison Group in the general vicinity of the Little Sheep Creek area are found in Ashbough Canyon in the Ruby Range, approximately 60 km to the north. Huh (1968, p. 24) described 850 feet (259 m) of Mission Canyon Limestone in Ashbough Canyon and recognized three important lithologic subdivisions: a lowermost crinoid-bryozoan calcarenite, a mixed crystalline limestone unit, and "an uppermost evaporite (?) type carbonate" unit.

Distribution and Lithology

The Mission Canyon Limestone is found in the northeastern corner of the thesis area, and as an exotic block in the Cretaceous Beaverhead Formation in the NE¼, Section 30, T. 15 S., R. 9 W. The exact thicknesses cannot be determined because the limestone is on the upthrown side of the Tendoy fault and is in fault contact with the Beaverhead Formation. The upper 130 m of the limestone were mapped where the Formation crops out in the NE¼, Section 34, T. 14 S., R. 9 W. The limestone forms a prominent resistant ridge, the dip slope of which extends south into the thesis area.

The contact between the Mission Canyon Limestone and the overlying Big Snowy Formation is covered in the thesis area. The contact is drawn along a narrow drainage stratigraphically and topographically beneath a ridge of limestone which contains Caninia, a Chesterian coral compatible with Big Snow assignment. Elsewhere in southwestern Montana the contact with the Big Snowy is sharp and disconformable (Scholten et al., 1955, p. 365; Huh, 1968, p. 27). It is generally marked by a break in slope between the resistant Mission Canyon limestone and the shales of the Big Snowy Formation. In the Tendoy Range the top of the Mission Canyon, where exposed, is marked by solution cavities filled with material derived from the overlying Big Snowy Formation.

The Mission Canyon Limestone in the Little Sheep Creek area is mostly covered but an adequate composite description can be compiled from the scattered outcrops. The lower 100 m of the formation are made up of a thick-bedded limestone which is brownish gray (5 YR 4/1) on fresh surfaces, but weathers pinkish gray (5 YR 8/1) to light gray (N 7). The limestone is coarsely crystalline and has a petroleum odor on freshly broken surfaces. The weathered dip slopes of the outcrop have crinoid columnals, bryozoa, and brachiopod fragments standing out in high relief.

The upper 30 m consist of laminated and thinly laminated, fossiliferous, micritic limestone with a petroleum odor interbedded with minor amounts of a thinly bedded coquinoid sparite. The laminated and thinly laminated micritic limestone is yellowish gray (5 Y 8/1) in outcrop, but on fresh surfaces the color is olive gray (5 Y 4/1). Weathered surfaces of the platy limestone are often congested with an assemblage of ramose and fenestral bryozoa, crinoid columnals, and productid brachiopods which appear to be restricted to the bedding planes. The coquinoid limestone forms a resistant unit approximately one meter thick. It is a light brownish gray (5 YR 6/1) on fresh surfaces and weathers a light gray (N 7) to yellowish gray (5 Y 8/1) color. Examination under a dissecting microscope reveals a well-sorted grainstone with brachiopod shell fragments, ramose bryozoa, crinoid columnals and echinoid spines aligned paralled to bedding.

The exotic block of limestone imbedded in the Beaverhead Formation in the NW¼, Section 30, T. 15 S., R. 9 W., is greatly fractured and is made up of finely crystalline sparry limestone interbedded with thin beds of calcareous siltstone and silty limestone. The silty limestone contains ramose and fenestral bryozoa, crinoid columnals, and brachiopods. One genus of brachiopod was identified as <u>Pugnoides</u> cf. <u>P.</u>

ottumwa (White) by M. Gordon of the United States Geological Survey, Washington, D. C. The range of Pugnoides helps assign the exotic block to Mississippian strata. Similar exotic blocks of limestone occur in the Beaverhead Formation on both sides of Little Sheep Creek 3 km north of the thesis area. Similar occurrences of exotic blocks of limestone in the Beaverhead were mapped by Scholten et al. (1955) elsewhere in the Tendoy Range. The exotic Mississippian limestone may represent large eroded slide blocks derived from the flanks of the uplifted Blacktail-Snowcrest anticlinorium.

The Big Snowy Formation

Scott (1935, p. 1023-1032) defined the Big Snowy Group from extensive exposures in the Big Snowy Mountains of central Montana. He assigned the Kibbey, Otter, and Heath Formations to the group. Weed (1892, 1899) was the first to describe the Kibbey sandstone and Otter shale, but prior to 1935 all of the Late Mississippian and Pennsylvanian strata of central Montana and northwestern Wyoming had been included in the Quadrant Formation of Peale (1893). In 1935 Scott raised the Kibbey and Otter to formational status and added the Heath Formation to complete a triumvirate for the Big Snowy Group. At that time he proposed that the term Quadrant not be applied to Mississippian strata of central Montana and that the name be restricted to Pennsylvanian rocks as defined at the Quadrant type section. Seager (1942, p. 863) added the Charles Formation to the base of the Big Snowy Group. Sloss (1952, p. 66-67) reassigned the Charles Formation to the Madison Group and later Sando and Dutro (1974, p. 1) suggested that the Charles be restricted to a formation of the Madison Group as it occurs in the subsurface in the Williston basin.

In 1959 Gardner redefined the Big Snowy Group to include, in ascending order, the Kibbey, Otter, Heath, Cameron Creek, Alaska Bench, and Devils Pocket Formations. Maughan and Roberts later (1967, p. B5) redefined the Big Snowy Group to include only the Kibbey, Otter, and Heath Formations. The Heath Formation was restricted to shales and interbedded limestone beneath a regional Late Mississippian to Early Pennsylvanian unconformity. As defined, the three formations comprise one sedimentary cycle and closely approximate a time stratigraphic unit which spans Chesterian time. The rocks above the unconformity were divided into three formations and assigned to the Amsden Group. These formations are, in ascending order, the Tyler Formation, the Alaska Bench Limestone, and the Devils Pocket Formation. These formations have counterparts in the type Amsden Formation in northern Wyoming (Maughan and Roberts, 1967, p. B11). The basal part of the Tyler Formation, the Stonehouse Canyon Member, yielded spores of an Early Pennsylvanian age, further substantiating the division of the strata into two groups as proposed by Maughan and Roberts (1967, p. B21).

Regional Distribution

The Big Snowy Group occupies parts of the Williston Basin and extends eastward from Montana into North Dakota and northwestern South Dakota (Easton, 1962). The extension of the Big Snowy Group into southwestern Montana by various authors is summarized by Maughan and Roberts (1967, p. B10). Changes in facies and thicknesses have not permitted direct correlation with the formations of the type area. Where the rocks cannot be mapped separately, the Big Snowy is of formational rank.

Scholten et al. (1955, p. 363) mapped an interval of reddish shaly and gypsiferous beds and interbedded black shales and dark limestones between the Mission Canyon and the Amsden Formations in the Lima region. They thought the strata may represent the Kibbey and Heath Formations, respectively, and maintained the group ranking of the Big Snowy for that interval.

Hildreth (1980) mapped 138 m of thin-bedded limestone of the Heath equivalent of the Big Snowy immediately north of the Clark Canyon Dam. Kiecker (1980, p. 19-25) mapped about 150 m of Big Snowy strata in Big Sheep Creek. Both authors chose to regard the Big Snowy as a formation rather than as a group. Moran (1971) and Murray (1973) did not map the Big Snowy in the Centennial Range; instead they found the Amsden Formation resting unconformably on the Mission Canyon Limestone. Because of the limited exposure in the Little Sheep Creek area the Big Snowy interval will be dealt with as a formation.

Distribution

The Big Snowy Formation is approximately 110 m thick and can only be located in the NE¼, Section 34, T. 14 S., R. 9 W. The exposures are very limited and restricted to a resistant limestone ridge in the lower half of the formation and to a gully exposing the mudstone of the upper part. The contact with the underlying Mission Canyon is a regional erosional unconformity (Maughan and Roberts, 1967; Scholten et al., 1955). In the thesis area it is covered but is taken to be in the drainage immediately down section from the resistant limestone. The contact with the overlying Amsden Formation is gradational (Scholten et al., 1955, p. 365) but covered in the thesis area.

Litholcgy

The Big Snowy Formation consists of a lower limestone unit and an upper black mudstone unit. The limestone, making up approximately the lower half of the formation, can be divided into two parts. The lower part consists of thin-bedded non-fossiliferous micrite with discontinuous, very thin beds of laminated silty limestone. The micrite has a petroleum odor and is olive brown (5 Y 5/6 - 4/4) on fresh surfaces, yellowish gray (5 Y 8/1) upon weathering. Above the thin-bedded micrite are very thick beds of a finely crystalline fossiliferous sparite that contain thin stringers of bedded chert. It also has a petroleum odor and its coloration is a medium dark gray (N 4) on fresh surfaces and light gray (N 7) on weathered surfaces. The top part of the lower limestone unit consists of a thin-bedded coquinoid limestone interbedded with a fossiliferous laminated silty limestone that weathers to flaggy and platy talus.

A gully running parallel to strike provides the only exposure for the approximately upper 45 m of the Big Snowy. Trenching in the gully bank unearthed a dark gray clayey mudstone in which bedding has been disrupted by the growth of authigenic gypsum crystals.

Correlation of the Mississippian Strata

Up to about 1960 the stratigraphy of the Mississippian strata in the northern Cordillera had been based on lithologic correlations. The first important contribution to the biostratigraphic correlation of the Mississippian strata across the Wasatch line was made by Sando and Dutro in 1960. They recognized five provisional zones (A, B, C_1, C_2, & D) based on certain generic associations of corals in the Madison

Group and Brazer Dolomite in northwestern Utah, western Wyoming, and southwestern Montana. At that time the Madison coral zones could not be recognized in the type Mississippian in the midcontinental region. Provisional correlation of the coral assemblages was made using associated brachiopods. The same authors (Dutro and Sando, 1963a) defined three more coral zones (E, F, & K) and four brachiopod zones in post-Lodgepole Limestone strata in southeastern Idaho. The strata had been incorporated in the loosely defined Brazer Limestone; however, Dutro and Sando (1963a, p. 1966) renamed the sequence of limestones and sandstones the Chesterfield Range Group. The coral zones could be correlated with other Late Mississippian strata in Utah, southwestern Montana, south-central Idaho, western Wyoming, and western Canada with the aid of associated brachiopods. The zones, designated A, B, C_1, C_2, D, E, F, and K, span most of the Mississippian Period from late Kinderhookian to Chesterian time. Dutro and Sando (1963b) recognized one late Mississippian coral zone and two brachiopod zones in the Big Snowy-Amsden interval in southwestern Montana. Sando (1967) provided a composite zonation for the entire Mississippian which was expanded to include twelve megafaunal zones based on corals and brachiopods. He recognized two major cycles of sedimentation: a Madison cycle and a post-Madison cycle, both of which were divided into six zones. Sando, Mamet, and Dutro (1969) incorporated the twelve megafaunal zones of Sando (1967) in a comparison with fifteen foraminiferal zones established for the northern Cordillera by Mamet and Skipp (*in* Sando, Mamet, and Dutro, 1969, p. E13). The comparison resulted in "consistent chronostratigraphic correlations of the Mississippian of the northern Cordilleran region both with the Cordilleran basin and with the type

Mississippian sequence," (Sando, Mamet, and Dutro, 1969, p. E1). The cosmopolitanism of the foraminifera provided a useful correlation with standard Carboniferous sequences in western Europe. By 1975 the fifteen foraminiferal zones had been expanded to twenty-one Carboniferous zones which included the late Chesterian, Morrowan, and Atokan ages. Gutschick, Sandberg, and Sando (1980) describe the sequence of events which had developed on the shelf margin and carbonate platform that had extended from Montana to Nevada during Mississippian time; the paleogeography is based on stratigraphic evidence and paleontologic correlations of 32 sections using Mamet's foraminifera zones, Sandberg's conodont zones, and coral zones modified after Sando, Mamet, and Dutro (1969).

Plate III presents a summary of Carboniferous stratigraphy as it occurs in central and southwestern Montana, and central and southeastern Idaho.

The Amsden Formation

The Amsden Formation was first defined by Darton (1904, p. 396) to include a sequence of red shales, limestones, and sandstones which lie between the Madison Limestone and the overlying Tensleep Sandstone in north and central Wyoming. The name is derived from the Amsden branch of the Tongue River on the eastern flank of the Bighorn Mountains of Wyoming. Since Darton's time the Amsden has been recognized in central and northwestern Wyoming, central and southwestern Montana, and in the subsurface in eastern Montana and part of the Dakotas. The Amsden Formation in Wyoming is now recognized to be time transgressive (Sando et al., 1975, p. A49) and straddles the Mississippian-Pennsylvanian boundary.

The fact that the Amsden Formation is time transgressive has been the source for much debate in the literature since 1906. Sando et al. (1975, p. A3-A17) provide a synthesis of the literature concerning the biostratigraphic and lithologic boundaries of the Amsden Formation in Wyoming since 1906, a process which will not be repeated here.

The Amsden Formation in Wyoming has been divided variably into two member, or three member schemes. Sando et al. (1975) divide the Amsden into three mappable members which are, in ascending order, the Darwin Sandstone Member (Blackwelder, 1918, p. 422), the Horseshoe Shale Member, and the Ranchester Limestone Member (Mallory, 1967, p. G14).

The Amsden was extended into central Montana by Scott (1935, p. 1029) who applied the name to red shales and limestones conformably overlying his newly defined Heath Formation of the Big Snowy Group. Beekly (1955) divided the Heath into two parts based on an unconformity within the formation. He retained the name Heath for the marine black shales and interbedded limestone beneath the unconformity. Beekly assigned black shales and interbedded sandstone above the unconformity to the Amsden Formation.

Mundt (1956, p. 1920-1925) recognized the Heath division proposed by Beekly but chose to subdivide the strata above the unconformity into three formations. They are, beginning with the oldest, the Tyler, the Alaska Bench, and the Amsden Formations. The Amsden was restricted to a sequence of cherty dolomites lying unconformably on the Alaska Bench Formation. The newly named Amsden strata could be correlated with the upper part of the type Amsden in Wyoming. He felt that the lower strata of the type Amsden pinch out northward and were not laterally correlative to similar units in Montana.

Gardner (1959, p. 335-337) did not recognize the unconformity above the Heath Formation and believed the entire Montana sequence to represent a period of continuous deposition. He discarded the name Amsden and assigned all of the units to the Big Snowy Group. The formations are, in ascending order, Scott's (1935) Kibbey, Otter, and Heath Formations, a newly named Cameron Creek Formation, the Alaska Bench Formation (as defined by Mundt, 1956), and a newly defined Devils Pocket Formation. Maughan and Roberts (1967, p. B5) once again recognized the widespread unconformity at the top of the Heath Formation. They also recognized the natural grouping of the rocks above and below the unconformity and chose to divide the strata formally into two groups: the Big Snowy below, and the Amsden above. The Amsden was divided into the Tyler, Alaska Bench, and Devils Pocket Formations.

Maughan and Roberts (1967, p. B19) recommended that strata approximately equal to the Amsden Group in western Montana and adjacent parts of Wyoming and Idaho be called the Amsden Formation.

The Amsden Formation, or Group, lies unconformably on the Mission Canyon Limestone or the Big Snowy in Wyoming and central Montana. The unconformity extends into parts of southwestern Montana. The area west of a line extending approximately northeast from Monida, Montana appears to have been an area of continuous deposition during late Chesterian to Atokan time (Sando et al., 1975, Figures 15-22).

The Big Snowy was deposited on a karst surface which had developed on the Mission Canyon Limestone during middle Meramecian time. Deposition was continuous in the area of the Tendoy Range and Armstead anticline from Big Snowy through Phosphoria time. Hildreth (1980) describes a transitional contact from pure limestone in the Big Snowy Formation to calcareous sands in the Amsden. Scholten et al. (1955, p. 366) considered the contact to be conformable in the Lima region. Local outcrops of a limestone conglomerate (Klecker, 1980, p. 27-28; Scholten et al., 1955) at the base of the Amsden mark the position of local erosional unconformities. In the Little Sheep Creek area the contact with the Big Snowy Formation is covered but is considered to be conformable.

The Amsden Formation is conformably overlain by the Quadrant Sandstone or its equivalent, the Tensleep Formation, wherever those relationships exist; in parts of central Montana the Amsden Group is unconformably overlain by the Ellis Group. The conformable contact with the Quadrant Sandstone can be observed in the Armstead anticline (Hildreth, 1980), Little Water Canyon (Klecker, 1980, p. 27), and in Little Sheep Creek.

Distribution and Topographic Expression

The Amsden Formation is located in the northeast corner of the thesis area in the NE¼, Section 34, T. 14 S., R. 9 W. It is approximately 180 m thick. The lower two-thirds of the formation are completely covered and form a very gentle grass-covered slope. The upper third is made up of more resistant limestone which is exposed along the northern side of the valley of the West Fork of Little Sheep Crreek immediately north of the map margin. A small outcrop occurs in the south side of a spur in the SE¼, NE¼, Section 34.

The contact with the underlying Big Snowy Formation is considered to be conformable as recognized regionally by Scholten et al. (1955), Sando et al. (1975), and Hildreth (1980). In Little Sheep Creek the upper part of the Big Snowy and the lower part of the Amsden are covered and the nature of the contact could not be observed. Characteristic lithologies of the basal Amsden in southwestern Montana would, by nature, be more resistant than the gypsiferous mudstone of the upper Big Snowy. The lower contact of the Amsden is assumed to coincide with a slight break in slope upsection from the only exposure of the upper Big Snowy mudstone.

The contact with the Quadrant Sandstone consists of an intertonguing relationship between thin-bedded limestone of the Amsden and thin-bedded quartz sandstone typical of the Quadrant. The contact is drawn at the last occurrence upsection of the thin-bedded limestone.

Lithology

Sloss and Moritz (1951, p. 2159) characterized the formation in southwestern Montana as consisting of a basal unit of poorly exposed, commonly red to purplish red shale, siltstone and shaly dolomite. The lower beds grade up into a sequence of fossiliferous limestones which are initially thin-bedded but become very thick-bedded towards the top of the formation. Scholten et al. (1955, Table 1) described 90 to 200 feet (27-61 m) of the Amsden Formation in the Lima region. The formation generally consists of thin-bedded limestone, calcareous, partly sandy shales, and a basal sandstone. Hildreth (1980) mapped as the Amsden Formation a 9 m thick interval of thin-bedded calcareous fine-grained sandstone between the Big Snowy limestone below and the Quadrant quartzite above. Both contacts are gradational and conformable. Klecker (1908, p. 31) described about 200 feet (61 m) of the Amsden in the Little Water Canyon area. There the formation consists of a lower limestone and red shale unit, a middle sandstone unit, and an upper limestone unit.

In the thesis area, only the upper 30 m of the Amsden Formation were exposed well enough for a general description. The upper part of the Amsden consists of thin-bedded, dense, finely crystalline spar that is generally unfossiliferous. Two fossiliferous horizons yielded a fossil hash of crinoid columnals, brachiopod fragments, and echinoid spines. The silt content of the limestones increases upsection as the Amsden grades into the Quadrant. The purer limestone beds are medium gray (N 5) to medium dark gray (N 4) on fresh surfaces and weather to light gray (N 7) and medium light gray (N 6). The silty limestone beds are light olive gray (5 Y 6/1) on fresh surfaces, yellowish gray (5 Y 8/1) and light gray (N 7) on weathered surfaces. The intervening quartz sandstone is calcareous, very fine-grained, moderate yellowish brown (10 YR 6/2 to 10 YR 5/4) on weathered surfaces, and very pale orange (10 YR 6/2) where fresh.

The Mississippian-Pennsylvanian Boundary

The controversy concerning the nomenclature of the Late Carboniferous strata in Wyoming and central Montana has evolved around the precise dating of the rocks and the position of the Mississippian-Pennsylvanian boundary. It could not be determined whether the Amsden was entirely Mississippian or entirely Pennsylvanian, or whether both periods were represented. Maughan and Roberts (1967) addressed the problem in their definitive study of the Amsden and Big Snowy Groups in central Montana. In 1975 Sando, Gordon, and Dutro were able to establish the temporal relationship of the Amsden fauna and helped to clarify the Amsden problem in Wyoming.

The biostratigraphy of the Amsden Formation has been summarized by Maughan and Roberts (1967) and Sando et al. (1975). What became clear to Sando et al. (p. A2) was that the Amsden strata are time transgressive, a concept which had eluded most stratigraphers until then. Mundt (1956, p. 1929) believed that the lower strata of the Amsden in Wyoming pinched out northward and were not continuous with the Montana strata. Maughan and Roberts (1967, p. B19, Plate 1) showed that the Big Snowy strata pinch out southward of exposures in the Big Snowy Mountains. The Big Snowy strata and the lower Amsden of Wyoming were thought to be time equivalents by Gardner (1959, p. 345). Sando et al. (1975, p. A49) concluded that simultaneous sedimentation occurred during Chesterian time (foraminifera zones 16i to 18, Plate III) in two shallow marine environments that were separated by an emergent area called the South Montana arch.

During latest Chesterian time (foraminiferal zone 19, Plate III) the Big Snowy basin became emergent, the South Montana arch ceased to exist, and the Wyoming sea transgressed northward into south-central Montana. The erosional unconformity at the top of the Heath Formation was created during this interval. Expansion of the Wyoming sea northward continued into the Early Pennsylvanian (Morrowan) initiating a new cycle of marine deposition on the Big Snowy Group. The cycle of marine deposition in the greatly expanded Wyoming sea continued into the early Atokan (foraminiferal zone 21, Plate III). Local sedimentation was interrupted in central Montana by an emergent island, the Montana uplift, which created the unconformity between the Alaska Bench and Devils Pocket Formations. A much more complete description of the stratigraphy and geologic history of Wyoming and immediate areas is given in Sando et al. (1975, p. A54-A66).

The Amsden Formation of Wyoming is time transgressive, sedimentation having continued uninterrupted through Chesterian time and into the Morrowan. Breaching of the emergent barrier between the Big Snowy and Amsden seas caused a cessation of sedimentation of the Big Snowy strata and continued spreading of Amsden sedimentation across most of Montana and the Dakotas. Brief subaerial exposure during the latest Chesterian created a hiatus between the Big Snowy and Amsden Groups in central Montana. Subsequent sedimentation of Amsden strata in central Montana occurred only during the Early Pennsylvanian; the Big Snowy-Amsden boundary coincides with the Mississippian-Pennsylvanian boundary. Hildreth (1980) discovered *Caninia* and *Cleiothyridina* in the lower Amsden in the Armstead anticline. These fossils indicate that Amsden sedimentation commenced during late Chesterian time. The Amsden Formation in southwest Montana is, therefore, time transgressive.

The Quadrant Formation

The first reference to the Quadrant Formation was made by Peale (1893, p. 32-43) who applied the name to rocks in the Three Forks area that occupied the same stratigraphic position as quartz sandstones on Quadrant Mountain in the northwest part of Yellowstone National Park. The strata so named were those lying between the Madison Group limestones below, and the Ellis Group above, including rocks of the Amsden Formation. Iddings and Weed had designated the type section to be on the southeast corner of Quadrant Mountain (*in* Scott, 1935, p. 1014). They had apparently used the name Quadrant during field studies in the

Gallatin Range but did not publish their work until 1899. By their definition, the Quadrant Formation included 401 feet (122 m) of sandstone with local interbeds of dense limestone. Scott (1935, p. 1017) reexamined the Quadrant at the type section and assigned the lower 109 feet (33 m) of red and purple shale, shaly limestone, limestone and minor sandstone to the Amsden Formation. He believed the Quadrant was of marine origin and was the westward correlative of the Tensleep Formation of Wyoming. Scott assigned the Asmden to the Chesterian interval and considered the Quadrant as "basal Pennsylvanian in age" (Scott, 1935, p. 1031). He did not recognize the Quadrant in central Montana but instead referred those rocks to the Big Snowy Group and the Amsden Formation.

Thompson and Scott (1941, p. 349-353) lowered the position of the base of the Quadrant Formation at the type section to include strata previously assigned to the Amsden which contained Pennsylvanian fusulinids. The beds below were recognized as Mississippian. They abandoned the Amsden name and tentatively called the Mississippian strata between the Madison Group limestone and the Qudrant, the Sacajawea Formation.

Sloss and Moritz (1951, p. 2164-2165) measured 2,662 feet (812 m) of Quadrant sandstone in Big Sheep Creek 17 km north of the thesis area and 1,700 feet (519 m) in the western part of the Snowcrest Range. At the type section the Quadrant is 290 feet (89 m) thick (Scott, 1935, p. 1017). They concluded that the dramatic decrease in thickness of the Quadrant in a west to east direction was caused by post-Pennsylvanian erosion. Maughan (1975, p. 288) recognized a westerly source for the Quadrant sandstone and postulated that the abrupt east-west change in thickness was caused by rapid subsidence of an embayment of the Cordilleran geosyncline in western Montana and onlap of the sands and finer-grained sediments onto a more stable and shallow shelf. During Late Pennsylvanian and Early Permian time the shelf area in central and eastern Montana was uplifted and most Late Pennsylvanian strata were removed prior to deposition of Permian sediments. Pre-Jurassic uplift and erosion removed all Pennsylvanian strata that once had covered northern Montana, and large parts of North and South Dakota (Maughan, 1975, p. 291).

Regional Distribution and Correlation

The Quadrant Formation is distributed across southwestern Montana, parts of western, central and southcentral Montana, and northwestern Wyoming. It is the lateral equivalent of the Tensleep Sandstone of Wyoming and the Wells Formation of southeast Idaho (Scott, 1935, p. 1019; Maughan and Roberts, 1975, Figure 13; Plate III). The Quadrant Formation intertongues with the Devils Pocket Formation between western and central Montana. In eastern Montana the two formations are mostly interbedded dolomite and mudstone and are indistinguishable from the lower part of the middle member of the Minnelusa Formation in North and South Dakota.

Throughout its areal extent, the Quadrant Formation conformably overlies the Amsden Formation or, in central Montana, the Devils Pocket Formation of the Amsden Group. Parts of the Quadrant were removed by erosion from central Montana prior to deposition of Permian strata (Maughan, 1975, p. 291). The contact with the overlying formations is, therefore, disconformable throughout most of Montana and northwestern Wyoming. Southwestern Montana was not affected by the Late Pennsylvanian-Permian uplift and continued deposition of sediments occured into the Permian (Maughan, 1975, p. 287). The conformable contacts were described by Scholten et al. (1955, p. 366), and Klecker (1980, p. 35), and recognized by me at Little Sheep Creek.

Distribution and Topographic Expression

The Quadrant Formation is distributed in an S-like pattern in the northeastern and eastern parts of the thesis area (Figure 6). There it is exposed in the cores and on the limbs of two folds gently plunging to the southwest (Plate I). It is also found on the overturned limb of a syncline in the southeastern corner of the thesis area.

The Quadrant Formation is of a homogeneous composition, is very thick, and resistant. It therefore underlies some of the higher elevations in the Little Sheep Creek area. The highest point in the Tendoy Range, Garfield Mountain (3,343 m), which is located three kilometers east of the thesis area, is underlain by Quadrant sandstone.

Thickness and Lithology

The Quadrant Formation is generally covered by blocky and slabby talus that are weathering products of the thin to thick, crossbedded, well-indurated sandstone. For this reason no attempt was made to described the formation in detail greater than necessary to be able to recognize it in the field. A few good, in-place exposures, however, were located to provide a general description.

The Quadrant Formation is composed of approximately 730 m of sandstone that is generally thick- to very thick-bedded, planar crossbedded with a texture that is variably very fine-grained, and medium-grained,

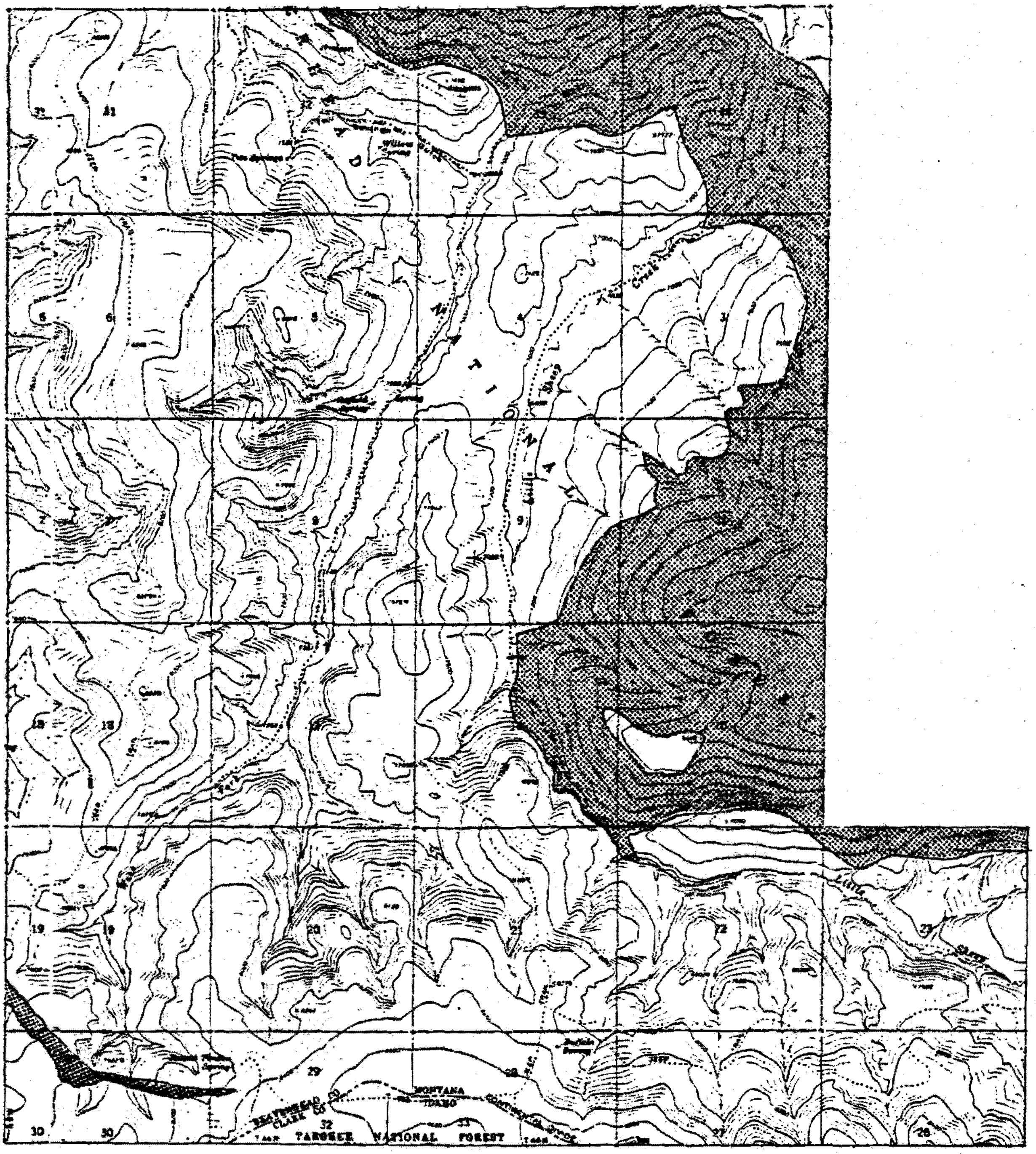


Figure 6, Distribution map of the Quadrant Formation, orange (10 YR 8/6) on fresh surfaces and weather light to medium gray (N 7 to N 5), pale to moderate yellowish brown (10 YR 6/2 to 10 YR 5/4) or moderate orange pink (10 R 7/4). The sandstone is a very well-indurated quartz arenite with subrounded to subangular quartz grains and minor chert. The cement is variably calcareous or siliceous.

Large-scale planar crossbedding is evident in all of the exposures; however, only one exposure was suitable for measuring the sedimentary structures. Crossbed sets range from 12° to 25° with an average of 19° from the horizontal. Paleocurrent direction was from the northeast. At this locality, the beds show a penecontemporaneous slump feature (Figure 7).

Fossils and Age

No fossils were found in the Quadrant Formation in Little Sheep Creek; the formation is generally unfossiliferous. Thompson and Scott (1941) described one species each of Wedekindella and Fusulinia at the type section about 14 feet (4 m) below the Phosphoria Formation. The fusulinids indicate a middle Desmoinesian age for the upper part of the Quadrant in northwestern Wyoming. Henbest (1956, p. 59, 62) assigned the Quadrant Formation a Desmoinesian age and correlated it with the upper part of the Tensleep Formation of northern Wyoming; The Tensleep also yielded Wedekindella fusulinids and is considered to be late Atokan to late Desmoinesian in age.

Over most of the regional extent of the Quadrant Formation, Late Pennsylvanian strata are not represented and the upper boundary of the Quadrant represents an unconformity. The overlying rocks range in age from Early Permian to Middle Jurassic and, locally, Cretaceous (Maughan,

Figure 7. Crossbedding in the Quadrant Formation with a penecontemporaneous slump. North side Middle Fork of Little Sheep Creek, NW¼, Section 22, T. 15 S., R. 9 W.

1975, p. 287). In extreme southwestern Montana, however, Permian strata lie conformably above the Quadrant Formation. Parts of the upper Quadrant probably of Virgilian age. No fossils were found in this interval, but the rocks are continuous with Virgilian strata in southeastern Idaho (Maughan, 1975, p. 289).

THE PERMIAN SYSTEM

The Permian System of southeastern Idaho and adjoining regions in Utah, Nevada, Wyoming, and Montana is made up of a complicated intertonguing of facies that represents continuous deposition in laterally shifting environments. In general, a chert-mudstone-phosphorite facies dominant in eastern Idaho and southwestern Montana intertongues with sandstone facies to the northeast and carbonate facies to the south and east. These rocks had been assigned names which were applied to the local stratigraphic interval without much consideration to the specific facies and their regional intertonguing relationship. The predominantly carbonate rocks lying between the Pennsylvanian Weber Sandstone and the Triassic Dinwoody or Woodside Formations in northeastern Utah were called the Park City Formation; the predominant cherts, phosphatic mudstones, and sandstones between the Quadrant and the Dinwoody Formations in Montana were assigned to the Phosphoria Formation. Also assigned to the Phosphoria were strata lying between the underlying Tensleep sandstone, or the Wells Formation, and the overlying Dinwoody or Woodside Formations in eastern Idaho, western Wyoming, and Rich County, Utah. Rocks occupying a similar stratigraphic position in northwestern Utah and northeastern Nevada were called the Gerster Formation (McKelvey et al., 1956, p. 2829, 2833, 2834).

In 1956, McKelvey and numerous co-workers, under the auspices of the United States Geological Survey, revised the Permian nomenclature in the central western states in order to provide a system which more closely follows "the American system of stratigraphic nomenclature" (McKelvey et al., 1956, p. 2838). The nomenclature they adopted is based on the recognition of lithic units and application of those names where the specific lithology is dominant. To quote from McKelvey et al. (1956, p. 2834):

> "The Phosphoria formation is best developed in eastern Idaho, northern Utah, western Wyoming, and southwestern Montana, but tongues of it extend over a much wider area. The Park City formation is best developed in north-central and eastern Utah and in southwestern and west-central Wyoming, but tongues extend to eastern Idaho and to Montana as well. The Shedhorn Sandstone is best developed in the general vicinity of Yellowstone Park, but tongues extend over much of southwestern Montana and northwestern Wyoming. The three formations can be regarded as end-member types that interfinger over much of the area."

This nomenclature is presently accepted and used by stratigraphers.

McKelvey et al. (1956, p. 2834) subdivided the three formations into eleven members, six of which were recognized in the Little Sheep Creek area. These are, in descending order:

Tosi Chert Member, Phosphoria Formation (Sheldon, _in_ McKelvey et al., 1956, p. 2851)

Retort Phosphatic Shale Member, Phosphoria Formation (Swanson, _in_ McKelvey et al., 1956, p. 2850)

Franson Tongue, Park City Formation (Cheney, _in_ McKelvey et al., 1956, p. 2842)

Rex Chert Member, Phosphoria Formation (Richards and Mansfield, 1912, p. 684)

Meade Peak Phosphatic Shale Member, Phosphoria Formation (McKelvey et al., 1956, p. 2845)

Grandeur Tongue, Park City Formation (Cheney et al., _in_ McKelvey et al., 1959, p. 12-15)

The Shedhorn Sandstone was not recognized in the thesis area.

Where members of one formation extend into an area dominated by another formation, the members are regarded as tongues within the second formation (McKelvey et al., 1956, p. 2835). For this reason the Permian rocks in the Little Sheep Creek area are mapped as the Phosphoria Formation, with recognition of the Grandeur and Franson Tongues of the Park City Formation.

The Phosphoria Formation

The Phosphoria name was first applied by Richards and Mansfield (1912, p. 684) to dark chert, phosphatic and carbonaceous mudstone, phosphorite, cherty mudstone and minor amounts of carbonate rock found in Phosphoria Gulch near Meade Park, Idaho. Two members were recognized and described: a lower phosphatic shale member about 200 feet (61 m) thick and the overlying Rex Chert member consisting of about 150 feet (46 m) of bedded chert, overlain by 100 feet (31 m) of mudstone and cherty mudstone. McKelvey et al. (1956, p. 2845, 2849) named the lower shale the Meade Peak Phosphatic Shale Member, and the mudstones above the Rex Chert, the Cherty Shale Member. The Meade Peak and Rex Chert names were extended to include beds in western Wyoming and southwestern Montana.

Three other members of the Phosphoria Formation which do not occur at the type section also were named. These are: the Retort Phosphatic Shale Member named in Montana by Swanson (<u>in</u> McKelvey et al., 1956, p. 2850), which can be extended into western Wyoming, the Tosi Chert Member named in western Wyoming by Sheldon (<u>in</u> McKelvey et al., 1956, p. 285) and extended into Montana, and the lower chert member which is found only in western Wyoming (Sheldon, <u>in</u> McKelvey et al., 1956, p. 2845).

The Park City Formation

The Park City Formation was first named by Boutwell (1907, p. 443-446) in the Wasatch Mountains of Utah for carbonate rock with minor sandstone and shale lying between the Weber Quartzite below, and the Woodside Formation above. The type area is in Big Cottonwood Canyon in the Park City mining district near Salt Lake City. At the type section, the carbonates are separated by a phosphatic shale unit which was traced by Richards and Mansfield (1912, p. 684-689) northward into the Phosphoria Formation and shown to be continuous with the Phosphoria. Cheney (in McKelvey, 1956, p. 2840) called the shale unit the Meade Peake Phosphatic Shale Tongue of the Phosphoria Formation. The upper carbonate unit was formally named the Franson Member by Cheney (in McKelvey, 1956, p. 2842); the lower carbonate unit was named the Grandeur Member by Cheney et al. (in McKelvey et al., 1959, p. 12-15).

Uranium and vanadium exploration during World War II, and post-war development of the phosphate industry, prompted the detailed mapping of phosphate-bearing rocks in the middle western states. The investigation was undertaken by the United States Geological Survey which published interim reports beginning in 1949 and culminated in 1964 with the publication of a detailed report on the Permian rocks in southwestern Montana by Cressman and Swanson (1964, p. 276; and references therein). Readers interested in a more detailed discussion of the Phosphoria Formation in southwestern Montana are referred to this excellent paper.

A section of the Phosphoria Formation was measured in the West and Middle Forks of Little Sheep Creek by Cressman, Wilson, and Tandy in 1949 (in Cressman and Swanson, 1964, p. 476-481). The section was described from natural exposures and two bulldozer trenches, the scars of which can still be seen today. The sides of the bulldozer trenches have since been covered by mass wasting so that the exposures created by mechanical means in 1949 are no longer available. The four members of the Phosphoria Formation and the two tongues of the Park City Formation described by Cressman and Swanson in Little Sheep Creek could, however, be recognized and described from numerous exposures in the thesis area.

Regional Distribution and Correlation

The chert and mudstone facies of the Phosphoria Formation are best developed in an area centered in eastern Idaho. Tongues of the formation extend into Montana, Wyoming, Utah, northeastern Nevada and northwestern Colorado (McKelvey et al., 1956; Wardlaw, 1980, p. 353) giving the formation a large area over which it is recognized. The intertonguing relationship between the Phosphoria and its lateral equivalents provides an integrated system of facies deposited during Permian time which covered much of the northern Rocky Mountain region and the northeastern Great Basin. The Phosphoria intertongues with the Shedhorn Sandstone to the northeast and the Park City Formation to the south and east (McKelvey et al., 1956). Equivalent Permian-age rocks in southeastern Wyoming and northern Colorado are the red beds of the Moenkopi, Satanka, Goose Egg, and Chugwater Formations (Miller and Cline, 1934, p. 281; Sheldon et al., 1967, p. 157; McKelvey et al., 1956, p. 2854). Equivalent rocks in eastern Wyoming, western South Dakota and southeastern Montana belong to the Minnekahta Limestone (Wardlaw, 1980, p. 359). In central and southern Utah and eastern Nevada equivalent age rocks are the Kaibab Limestone, Plympton Formation, Murdock Mountain Formation and the Gerster Limestone (Wardlaw, 1980, p. 354-359).

North and east of the thesis area in Montana and Wyoming the Phosphoria Formation overlies the Quadrant and Tensleep Formation with an unconformity marked by a conglomerate or an erosional surface. In other places paleontological evidence indicates an absence of Late Pennsylvanian or Early Permian rocks (Cressman and Swanson, 1964, p. 298-299; Sheldon et al., 1967, p. 157). The Permian rocks in most of the region are disconformably overlain by Triassic rocks, or in central Montana where pre-Jurassic erosion stripped away older rocks, the Phosphoria Formation is disconformably overlain by Jurassic strata (Sheldon et al., 1967, p. 169).

Distribution and Topographic Expression

The Phosphoria Formation is a prominent and widespread formation in the Little Sheep Creek area that parallels the Quadrant Formation. The chert and mudstone facies are non-resistant and poorly exposed. The carbonate Grandeur and Franson Tongues of the Park City Formation are resistant, however, and were the basis for delineating the outcrop pattern of the Phosphoria Formation in the field. In the eastern part of the thesis area the Phosphoria forms a broad grass-covered slope on the dip-slope of the tree-covered Quadrant Formation. It is also the resistant unit underlying the ridge immediately north of Two Springs Gulch in the northern part of the area (Plate I). A tectonically thinned part of the Phosphoria is found on the southern limb of the overturned syncline in the southwestern part of the thesis area.

The Grandeur Tongue of the Park City Formation is transitional with the underlying Quadrant Formation and marks a conformable contact. The contact of the Phosphoria with the overlying Dinwoody Formation is covered; the exact nature of the contact could not be determined in the field. The nature of the Permian-Triassic boundary is discussed in a subsequent section.

Thickness and Lithology

The Phosphoria Formation is approximately 200 m thick in the Little Sheep Creek area. Cressman and Swanson (1964, p. 476-481) described 487 feet (149 m) of the Phosphoria in Little Sheep Creek and recognized six of the eleven subdivisions of the Permian System described by McKelvey et al. (1956).

The Grandeur Tongue. Cressman and Swanson measured about 345 feet (105 m) of the Grandeur Tongue of the Park City Formation in Big Sheep Creek, but only the upper 167 feet (51 m) in Little Sheep Creek; the lower half and the contact with the Quadrant Formation are covered at their section location. The contact with the Quadrant Formation and the lower 30 m of the Grandeur Tongue can be observed on the south canyon wall in the SE¼, NE¼, SE¼, Section 16, T. 15 S., R. 9 W. The contact is taken to be the first appearance of a micritic dolomite that is yellowish gray (5 Y 7/2) on a fresh surface, weathering to pale yellowish brown (10 YR 6/2).

The Grandeur Tongue is generally micritic to very finely crystalline dolomite interbedded with limestone, siltstone, and fine-grained sandstone. Outcrops of the dolomite are yellowish gray (5 Y 7/2) or very pale orange (10 YR 8/2) on fresh surfaces, and weather very light gray (N 8) and dark yellowish orange (10 YR 6/6) to yellowish gray (5 Y 7/2). The dolomite varies from being very thin-bedded to very thick-bedded.

The interbedded limestone is a micritic to finely crystalline spar that is generally pale yellowish brown (10 YR 6/2) and light olive gray (5 Y 6/1) on fresh surfaces, weathering light gray (N 7) to yellowish gray (5 Y 8/1). The bedding characteristics vary from being very thin-bedded to very thick-bedded. The limestone is nearly always fossiliferous with fenestral and ramose bryozoans, crinoid columnals up to 8 mm in diameter, and productid brachiopods. The top of the Grandeur Tongue is marked by a thin-bedded medium crystalline limestone that is yellowish gray (5 Y 7/2) on fresh surfaces, weathering light olive gray (5 Y 6/1). A noticeable characteristic of the carbonates of the Grandeur Tongue is the presence of nodular chert and some minor thin (less than 30 cm) discontinuous beds of chert (Figure 8). The chert is pale to dark yellowish brown (10 YR 6/2 to 10 YR 5/4). The sandstones are very fine- to fine-grained calcareous quartz arenites that are generally well-sorted with subangular to angular grains. A sandstone bed in the upper part of the unit contains reddish brown iron oxide spots (about 2% of the rock) which appear to be alteration products of pyrite. The outcrops are generally very pale orange (10 YR 8/2) and pale yellowish brown (10 YR 6/2) on weathered surfaces whereas on fresh surfaces they are yellowish gray (5 Y 7/2) to pale grayish orange (10 YR 8/4). The sandstone is thin- to thick-bedded with some beds having thin, planar laminations and cross laminations. One exposure of sandstone contains symmetric ripples that have a bi-directional flow pattern oriented N 34°W.

About 70 m above the base of the Grandeur Tongue there occurs a calcareous laminated to very thin-bedded siltstone that is pale to moderate reddish brown (10 R 5/4 to 10 R 4/6) with mottled areas that are dark yellowish orange (10 YR 6/6). This siltstone unit is approximately 25 m thick and forms a distinct reddish orange swash of color on the Phosphoria slopes (Figure 9). Lower in the section and more intimately associated with the carbonates, the siltstones form laminated interbeds that are very pale orange (10 YR 8/2).

The Franson Tongue. This unit is only 43 m thick in Little Sheep Creek as measured by Cressman and Swanson (1964, p. 477-478). The base of the unit is made up of about 12 m of thick-bedded finely crystalline very pale orange (10 YR 8/2) dolomite interbedded with very thick-bedded limestone that contains ramose bryozoans. The upper part of the Franson Tongue consists of thin- to thick-bedded limestone that has a fine to medium crystalline texture. The outcrops are generally light gray (N 7) to light olive gray (5 Y 6/1) on fresh surfaces, weathering very light gray (N 8) and pale yellowish brown (10 YR 6/2). The rocks are fossiliferous, containing sparse to nearly coquinoid populations of ramose bryozoans, crinoid columnals, and brachiopods. Less frequently the limestones contain glauconite pellets and chert, the latter as both nodular and thin discontinuous beds. The chert in both carbonate tongues of the Park City Formation is of replacement origin (Cressman and Swanson, 1964, p. 296).

The Meade Peak Phosphatic Shale Member. This unit is about 4 m thick (Cressman and Swanson, 1964, p. 477-478) in Little Sheep Creek and is poorly exposed. The member consists of olive gray (5 Y 4/1) laminated and shaly siltstone, and thin-bedded, medium sand-size, oolitic phosphorite that contains phosphatic nodules 1-3 mm in diameter. The phosphorite is dusky yellowish brown (10 YR 2/2) on fresh surfaces and weathers to a dark yellowish brown (10 YR 4/2). The Meade Peak shale

Figure 8. Chert nodules extracted from dolomite in the Grandeur Tongue of the Park City Formation within the Phosphoria Formation; jeep trail NE¼, Section 32, T. 14 S., R. 9 W.

Figure 9. View looking west across the Middle Fork of Little Sheep Creek from the SE¼, Section 3, T. 15 S., R. 9 W. Bulldozer cut right center on the distinctive red siltstone of the Grandeur Tongue, Park City Formation. Two Springs Gulch center left, Gallagher Gulch center right, White Pine Ridge right center.

overlies the Grandeur Tongue of the Park City Formation and lies beneath the Rex Chert Member of the Phosphoria Formation.

The Rex Chert Member. This member consists of interbedded chert and sandstone with minor amounts of siltstone and cherty mudstone approximately 18 m thick. The chert is thin- to thick-bedded and ranges in color from olive gray (5 Y 3/2) to dark gray (N 3) and brownish black (5 YR 2/1). The chert is not well exposed in Little Sheep Creek but the angular weathering products form a pebbly gravel that crunches underfoot.

The sandstone is fine- to medium-grained with some beds being very fine-grained. The sandstone is of the "salt-and-pepper" variety being composed of grains of quartz and dark chert grains with the latter comprising 15-45% of the rock. Approximately 5-10% of the detritus is feldspar, and careful observation shows the presence of glauconite pellets and dark phosphatic pellets locally. Consequently, the color of the rock can be light or dark; specifically, the rock is pale yellowish orange (10 YR 8/6), medium gray (N 5), brownish gray (5 YR 4/1), or pale grayish blue (5 PB 6/2). The sandstone is generally thin-bedded, but it can also show very thick-bedded, featureless bedding characteristics.

The siltstone and cherty mudstone are variously laminated and fissile, weathering out to shale chips that are dark yellowish orange (10 YR 6/6), and pale to moderate yellowish brown (10 YR 5/4).

The Retort Phosphatic Shale Member. This member of the Phosphoria Formation overlies the Franson Tongue of the Park City Formation and underlies the Tosi Chert Member of the Phosphoria Formation. The Retort

Shale Member consists mostly of siltstone and lesser amounts of phosphatic sandstone approximately 9 m thick. The siltstone is calcareous, laminated or fissile, and contains varying amounts phosphatic material that give the rock a dirty brownish color. Weathered shards of siltstone are generally light olive gray (5 Y 5/2) in color. The sandstone is very fine- to fine-grained, thin-bedded, calcareous, and contains varying amounts of phosphatic and glauconite pellets. The rock is generally moderate yellowish brown (10 YR 5/4) to brownish gray (5 YR 4/1) on fresh surfaces and weathers pale and dark yellowish brown (10 YR 6/2 and 10 YR 4/2).

The Tosi Chert Member. This unit overlies the Retort Phosphatic Shale Member of the Phosphoria Formation and underlies the Triassic Dinwoody Formation. The Tosi Chert consists of about 30 m of olive gray (5 Y 4/1), thin-bedded chert, laminated cherty siltstone, laminated silty mudstone, and minor amounts of medium-grained sandstone. The sandstone is a well-sorted quartz arenite with subangular grains and about two percent dark chert. It is thinly bedded and is olive gray (5 Y 4/1) in color. The siltstone is grayish red (10 R 4/2) to blackish red (5 R 2/2) in color, the silty mudstone is light olive gray (5 Y 6/1), and the cherty siltstone is moderate yellowish brown (10 YR 5/4).

Fossils and Age

Miller and Cline (1934, p. 281-283) considered the Phosphoria Formation to be Middle Permian (Guadalupian) in age. They based their conclusions on an assemblage of nautiloid and ammonoid cephalopods found in the "goniatite beds" of what is now considered the Meade Peak

Phosphatic Shale Member. The collection was made below the Rex Chert Member, about 150 feet (46 m) above the Wells Formation in the Sublette Range of western Wyoming. The Phosphoria was considered correlative with the Word Formation of Texas.

Frenzel and Mundorf (1942, p. 675-768) discovered the fusulinids, Schwagerina laxissima and Pseudoschwagerina montanensis, in the basal limestone of the Phosphoria (now the Grandeur Tongue of the Park City Formation) about 3 m above the Quadrant Formation near Three Forks, Montana. The fusulinids indicate a Wolfcampian (Early Permian) age and can be correlated with the Wolfcamp and Huerco Formations of Texas. The time interval between the Quadrant and Phosphoria Formations represents a hiatus in the Three Forks area that spans the Late Pennsylvanian.

King (1930, p. 30-33) found brachiopods in the Meade Peak Member and the Franson Member that could also be correlated with the Word Formation of Texas. King noted that the brachiopod fauna of the chert and carbonate facies differed from the phosphatic shale facies, but that "together the faunas of the two members form a rather varied aggregation of forms of similar stratigraphic affinities" (King, 1930, p. 31). This relationship was further substantiated by a more definitive study of Permian faunas in the Middle Rocky Mountain region by Yochelson in 1968 (in Boyd and Maughan, 1973, p. 305). The species of brachiopods could not provide a time bracket for the Phosphoria, but lithostratigraphic relationships combined with the biostratigraphy provided an Early to Middle Permian age for the Phosphoria Formation.

Clark et al. (1977, p. 656) found Neogondolella postserrata conodont fauna in the upper part of the Gerster Formation of northern Utah which indicate a late Guadalupian age for that formation. The Gerster Formation encompasses the youngest known Permian rock in the northeastern Great Basin and Rocky Mountain region. The Gerster is overlain by the Dinwoody Formation which bears a conodont fauna including *Anchignathodus sarcicus*, a species which is confined to the Early Triassic (early Scythian) in several parts of the world (Clark et al., 1977, p. 657). A similar stratigraphic relationship exists between the Phosphoria and Dinwoody Formations in southwestern Montana. The time interval represented by the Gerster Formation spans middle Wordian to early Capitanian (early Guadalupian) time and is time-equivalent to parts of the Rex Chert, the Retort Phosphatic Shale, and an unnamed upper siltstone member of the Phosphoria Formation (Clark et al., 1979; Wardlaw and Collinson, 1979, p. 157; Wardlaw, 1980, p. 354). The stratigraphic relationship of the Phosphoria with the underlying Quadrant Formation in southwestern Montana indicates that deposition continued from the Pennsylvanian into Middle Permian (middle Guadalupian) time.

Fossils were found only in the Grandeur and Franson Tongues of the Park City Formation in the Little Sheep Creek area. The fossils include ramose and fenestral bryozoans, circular crinoid columnals, productids, and spiriferids. One genus of brachiopods has been tentatively identified by me (Shimer and Shrock, 1943, p. 349) as *Dictyoclostus*.

Boyd and Maughan (1973, p. 395) focused their attention on defining the Permian-Triassic boundary in the northern Rocky Mountain region. They noted that historically there had been a general lack of interest in the boundary because of marked contrasts in the resistivity of the rocks above and below the contact. This observation may be valid in parts of Idaho, Utah, and Wyoming where the Dinwoody Formation overlies carbonates of the Park City Formation; however, in southwestern Montana and western Wyoming, where the lithologic variation is not that great, the nature of the contact is obscure. Consequently, some stratigraphers have described the Permian-Triassic boundary as disconformable and others have considered it as gradational (Boyd and Maughan, 1973, p. 310).

In southwestern Montana both the upper part of the Phosphoria Formation and the basal Dinwoody Formation consist of nonresistant mudstones. The contact is rarely exposed and its position is derived from slight variations in the composition of the mudstones. Characteristics of the Permian beds such as sponge spicules, chert, glauconite grains, and phosphatic pellets are not present in the Dinwoody strata (Cressman and Swanson, 1964, p. 354-355). In contrast, the basal Dinwoody mudstone is calcareous and contains a greater proportion of clay than do the Phosphoria strata.

Love (1948), referred to by Moritz (1951, p. 1784), stated that there is a very slight discordance between the Dinwoody and Phosphoria Formations observable on a regional scale. Newell and Kummel (1942, p. 941) recognized three subdivisions of the Dinwoody Formation in western Wyoming and southeastern Idaho based on the relative abundance of _Lingula_ and _Claraia_. They were able to demonstrate an onlapping sequence northeastward into Wyoming where progressively younger rocks in the Dinwoody overlie the Phosphoria. Sheldon et al. (1967, _in_ Boyd and Maughan, 1973, p. 311) argued that the apparent onlap can also be explained as a facies change caused by contemporaneous deposition of various units in different areas under regressive conditions. Wardlaw

(1980) used brachiopod and conodont biostratigraphy to reconstruct a similar syndepositional model for the Permian strata in the same area. However, biostratigraphic control of the Triassic rocks does not exist in enough detail to confirm the interpretations of Newell and Kummel or Sheldon et al.

What evidence that does exist indicates that there is a hiatus which possibly represents late Capitanian and all of the Ochoan time in the Late Permian. In the southern Wasatch Mountains, Utah, the Woodside Formation lies directly on Permian strata with an angular unconformity. North and northwest of the Wasatch Mountains the Woodside intertongues with the Dinwoody so that the lower Woodside and upper Dinwoody Formations are time-equivalent (Kummel, 1954, p. 170). A hiatus with an erosional unconformity is implied between the Permian and Triassic rocks.

In southwestern Montana the boundary was considered to be conformable by Scholten et al. (1955, p. 366) and Kummel (1954, p. 168). Moritz (1951, p. 1785) quotes Sloss who suggested that the lower Dinwoody strata may be Permian and may actually represent continuous sedimentation from the Permian into the Triassic. Prompted by such a statement Kummel (1954, p. 168) looked for, and found, Scythian ammonites and Claraia in the Dinwoody within five feet of the Phosphoria-Dinwoody contact in the same area. Kummel still felt the contact to be conformable in southwestern Montana and southeastern Idaho. Klecker (1980, p. 52) quotes B. R. Wardlaw and suggests that the thin strata between positively identified Permian and Triassic strata may represent continued deposition across the boundary. Wardlaw suggested that the Phosphoria basin was filled to near base level during the Middle Permian and that if sedimentation did continue into Late Permian time, the volume was small because of the establishment of a profile of equilibrium across southwestern Montana. Bypassing of sediment, or erosion prior to Dinwoody deposition, would have left very little of this sediment. Klecker concluded that the Phosphoria-Dinwoody contact within southwestern Montana would therefore be a disconformity.

In 1977, Clark et al. further defined the Permian-Triassic boundary in the Terrace Mountains of northeastern Utah using conodont biostratigraphy. They found a Neogondolella postserrata fauna in the uppermost beds of the Gerster Formation. This fauna can be correlated with the Lamar Limestone of Texas which is Amorassean (late Guadalupian) in age. The lowermost beds of the Dinwoody Formation contain a Griesbachian (early Scythian) conodant fauna, including Anchignathodus typicalis. The Gerster Formation represents the youngest Permian strata in the northeastern Great Basin (Clark et al., 1977, p. 655). Clark et al. found a Guadalupian age brachiopod fauna in the upper part of the Gerster, one species of which is found in the uppermost beds of the Phosphoria Formation in Montana and Wyoming. This correlation between the uppermost beds of the Gerster and Phosphoria Formations has been confirmed by Wardlaw (1980, p. 358) based on both conodont and brachiopod faunas.

The preceding discussion has involved only a summary of some of the most pertinent literature concerning the Permian and Triassic Systems. Though inconclusive it does indicate a widespread hiatus across southwestern Montana, southeastern Idaho, western Wyoming and northwestern Utah that involved either non-deposition or erosion during Late Permian time.

THE TRIASSIC SYSTEM

The Triassic System of southwestern Montana consists of three mutually conformable formations. They are, in ascending order, the Dinwoody, the Woodside, and the Thaynes Formations. The Dinwoody unconformably overlies the Permian Phosphoria Formation, and the Thaynes Formation is overlain by the Gypsum Spring Tongue of the Twin Creek Formation by an unconformity spanning Middle Triassic to Early Jurassic time.

The Dinwoody Formation

Condit (1916, p. 263) first applied the name Dinwoody to 200 feet (60 m) of shale forming the upper part of the Embar Formation in Dinwoody Canyon in the Wind River Range near Dubois, Wyoming. Blackwelder (1918, p. 425) defined the formation as consisting of 200 feet (60 m) of gray siltstones and shales with thin, brown limestones near the base. The upper limits of the Dinwoody were considered to be the change in color from gray siltstones and shales to red siltstones and shales typical of the Chugwater Formation. Newell and Kummel (1942, p. 941) discovered that even within Dinwoody Canyon the color transition was local, discontinuous, and was not a suitable marker for the top of the Dinwoody Formation. They restricted the Dinwoody at the type locality to a 90 foot (27 m) sequence of dominantly silty strata between the Phosphoria Formation and the top of the resistant siltstones located about half-way toward the top of the original Dinwoody.

Regional Distribution and Correlation

The Dinwoody Formation occurs wherever Early Triassic rocks are found in southwestern Montana, western Wyoming, northern Utah and northeastern Nevada (Newell and Kummel, 1942, p. 942; Reeside et al., 1957, p. 1479; Collinson and Hasenmueller, 1978, p. 177; Wardlaw and Collinson, 1979, p. 157). Its maximum measured thickness is 2,444 feet (745 m) southeast of Fort Hall, southeastern Idaho, at the apex of a northeast-trending depositional trough (Kummel, 1960, p. 233, Figure 1). The Dinwoody thins markedly north, east, and south where it intertongues with, and is transitional to, red shales and sandstones of the Woodside and Chugwater Formations (Figure 10). The thinning of the Dinwoody, and the general absence of any Triassic strata in central-western and northern Montana, are caused by pre-Jurassic erosion rather than non-deposition (Moritz, 1951, p. 1798).

The Dinwoody Formation is correlative with the Candelaria Formation of central Nevada (Collinson and Hasenmueller, 1978, Figure 2), the Moenkopi Formation in the eastern Uinta Mountains of Utah, and the Lykins Formation of northeastern Colorado (Reeside et al., 1957, p. 1478).

Distribution and Topographic Expression

The Dinwoody Formation is a widespread unit that follows the outcrop pattern of the Phosphoria Formation. It is also exposed in the axis of a minor anticline in the SE¼, Section 17, T. 15 S., R. 9 W. and along the limbs, and in the axis, of folded strata in the southwest corner of the map (Figure 11).

The lower fissile calcareous siltstone of the Dinwoody is nonresistant and forms gentle slopes. The limestones and calcareous siltstones of the upper Dinwoody form discontinuous narrow ledges that are typically thin-bedded and "chocolate brown" in color.

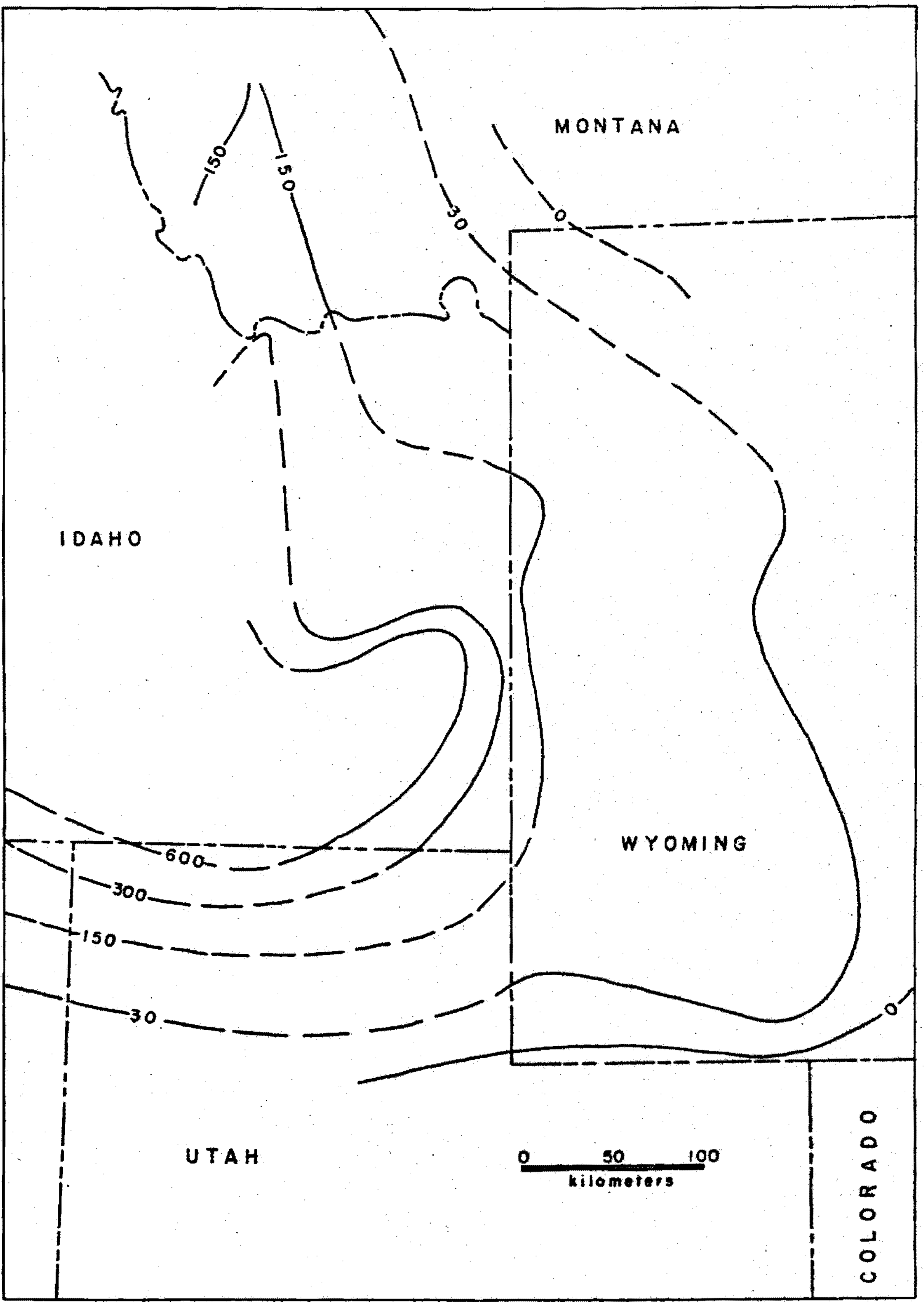


Figure 10. Isopach Map of the Dinwoody Formation (modified from Moritz, 1951; Kummel, 1960; Collinson and Hasenmueller, 1978). Isopachs in meters.

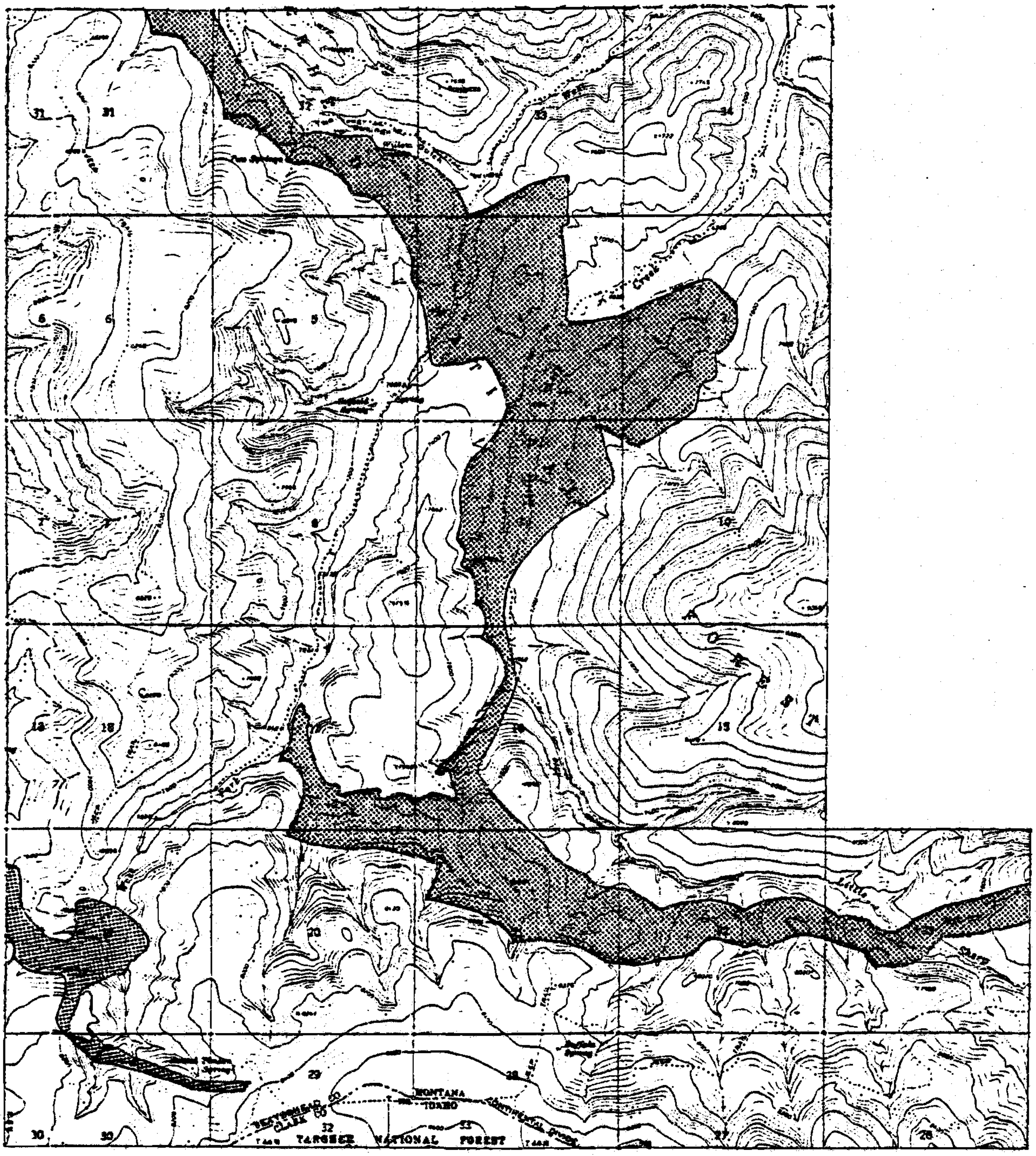


Figure 11. Distribution map of the Dinwoody Formation.

Thickness and Lithology

The Dinwoody Formation in southeastern Idaho and western Wyoming was divided into three units by Newell and Kummel (1942, p. 941). The lowest division, a basal siltstone, is geographically limited to the eastern flank of the geosynclinal area. The second division was restricted to thin, dense limestones, and dark gray to olive shales that contain large numbers of *Lingula borealis* Bittner. The upper division consists of resistant, calcareous, olive to brown siltstone interbedded with hard, gray limestone that contain an abundance of *Claraia stachei* Bittner. The upper two divisions of the Dinwoody Formation were referred to as the *Lingula* and *Claraia* zones, respectively.

Moritz (1951, p. 1787-1788) could distinguish the three subdivisions of the Dinwoody in southwestern Montana but found that they did not represent readily mappable units. Instead, he proposed a twofold subdivision. The lower unit, the Shale Member, consists of laminated and very thin-bedded siltstones with discontinuous interbeds of limestone. The upper unit, the Limestone Member, consists of gray to brown argillaceous limestones, siltstones and interbedded olive to brown shales equivalent to the siltstone-limestone facies of Kummel (1957).

This twofold division of the Dinwoody Formation can be extended south into the westernmost exposures of the formation in southeastern Idaho (Kummel, 1957, p. 442).

In the Little Sheep Creek area the Dinwoody Formation is approximately 256 m thick and is divisible into a lower fissile calcareous siltstone member and an upper siltstone-limestone member.

The lower member is approximately 30 m thick and consists of laminated and very thin-bedded calcareous siltstone with isolated interbeds of thinly bedded silty limestone. The calcareous siltstones are nonresistant, weathering to shaly and flaggy shards that are dark yellowish brown (10 YR 4/2) and grayish orange (10 YR 7/4) on broken surfaces, and weather pale to moderate yellowish brown (10 YR 6/2 - 10 YR 5/4).

The silty limestone is composed of finely crystalline sparry calcite and is pale yellowish brown (10 YR 6/2) on fresh surfaces and moderate yellowish brown (10 YR 5/4) when weathered. Except for stratigraphic position, it is indistinguishable from the silty limestones in the upper member of the Dinwoody.

The upper member of the Dinwoody Formation consists of about 226 m of interbedded calcareous siltstones and limestones in nearly equal proportions and, to a lesser extent, interbeds of silty limestone.

The calcareous siltstone is generally light gray (N 7) and yellowish gray (5 Y 8/1) on fresh surfaces and weathers pale to moderate yellowish brown (10 YR 6/2 - 10 YR 5/4). It is very thin- to thin-bedded (1-18 cm) with microstructures on weathered surfaces consisting of horizontal thin laminations, cross-laminations, and festooned trough cross-laminations, the latter with sets 2-3 cm thick. Near the top of the Dinwoody, beneath the contact with the Woodside Formation, the calcareous siltstone contains interbeds of dolomite and thin-bedded and laminated very fine-grained sandstone.

Estimation of the calcite-quartz silt mixture of the siltstones is difficult petrographically because of the very fine-grained texture of the rock, the complete cementation with calcite, and the partial replacement of the quartz with calcite. The estimated quartz silt component is about 80-95%; however, modal analyses of a representative sample yield the following apparent silty limestone composition:

Calcite	64%
Monocrystalline unstrained quartz	27
Monocrystalline strained quartz	3
Heavy minerals	4
Fossil fragments	2
	100%

For the purpose of classifying calcareous siltstones in this thesis, a siltstone is defined as that rock which contains detrital quartz grains and accessory minerals in the Wentworth silt-size range which are in framework support, or constitute at least 30% of the rock. An allochemical rock with a quartz silt content of 5-30% will be classified as a silty limestone. A quartz silt content of less than five percent in a limestone will not bear on the classification of the rock.

The calcareous siltstones consist of well-sorted angular to subangular grains of quartz with calcite cement. The sizes of the quartz grains range from medium silt to very fine sand with the mode falling in the medium silt range. Accessory minerals include biotite, muscovite, zircon, plagioclase, hematite, and collophane as pellets and as Lingula fragments. Of minor occurrence are chlorite, tourmaline, and staurolite.

Laminations in the calcareous siltstones are the result of textural and compositional changes in the rock. The laminations are generally very thin (up to 1 mm) and consist of quartz silt-rich laminations with varying admixtures of shell fragments and collophane and glauconite pellets. The quartz laminations alternate with concentrations of thin pelecypod shell fragments aligned parallel to bedding or with heavy mineral concentrations.

Many outcrops of calcareous siltstone display sedimentary microstructures; however, some beds are thoroughly homogenzied by bioturbation. Trails found on the surface, and disruption of the laminations along bedding planes, indicate burrowing activity restricted to the horizontal plane.

The limestones are well-indurated, finely crystalline rocks that are lighter in color than the calcareous siltstones. Fresh surfaces vary in color from white (N 9) to the lighter shades of gray (N 8-6), as well as yellowish gray (5 Y 8/1) and pale yellowish brown (10 R 6/2). The weathered surfaces are light gray (N 7), light olive gray (5 Y 6/1), and pale to moderate yellowish brown (10 YR 6/2 - 5/4). The limestones are very thin- to thick-bedded with most outcrops being thin-bedded (20-45 cm).

Petrographically most of the limestones are generally well-washed and well-sorted allochemical rocks that have undergone neomorphic recrystallization to the extent of nearly obliterating the original composition and texture of the rock; outcrops of the fine crystalline rocks are easily mistaken for silty limestones in the field. The limestone classification and characteristics are summarized in Table 1.

The bioclastic material consists predominantly of pelecypod shell fragments, the inarticulate brachiopod Lingula, echinoid plates and spines, and, rarely, gastropod fragments. The pelecypod shells occur as fragments up to 7 mm long that are generally aligned subparallel to bedding. The weathered anti-dip surfaces of pelecypod coquinas have a crenulated appearance because of the alignment of the shell fragments. Microscopically the original aragonite of the pelecypod shell structure

Table 1

Limestone Classification
of the
Dinwoody Formation

Classification (Folk, 1962)	Characteristics
1. Packed biopseudosparite	65% pelecypods, 25% *Lingula* fragments in grain support, 10% matrix
2. Intramicrosparite	Angular fragments of pelmicrite in a finely crystalline spar matrix
3. Packed biopelmicrosparite	Allochems include 45-60% pelecypods, 36-45% pellets, 10% silty micrite intraclasts in grain support in a microsparry matrix, homogeneous
4. Packed pelsparite	75% pellets up to 0.25 mm diameter, 15% shell fragments, and coated grains, 10% matrix of microspar
5. Sparse biomicrosparite	35-45% fragments of *Lingula*, echinoid plates and spines, pelecypods, and gastropod steinkern in microspar-pseudospar matrix support

has been replaced by calcite spar. Micrite rims outline the original shell surfaces; commonly neomorphic recrystallization has extended the growth of calcite crystals across micrite shell boundaries. One sample contains neomorphosed shell fragments along with fragments which have retained their original shell structure; the microstructure is typical of pelecypods (Figure 12; Majewski, 1969, Plate 61). Lingula occurs as phosphatic fragments up to 1.5 mm long.

Gastropod shell fragments are generally partially fragmented and chambered, with the aragonitic shelly material having been replaced with calcite spar. Most chambers are filled with micrite or silty micrite which form distinct steinkerns (Figure 13) in the absence of a shell wall. The micrite material of the steinkerns has not undergone the extensive neomorphism of the micrite in the interparticle matrices.

Intraclasts of micrite or silty micrite form one to five percent of the allochemical rocks. The clasts are generally elliptical, well rounded, and have not been affected by the neomorphism which has altered the rocks in which they occur.

The pelbiomicrosparites are thinly laminated with the laminations consisting of concentrations of packed pellets (60%) and pelecypod fragments (40%) alternating with concentrations of packed pelecypod fragments.

Intermediate in composition between the calcareous siltstones and the limestones are the silty limestones of the Dinwoody Formation. The silty limestones are pale yellowish brown (10 YR 6/2) or yellowish gray (5 Y 8/1) on fresh surfaces, and weather pale to dark yellowish brown (10 YR 6/2 to 10 YR 4/2). They are generally thin-bedded and contain discontinuous interbeds of very thin-bedded calcareous siltstones and

Figure 12. Photomicrograph, sparse biomicrosparite. L, Lingula shell fragments; P, recrystallized pelecypod shell fragment showing internal structure; M, recrystallized pelecypod shell with a micrite rim. Sample Kmb 79-51, Dinwoody Formation, NE¼, NE¼, Section 30, T. 15 S., R. 9 W. Plane light.

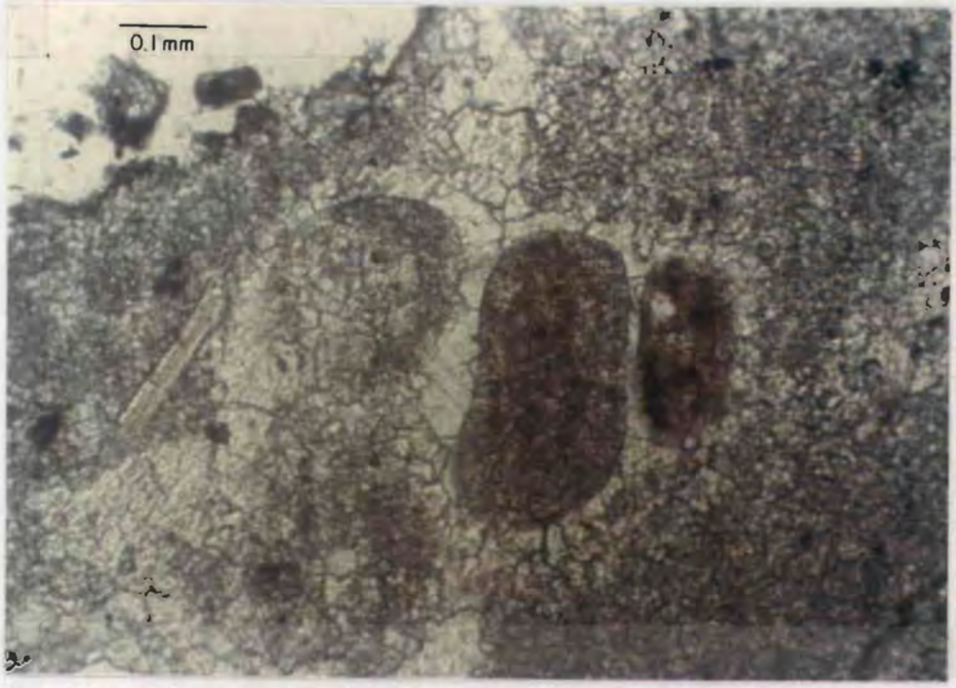


Figure 13. Photomicrograph, gastropod shell recrystallized to clear sparry calcite and filled with darker micrite. A mold of the gastropod chamber would be retained as a steinkern upon dissolution or the breaking of the gastropod shell. Dinwoody Formation, E½, Section 23, T. 15 S., R. 9 W. Plane light.

limestones. The thin beds of silty limestone also exhibit horizontal laminations, ripple laminations, and trough cross-laminations. Vertical burrows are evident on the surface of some weathered rocks.

Petrographically the silty limestones exhibit a zonation of allochems and quartz silt that reflect dynamic changes in the environment of deposition and changes with the source of the detritus. The cross-laminations visible on the weathered surfaces of the silty limestones are accentuated in part by thin discontinuous layers of 1) concentrated quartz silt, 2) mixed zones of quartz silt and biotic fragments, 3) mixed zones of quartz silt, pellets, and micrite intraclasts, 4) concentrated pelecypod and Lingula shell fragments, and 5) biotic fragments in micrite matrix support. Accessory minerals include zircon, tourmaline, muscovite, and collophane. Some quartz grains have a hematite coating. A majority of the intraclasts are micrite; however, some intraclasts are also an argillaceous micrite, a silty micrite or a calcareous and hematite stained siltstone. The lithoclasts are well rounded and range in size from submacroscopic clasts to clasts as large as 20 mm by 11 mm in cross-section.

The contact with the underlying Permian Phosphoria Formation is disconformable and has been discussed in a previous section. The contact with the overlying Woodside is conformable and gradational. Where observed, the contact is considered to be coincident with the change in lithology from a predominance of light gray calcareous siltstone of the Dinwoody to the fissile siltstone of the Woodside Formation.

Fossils and Age

The end of the Permian marked a worldwide diminution of some classes of marine invertebrates and the total extinction of others. The result is that the Early Triassic seas were impoverished in fauna; there was an abundance of individuals which represented a restricted number of classes. Four crinoid genera are known from the Early Triassic formations: *Encrinus*, *Pentacrinus*, *Isocrinus*, and *Balanocrinus*. The number of brachiopod genera was reduced to *Lingula*, *Orbiculoidea*, *Spiriferina*, *Rhynchonella*, and *Terebratula* (Kummel, 1957, p. 438).

The mollusca make up the principal fauna of Early Triassic sedimentary rocks with pelecypods and cephalopods, especially ammonites, making up the largest percentage of species and numbers of individuals. *Pseudomonitis*, *Claraia*, *Eumorphotis*, *Girvilleia*, *Anodontophora*, and *Myalina* are the most common of the pelecypod genera. Gastropods are represented by the genus *Bellerophon* (Kummel, 1957). Approximately 130 genera of ammonites are known from the Early Triassic; however, Newell and Kummel (1942, p. 950) discovered only two subgenera of *Ophiceras* in the Dinwoody Formation of southeastern Idaho and southwestern Wyoming. Moritz (1951, p. 1801) found poorly preserved and unidentifiable remains of ammonoids in the Dinwoody in southwestern Montana. Scholten et al. (1955, Table 1) identified *Meekoceras* and three species of *Ophiceras* in the same region.

The Dinwoody Formation in the Little Sheep Creek area is fossiliferous; however, the fossils are generally very fragmentary or occur only as impressions on bedding planes of the calcareous siltstone. Petrographic examination of the Dinwoody rocks reveals an abundance of pelecypod shell fragments, *Lingula*, echinoid plates and spines, and a few gastropods. The numbers of specimens are large, but the diversity of species appears to be small. Differentiation other than according to phylum and class is difficult. Lingula is a very common fossil (Shimer and Shrock, 1959, p. 285) found as impressions on calcareous siltstones and as phosphatic shell fragments in thin section (Pettijohn, 1975, p. 431). Eumorphotis could also be identified from well-preserved impressions in siltstone (Moore, 1969, p. N 337). The ammonite Ophiceras (Shimer and Shrock, 1959, p. 569) was tentatively identified by me from Dinwoody float found east of the thesis area.

Correlation of the Dinwoody Formation with standard Early Triassic Stages of Greenland, India, Siberia, and the Alps must be made on the basis of pelecypod fauna (Moritz, 1951, p. 1801-1802). The Claraia zone of Newell and Kummel (1942, p. 941) can be correlated with the Otoceras zone of India and Greenland and the Claraia zone of Siberia (Moritz, 1951, p. 1802) which are standard faunal zones of the Scythian (Early Triassic; Newell and Kummel, 1942, p. 950).

Clark et al. (1977) found a well-developed conodont fauna in the Dinwoody Formation in northwestern Utah, including Anchignathodus isarcicus, which is a species confined to the earliest Scythian. As discussed earlier, the Permian-Triassic boundary as represented by the Gerster-Dinwoody contact in northwestern Utah has been correlated with the Permian-Dinwoody contact in southwestern Montana and adjacent areas.

Based on pelecypod and conodont zones, the Dinwoody Formation is early Scythian (Early Traissic) in age.

Environments of Deposition

Deposition of Dinwoody sediments in southwestern Montana occurred on the northwestern margin of an embayment of the Cordilleran miogeosyncline (Figure 10). The embayment during Dinwoody time extended into western Wyoming at least as far as the Wind River Range. The siltstone-limestone facies described for the Dinwoody Formation in the thesis area is the most widespread facies of the formation (Kummel, 1957, p. 456) and is always associated with marginal areas of the miogeosyncline. The fauna of the Dinwoody Formation includes shallow-water forms, and coupled with the near-uniform deposition of the siltstone-limestone facies, indicate very broad equable conditions for sedimentation with even rates of subsidence. A discussion involving the fauna, lithology, and bedding characteristics of the Dinwoody strata will help to illuminate more specifically the conditions under which sedimentation occurred at Little Sheep Creek.

The faunal, lithologic, and bedding characteristics of the Thaynes Formation are similar to the Dinwoody Formations. For this reason most of the discussion presented for the Dinwoody environments applies to the Thaynes, and for purposes of brevity will not be repeated later.

Lingula is characteristic of nearshore faunas with low species diversity (Raup and Stanley, 1971, p. 208). The Dinwoody fauna as a group are great in numbers but of a low species diversity. Rather than facies control, Kummel (1957, p. 438) and Perry and Chatterton (1979, p. 307) have attributed this homogeneity to the widespread extinction of many large groups of animals at the end of the Permian. Even though Lingula can tolerate wide salinity ranges (McKerrow, 1978, p. 99) the associated fauna describe a picture of near-normal marine salinity in shallow, well-aerated waters; ammonites were stenohaline, and inhabited clear waters; echinoderms are sensitive to reductions in salinity to less than 27 parts per thousand (McKerrow, 1978, p. 238); some brachiopods can tolerate muddy conditions but generally they cannot live in areas where mud and silt are being actively deposited (Rudwick, 1970, p. 159). Bivalves are both infaunal and epifaunal and generally prefer clear waters with less than 20% suspended sediment (Raup and Stanley, 1971, p. 205). More specifically, Eumorphotis (described in Little Sheep Creek), Pseudomonitis, Myalina, and Bellerophon (described elsewhere in the Dinwoody) are reef-foreslope and reef-top animals which live on a variety of substrates from shell beds (Myalina) to lime mud (Bellerophon; McKerrow, 1978, p. 161, 167, 187). These reef-like bioherms generally form in well aerated, turbulent, and clear waters.

Micritization of carbonate substrates such as molluscan shell fragments is performed by endolithic algae (Bathurst, 1975, p. 381) that are dominantly photosynthetic and must live within the photic zone (Kobluk and Risk, 1977, p. 1070). Assuming clear waters, photosynthesis can take place to a depth of about 100 m (Raup and Stanley, 1971, p. 199). Colonization and establishment of the algae on the substrates require quiet conditions at, or just below, the sediment-water interface and have been observed today at depths up to 20 m (Kobluk and Risk, 1977, p. 1074, 1080); however, Kobluk and Risk do suggest that the process can take place at greater depths.

Assuming that the bivalves occupied bioherms, upon expiration of the animal, the shells would necessarily have been removed to a quiet environment to be micritized. Bioherm-like masses were not observed at Little Sheep Creek, but loosely packed pelecypod shell fragments in a microspar matrix (sparse biomicrosparite) were. This facies has been described (Wilson, 1975, p. 65) from shelf environments with open circulation at, or just below, wave base. The shells in the more densely packed limestone (packed biopseudosparite) are fragmentary, never whole, but sometimes merely disarticulated. This microfacies is associated with shoal environments in agitated waters within wave base; the shell fragments collect in swales in proximity to shoals (Wilson, 1975, p. 65). The swales may channel tidal or channel currents in which the shells would be winnowed of fine material, redistributed, and laid down as traction deposits.

Intraclasts of micrite, silty micrite, and argillaceous micrite are derived from nearby areas of quiet low energy sedimentation. Steinkerns are developed by sediment sifting into gastropod chambers as the mollusc is rolled by gentle currents over very fine-grained sediments (Bathurst, 1975, p. 364). The sediment is cemented and forms a mold of the gastropod upon dissolution of the shell wall. Horizontal burrows and bioturbation in the calcareous siltstones also indicate slow sedimentation and relatively quiet conditions in subtidal environments (Perry and Chatterton, 1979, p. 308). The packed pelsparite with an admixture of coated grains indicates currents strong enough to agitate the sediment but incapable of destroying the fragile fecal pellets. These conditions occur in restricted marine shoals (Wilson, 1975, p. 67) affected by gentle currents.

The faunal evidence indicates that the Dinwoody sea was shallow, with slow sedimentation in an environment which had bioherm-like mounds built up to wave base which formed protective barriers for back-reef quiet water sedimentation. Currents served to transport and concentrate lenses of shelly debris in an area dominated by quartz silt sedimentation.

The picture is not complete, however, because bioherm-like masses and extensive mudstones were not observed at Little Sheep Creek. Furthermore, bedding characteristics indicate a much more active and energetic process of sedimentation in the area. Sedimentation in ancient epeiric and miogeosynclinal seas is commonly interpreted to have been "shallow water" shelf-like deposition based solely on faunal evidence. Authors such as Shaw, and Mazzullo and Friedman (cited in Klein and Ryer, 1978, p. 1050) have previously stated that epeiric seas were tideless, and that only the shelf margins were influenced by tides. Klein (1977, p. 1-7) and Klein and Ryer (1978, p. 1050-1051), citing oceanographic studies of sedimentation in Recent shelf seas, and using ancient analogues, concluded that shallow seas were subjected to wide tidal ranges with the prevailing depositional process controlled by a tidal current system. There is a direct correlation between shelf width and tidal influences; wide shelves enhance tidal ranges and tidal current intensity.

Bedding characteristics and sedimentary structures in the Dinwoody Formation (Table 2) indicate deposition of the sediments by tractive currents. The fauna indicate shallow water sedimentation and the great areal extent of the siltstone-limestone facies of the Dinwoody indicates broad equable shallow shelf sedimentation. The equable shallow marine conditions of sedimentation with probable normal salinities over such a large region indicate that there was adequate circulation of the water to maintain these conditions. It is conceivable, therefore, that sedimentation was controlled by tidal processes during Dinwoody time.

Repeated beds, each about 25 cm thick, of calcareous siltstone that show a modest fining upward coincident with a change in bedding from bedded siltstone with horizontal laminations and ripple cross-lamina-

TABLE 2. Dinwoody Formation: Sedimentary Features and Environmental Associations

Sedimentary Features	Environmental Associations	References
Horizontal laminations	Sedimentation of suspension clouds in current velocities below genesis of ripples	Reineck and Singh, 1975
Ripple laminations	Migration of asymmetrical current ripples moving in a uniform direction over a sandy surface; shallow shelf (low flow regime)	Wilson, 1975; Reineck and Singh, 1975
Trough cross-laminations	Scour-and-fill formed in moderately strong currents common in outer seaward parts of shelves; or tidal or longshore currents	Wilson, 1975; Reineck and Singh, 1975
Bioturbation; horizontal burrows	Slow sedimentation, thorough homogenization of sediment by burrowing organisms	Pettijohn and Potter, 1964; Wilson, 1970
Vertical burrows	Escape structures in areas of high rates of sedimentation; or infaunal dwellings	Pettijohn and Potter, 1964; Reineck and Singh, 1975
Clastics and carbonates in well segregated beds	Open platform when associated with stenohaline organisms	Wilson, 1970
Mud (silt) deposits that contain layers of coarse silt or fine sand	Shelf sedimentation containing storm sand layers	Reineck and Singh, 1975
Swell lag deposits	Concentrations of whole shells convex up on bedding planes as the result of passing storm swells	Brenner and Davies, 1973

tions upward to horizontal laminations with shaly partings. The contact between the shaly siltstone below and the bedded siltstone above is abrupt. A thin (2 cm) bed of fossiliferous limestone is interbedded with the siltstone. The graded bedding and the changes in sedimentary structures indicate a waning current regime for each cycle. This type of sedimentation is typical of shelf deposits (Wilson, 1975, p. 77) and can be attributed to storm surges, or to periodic waning tidal currents.

The most commonly observed origin for millimeter-thick horizontal laminations in fine-grained sandstones and siltstones is by the wash-and-backwash action of waves on a beachfront. The products of such wave action are thin laminations that show a reverse grading of quartz and heavy mineral grains. Quartz-rich and heavy mineral-rich laminations have been described in the calcareous siltstone and silty limestone of the Dinwoody Formation; however, the reverse grading cannot be demonstrated. Reineck and Singh (1975, p. 106) do suggest that horizontal laminations can be developed in shoal and nearshore environments by sedimentation of suspension clouds of silts and muds in current velocities less than necessary to generate ripples. That there is no direct evidence for sub-aerial exposure of Dinwoody sediments at Little Sheep Creek, and that the limestones are also laminated helps to support the hypothesis of a total subaqueous environment for the Dinwoody.

The production of suspension clouds may be enigmatic. They could be produced by particularly strong spring tides or by storms. Both sources are likely. The greater volume of water and the stronger currents associated with spring tides may be enough to resuspend the finer-grained sediments from the shelf bottom (Wilson, 1975, p. 81).

Storm-generated sedimentation is indicated by whole disarticulated bivalve shells found convex up on bedding planes of some calcareous siltstones. Similar whole shell cocquinoid mudstones were described by Brenner and Davies (1973, p. 1692-1694) from Late Jurassic strata of Montana and Wyoming. The unbroken shells and fine-grained sediments are indicative of quiet environments generally below wave base. The concentration of the shells along bedding planes indicates short, intermittent periods in which the shells are concentrated without being trqnsported great distances. These deposits are called swell lags by Brenner and Davies and are caused by high amplitude marine swells. A pressure gradient created by the passing swell lifts the sediment from the sea floor; the heavier particles settle out first as the swell passes. Repeated passage of swells will result in many similar cycles producing concentrations of shell-lag deposits. Powers and Kinsman (cited in Brenner and Davies, 1973, p. 1694) described swell lags in Recent North Atlantic sediments at depths of 15-45 m. Swells of such magnitude are probably storm-generated.

Lithoclasts of micrite in the intramicrosparite are probable tidal channel lag deposits (Wilson, 1975, p. 82; Reineck and Singh, 1975, p. 366; Klein, 1977, Table I). Intraclasts of silty micrite, micrite, and argillaceous micrite present in varying quantities in nearly all the rocks examined petrographically indicate a possible closely associated tidal flat environment. Distances are relative and cannot be quantified. The clasts are commonly submacroscopic and well rounded, indicating transport beyond the tidal flat onto the shelf.

Sedimentation during Dinwoody time in the Little Sheep Creek area occurred in a shallow, clear, well oxygenated shelf environment of depths probably less than 50 m. Paleogeographic reconstruction indicates that during the Triassic, southwestern Montana lay within the latitude of the northeastern trade winds, north of the paleo-equator, and that the climate was hot and arid (Peterson, 1978). The craton at that time was a low-lying feature; the sediments were fine-grained, and the rate of sedimentation was low. The source of the quartz silt and very fine-grained sediment in the Dinwoody may be from the slow discharge of rivers from an arid hinterland and may also be wind-transported. The few paleocurrent directions available from the Thaynes and Dinwoody Formations indicate both a northwest and a southeast transport direction with dominant transport to the southeast.

A regression occurred at the close of Dinwoody time with an advance of the shoreline westward concurrent with a change in sedimentation from a shallow tide-dominated shelf to the intertidal system of the Woodside Formation. The proximity of the shoreline at the demise of Dinwoody sedimentation is marked by the interbeds of fine grain sandstone and dolomite in the calcareous siltstone.

The Woodside Formation

The Woodside Formation was named by Boutwell (1907, p. 446) from exposures in Woodside Gulch in the Park City mining district of north-central Utah. At the type section the Woodside Formation consists of 1,180 feet (360 m) of fine-grained, dark red shale with local fine-grained sandstone and light colored greenish gray shales. Mansfield (*in* Moritz, 1951, p. 1791) extended the Woodside terminology into Idaho to include all of the sedimentary rocks between the Phosphoria Formation and the basal, *Meekoceras*, zone of the Thaynes Formation.

As defined, Mansfield's Woodside included beds correlative with the Dinwoody Formation of the Wind River Range, Wyoming. In 1942 Newell and Kummel (p. 945-947) restricted the Woodside strata in southeastern Idaho to units which bear a lithologic similarity to the type Woodside. In western Wyoming, southeastern Idaho, and southwestern Montana the Woodside forms distinctive red beds that separate the Dinwoody and the Thaynes Formations (Newell and Kummel, 1942, p. 942). The lithology of the type section remains relatively constant throughout the area of distribution of the Woodside Formation.

Moritz (1951, p. 1791) found the best exposures of the Woodside in southwestern Montana to be in the Centennial and Gravelly Ranges. Exposures less than 200 feet (61 m) thick were found in Little Water Canyon and Little Sheep Creek in the Tendoy Range. Kummel (1954, p. 171) described the Woodside as consisting of only "several thin, intercalated red shale and siltstone beds that represent feather-edges of tongues of the Woodside Formation." Scholten et al. (1955, p. 366-367) mapped less than 125 feet (38 m) of red, maroon, and gray shales and siltstones of the Woodside Formation in the Tendoy Range. Klecker (1980, p. 79) mapped, as the Woodside Formation, 61 m of "silty, algal-laminated dolomites, and dolomitic limestones interbedded with thin-bedded and laminated calcareous siltstones and very fine-grained sandstones, with subordinate fossiliferous limestone."

Regional Distribution and Correlation

The Woodside Formation is thickest at the type section and thins markedly in all directions. From the type area the Woodside can be traced northeast into Wyoming, north into Idaho and southwestern Montana, and into northeastern Utah. The Woodside Formation intertongues with the Dinwoody Formation along its western limit of exposure in an arcuate belt trending southward from southwestern Montana, along the Idaho-Wyoming border and west across northern Utah. West of the type area the amounts of red shales and siltstones of the Woodside decrease abruptly as the formation intertongues with the Dinwoody. The upper Dinwoody is equivalent to the lower Woodside at the type locality (Kummel, 1954, p. 168, 171).

The Woodside can be traced into western Wyoming where it grades laterally into the Chugwater Formation. In central Wyoming the Dinwoody is overlain by red shales and siltstones of the Red Peak Member of the Chugwater Formation. This member has the same lithologic character as the Woodside but it is equivalent only to the upper half of the type Woodside. South and east of the type area, in the eastern Uinta Mountains, the Woodside grades down into red beds which are commonly included in the Moenkopi Formation (Kummel, 1954, p. 171). As the Thaynes Formation thins eastward it brings the Woodside into contact with the Ankareh Formation. In the Uinta Mountains the Ankareh and Woodside Formations consist of red shales and siltstones; as the Thaynes pinches out the demarcation between the Ankareh and Woodside Formations becomes obscure. Thomas and Krueger (1946, p. 1266-1268) chose to refer the red beds to the Woodside Formation where the Thaynes was absent in the Uinta Mountains.

In extreme southwestern Montana the Woodside is conformably overlain by the Thaynes Formation. However, the upper Thaynes is missing; the contact with the overlying Jurassic and Cretaceous beds being an erosional unconformity. North and east the Thaynes pinches out bringing the Woodside directly in contact with the overlying upper Mesozoic strata. In the Greenstone Mountain area of the Pioneer Mountains north of Dillon, the Woodside is overlain by the Cretaceous Kootenai Formation (Sharp, 1970, p. 58). Northeast in the Madison Range, the Woodside is missing entirely and the Dinwoody Formation is overlain by the Jurassic Ellis Group (White, 1974, p. 63).

Thickness, Lithology, and Distribution

The Woodside Formation at Little Sheep Creek is generally poorly exposed, yet is one of the most noticeable units in the area. This presents an enigma, but one that is justified because the siltstones which form the bulk of the formation weather to form pale yellow and red soils. The distribution of the Woodside is depicted in Figure 14.

The formation is entirely conformable with the Dinwoody Formation below, and the Thaynes above. Both contacts are transitional and represent changes in sea level during the Woodside time. These changes are reflected in subtle lateral shifts in facies observable in the Little Sheep Creek area which which are also represented regionally. Where exposed the contact between the Woodside and Dinwoody Formations is taken to be the change in rock type from the "chocolate brown" calcareous siltstones of the Dinwoody to a predominance of laminated siltstones and carbonates but include fenestral and laminated limestones and dolomite. The contact with the overlying Thaynes Formation is drawn at the base of a thick ledge-forming sequence of calcareous siltstones with intercalated thin beds of limestone.

It should be clear that because the contacts are transitional,

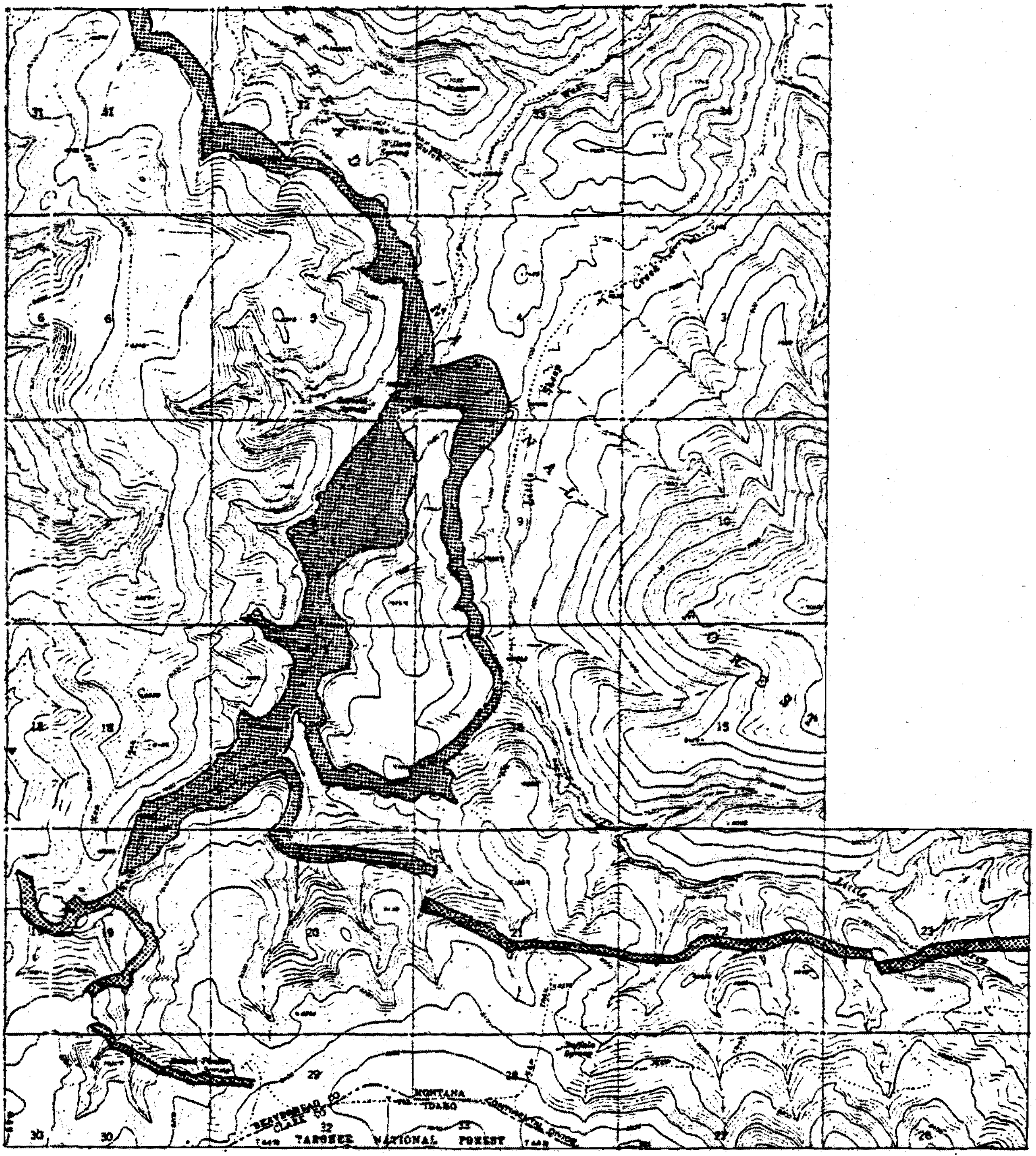


Figure 14. Distribution map of the Woodside Formation.

facies typical of one formation were found a short distance above or below the contacts in the adjacent formation. Where exposures were very limited, the position and nature of the contact was often difficult to locate. The best exposures of the Woodside-Thaynes contact can be observed on the left side of the valley of the West Fork of Little Sheep Creek in Sections 17, 18 and 19, and on the right side of the valley in the NE¼, Section 8, T. 15 S., R. 9 W. Exposures of the Dinwoody-Woodside contact are located in the SE¼, SW¼, Section 17, and in the W½, NE¼, Section 19, T. 15 S., R. 9 W.

The Woodside Formation is 40 m thick at Little Sheep Creek and is made up of fissile, very thin-bedded and laminated siltstones with minor very fine-grained sandstone, interbedded with thin beds of limestone, silty limestone, and dolomite.

A noticeable feature of the Woodside is the red splash of color formed by the weathered siltstones. Colors of freshly exposed siltstone are quite variable. Specifically, the colors are white (N 9), very light gray (N 8), yellowish gray (5 Y 8/1), grayish yellowish orange (10 YR 7/6), grayish yellow green (5 6 Y 7/2), pale to moderate brown (5 YR 5/2 - 4/4), and moderate reddish brown (10 R 4/6). The light colored rocks weather to similar hues and light greenish gray (5 GY 8/1) whereas the darker colored rocks weather greenish gray (5 GY 6/1) and moderate yellowish brown (10 YR 5/4). The siltstones are predominantly calcareous but local exposures are non-calcareous.

Petrographic analyses of several samples of the siltstone were made and two representative examples were selected for discussion: a calcareous siltstone and an intraclastic fossiliferous calcareous siltstone. Modal analysis of the calcareous siltstone provided the following information:

Monocrystalline quartz (19% strained, 81% unstrained)	43%
Calcite as iron oxide stained spar cement	49
Calcite (detrital)	3
Others, in decreasing order: muscovite, hematite, plagioclase, phosphatic shell fragments, tourmaline (?), and zircon	3
Porosity	2
	100%

83% of the detrital quartz is 20-40 microns in diameter, 13% less than 10 microns

Classification: Well-sorted calcareous quartz siltstone

The detrital grains are in framework support. Many of the quartz grains and some collophane pellets have irregular serrated boundaries with the adjacent calcite, indicating replacement by the latter. The detrital calcite grains are well-rounded, coarse silt-sized grains outlined by a dust ring, with rhombic or irregular syntaxial overgrowths.

Laminations in the siltstone range from 250 to 500 microns in thickness and are the result of alternations in couplets of coarse and fine silt. The coarse silt layers consist of mixtures of angular quartz silt, detrital calcite, and collophane pellets. The fine silt layers consist of a greater concentration of quartz, fewer detrital calcite grains and a concentration of muscovite and heavy minerals. This type of graded bedding is typical of intertidal flats and beaches and can be attributed to deposition from waning tidal currents (Reineck and Singh, 1975, p. 104).

The heavy minerals are a mature assemblage of stable detrital grains that are fine silt-size and rounded, indicating a period of long sediment transport, or a source of reworked sediment. The ultimate source of the detritus was from acid igneous rocks (zircon, tourmaline, muscovite, plagioclase) and metamorphic terrain (strained quartz). The genesis of sedimentary phosphates is discussed in a succeeding section on the Thaynes Formation.

Additional observations in siltstone thin-sections are rare pelecypod shell fragments with micrite rims recrystallized to columnar sparry calcite, micrite intraclasts, ooliths, and detrital calcite with isopachous rims of bladed calcite. The micrite intraclasts and ooliths make up about three percent of one lamina and provide two interesting observations. The first is a lithoclast that is well rounded and penetrated by an angular quartz grain, indicating that the intraclast was soft when deposited. The second is that the nucleus of an oolith consists of a micrite clast that has traces of algal borings (Scholle, 1978, p. 127) which are restricted to the nucleus and do not extend into the concentric layers of micrite around the nucleus. The nucleus was probably an intraclast of micrite which had been temporarily laid down in an environment quiet enough to be colonized by boring algae (Kobluk and Risk, 1977, p. 1080). A change in the environment, caused by a shift of sea level or a storm, may have redistributed the intraclast into a shoaling environment during which it was rounded, and around which calcite accreted (Wilson, 1975, p. 66).

The fossiliferous intraclastic calcareous siltstone contains echinoid plates and spines, pseudopunctate brachiopods, Lingula shell fragments, pelecypods, gastropods, rare disarticulated ostracods, collophane pellets, and intraclasts in a hematite-stained calcite cement.

The echinoid spines contain collophane as intraparticle pore cement. Collophane was observed adhering to one side of a punctate brachiopod shell. Both examples represent micro-reducing diagenetic environments conducive to the precipitation of collophane (Pettijohn, 1975, p. 428). Collophane also occurs as minor cement.

The quartz silt shows variations in the degree of packing and is either floating in the calcite matrix or is in framework support. The matrix is an argillaceous microspar that is stained yellow with limonite. Several of the molluscs and one large interparticle pore are filled with clear calcite spar in two forms (Figure 15). The first is a microspar lining one part of the pore space. This represents a geopetal structure with an orientation consistent with the orientation of the rock in outcrop. The microspar represents neomorphism of micrite that filtered into the sheltered pore space after deposition of the sediment. The interparticle pore may be a burrow, or it may have been created by solution of an allochem. The second type of calcite is a clear equant spar that has occluded most of the porosity. Mollusc shells either still exhibit their primary microstructures or have been completely replaced with spar.

That the iron oxide is restricted to the matrix and is excluded from the pore-filling cement indicates that the iron oxide staining is a diagenetic product and occurred before cementation of the rock and recrystallization of the micrite. An extended discussion of the source of iron oxide cement is included in the section on the Thaynes Formation lithology.

The fossils are fragmented but otherwise not very worn. The admixture of micrite, silt, and various sizes of fossil fragments provides

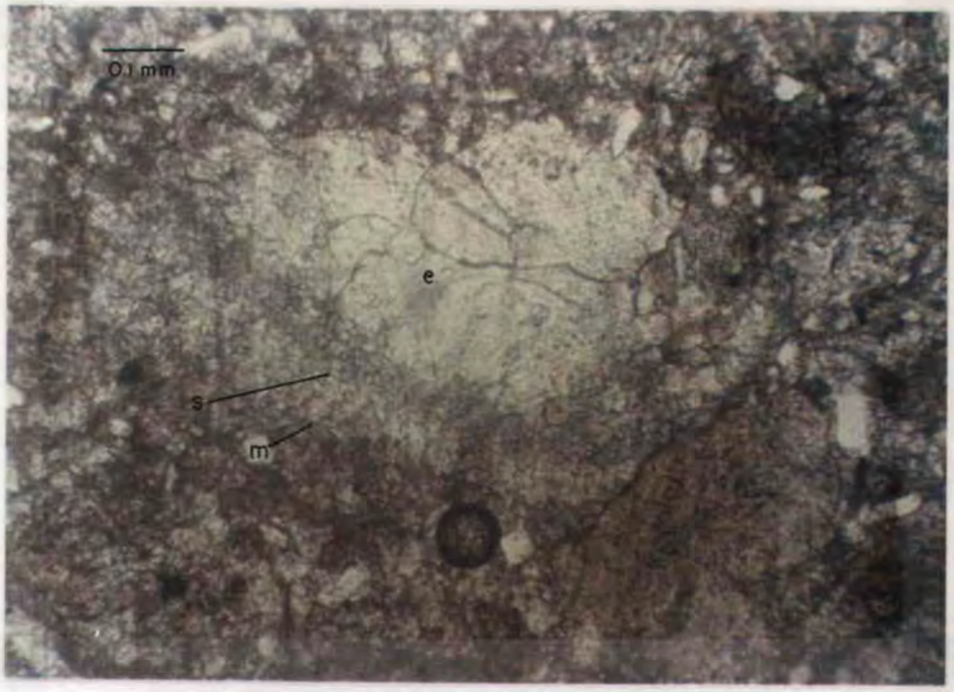


Figure 15. Cementation in calcareous siltstone, Woodside Formation. Photomicrograph showing a layer of micrite (m) lining bottom of pore, scalenohedral spar (s) normal to lower pore wall, equant spar (e) filling the pore. Sample Kmb 79-12a, NE¼, NW¼, SW¼, Section 17, T. 15 S., R. 9 W. Plane light.

a picture of a very poorly sorted sediment that was deposited quickly before winnowing of the finer sediment could occur. The intraclasts are angular to rounded fragments of micrite, argillaceous micrite, and fine-grained calcareous siltstone. Restricted circulation on a shelf, or a tidal flat environment, is indicated in this rock by the poorly sorted mixed fauna, including ostracods, and the silty terrigenous component with the hematite stain (Wilson, 1970, p. 232; Wright, 1973, p. 30).

The carbonate beds of the Woodside Formation are generally lenticular thin-bedded, or laminated limestone, silty limestone, intraclastic conglomerate, and dolomite; they rarely exceed a half-meter in thickness. The carbonates are light colored rocks variably grayish yellow (5 Y 8/4), very light gray (N 8), or pale yellowish brown (10 YR 6/2) on fresh surfaces, and grayish orange (10 YR 7/4), greenish gray (5 GY 6/1), and pale yellowish brown (10 YR 6/2), on weathered surfaces.

The limestones can be classified as mixed, packed oolitic and bioclastic sparites, oolitic intramicrosparites, and fenestral algal laminated silty micrite. The dolomite is a finely crystalline spar. Ghosts of shelly debris outlined by micrite rims indicate that the carbonate was originally a well-washed pelecypod limestone very similar to biomicrosparites described elsewhere from the Triassic formations.

The outcrops of dolomite at Little Sheep Creek are small, local, and thus incomplete; the origin of the dolomite, therefore, is difficult to assess. The arid climate of the region postulated by Peterson (1978) and evident in the correlative Chugwater, Ankareh, and Moenkopi Formations does provide an environment for the evaporative reflux model of Deffeyes et al. (1965) and the development of dolomite crusts in supratidal environments described by Shinn et al. (1965). Dolomite rhombs do occur sporadically in the limestones and siltstones of the Woodside Formation; in the field the slow reaction of some siltstones to dilute hydrochloric acid was interpreted to be the result of dolomitization. That the dolomite is not restricted to a particular lithology helps to rule out the origin of the dolomite as a supratidal crust (Shinn et al., 1965). The reflux models of Deffeyes et al. (1956) and Shinn et al. (1965) are developed around the creation of hypersaline solutions in a supratidal environment with high magnesium-calcium ratios because of excessive evaporation in a tropical climate. These supratidal environments are coincident with the development of extensive gypsum and anhydrite beds. Evaporite beds, or solution breccias, were not discovered in the Woodside Formation in Little Sheep Creek.

Cementation of the Woodside sediments indicates one, and possibly two, episodes of diagenesis in a freshwater phreatic zone. It is probable that dolomitization occurred according to the "dolomite bus" model of Folk and Land (1975). They have shown that dolomitization of carbonate rocks will take place with low magnesium-calcite ratios when crystallization is slow and not inhibited by foreign ions. These conditions can be met by the dilution of normal marine waters with fresh water subaerially or in the mixing area between freshwater and marine phreatic zones. Dilution of the marine waters does not alter the magnesium-calcium ratio, but it will remove inhibiting ions and allow slow crystallization of dolomite.

The dolomite bus model has been used by Dunham and Olson (1978) to help explain the geographical restriction of dolomites in Paleozoic strata in the Cordilleran miogeosyncline of Nevada. Migration of a coastline lens of fresh groundwater because of seasonal changes in precipitation, or because of changes in sea level, could be the mechanism needed for the mixing of freshwater and marine phreatic solutions. It is possible, therefore, that dolomitization in the Woodside Formation occurred according to the dolomite bus model of Folk and Land (1975).

The algal limestone is thinly laminated with fenestrae aligned parallel to laminar planes. The laminae consist of alternations of layers which contain approximately 40% quartz silt and micrite and layers of micrite with one to three percent quartz silt. The fenestrae account for about a 25% primary porosity, over half of which has since been occluded by equant calcite spar cement. The micrite has undergone partial aggrading neomorphism to microspar (Figure 16). The uneven, crenulated laminae, the fenestrae, and the close association of interbedded intertidal mudstones indicate the most likely origin of the laminae to be by the entrapment of micrite by mucilaginous algal mats in an intertidal environment. The algal limestones can be classified according to Logan et al. (1964, p. 78) as spaced laterally linked hemispheroids (LLH-S) with a microstructure of close laterally linked hemispheroids (LLH-C).

The oolitic intramicrosparites (Figure 17) contain variable quantities of micrite intraclasts, pellets, pelletal and oolitic grapestone, and ooliths. The intraclasts of one sample consist of angular fragments of micrite ranging in size from 0.5 to 14 mm in the longest dimensions, and thin rip-ups of micaceous clay.

A modal analysis of one intraclastic conglomerate shows that the rock consists of 25% rounded intraclasts, two-thirds of which are silty calcareous claystone, with the remaining third being claystone and cal-

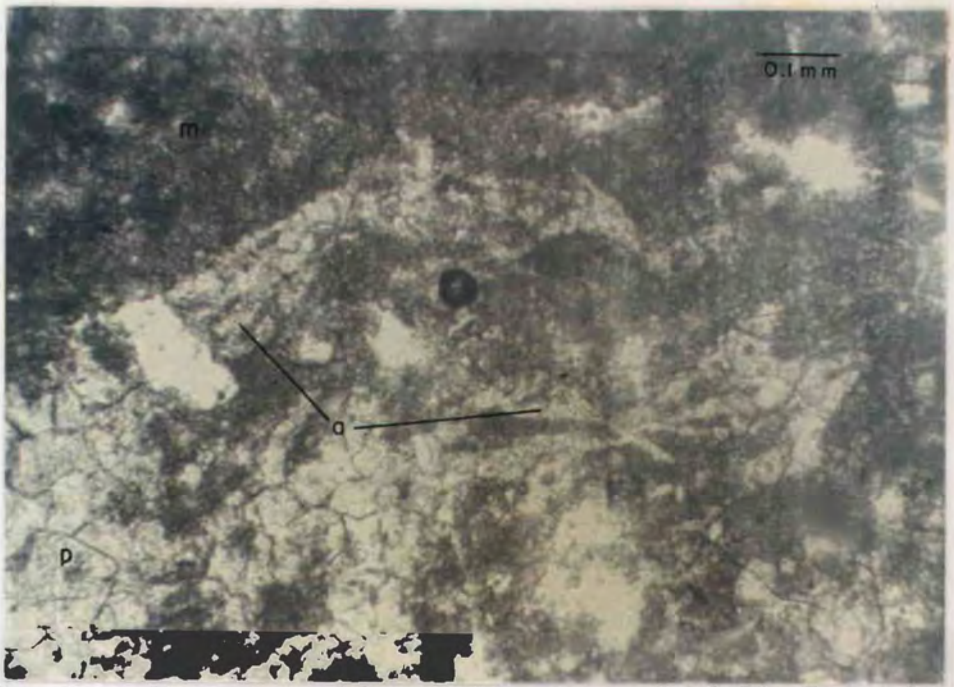


Figure 16. Aggrading neomorphism in limestone, Woodside Formation. Photomicrograph showing clotted appearance of microspar (m), pseudospar (p), and acicular neomorphic spar (a) as single crystals and in a radiating pattern. Sample Kmb 79-141, NW¼, NE¼, Section 30, T. 15 S., R. 9 W. Plane light.

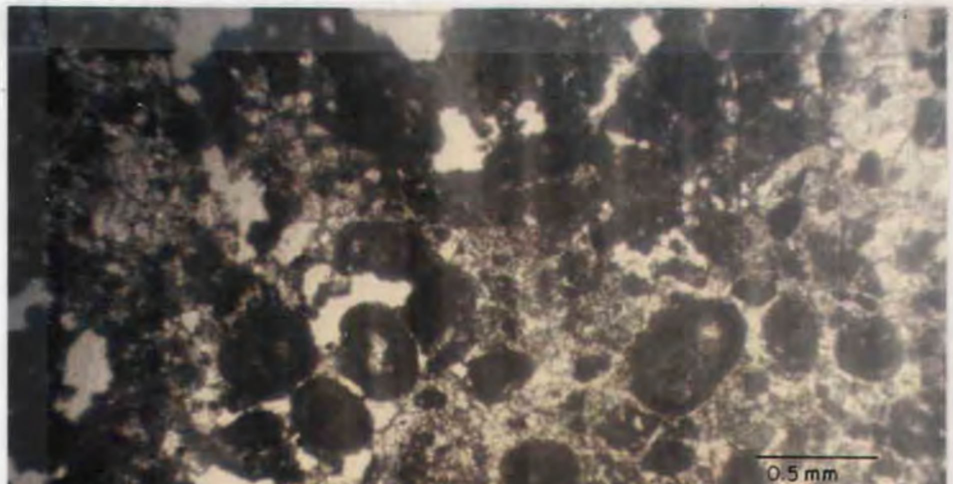


Figure 17. Photomicrograph, oolitic intramicrosparite of the Woodside Formation. Sample 79-91, NW½, NE¼, Section 17, T. 15 S., R. 9 W. Plane light.

careous siltstone. Fourteen percent of the rock consists of angular quartz silt, heavy minerals, and a few shell fragments. Porosity accounts for two percent of the rock. The matrix consists of quartz silt mixed with micrite, the latter having been subjected to aggrading neomorphism. Interparticle porosity has been reduced by calcite spar cement. Less than one percent of the rock consists of silica cement.

The intraclasts are possibly chips of mud derived from tidal flats, levees, and algal mats that are swept away by tidal currents and accumulate in tidal channels and pot holes as lag deposits (Wilson, 1975, p. 82).

The mixed oolitic and biotic sparites have varying mixtures of both components in framework support with micrite intraclasts making up less than five percent of the allochems (Figure 18). The shells are thin fragments less than 2 mm in length. Some shells have not been broken, but merely disarticulated, and are as much as 6 mm long. All of the shelly debris is aligned parallel to bedding. Some shells have been totally micritized by filamentous algae (Bathurst, 1975, p. 381-387). Others have only a micrite rim, but the original aragonite has been replaced with calcite spar. A majority of the shell fragments are most likely those of pelecypods with an admixture of angular Lingula fragments. Compaction of the sediment in places has broken the thin micritized shell walls.

The ooliths are well sorted, circular to elliptical in cross section, with an average diameter of 0.5 mm. The nuclei are either shelly fragments or micrite intraclasts.

The porosity of the oobiosparites is about 15% and can be classified as interparticle, intraparticle, and moldic (Choquette and Pray,

Figure 18. Photomicrograph, mixed oolitic biosparite of the Woodside Formation. Sample 79-114, NE¼, Section 5, T. 15 S., R. 9 W. Plane light.

1972). The cement is a finely crystalline equant spar that fills in the interstices between the allochems. Larger equant calcite spar forms interparticle cement in some of the ooliths.

The ooliths and pelecypod shell fragments were deposited in a shoal environment in agitated waters as well-washed, moderately well sorted sediment (Wilson, 1975, p. 66). Cementation has involved the precipitation of calcite spar as an interparticle pore cement coincident with dissolution and replacement of the aragonite shells. Selective dissolution of many ooliths created a substantial porosity which has been partially filled in with larger equant, and scalenohedral spar. Careful observation shows that the interparticle pores are lined with a fringe of isopachous blades of calcite oriented perpendicular to the pore wall. The rest of the pore is partially or completely filled with larger equant calcite crystals (Figure 18).

Calcite diagenesis in the mixed oolitic biosparites is indicative of the sequence of diagenetic events observed in the Triassic strata. A summary of the possible diagenetic sequence involves:

1. Micritization of allochems by boring algae (Bathurst, 1975, p. 381-387) prior to, or during, final sedimentation.

2. Precipitation of isopachous aragonite (?) and calcite normal to interparticle pore walls in a marine phreatic zone (Bathurst, 1975, p. 364, 453, 499; Longman, 1980, p. 464).

3. Freshwater phreatic solution of aragonite shells and replacement with scalenohedral and equant calcite spar; syntaxial overgrowth on single calcite grains; occlusion of porosity with equant calcite spar; neomorphic recrystallization of micrite to microspar or pseudospar (Bathurst, 1975, p. 325-333; Longman, 1980, p. 473-477; Buchbinder and Friedman, 1980, p. 401).

4. Formation of solution porosity (vugs) that transcend allochemical boundaries in a freshwater vadose zone (Longman, 1980, p. 468-473).

5. Filling in of vugs with equant calcite spar in the freshwater phreatic zone (Longman, 1980, p. 481).

A bed-by-bed description of the Woodside Formation could not be made because of inadequate exposures. Descriptions of several partial sections were made, however, and one is illustrated in Figure 19.

Fossils and Age

No fossils were identified from the Woodside strata. Petrographic examination of the rocks reveals an assemblage very similar to the underlying Dinwoody and the overlying Thaynes Formations except for the addition to ostracods.

The Woodside Formation is Early Triassic (Scythian) in age.

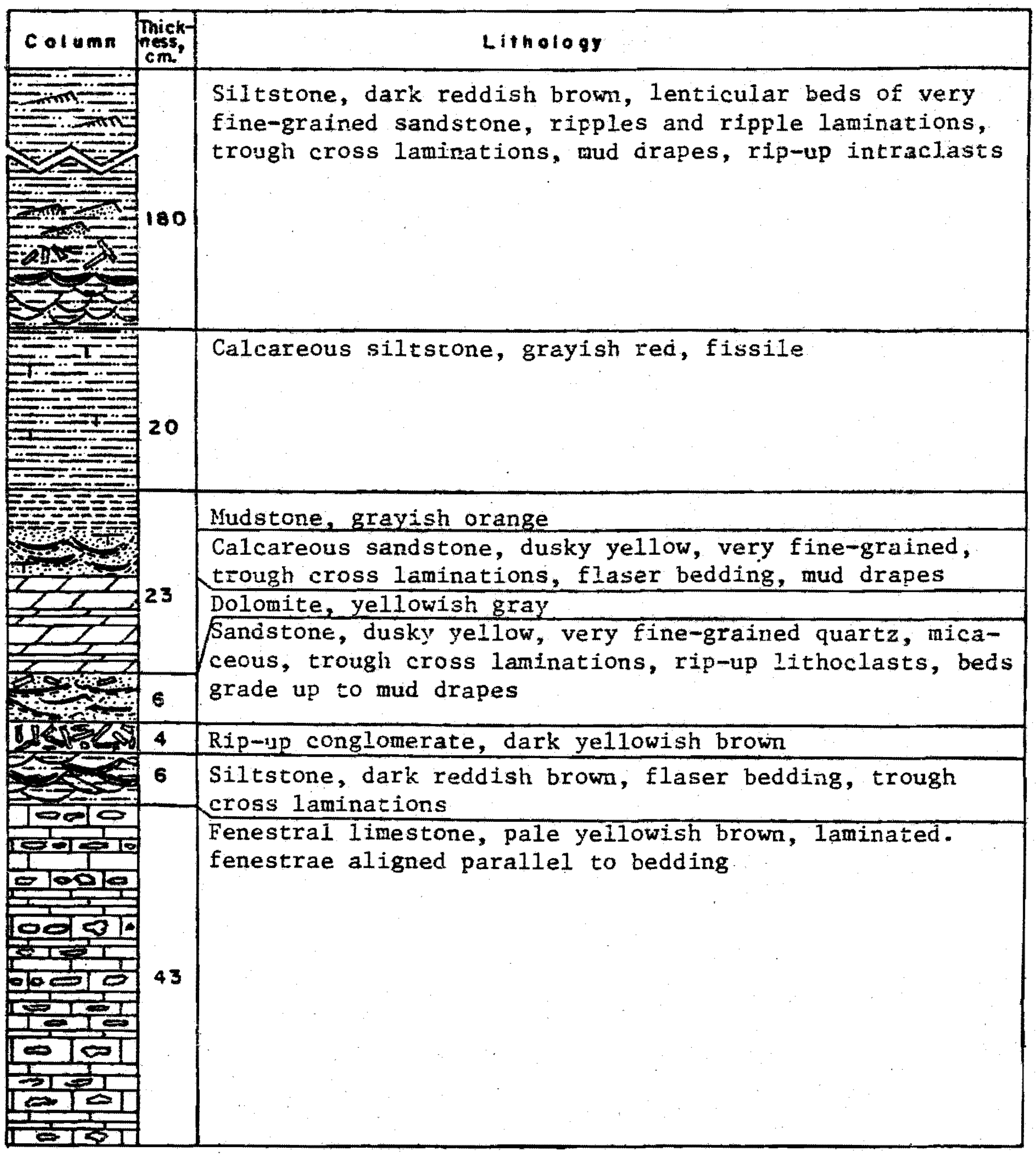
Environments of Deposition

Ginsburg (1975, p. 92) summarizes four categories of sedimentary structures which are diagnostic of siliciclastic tidal flat deposits:

1. Rapid reversals of depositing currents: herringbone cross-stratification, reactiviation surfaces.
2. Small scale alternations in slack water and strong current: flaser and lenticular bedding, clay (mud) drapes over ripples and sand waves.
3. Intermittent subaerial exposure: desiccation cracks, rain imprints, plant roots, evaporites, peat or coal beds, tracks of vertebrates (algal laminations).
4. Alternating erosion and deposition: channels, scours, mud chips, winnowed sand beds and lenses.

(Parentheses are mine)

Table 3 lists the sedimentary features and environmental associations of the Woodside Formation. Comparison of the elements in the table with Ginsburg's diagnostic sedimentary structures for tidal flats indicates that the Woodside sediments were deposited in a tidal



Thickness, cm.	Lithology
180	Siltstone, dark reddish brown, lenticular beds of very fine-grained sandstone, ripples and ripple laminations, trough cross laminations, mud drapes, rip-up intraclasts
20	Calcareous siltstone, grayish red, fissile
23	Mudstone, grayish orange
	Calcareous sandstone, dusky yellow, very fine-grained, trough cross laminations, flaser bedding, mud drapes
	Dolomite, yellowish gray
6	Sandstone, dusky yellow, very fine-grained quartz, micaceous, trough cross laminations, rip-up lithoclasts, beds grade up to mud drapes
4	Rip-up conglomerate, dark yellowish brown
6	Siltstone, dark reddish brown, flaser bedding, trough cross laminations
43	Fenestral limestone, pale yellowish brown, laminated. fenestrae aligned parallel to bedding

Figure 19. A partial stratigraphic column representative of the Woodside Formation measured in a stream bank, east side of drainage, SE¼, Section 19, T. 15 S., R. 9 W.

TABLE 3. Woodside Formation: Sedimentary Features and Environmental Associations

Sedimentary Features	Environmental Associations	References
Horizontal laminae	Tidal current bedload transport	Klein, 1971; 1977; Kumar and Sanders, 1975
Primary current lineations	Upper flow regime when found in combination with horizontal bedding; common on beaches and produced by backwash	Reineck and Singh, 1975
Asymmetric ripples with ripple cross-laminae on smooth surfaces	Small-current ripple laminations	Reineck and Singh, 1975
Asymmetric ripples with ripple cross-laminae developed on irregular substrata	Wave-generated ripples created by reworking of non-cohesive sediments during emergence on a tidal flat	Reineck and Singh, 1975
Flat-topped current ripples	Late stage emergent ebb flow with sudden changes in flow directions at extremely shallow water depths	Klein, 1971
Herringbone cross-laminae	Reversing tidal current bedload transport	Klein, 1977
Horizontal laminae grading up into in-phase climbing ripples and then into migrating ripple laminae	Rapid reduction of sediment load and flow regime during late stage emergent ebb flow	Klein, 1971; Reineck and Singh, 1975
Mud drape on ripples; flaser bedding	Alternation of tidal current bedload sediment transport with mud suspension deposition during slack water periods	Klein, 1977

TABLE 3, Continued

Sedimentary Features	Environmental Associations	References
Interference ripples	Late stage emergence run-off producing changes in flow direction, or wind-generated ripples superimposed on current ripples	Reineck and Singh, 1975; Klein, 1977
Trough cross-laminations	Deposition from migrating small current and wave ripples	Reineck and Singh, 1975; Klein, 1977
Lenticular bedding; isolated very fine-grained sandstone in mudstones	Alternation of tidal current bedload transport with suspension settlement during slack water periods	Wilson, 1970; Reineck and Singh, 1975
Flat-pebble rip-up conglomerate	Tidal channel lag deposits filling channels and pot holes	Wilson, 1975
Fenestral algal limestone and stromatolites (LLH structure)	Developed on continuous mats in protected locations or reentrant bays and behind barrier islands. Fenestrae developed by desiccation and involve inundation and maximum exposure	Logan et al., 1964; Wilson, 1975; Woods and Brown, 1975
Angular micrite intraclastic grainstone with pellets	Middle to upper intertidal zone with moderate to extreme exposure and desiccation, i.e., natural levees, supratidal storm lag deposits	Wilson, 1975; Woods and Brown, 1975; Ginsburg and Hardie, 1977
Mudcracks	Subaerial exposure and desiccation of tidal flats	Klein, 1971; Wilson, 1975

TABLE 3, Continued

Sedimentary Features	Environmental Associations	References
Slump features and small scale fault planes	Emergence at low tides of tidal point bar deposits subject to very rapid deposition and high water content of fine-grained sediments	Reineck and Singh, 1975
Clastics and carbonates in well segragated beds	Restricted tidal platforms when associated with limited fauna or other tidal flat features	Wilson, 1970
Oolites	Shoal environment in agitated waters, originates through water movement on oolite shoals, beaches, and tidal bars	Wilson, 1975
Steinkernen	Gentle agitation of gastropods over fine-grained sediment in intertidal flats and lagoons. Gastropod chambers are filled with cemented sediment with subsequent loss of the shell wall	Bathurst, 1975

flat environment with offshore shallow water carbonate shoals and open shallow shelf environments.

The Thaynes Formation

Boutwell (1907, p. 448) first described and named the Thaynes Formation after Thaynes Canyon in the Park City mining district of north-central Utah. At the type locality the Thaynes consists of 1,190 feet (363 m) of limestone, calcareous and siliceous sandstone, and shale. Each rock type is represented repeatedly throughout the formation; however, Boutwell was able to group the rocks into two members dependent on the dominant lithology. They are a lower sandstone member and an upper limestone member separated by a middle red shale unit. The middle red shale unit persists eastward from the type area into the Uinta Mountains and was thought to represent a westward extending tongue of the Ankareh Formation.

The basic two-fold division of the Thaynes Formation persists throughout its regional distribution even though locally several authors have chosen to further subdivide the Thaynes. In 1916 Mansfield (in Kummel, 1954, p. 172) raised the Thaynes to group status near Fort Hall in east-central Idaho. At that locality the Thaynes is 5,525 feet (1,685 m) thick and was divided, in ascending order, into the Ross Fork Limestone, the Fort Hall Formation, and the Portneuf Limestone. Kummel recognized that the two lower units are local and distinctive only in the Fort Hall area, whereas the Portneuf Limestone can be traced into Wyoming as a member of the Thaynes Formation (Kummel, 1954, p. 172). The group status of the Thaynes was discarded by Kummel.

In the same paper Kummel (1954, p. 172-173) recognized seven major rock units in the Thaynes of southeastern Idaho. The seven divisions later were useful in helping to define the axis of the Cordilleran miogeosyncline during Early Triassic time (Kummel, 1957).

Moritz (1951, p. 1794) divided the Thaynes Formation in southwestern Montana into three members: upper and lower limestone members separated by a sandstone member. These three members are covered by the Snake River lava plain to the south and cannot be traced into Idaho (Kummel, 1960, p. 236).

Regional Distribution and Correlation

The Thaynes Formation is distributed over a wide area including northern Utah, east-central Nevada, eastern Idaho, western Wyoming, and southwestern Montana. The thickness and distribution of the Thaynes are very similar to the Dinwoody Formation except that the eastward extent of the Thaynes is limited to western Wyoming and it extends farther south into northern Utah (Figure 20; Kummel, 1954, p. 171; 1960, p. 235; Collinson and Hasenmueller, 1978, p. 177).

The Thaynes Formation intertongues with red beds of the Chugwater Formation in western Wyoming and with the red beds of the lower part of the Ankareh Formation in the western Uinta Mountains (Kummel, 1954, Figure 18; Reeside et al., 1957, p. 1479). A lack of fossils in the Triassic red beds of Wyoming, Utah, and Colorado make biostratigraphic correlations difficult. Stratigraphic positions of the rock units place the Thaynes Formation as a correlative of the Chugwater Formation of central Wyoming, the lower part of the Ankareh Formation in the western Uinta Mountains, the upper Moenkopi Formation in the eastern Uinta Mountains, southwestern Utah, and southern Nevada, the upper Lykins Formation

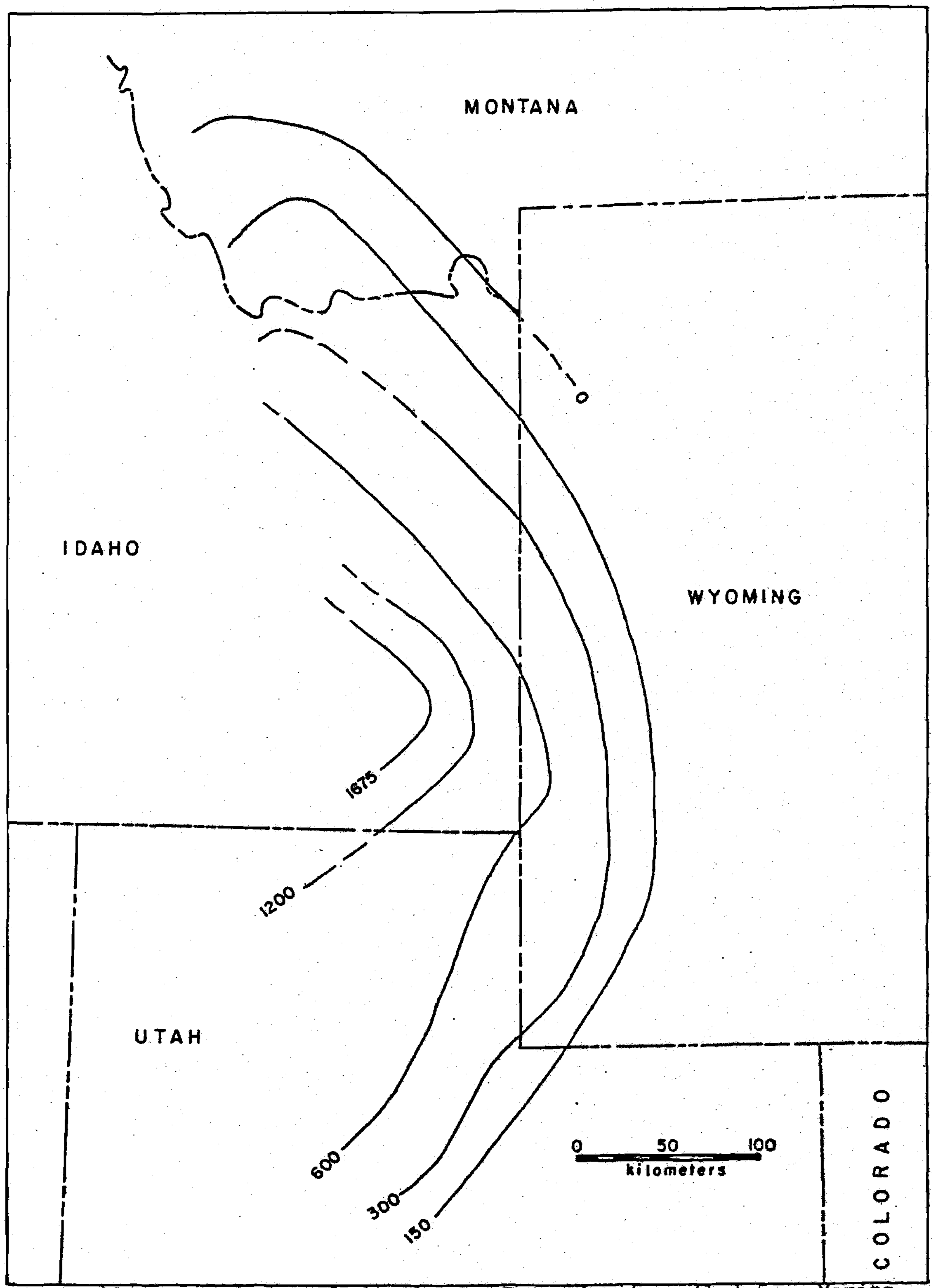


Figure 20. Isopach map of the Thaynes Formation (compiled from Moritz, 1951, Kummel, 1960, Collinson and Hasenmueller, 1978). Isopachs in meters.

of northeastern Colorado, and the upper part of the Spearfish Formation of South Dakota and eastern Montana (Moritz, 1951, Table I; Kummel, 1954, Figure 18; Reeside et al., 1957, p. 1479; Collinson and Hasenmueller, 1978, Figure 2).

Distribution and Topographic Expression

The Thaynes Formation is the most widespread Triassic formation in the thesis area. Ample exposures are provided by folding and thrusting of the unit in the southwestern part of the area and by Recent incising of valleys by the Middle and West Forks of Little Sheep Creek. The limestones and calcareous siltstones of the Thaynes weather to form discontinuous ledges and low cliffs on steep valley sides. Elsewhere (Figure 21) the Thaynes forms high ridges and spurs between drainage systems tributary to the Middle Fork.

Early Jurassic erosion removed the entire Triassic strata and parts of the underlying formations from most of western Montana (Scholten et al., 1955, p. 367). In southwestern Montana an indeterminable thickness of the upper Thaynes was removed so that the contact with the overlying Ellis Group, or, locally the Gypsum Spring Tongue of the Twin Creek Formation, is entirely unconformable. Locally, parts of the Thaynes have also been removed by pre-Beaverhead, by pre-Round Timber, and by Recent erosion.

The Thaynes Formation is also an incompetent unit involved in the multiple episodes of deformation that occurred in the area during the Laramide orogeny. The excellent exposures of the formation in Sections 19 and 20 in the southwestern part of the area are the result of axial thickening of the Thaynes along several folds, and a syntectonic high-

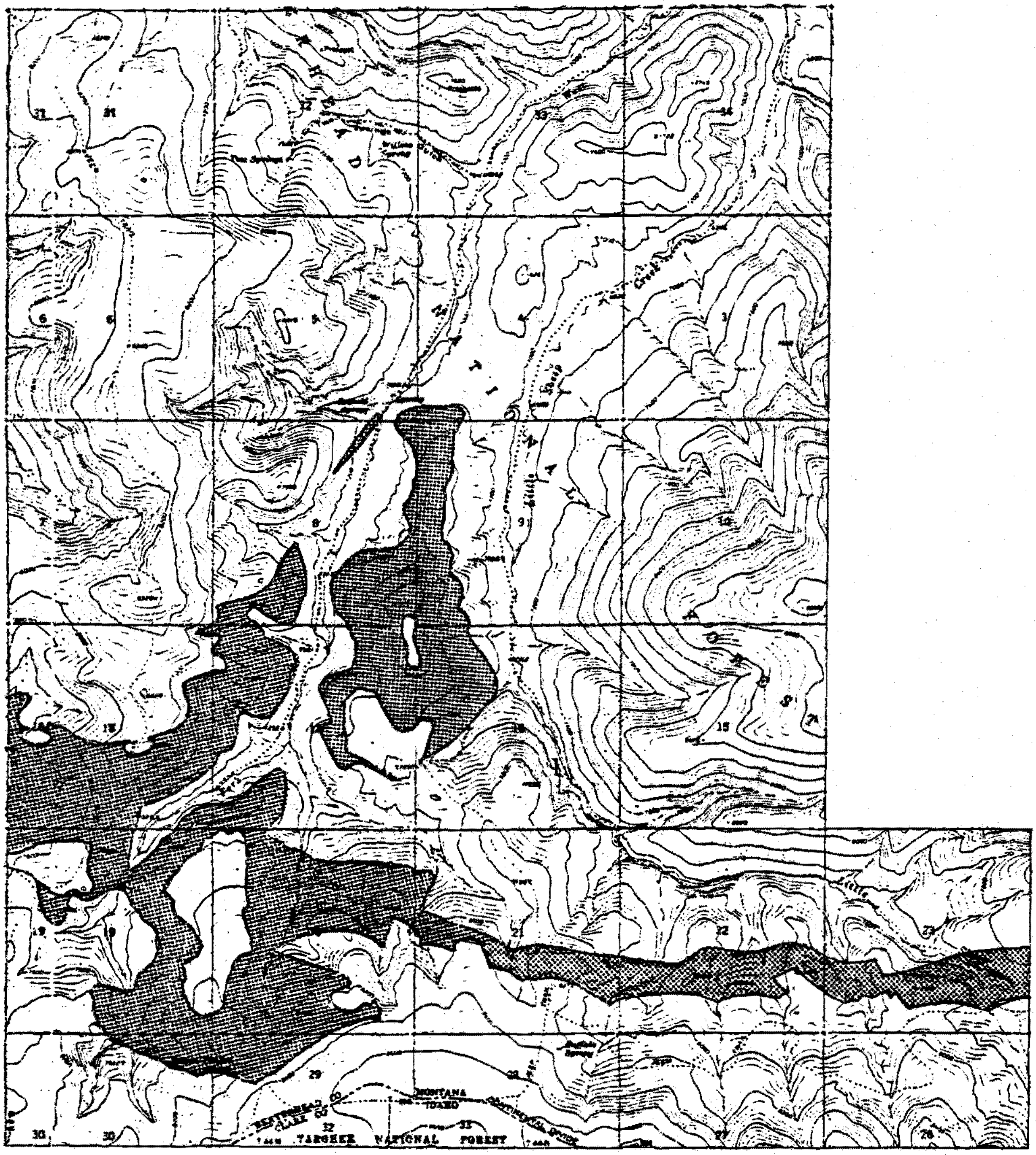


Figure 21. Distribution map of the Thaynes Formation.

angle reverse fault (Plate II; cross-section A-A').

A bed-by-bed description of the Thaynes Formation was not attempted in the field for the reasons stated above. A composite description is given in the following section.

Thickness and Lithology

The Thaynes Formation is unconformably overlain by the Gypsum Spring Tongue of the Twin Creek Formation, or the Beaverhead Formation, or the Round Timber Limestone Member of the Medicine Lodge Beds in the thesis area. As discussed in the section on the Woodside Formation, the Thaynes conformably overlies the Woodside. The thickness of the Thaynes Formation is approximately 207 m.

The three-fold subdivision of the Thaynes Formation described by Moritz (1951) and which was recognized by Scholten et al. (1955, p. 366) and Klecker (1980, p. 97) is not present with such apparent acuity in Little Sheep Creek. Instead, the Thaynes is made up of a sequence of interbedded limestones, silty limestones, calcareous siltstones, and very fine-grained sandstones. An approximation of the three members can be made, however, because the lower and upper thirds of the formation are predominantly limestone and silty limestone, whereas the middle third is mostly calcareous siltstone and silty limestone.

The limestone in the upper part of the Thaynes Formation is a sparse to packed biosparite that is usually pale yellowish brown (10 YR 6/2) on fresh surfaces with weathered outcrops typically light gray (N 7) or grayish orange (10 YR 7/4). The limestone is very thin-bedded and forms thin, resistant, step-like outcrops interbedded with laminated thin beds of silty limestone and calcareous siltstone. Etched out in high relief on weathered surfaces of the limestone are circular crinoid columnals, Pentacrinus, pelecypod shell fragments, and echinoid plates and spines. Nodular chert and discontinuous beds of chert up to 4 cm thick are found in the limestone of the middle and upper parts of the formation.

Microscopic examination of the limestone reveals that the echinoid plates and spines have been subjected to three diagenetic events: amorphous glauconite has developed in the intraparticle pores, syntaxial overgrowth of calcite has occurred, and the echinoid debris is being selectively replaced by chalcedony (Figures 22, 23, 24). Gastropods, not observed in hand sample, occur in thin section and some are filled with collophane (Figure 24); the shape of the chambers has been retained as steinkernen wherever the molluscan shell is no longer present. Pelecypods are disarticulated, but nearly whole, being as much as 7 mm long. They have micrite rims and the shelly material has been replaced with clear scalenohedral and equant sparry calcite.

Collophane is found in gastropod chambers as pellets with minor amounts of quartz silt and calcite inclusions, and as ooliths. Glauconite occurs as pellets and as intraparticle cement in echinoid debris.

Modal analysis of one sample of limestone provided the following composition:

Pelecypods	24%
Echinoids (plates and spines)	16
Brachiopods (including Lingula)	1
Collophane	2
Quartz silt	4
Silty limestone lithoclasts	1
Calcite spar	41
Chalcedony	2
Microspar	4
Porosity	5
	100%

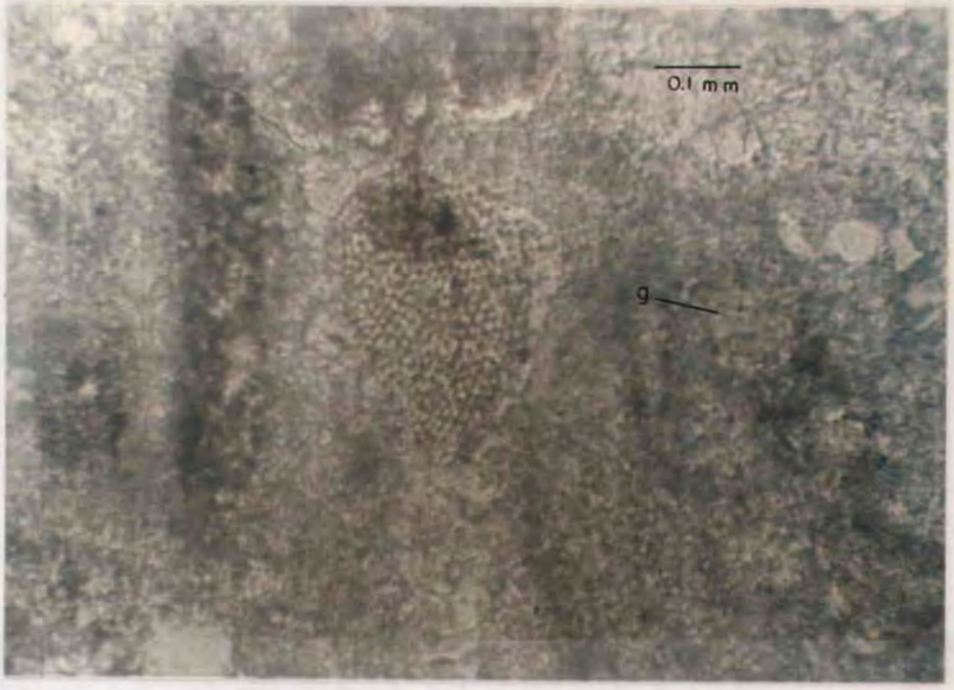


Figure 22. Photomicrograph of glauconite filling in intraparticle pores of an echinoid plate (e), glauconite (g) pellet on right. Sample Kmb 79-180, Thaynes Formation, SE¼, NW¼, Section 20, T. 15 S., R. 9 W.. Plane light.

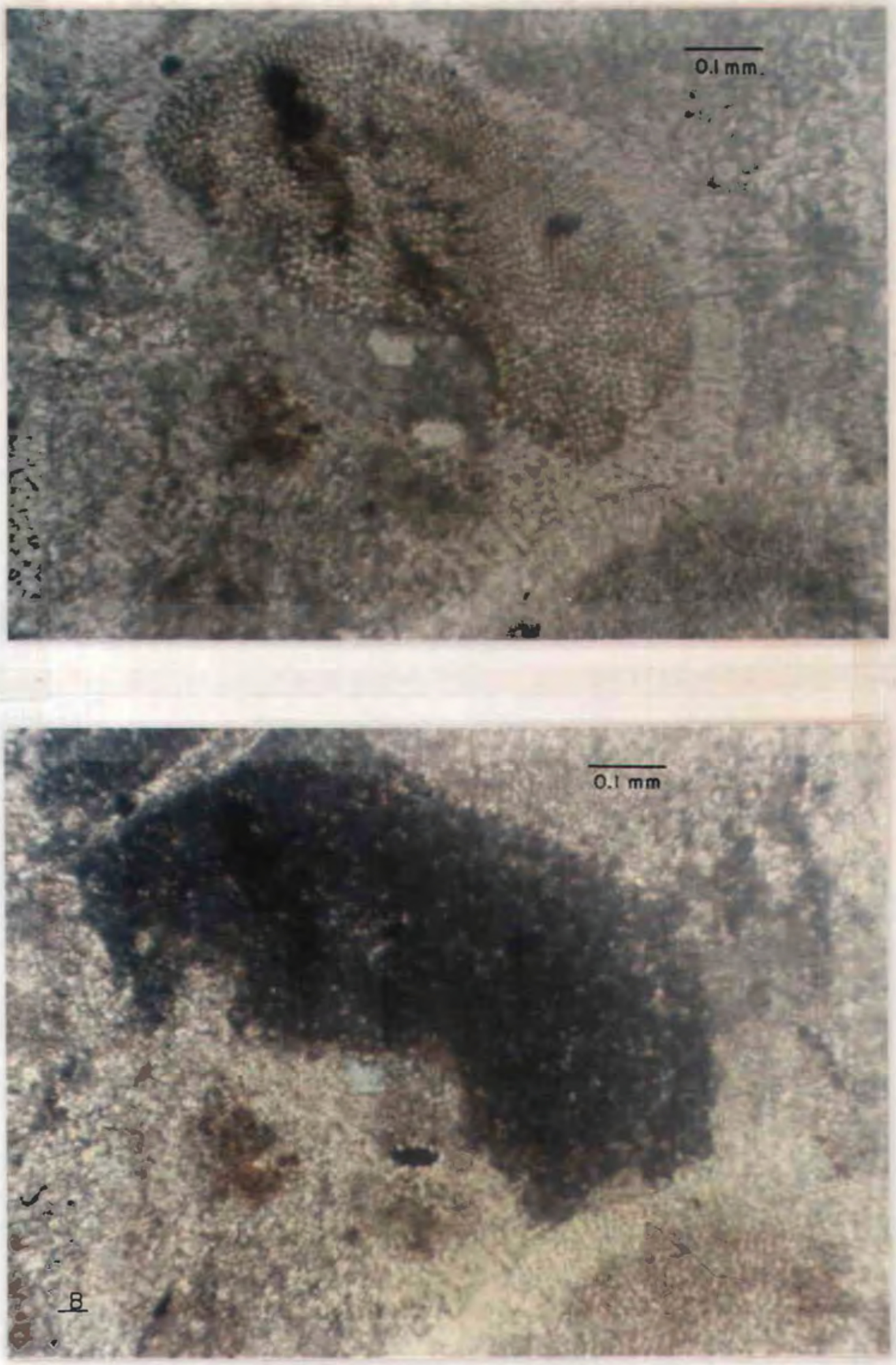


Figure 23 a. Photomicrograph showing syntaxial overgrowth of calcite on an echinoid plate. Sample Kmb 79-180, Thaynes Formation, SE¼, NW¼, Section 20, T. 15 S., R. 9 W. Plane light.

b. Same, crossed nicols.

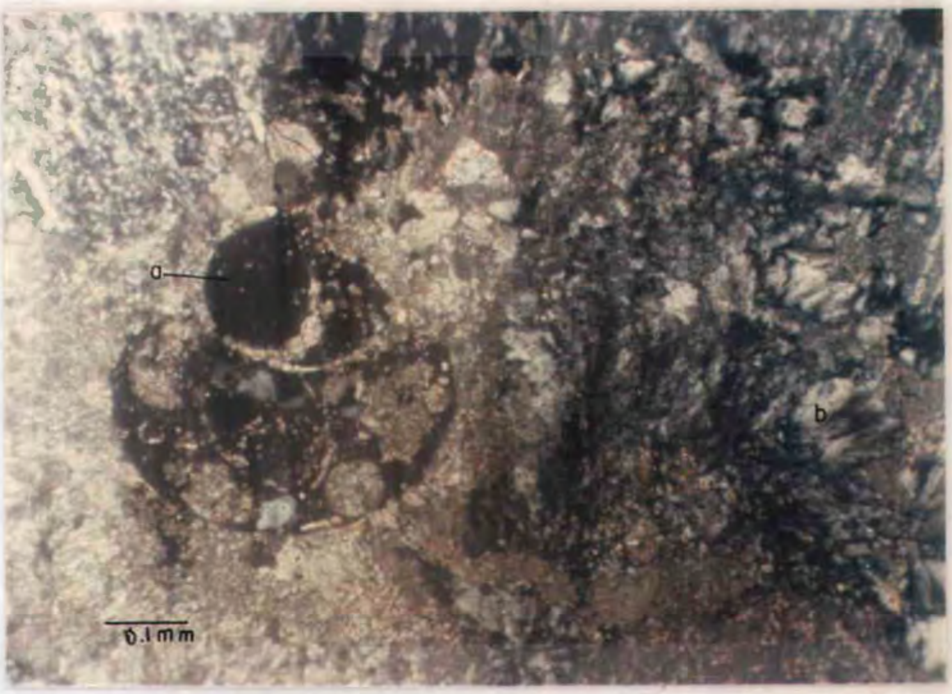


Figure 24. Photomicrograph of a) gastropod shell filled with collophane, quartz silt, and calcite, b) secondary silica cement as chalcedony. Sample Kmb 79-180, Thaynes Formation, SE¼, NW¼, Section 20, T. 15 S., R. 9 W. Crossed nicols.

The sparry calcite occurs as cement occluding intraparticle pores, interparticle voids, and in small fractures. Isopachous scalenohedral calcite spar is found oriented normal to pore walls. In completely occluded pores the scalenohedral spar meets across the center of the pore with a serrated suture line; larger pores are filled with clear equant spar. Calcite also occurs as syntaxial overgrowths on echinoid and crinoidal calcite crystals. The microspar appears to be neomorphosed micrite. In the biosparite it is not possible to determine if the finer crystalline spar is a cement, or a product of neomorphic recrystallization. The porosity generally occurs as a secondary solution of sparry calcite cement and is confined mostly to pelecypod and echinoid fragments.

The lower limestone member is a packed biomicrosparite with varying admixtures of intraclasts. The limestone is pinkish gray (5 YR 8/1), pale yellowish brown (10 YR 6/2) or pale grayish orange (10 YR 8/4) on fresh surfaces, and weathers moderate to dark yellowish brown (10 YR 6/2 - 5/4) and yellowish gray (5 Y 7/2). The biotics consist of whole disarticulated or broken pelecypod shells, gastropods, pseudopunctate brachiopods, <u>Lingula</u>, crinoid ossicles, and echinoid plates and spines. The pelecypod debris is generally aligned subparallel to bedding and forms most bioclastic material. The pelecypods also exhibit micrite rims and complete recrystallization of the original aragonite to calcite. Further aggrading neomorphism has caused the recrystallized pseudospar to transcend some micrite rims. Gastropods are less common, but where they occur they are either filled with silty micrite or with clear sparry calcite cement. Steinkernen of the silty micrite are found preserving the internal molds of the gastropods when the confining shell is no longer present. Micrite confined to the lower part of gastropod chambers forms a geopetal structure corresponding to the original orientation of the bedding. The echinoid debris occurs as single calcite crystals with syntaxial overgrowths. The intraclasts are rounded angular chips of micrite or argillaceous siltstone.

The limestones of the Thaynes are similar to those found in the Dinwoody Formation with at least one minor variation which occurs in the lower limestone of the Thaynes. It is the presence of hematite restricted mostly to the allochems and argillaceous siltstone intraclasts.

Figure 25 is a photomicrograph of a limestone found in the upper Woodside near the transitional contact with the Thaynes Formation. Even though not in the Thaynes, the faunal composition and texture are similar to the Thaynes limestone and are used here to illustrate two points. The first is the hematite cement previously mentioned; the second concerns the algal borings found in biotic-fragments and micrite intraclasts (Figure 26). The latter have been conveniently outlined by the hematite cement. The cement and the algal borings are excellent expositors of the history of this limestone and similar limestones found in the lower Thaynes.

The algal borings are restricted to intraclasts of microspar and calcareous siltstone, and to calcite fossil fragments (Figure 27). The hematite is found as intraparticle cement 1) in echinoid plates and spines, 2) in the pseudopunctae of brachiopod shell fragments, 3) in algal borings, 4) as a cement restricted to the intraclasts, and 5) disseminated to a limited extent as intergranular cement in the sparry calcite matrix.

T. R. Walker discusses the production of red beds in both modern

Figure 25. Photomicrograph of limestone in the Woodside Formation near the Woodside-Thaynes contact showing red hematite cement concentrated in intraparticle pores of echinoid plates (e), siltstone intraclasts (i), in calcite lithoclasts (c), and as local zones of interparticle cement in calcite spar (ic). Sample Kmb 79-106, NE¼, Section 8, T. 15 S., R. 9 W. Plane light.

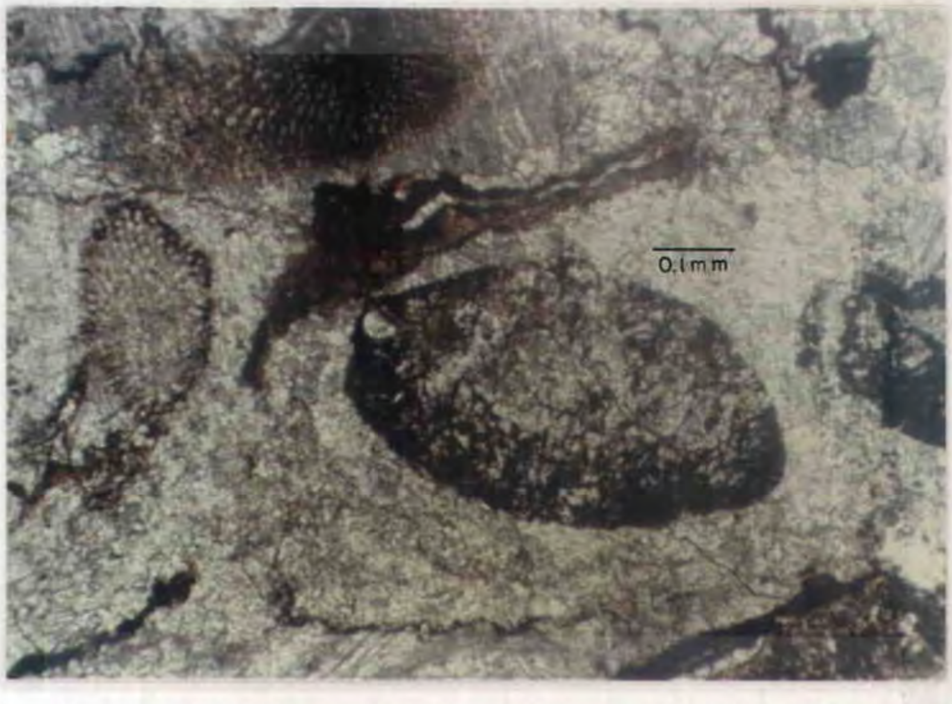


Figure 26. Greater magnification of calcite clast (c) in Figure 25 showing algal borings. Plane light.

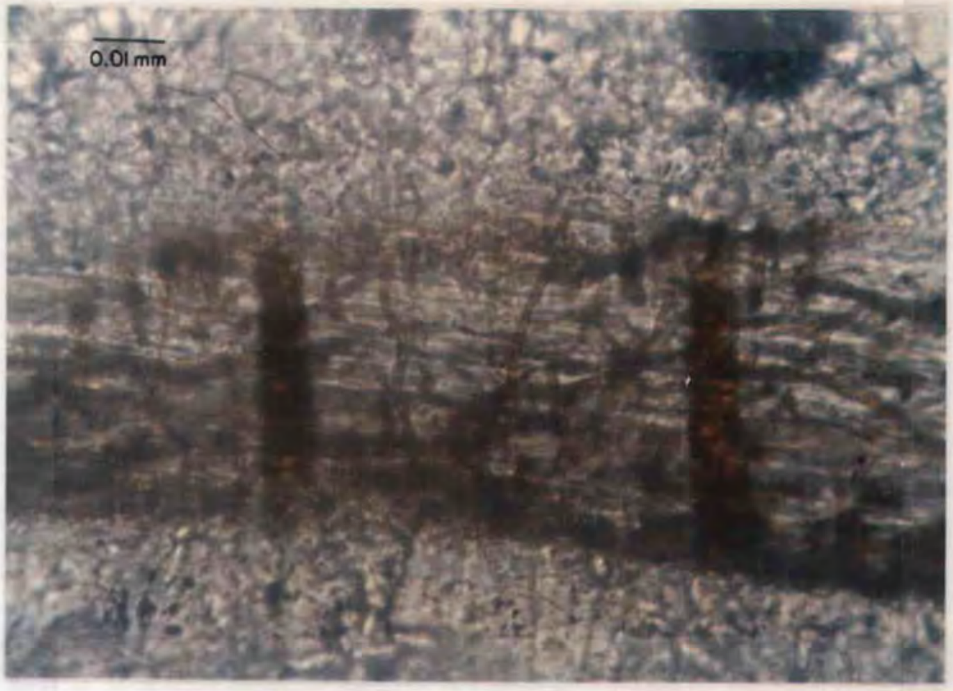


Figure 27. Photomicrograph of a punctate brachiopod showing original shell wall structure and algal borings. Punctae and borings are outlined by a faint lining of hematite (?) cement. Sample 79-38, Thaynes Formation, NW¼, NE¼, Section 19, T. 15 S., R. 9 W. Plane light.

tropical (1974) and modern arid (Walker and Honea, 1976) climates and in the geologic record. He found that the development of the red pigment, hematite, is not a function of climate; instead, it is a product of intrastratal alteration of iron-bearing minerals, under conditions in which the chemistry of the interstitial waters favors the formation and preservation of hematite (Walker, 1976, p. 279). The creation of red beds is dependent on two factors. The first is a source of iron, usually in the form of iron-bearing silicate minerals. The second is an environment unstable with respect to the silicates and yet stable for hematite. Hydrolysis of unstable source minerals releases ferrous ions which may be carried away in pore water solution or, if the eH and pH of the water lie in the stability field for ferric oxides, hematite, or a precursor oxide that ultimately ages to hematite, may be precipitated (Walker, 1976, p. 247). Such interstitial waters are generally oxygenated (positive eH) and slightly alkaline. The potential for the development of red beds can remain for a geologically long time (Walker, 1976, p. 245); the process can be reversed if there is a change in the eH or pH of the interstitial water.

The source of iron in such fine-grained sediments as those found in the Woodside Formation is from biotite, detrital magnetite, and clay minerals concentrated in the finer sediment fractions (Walker, 1969, p. 279). I postulate that one source of iron in the Thaynes limestone is glauconite, examples of which are found as unaltered pellets and intraparticle pore cement higher in the Thaynes Formation. Another source is the clay minerals in the argillaceous intraclasts.

The lining of algal borings with hematite indicates that the clastic material was colonized with endolithic algae prior to the precipitation of the glauconite. Most endolithic algae are photosynthetic and show a species zonation from the supratidal to the light compensation level (Kobluk and Risk, 1977, p. 1070) with the greatest varieties of species existing in a depth range of 0-25 m. Kobluk and Risk studied the rates of micritization of limestone substrates by endolithic algae in the warm tropical waters off the Bahamas. They found that complete infestation to a depth of 30 microns occurred within about 30 days. Micritization of the algal borings proceeded soon after vacation, and was completed within three years. The colonization of limestone substrates by endolithic algae requires relatively quiet periods of time of short duration (geologically) at shallow depths in order for either constructive or destructive micritization to occur. That the allochems in Figure 26 have been bored but not micritized indicates that the process was halted prior to completion of micritization.

It is possible, therefore, that the allochems in the Thaynes were deposited temporarily in a quiet environment, colonized with endolithic algae and then swept away to be deposited elsewhere in close association with clayey muds. That the biosparites typical of the Thaynes are interbedded in the intertidal red beds of the Woodside Formation indicates that the process was entirely possible. The mixture of bored allochems and non-bored angular chips of silty claystone in Figure 25 indicates that these elements were redistributed and deposited as sparse to packed biointramicrites, possibly as tidal channel lag deposits.

Aggrading neomorphism has recrystallized the micrite to microspar or finely crystalline pseudospar. Little is known of neomorphic recrystallization except that it is a wet transformation process requiring water (Bathurst, 1975, p. 476). The present state of neomorphism in the limestones was reached after the development of the red pigment. There are three lines of evidence for this. The first is the syntaxial overgrowth of calcite on echinoid plates and spines found in the lower and upper limestones of the Thaynes; the hematite is confined to the original calcite crystal and does not occur as inclusions within the overgrowth, or as a lining at the margins of the syntaxial crystal. The second is the random and patchy development of hematite as intergranular cement between spar crystals. Third is the varying degrees to which the hematite pigment has developed in the allochems. The color developed ranges from light brown (5 YR 5/6) and transparent to opaque and moderate brown (5 YR 4/4) in reflected light. This change in color may represent varying degrees to which oxidation has occurred, or a reversal of the process with reduction of the hematite and loss of pigment. The high degree to which the red pigment has developed leads me to believe that an oxidizing environment has prevailed in the limestone. Slight mobilization of ferrous ions and crystallization of hematite as intergranular cement has also occurred.

Diagenesis, therefore, has first involved precipitation of hematite as the red pigment and second, neomorphic recrystallization of the calcite in the limestone. Hematite is stable at a pH which is below the stability field for calcite so that the diagenetic sequences just described are physically and chemically possible (Pettijohn, 1978, Figure 11-27).

The nodular and discontinuous beds of chert as depicted in Figure 28 are secondary. There is a gradational boundary between the microcrystalline quartz and limestone, with small isolated blebs of chalcedony in the limestone. Relict outlines of fossils occur in the micro-

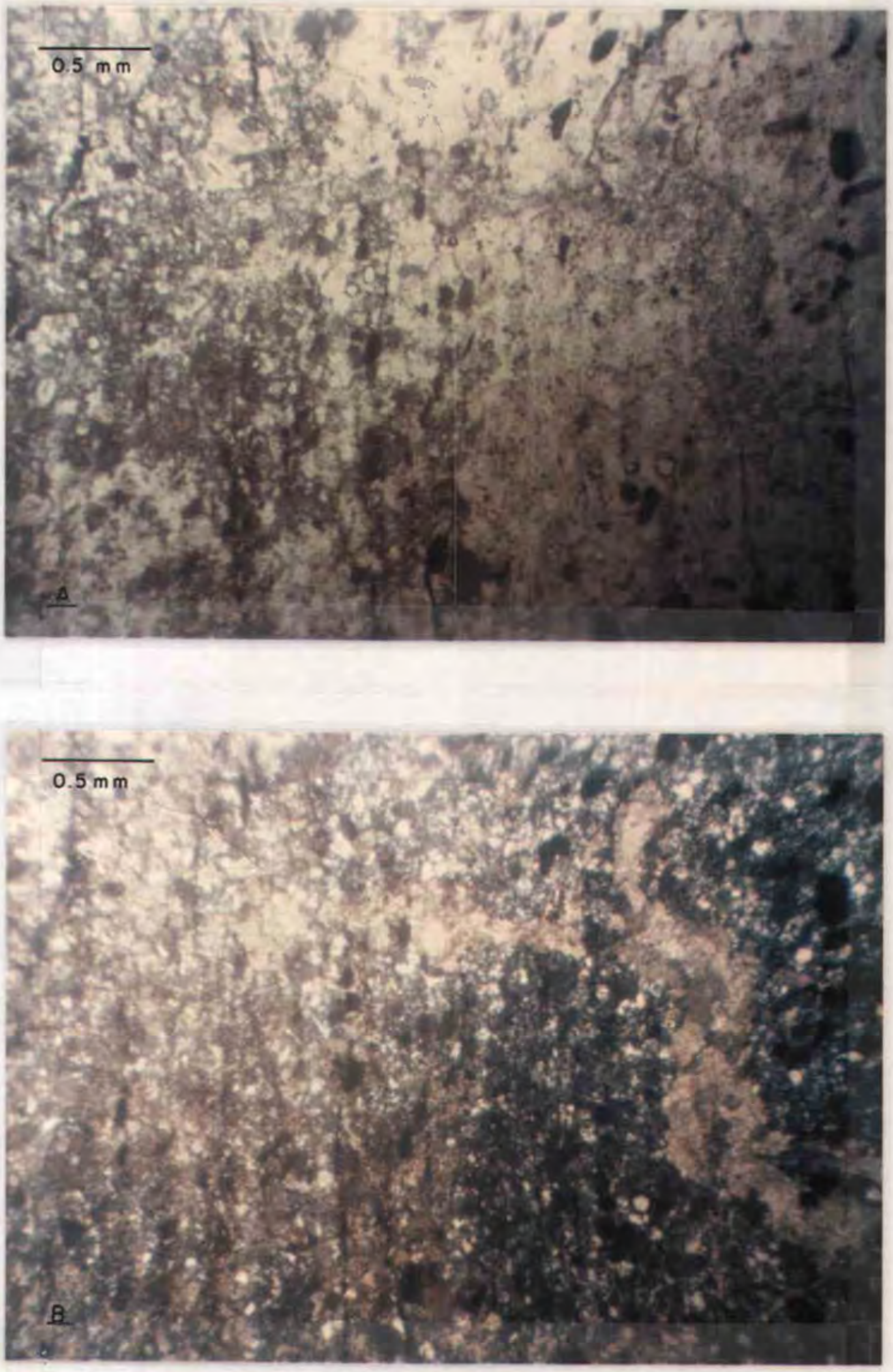


Figure 28. Photomicrograph showing secondary origin of chert in Thaynes limestones. a) Plane light. Note even distribution of quartz silt grains and argillaceous material across the slide. b) Crossed nicols showing chert on the right and chert disseminated in the calcite on the left. Sample 79-188, SE¼, NE¼, Section 20, T. 15 S., R. 9 W.

crystalline quartz. Furthermore, the distribution of fossils and hematite is the same across the boundary.

The siltstones and very fine-grained sandstones of the Thaynes Formation are pale yellowish brown (10 YR 6/2) and light grayish orange (10 YR 8/4) on fresh surfaces and weather moderate yellowish brown (10 YR 5/4), very pale orange (10 YR 8/2), or light grayish orange (10 YR 8/4). The texture and composition of these clastic rocks varies little between the two classes of sediment. Modal analysis of a calcareous siltstone provided the following parameters:

Monocrystalline unstrained quartz	27%
Monocrystalline strained quartz	11
Collophane	3
Heavy minerals	4
Fossils	1
Porosity	1
Calcite	53
	100%

Collophane occurs as pellets and as *Lingula* shell fragments. Heavy minerals are generally subangular to subrounded and are smaller than the quartz. They include zircon, tourmaline, staurolite (?) and oxidized opaque minerals, possibly pyrite or hematite. Glauconite pellets were not counted but are found in the siltstone and sandstones. The quartz grains are angular to subangular and are well sorted. The fossils consist of rare disarticulated brachiopods and some gastropods. Finely crystalline calcite, possibly detrital, forms only six percent of the sample. The rest of the calcite is a sparry calcite cement that is stained with a hematite pigment. The hematite also is concentrated in small areas giving the samples a finely mottled appearance with small pinpoints of reddish stain. Very small amounts of silica cement are associated with the hematite concentration and some silica was observed replacing calcite, consuming fossils, quartz grains, and incorporating the hematite stain.

Fine laminations of about 1 mm thickness or less of some of the siltstones are formed by concentrations of the heavy minerals, and alternations of quartz silt and micrite. The laminations have been developed by variations in traction current velocity, as opposed to the algally induced entrapment of suspended detrital grains described for the Woodside fenestral algal limestones.

One sample containing both siltstone and limestone helps to exemplify the intimate relationship between the interbedded units of the Thaynes and their origins as tractive-formed deposits (Figure 29). The siltstone shows a normal grading from a well sorted calcareous siltstone in framework support to a less well packed calcareous siltstone in matrix support. Also increasing upward is the abundance of muscovite and biotite, and hematite stain. The increase in hematite may reflect an increase in the amount of detrital clay and altered biotite concentrated in the upper part of the sediment as it settled out of suspension. The boundary between the two lithologies is marked by a wavy, irregular layer less than 2 mm thick of siltstone and claystone in a hematite cement. The overlying limestone is a biosparite that contains angular intraclasts of the hematite-bearing and argillaceous siltstone within 5 mm of the contact.

The normal grading in the siltstone with the thin layer of mud at the top of the sequence indicates possible settling of sediment from storm or tide-generated suspension clouds. That the lower third of the Thaynes Formation is predominantly a limestone alternating with thin beds of calcareous siltstone supports the hypothesis (Reineck and Singh,

Figure 29. Moderate yellowish brown calcareous siltstone overlain by light gray limestone above a local erosional surface. Lower Thaynes Formation, west side of the Middle Fork of Little Sheep Creek, SE¼, Section 18, T. 15 S., R. 9 W. Scale is 26 mm.

1975, p. 305) of storm-generated siliciclastic sedimentation on a carbonate shelf. The argillaceous rip-ups in the lower part of the overlying biosparite, and the well-washed nature of the limestone indicate deposition of the limestone by tractive currents. The currents may be tide-generated or storm wave-generated.

The silty limestones of the Thaynes represent a mixing of the two end member lithologies of siltstone and fossiliferous limestone.

Authigenic Shelf Minerals

Phosphate. The most common sedimentary phosphate is a carbonate fluorapatite collophane (Pettijohn, 1978, p. 420; Cook, 1976, p. 512). Rocks containing more than 19.5% phosphate (about 50% as apatite) are defined as phosphorites; if they contain more than 7.8% phosphate (about 20% as apatite) they are described as phosphatic (Cressman and Swanson, 1964, p. 282). The abundance of collophane in the Triassic rocks of the Little Sheep Creek area does not exceed 2%; however, because the mineral is ubiquitous in nearly all the samples examined petrographically, a brief discussion about phosphorites will be included here.

Collophane is an amorphous mass that is a pale yellow or pale brownish yellow in plain light. Between crossed nicols the mineral is isotropic, or sometimes it exhibits a weak birefringence. It occurs as phosphatic shell fragments, as intraparticle cement in echinoderm plates and spines, and as a cement in the more silty fractions of the rocks. It also occurs rarely as ooliths, and more commonly as pellets in which are often found silt-sized grains of quartz, calcite, and mica, indicating an accretionary origin for the pellets. Embayments along the edges of some collophane pellets by calcite indicate replacement of the phosphate by calcite.

Modern pelletal and nodular phosphorites are forming in normal marine waters, generally along the western margins of continental land masses between the 40° lines of latitude. Landward the water is commonly hypersaline, the climate is arid, and there is little terrigenous influx of sediment. The source of phosphorous is generally by the upwelling of cold nutrient-rich waters associated with divergent surface currents in the tradewind belts of the world. Paleogeographic reconstruction of ancient phosporite deposits using paleomagnetic data indicates that the geographical location of ancient phosphorites is analogous to those of the present.

The development of phosphates today occurs in bottom sediments of marine waters less than 500 m deep in which is developed a prolific biota. The phosphate-rich waters and the enhanced biological activity are a cause-and-effect relationship which enhances the production of phosphorites. The decomposition of organic matter at, or below, the sediment-water interface creates a low pH and a high alkalinity which causes the leaching of phosphate from the organic remains and its concentration in the interstitial waters. The interstitial concentration of phosphate can be 10 to 100 times greater than the overlying marine waters. It becomes apparent that even though the marine waters can be well oxygenated, the local diagenetic conditions under which phosphates can develop must be organic-rich and anoxic.

A pH of 7.1 to 7.8 and a high alkalinity are the requirements for the precipitation of apatite. Under these conditions the phosphatization of siliceous oozes, terrigenous clays, fossil fragments, coprolites, and calcareous sediments may occur. Further upgrading of the phosphate content and the development of pelletal and nodular phosphorites is made by the winnowing of the sediments by bottom currents, or by the lowering of sea levels.

To reiterate, the collophane in the Triassic rocks occurs as phosphatic shell fragments, as pellets and ooliths, and as a cement. Some inarticulate brachiopods incorporate phosphate into their shells as a physiological process (Pettijohn, 1978, p. 427). The silty and bioclastic nature of the Triassic sediments indicates a relatively high rate of terrigenous sediment influx, a prolific biota including fungi and photosynthetic algae, and active bottom currents. Blatt, Middleton, and Murray (1980, p. 588) suggest that even though terrigenous sediment input may be high, phosphates can accumulate on topographic highs such as bank tops, ridges, deep hills, and parts of the mainland shelf that are shallower than their surroundings.

The Triassic seas occupied approximately the same geographical position as the Late Paleozoic seas, especially those during the deposition of Phosphoria phosphorites. Even though phosphorites have not been described in the Thaynes, Woodside, or Dinwoody Formations, conditions did prevail in which local phosphorite deposits could develop and be distributed by currents as intraclastic debris in the shallow shelf sediments. The conditions included:

1) Shallow marine shelf located west of a continental land mass between the continent and a deeper miogeosynclinal basin.

2) Paleomagnetic latitude within 30° N, which today is found in the easterly tradewind belt, and the possible cause of divergent upwelling off of the coast (Peterson, 1978; Miller, 1966, p. 84).

3) Landward grading of the marine environment into hypersaline, arid, intertidal deposits with an arid hinterland represented by the lateral Triassic correlatives in Montana, Wyoming, southeastern Idaho and Utah of the Chugwater, Ankareh, Moenkopi and Woodside Formations.

4) Adequate biota indicating well oxygenated waters.

5) Broad shallow marine shelf.

Glauconite. Glauconite is a microcrystalline silicate mineral rich in iron and potassium (Pettijohn, 1975, p. 425). It generally occurs as roughly elliptical polylobate granules or pellets averaging 0.5 mm in diameter deposited in current agitated waters. It is a major constituent of greensands and has also been described in calcarenites.

The mineral is thought to form by the alteration of biotite or by the hydration of silica and the subsequent absorption of bases and loss of alumina (Pettijohn, 1975, p. 426-427). Sea water is essential for the development of glauconite and the process is facilitated by organic matter. Glauconite can, therefore, develop from fecal pellets, clayey substances filling cavities of foraminifera, radiolaria, and tests of other organisms, or from silicate minerals such as the feldspars, micas, and from volcanic glass (Pettijohn, 1975, p. 427).

Glauconite is forming today in open marine environments with slightly reducing conditions at depths of 18 to 730 m (Pettijohn, 1975, p. 426). Glauconite is also forming in well-oxygenated waters; however, it is being precipitated in the tests of marine organisms where the decay of organic material creates a very local micro-reducing environment conducive to the precipitation of the mineral. Glauconite is sometimes found as a continuous encrusting layer which "suggests a continuous reducing environment beneath perhaps a continuous gelatinous film known as pelogleas deposited on subaqueous surfaces today by green, or blue-green algae." (Bathurst, 1975, p. 411)

Fossils and Age

The Thaynes Formation in southeastern Idaho contains five ammonite zones which are, beginning with the oldest, the *Meekoceras*, *Anasibirites*, *Tirolites*, *Columbites*, and *Prohungarites* zones (Kummel, 1954, p. 184). The *Meekoceras* zone is the most widespread in the Thaynes and can be found marking the lower part of the formation in southeastern Idaho, northwestern Utah, western Wyoming, and southwestern Montana (Kummel, 1954, p. 171). Moritz (1951, p. 1802) found poorly preserved and unidentifiable ammonite molds in the lower Thaynes. These faunas are correlative with world-wide Smithian to Spathian (Early Triassic) stages. Perry and Chatterton (1979, p. 307-310) correlate the widespread occurrence of brachiopods *Rhynchonella triassicus* (Girty) and *Protogusarella smithi* with a distinctive conodont fauna found in the overthrust belt of Wyoming which bears *Platyvillosus asperatus* Clark, Sincavage, and Stone, *Neospathodus triangularis* (Bender), and *Neogondolella jubata* Sweet. These conodonts are indicative of an early Spathian age (Perry and Chatterton, 1979, p. 307; Collinson and Hasenmueller, 1978, Figure 1).

Fossils are poorly preserved in the Thaynes Formation. Petrographic analysis of the carbonates reveals an assemblage similar to the Dinwoody Formation. These include *Lingula*, pseudopunctate brachiopods, pelecypods, crinoid columnals, gastropods, and echinoid plates and spines. The most common fossils are the pelecypod shell fragments and the echinoderms. Very common to the upper limestones are the star-shaped ossicles of *Pentacrinus* columnals mixed with circular crinoid ossicles and echinoid spines. Molds and impressions of whole disarticulated pelecypod shells lying convex up can be found on weathered bedding planes of calcareous siltstones.

Besides Pentacrinus, Pugnoides (Scholten et al., 1955, Table 1), Eumorphotis and Rhynchonella (Klecker, 1980, p. 96) have been identified in the Thaynes Formation of southwestern Montana.

Environments of Deposition

Transgression of the Early Triassic sea at the close of Woodside time created environments similar to those found in the Dinwoody. Faunal and lithologic associations of the Thaynes indicate well aerated, shallow, clear waters of normal marine salinity over a broad shallow shelf dominated by tidal currents and storm surges.

Of the fauna described in the Thaynes, the ammonites are stenohaline requiring clear waters, Protogusarella and Rhynchonella have an affinity for colonizing silty argillaceous lime muds (Perry and Chatterton, 1979, p. 308) and Pentacrinus was a suspension feeder inhabiting shallow turbulent marine seas surrounding low islands or on submarine topographic highs (McKerrow, 1978, p. 225).

The Thaynes strata are generally laminated to thin-bedded; however, the limestone spar is only very thin- to thin-bedded and locally exhibits horizontal thin laminations on weathered surfaces. The silty limestones, calcareous siltstones, and very fine-grained sandstones are generally laminated and very thin-bedded with microstructures of asymmetric ripple cross-laminations, in-phase and in-drift climbing ripple laminations, scour-and-fill erosional features (Figure 30), primary current lineations, and horizontal and vertical burrows. The sedimentary features and their environmental associations are listed in Table 4. All of the structures are described from the lower third of the

Figure 30. Lower Thaynes Formation. a) Horizontal laminations grading up into in-drift climbing ripple laminations developed above a small erosional scour-and-fill structure in very fine-grained sandstone, SW¼, Section 17. b) Scour-and-fill in very fine-grained sandstone, SE¼, Section 18, T. 15 S., R. 9 W.

TABLE 4. Thaynes Formation: Sedimentary Features and Environmental Associations

Sedimentary Features	Environmental Associations	References
Horizontal laminations	Tidal current bedload transport; sedimentation from storm wave-generated suspension clouds; slow sedimentation in deeper part of shoreface below normal wave base	Reineck and Singh, 1975; Klein, 1971; 1977; Kumar and Sanders, 1975
Primary current lineations	Upper flow regime when found in combination with horizontal bedding; common on beaches and produced by backwash	Reineck and Singh, 1975
In-phase climbing ripple laminations; horizontal laminations grading up into in-drift climbing ripples	Rapid reduction of sediment load and flow regime during ebb tide or as storm layers	Reineck and Singh, 1975; Klien, 1971
Asymmetric ripple cross-laminations	Small current ripple laminations	Reineck and Singh, 1975
Trough cross-laminations	Deposition from migrating small current (and wave) ripples	Reineck and Singh, 1975; Wilson, 1975

Thaynes and, except for horizontal laminations and local trough cross laminations, are lacking in the upper two-thirds of the formation.

Deposition of the lower Thaynes sediments was dominated by tractive processes with short-lived influxes of siliciclastic sediments. As brought out in the discussion of Dinwoody environments, the shallow Triassic seas were dominated by tidal processes. There is no evidence for subaerial exposure during Thaynes time, so it can be postulated that the tractive currents were mostly tidal and were probably augmented by storms and longshore currents. Except for some beach environments (Table 4), sedimentation was almost entirely subaqueous.

The monotonous repetition of thin-bedded Pentacrinus-bearing sparse and packed biosparites of the upper Thaynes indicates deposition in a steady state environment. Pentacrinus apparently preferred to live in shallow turbulent marine conditions (McKerrow, 1978, p. 225). The disarticulated and fragmentary fossils in the packed biosparites support a high energy environment such as shallow shoals (Wilson, 1975, p. 65). Intercalations of fine-grained siliciclastic sediments within the limestone beds indicate intermittent storm-generated sedimentation (Reineck and Singh, 1975, p. 107, 306) and influxes of terrigenous quartz silts during times of high runoff from the craton. Trough cross-laminations were created by deposition from migrating currents, possibly tidal currents, and the horizontal laminations may be the result of tidal current bedload transport, deposition from storm suspension clouds, or from slow sedimentation below wave base. The sparsely packed biomicrosparites add support to sedimentation in quiet areas such as open circulation shelf environments where micrite could not be effectively winnowed from the shell debris (Wilson, 1975, p. 65).

THE JURASSIC SYSTEM

The Jurassic System at Little Sheep Creek is represented, in ascending order, by the Gypsum Spring Tongue of the Twin Creek Formation, the Sawtooth and Rierdon Formations of the Ellis Group, and the Morrison Formation.

The Twin Creek Formation: Gypsum Spring Tongue

The Twin Creek Formation was first defined by Veatch (1907, p. 56) for exposures of marine limestone, shale, and sandstone in the vicinity of Twin Creek between Sage and Nugget in western Wyoming. Imlay (1953a) further defined the formation by dividing it into seven mappable members, designated A-G, that could be recognized in Idaho, Wyoming and Utah. Oriel (1963) formally named Member A the Gypsum Spring Member. The remaining five members were named by Imlay in 1967 in his publication on the distribution, lithology, and faunal content of the Twin Creek Formation. The six members are, in ascending order, the Gypsum Spring, Sliderock, Rich, Boundary Ridge, Watton Canyon, and the Leeds Creek Members.

Regional Distribution and Correlation

The Gypsum Spring Tongue of the Twin Creek Formation at Little Sheep Creek is a 73 m sequence of interbedded pale yellow sandstone, limestone, and reddish brown siltstone. The presence of this unit in southwestern Montana is anomalous and problematic; a red bed sequence at the base of Jurassic strata had not previously been described from the Madison Range (Ray, 1967; Rosé, 1967; Lauer, 1967; Witkind, 1969), the Centennial Range (Moran, 1971; Murray, 1973), the Gravelly Range

(Mann, 1954), the Pioneer Range (Peters, 1970; Sharp, 1970), and from the Tendoy and Beaverhead Ranges (Scholten et al., 1955; Klecker, 1980). Moritz (1951, p. 1804) found the Sawtooth Formation resting unconformably on the Thaynes Formation in southwestern Montana. A measured section of the Ellis Group taken about 3 km east of the thesis area along Little Sheep Creek (Moritz, 1951, p. 1810) included 140 feet (43 m) of calcareous shale and argillaceous limestone in the Sawtooth Formation.

The northernmost extent of the Gypsum Spring Member had previously been described from Red Mountain, Teton County, Idaho (Imlay, 1967, Figure 1, Table 2), where it is 12 feet (7 m) thick. The member extends in a narrow band parallel to the Idaho-Wyoming border southward for 170 miles (272 km) to the southern end of the Wasatch Range and the western end of the Uinta Mountains. The Gypsum Spring Member thickens westward in an irregular manner from an average of 75 feet (23 m) in a line extending roughly north-south through Kemmerer, Wyoming to approximately 400 feet (122 m) a few kilometers east of Salt Lake City, Utah, and Idaho Falls, Idaho (Imlay, 1953a, p. 54; 1967, p. 4).

Sediments of the Twin Creek Formation were deposited in a north-south-trending trough whose western limit is difficult to determine because of paucity of Jurassic exposures westward. The trough is part of an embayment that spread east and south from the Pacific Coast region across the northernmost part of the United States and adjoining parts of Canada (Imlay, 1967, p. 53; Wright, 1973, p. 30). This Middle Jurassic (Bajocian) seaway extended east to western South Dakota, and south into eastern Idaho and western Montana.

Sedimentation was widespread and uniform. Even though normal marine conditions during the Bajocian prevailed only as far east as western Alberta and western Montana (Imlay, 1967, p. 57) the remaining marine environment southward was predominantly a site for deposition of red beds and evaporites in a restricted marginal intertidal-coastal environment (Wright, 1973, p. 33). The source of sediments was from the east, south, and west (Imlay, 1967, p. 57; Wright, 1973, p. 33) with circulation of marine waters and restriction of sedimentation affected by emergent areas, or islands, such as the Belt arch in central Montana (Imlay et al., 1948).

Not withstanding the nomenclature based on geographic location, sedimentation in the Bajocian seaway was rather uniform over a very large area so that correlations can be made on both lithostratigraphic and biostratigraphic criteria. The Gypsum Spring Member of the Twin Creek Formation can be correlated with the lower part of the Gypsum Spring Formation of north-central Wyoming, the lower red bed member of the Piper Formation of central Montana, and the lower part of the Sawtooth Formation of northwestern Montana (Cobban, 1945, p. 1272-1275; Imlay et al., 1948; Imlay, 1967, p. 19-21, Figure 10).

The lower red bed member of the Piper Formation can be traced as far west as a line running south from Liberty County, Montana, to the northwest corner of Yellowstone National Park. Some sections of Middle Jurassic strata in and near Yellowstone National Park contain beds that are intermediate in lithology between the Piper and Sawtooth Formations (for details see section on the Ellis Group); in this region the red beds of the Piper grade westward into the marine beds typical of the Sawtooth. Because red beds beneath the Sawtooth Formation have not previously been described in southwestern Montana, west of Yellowstone National Park, it does not seem possible that the red beds in Little

Sheep Creek can be a westward extending tongue of the lower Piper Formation.

The Sliderock Member of the Twin Creek Formation consists of brownish gray calcareous silty shale overlain by blocky limestone containing oyster fragments (Wright, 1973, p. 21; Imlay, 1953a, p. 54; 1967, p. 22). This member can be correlated by lithology and fauna with the Sawtooth Formation in Little Sheep Creek. That both the Sawtooth Formation of Little Sheep Creek and the Sliderock Member of the Twin Creek Formation are underlain by red beds may be more than coincidence. The north-south-trending trough into which the Twin Creek sediments were deposited probably did not begin to develop until the beginning of sedimentation of the Sliderock Member. The deposition of the red beds continued across a broad restricted marine environment in southeastern and southern Montana (lower Piper Formation), central Wyoming (Gypsum Spring Formation), and eastern Idaho and extreme southwestern Montana (Gypsum Spring Member).

The preceding reasoning may be cyclical, but it does appear that the Twin Creek Formation can be extended into southwestern Montana. Well-ingrained and well-established nomenclature within restricted geographical boundaries precludes the logic of nomenclature based on lithologic and faunal continuity. For this reason I propose that the Middle Jurassic red bed sequence of Little Sheep Creek discussed herein be called the Gypsum Spring Tongue of the Twin Creek Formation.

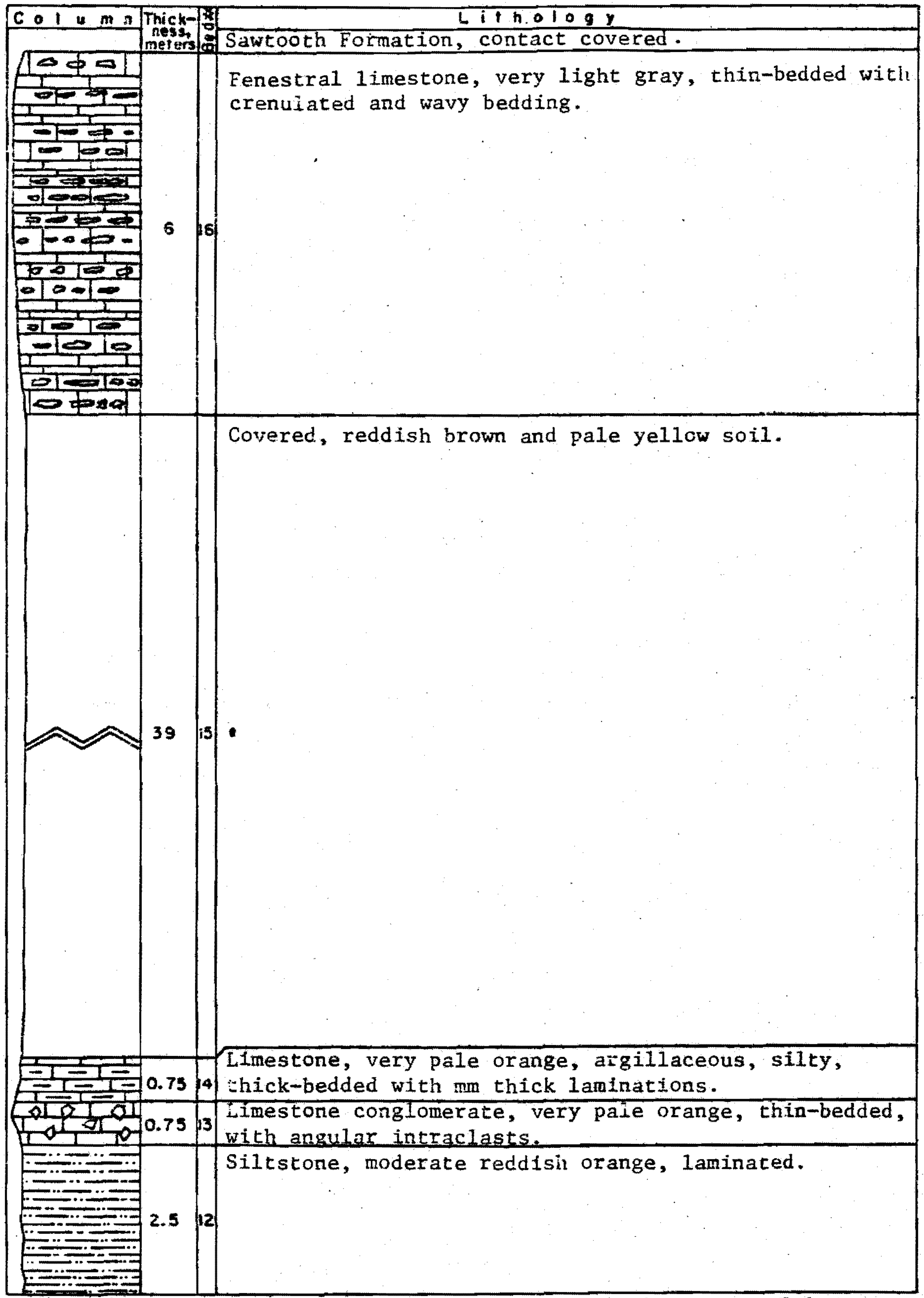
Not withstanding a little confusion when the sequence was discovered in the field, I first thought the red beds might represent the Lanes Tongue of the Triassic Ankareh Formation which had been described extending into the Thaynes Formation in northeastern Idaho (Kummel, 1957, Plate 6). The Lanes Tongue contains red shales and siltstones with subordinate white calcareous sandstones (Kummel, 1957, p. 175); limestones are conspicuously lacking from Kummel's description. The red beds in Little Sheep Creek are made up of light colored sandstones and red siltstones (Figure 31). They also contain conspicuous beds of vuggy limestone similar to the "yellow honeycomb limestones" in the Gypsum Spring Member (Figure 31, Bed #3; Imlay, 1967, p. 17). Also similar are limestone breccias and chert-bearing limestone, the latter containing Pentacrinus crinoid columnals and thin unidentifiable shell fragments. The fossils are not conclusive but until better collections can be made I feel that my correlations are sufficient, based on lithologic similarities.

Distribution and Topographic Expression

The Gypsum Spring Tongue of the Twin Creek Formation crops out parallel to the Thaynes Formation in the south-central and southeastern parts of the thesis area in a line extending from the central part of Section 20 and across the southern halves of Sections 21, 22 and 23, T. 15 S., R. 9 W. (Plate I). The unit forms an inconspicuous slope upsection from the resistant step-like pattern of the thin-bedded limestones of the upper Thaynes Formation. The limestones and calcareous sandstones form resistant ribs where soil cover is thin; the red siltstones are nonresistant and are covered with a reddish soil.

Lithology

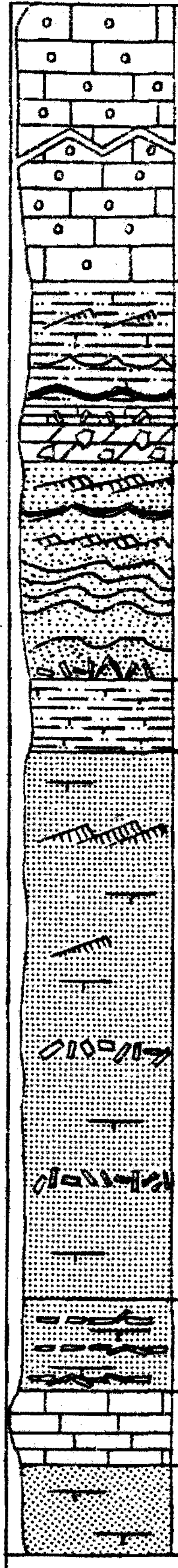
The Gypsum Spring Tongue consists of approximately 73 m of interbedded siltstone, limestone, and sandstone, and one thin bed of dolomite. A section was measured in the E½, Section 20, T. 15 S., R. 9 W.



Column	Thickness, meters	Bed No	Lithology
			Sawtooth Formation, contact covered.
	6	16	Fenestral limestone, very light gray, thin-bedded with crenulated and wavy bedding.
	39	15	Covered, reddish brown and pale yellow soil.
	0.75	14	Limestone, very pale orange, argillaceous, silty, thick-bedded with mm thick laminations.
	0.75	13	Limestone conglomerate, very pale orange, thin-bedded, with angular intraclasts.
	2.5	12	Siltstone, moderate reddish orange, laminated.

Figure 31. Stratigraphic column of the Gypsum Spring Tongue Twin Creek Formation. Section located in the SE¼, SE¼, NE¼, Section 20, T. 15 S., R. 9 W.

Figure 31 (continued)



6	11	Pelletal limestone, white, very thin-bedded, isolated small scale trough cross laminations, bioturbated.
1.7	10	Siltstone, dark reddish brown, calcareous, laminated ripple cross laminations, interference ripples, clay drapes, sharp crested oscillatory wave ripples, wavy bedding.
0.25	9	Limestone, grayish orange pink, angular intraclasts of micrite and packed oosparite, very thin-bedded.
0.5	8	Dolomite conglomerate, very light gray, angular pebbles of dolomite, thin-bedded.
3	7	Sandstone, moderate reddish brown, very fine-grained, very thin- to thin-bedded, mud drapes, symmetrical and asymmetrical ripples, rip-up intraclasts; truncated ripples with horizontal laminations above the truncated surface that grade up into asymmetrical laminae-in-drift climbing ripples, above which are small scale cross laminations. Current direction S76°E.
1	6	Siltstone, very pale orange, calcareous, thin-bedded, horizontal thin laminations.
7.5	5	Sandstone, dark yellowish orange, calcareous, very fine-grained, thin-bedded, ripple laminations, layers of siltstone intraclasts.
1.25	4	Sandstone, very pale orange, very fine-grained, calcareous, layers of silty intraclasts.
1	3	Argillaceous limestone, moderate yellowish brown, thin-bedded, honeycomb weathered, *Lingula* fragments.
1.25	2	Sandstone, grayish orange, very fine-grained, calcareous, thin-bedded, bioturbated.
	1	Thaynes Formation, contact covered.

Details are presented in Figure 31.

The siltstones are moderate reddish orange (10 R 6/6), dark reddish brown (10 R 3/4) or very pale orange (10 YR 8/2) on fresh surfaces, weathering to the same colors or grayish orange (10 YR 7/4). Outcrops are variably calcareous, laminated to thin-bedded, and contain microstructures including horizontal laminations, ripple cross-laminations, wind- and current-generated interference ripples, sharp-crested oscillatory wave ripples, clay drapes, and wavy bedding.

The limestones are white (N 9), light gray (N 8), or are the lighter hues of red (10 R) or yellow-red (10 YR). The limestones consist of argillaceous and silty medium to coarsely crystalline sparry calcite; packed intrasparite with angular intraclasts of micrite and silty micrite in framework support; silty packed pelsparites with pellets (90%), and grapestones, coated grains and oolites (Figure 32); packed edgewise rip-up limestone conglomerate with intraclasts of micrite in an oolitic micrite matrix. The limestones are generally thin-bedded; some units are very thin-bedded and some are thick-bedded. Weathered surfaces of the limestones exhibit various microstructures such as crenulated and wavy bedding (Bed #16, Figure 31) and thin horizontal laminations. The silty packed pelsparite is very thin-bedded, generally homogeneous in texture, but exhibits isolated small-scale (5-8 mm deep) trough cross-laminations which, microscopically, are filled with alternating lamina (less than 0.5 mm thick) of quartz silt and pellets. Fossils are very rare, consisting of only small <u>Lingula</u> shell fragments.

The sandstones are generally the light yellow red hues (10 YR) with one bed (#7) being a moderate reddish brown (10 R 4/6) and grayish red

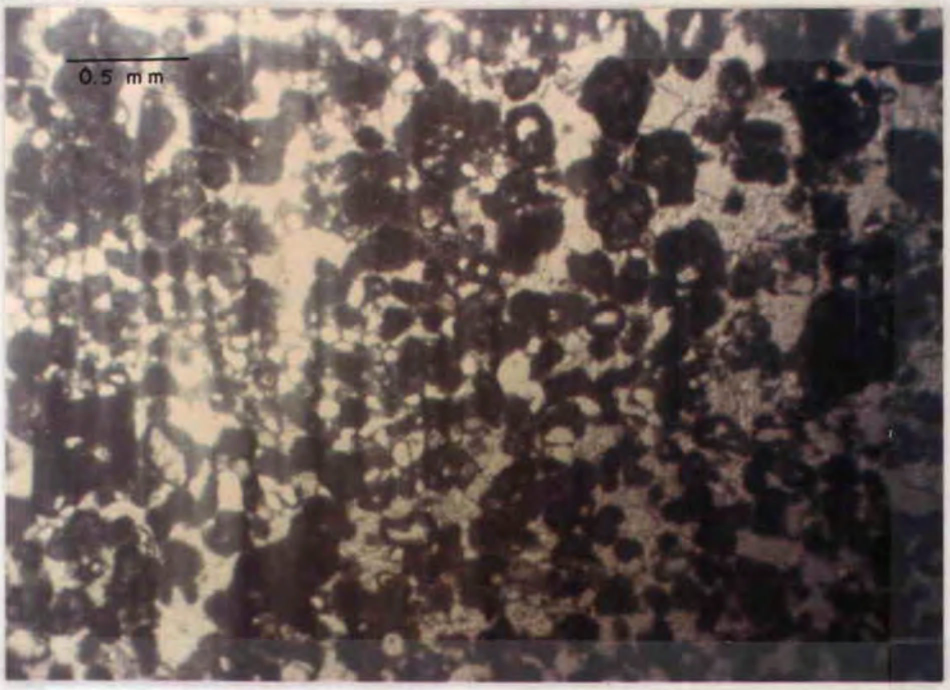


Figure 32. Photomicrograph of silty packed pelsparite (Bed #11, Figure 31), Gypsum Spring Tongue of the Twin Creek Formation. Plane light.

(10 R 4/2) with light bluish gray (5 B 7/1) mottling. They are very fine-grained, well-sorted rocks composed of angular grains of quartz and rare quantities of siltstone intraclasts, muscovite, and zircon in framework support with calcite cement. The sandstones are very thin- to thin-bedded with structures consisting of symmetrical and asymmetrical ripples, single layer concentrations of edgewise siltstone intraclasts, interference ripples (?), truncated ripples above which are horizontal laminations that grade up into asymmetric lamina-in-drift climbing ripples and then migrating ripple cross-laminations (Figure 33).

The dolomite is very light gray (N 8) on fresh surfaces and weathers pale greenish yellow (10 Y 8/2). The rock consists of angular granule- to pebble-size clasts of dolomite micrite in framework support with calcite cement. Laminated birdseye dolomite was observed elsewhere in the unit with the fenestrae filled with quartz and calcite.

In the upper part of the Gypsum Spring Tongue in the SE¼, Section 22, T. 15 S., R. 9 W. is a resistant outcrop of limestone containing rounded chert pebbles. The limestone is pale yellowish brown (10 YR 6/2) on fresh surfaces and weathers very pale orange (10 YR 8/2). The chert is very light gray (N 8) and occurs as rounded oval pebbles up to 3 cm in the long dimension and as angular granules and pebbles up to 1 cm in the greatest dimension. Total thickness of the unit could not be determined, but it could be traced laterally westward to the SE¼, Section 21. Over this distance of approximately 1.5 km the rock grades from the chert pebble-bearing limestone sand to a fine-grained to dense limestone that contains angular granule size fragments of chert. The limestone matrix is a poorly sorted, well-washed sandy pelbiointrasparite (Figure 34). The pellets are circular to oval

Figure 33. a) Very fine-grained sandstone of the Gypsum Spring Tongue, Twin Creek Formation showing edgewise siltstone intraclasts, symmetrical and migrating asymmetrical ripple laminations. b) underside of sample showing layer of concentrated siltstone intraclasts. Bed #7 (Figure 31). Gypsum Spring Tongue, Twin Creek Formation.

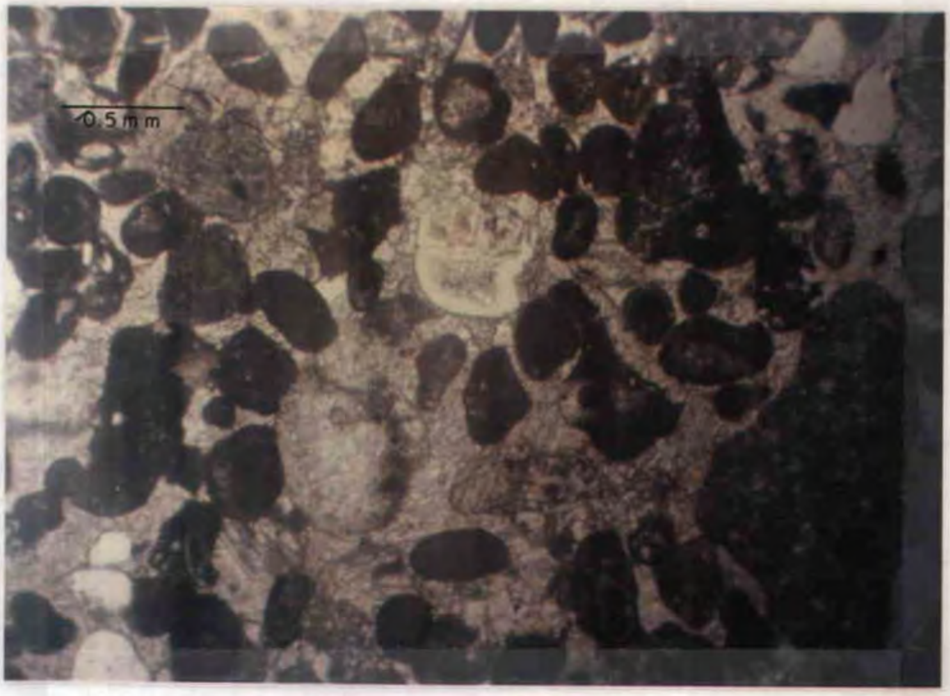


Figure 34. Photomicrograph of the sandy pelbiointrasparite matrix of the chert pebble lime sandstone. Sample Kmb 79-239, Gypsum Spring Tongue, Twin Creek Formation, SW¼, SE¼, Section 22, T. 15 S., R. 9 W. Plane light.

grains with an approximate mean diameter of 0.4 mm and consist of micrite, coated grains of biotic fragments, and collophane. The fossils are crinoid ossicles and echinoid plates (the latter occur as whole plates or as well-rounded detrital grains), and rare _Lingula_ shell fragments. Syntaxial overgrowth of calcite on the echinoderm fragments forms a cement which is poikiloblastic with the surrounding detritus. The intraclasts consist of micrite, silty micrite, and rare oosparite. The siliciclastic debris consists of medium sand size, subrounded to rounded grains of unstrained monocrystalline quartz and chert. Chert and chalcedony also selectively replace fossil fragments. Microscopically, the chert pebbles consist of finely disseminated cryptocrystalline quartz replacing micritic dolomite. This type of chert was observed locally as nodules in dolomite in the Permian Grandeur Tongue of the Park City Formation (Figure 8). Permian rocks were exposed during pre-Jurassic erosion northeast of the thesis area in Madison and Park Counties (Moritz, 1951, Figure 18), and may be the source of the chert pebbles. The fining westward of the sediments in this horizon indicate an easterly or northeasterly source for the pebbles.

The upper and lower contacts of the Gypsum Spring Tongue are covered. The lower contact is drawn at the first up-section occurrence of a very fine-grained calcareous sandstone above the thin-bedded _Pentacrinus_-bearing limestone of the Thaynes Formation. The upper contact is marked by a fenestral limestone approximately 5 m thick.

The lower contact represents a hiatus which spans the Middle and Late Triassic and the Early Jurassic. The contact between the Gypsum Spring Member and the Slide Rock Member of the Twin Creek Formation in Idaho is sharp but conformable (Imlay, 1967, p. 18). It is assumed the relationship between the Gypsum Spring Tongue and the Sawtooth Formation is also conformable at Little Sheep Creek.

Fossils and Age

The Gypsum Spring Tongue is sparsely fossiliferous. *Pentacrinus* columnals, echincid plates and spines, and *Lingula* shell fragments were the only fossils observed. This sparse fossil content prevails in the Gypsum Spring Member of the Twin Creek Limestone in Idaho and Wyoming with the addition of *Camptonectes* fragments (Imlay, 1967, p. 19). The Gypsum Spring Member in Idaho and Wyoming is middle Bajocian in age (Imlay, 1967, p. 19). Lateral correlation with the Gypsum Spring Tongue in Little Sheep Creek also places a middle Bajocian age on the unit in the thesis area.

Environments of Deposition

The rock types and bedding characteristics of the Gypsum Spring Tongue of the Twin Creek Formation indicate that deposition of sediments occurred in a restricted intertidal environment. The sedimentary features and their environmental associations are summarized in Table 5. The discussion concerning depositional environments of the Woodside Formation applies to the Gypsum Spring Tongue and will not be repeated here; the reader is directed to that discussion on paleointertidal sedimentation commencing on page 101.

An intertidal to supratidal environment that was much more restrictive during Middle Jurassic time than the Early Triassic is indicated by the presence in the Gypsum Spring Tongue of birdseye dolomite, dolomite conglomerate, brecciated limestone, and honeycomb-weathered limestone.

Honeycomb-weathered limestone is indicative of selective solution

TABLE 5. Gypsum Spring Tongue, Twin Creek Formation: Sedimentary Features and Environmental Associations

Sedimentary Features	Environmental Associations	References
Horizontal laminations	Tidal current bedload transport	Klein, 1971; 1977
Symmetric ripples	Oscillatory wave-generated ripples created by reworking of non-cohesive sediments during emergence on tidal flats	Reineck and Singh, 1975
Asymmetric ripples with ripple cross-laminations	Small current-generated ripples	Reineck and Singh, 1975
Wind- and current-generated interference ripples	Late stage emergence of tidal flats; wave-generated ripples superimposed on small current ripples	Reineck and Singh, 1975
Flat-topped ripples	Late stage emergent ebb flow with sudden changes in flow directions at extremely shallow water depths	Klein, 1971
Mud drapes over ripples	Tiday slack water mud suspension deposition	Klein, 1971
Trough cross-laminae	Deposition from migrating small current- and wave-generated ripples	Wilson, 1970; 1975; Reineck and Singh, 1975
Edgewise rip-up conglomerate	Tidal channel lag deposits	Wilson, 1975

TABLE 5, Continued

Sedimentary Features	Environmental Associations	References
Truncated ripple laminations above which are horizontal laminations that grade up into asymmetric laminae-indrift climbing ripples, above which are small-scale ripple cross-laminations	Tidal scour during submergence with deposition of suspended sediment from a rapidly declining flow regime with a reduction of sediment load	Klein, 1971; Reineck and Singh, 1975
Bioturbation	Slow sedimentation; thorough homogenization of sediment by burrowing organisms	Pettijohn and Potter, 1964; Wilson, 1975
Honeycomb weathering in limestones	Selective dissolution of soluble minerals in limestone; suggestive of dissolution of evaporite deposits	AGI Glossary, 1977

of very soluble minerals in a less soluble matrix after lithification has occurred (AGI Glossary, 1977, p. 336). Imlay (1967, p. 19) was able to trace brecciated limestone of the Gypsum Spring Member of the Twin Creek Formation eastward where it is replaced by gypsum masses in lower parts of the Gypsum Spring Formation. It is possible that the limestone breccia and the honeycomb limestone in Little Sheep Creek represent the former presence of gypsum which has been removed by solution.

The source of the Gypsum Spring Tongue dolomite may differ from that for the Woodside Formation. The process may involve the evaporite-reflux model of Defeyes et al. (1965) or the creation of a dolomite crust on a supratidal algal mudflat similar to that described by Shinn et al. (1965) on Andros Island in the Bahamas. Paleolatitude and paleoclimatic reconstruction places southwestern Montana in a semi-arid to subhumid climate during the Middle Triassic (Peterson, 1978), climatic conditions which are similar to penecontemparaneous development of dolomite today.

The association of dolomite, brecciated limestone, birdseye dolomite, and stromatolitic and pelletal limestone helps to support the evaporite-reflux model. In this model dolomite is precipitated from interstitial pore waters in which the magnesium/calcium ratio has been increased by the evaporation of the seawater and precipitation of gypsum. The association of gypsum and dolomite is, therefore, necessary in order to recognize the evaporite-reflux model in fossil tidal flats. A similar association can be made with the sedimentary rocks of the Gypsum Spring Tongue. In further support of an intertidal flat environment are the algal stromatolites (Logan et al., 1964), the birdseye structures (Shinn, 1968), the pelletal limestones (Shinn et al., 1965), and the environmental associations of the bedding characteristics listed in Table 5. Dolomite crusts subjected to intermittent wetting and drying in an intertidal-supratidal environment would develop desiccation polygons which would collect on the back levee slopes (Ginsburg and Hardie, 1975, p. 205) and as tidal channel lag deposits (Wilson, 1975, p. 82). The dolomite breccia found in the thesis area may have a similar origin, or it may be micrite rip-ups related to tractive current activity.

THE ELLIS GROUP

Peale (1893) designated as the Ellis Formation those rocks in the Three Forks area of Montana which lie between the Quadrant Formation and the Cretaceous rocks. He did not describe the rocks, and he did not specify a type section. The formation is named after old Fort Ellis, 50 miles (80 km) north of Yellowstone National Park. Cobban et al. (1945, p. 451-453) described 297 feet (91 m) of the Ellis Formation seven miles southeast of Bozeman, Montana and designated the type section to be in Rocky Canyon. Based on investigations in the Sweetgrass arch in north-central Montana, Cobban (1945, p. 1263) raised the Ellis to group status and subdivided it into three formations of marine origin. In ascending order they are the Sawtooth, Rierdon, and the Swift Formations. In Montana the Ellis Group unconformably overlies strata which range from Mississippian to Early Jurassic in age (Imlay et al., 1948). It underlies the Late Jurassic Morrison Formation or the Early Cretaceous Kootenai Formation (Cobban, 1945).

Stratigraphic terminology varies for the Jurassic marine rocks; formation names change both with geography and with regional changes in lithology. In central and southern Montana east of the Sweetgrass-Big

Belt line of uplift, the Ellis Group comprises, in ascending order, the Piper, the Rierdon, and the Swift Formations (Imlay et al., 1948). In a stratigraphic section at the northern end of the Teton Range, Wyoming, Edmund (1951, p. 8) included, in ascending order, the Nugget sandstone, the Twin Creek limestone, and the Stump sandstone as members in the Ellis Formation. Mann (1954, p. 25-30) described only 13 to 41 feet (4 to 14 m) of Ellis in the Gravelly Range without differentiating the formations. Prior to 1951 the marine Jurassic rocks of southwestern Montana had been mapped as the Ellis Formation. Moritz (1951, p. 1803) recognized the three subdivisions of Cobban (1945, p. 1267) and suggested that the Ellis Formation be raised to group status in southwestern Montana and that the discrete formations be recognized.

The Sawtooth Formation

Cobban (1945, p. 1270) defined the Sawtooth Formation as consisting of a basal fine-grained sandstone, a middle unit of dark gray shale containing a few thin dark limestone layers, and an upper unit of highly calcareous siltstone. The formation is named after exposures in the Sawtooth Range of Montana and the type section is in Rierdon Gulch where it is 136 feet (42 m) thick (Cobban, 1945).

In southwestern Montana the Sawtooth Formation is restricted to the central and southern parts of the Tendoy Range (Scholten et al., 1955, p. 367) and the western part of the Centennial Mountains (Moran, 1971, p. 102-107). In the Tendoy Range the formation is 110 feet (34 m) thick in Little Water Canyon and attains its maximum thickness of 140 feet (43 m) along the Middle Fork of Little Sheep Creek approximately 4 km east of the thesis area (Moritz, 1951, p. 1804).

Regional Distribution and Correlation

The Sawtooth Formation is thickest in a trough which extends from immediately southwest of the Sweetgrass Hills, northern Montana, southward along the Rocky Mountain front (Cobban, 1945, p. 1266) to the Idaho-Wyoming border. Thicknesses range from a feather edge near the south arch of the Sweetgrass uplift and the Belt arch uplift to nearly 230 feet (70 m) south of Glacier National Park (Imlay et al., 1948) and nearly 600 feet (183 m) in the subsurface in southern Alberta (Imlay, et al., 1948; Wier, 1949).

The Piper Formation of central and eastern Montana is characterized by a lower red bed-gypsum member, a middle gray shale-limestone-dolomite member, and an upper red bed-gypsum member (Imlay, 1952, p. 968). The members grade into each other vertically and laterally and thicken westward as they grade into the dark marine rocks of the Sawtooth Formation in northern Montana and in the vicinity of Yellowstone National Park (Imlay et al., 1948).

As will be discussed in a succeeding section the age of the Sawtooth Formation has been determined to be Middle Jurassic (Imlay, 1952, p. 969). The Nugget sandstone, which was included in the Ellis Formation by Edmund (1951), has been determined to be of Early Jurassic age (Mansfield, 1952, p. 35) and underlies the Gypsum Spring Formation in central Wyoming (Love et al., 1947). As previously stated, a marked erosional unconformity exists between the Thaynes Formation and the Ellis Group in southwestern Montana. It is conceivable, therefore, that the Nugget sandstone was erroneously included as a member of the Ellis Formation by Edmund.

In western Wyoming, northern Utah, and southeastern Idaho, the

Sawtooth is correlative with the lower part of the Twin Creek Limestone (Thomas and Krueger, 1946, p. 1275) which is Middle Jurassic in age (Imlay, 1945, p. 1020). The Twin Creek Limestone consists of seven mappable members; the lowest member is equivalent to the Gypsum Spring Formation (Imlay, 1950, p. 37-45). The base of the Piper Formation includes equivalents of the type Gypsum Spring Formation of central Wyoming, whereas the upper part includes beds that have been placed in the lower part of the "Lower" Sundance in the Wind River basin and in central Wyoming (Imlay et al., 1948).

Distribution and Topographic Expression

The Sawtooth Formation commonly forms gentle grass-covered slopes on which are developed a thick and light colored soil. It crops out parallel to the Thaynes Formation in the south-central and southeastern parts of the thesis area in a line extending from the E½, Section 20 through Sections 21, 22, and 23, T. 15 S., R. 9 W. (Plate I). A high angle reverse fault has exposed a partial section of the Sawtooth and Rierdon Formation in a gully in the SE¼, Section 20, T. 15 S., R. 9 W. The outcrop pattern is disrupted by a series of faults in the SE¼, Section 22 and the SW¼, Section 23, T. 15 S., R. 9 W. The formation is generally poorly exposed but samples and descriptions could be taken from stream banks and exposures on ridges protected by the overlying Rierdon limestone.

Thickness and Lithology

Moritz (1951, p. 1809-1810) described sections of the Sawtooth Formation in Little Water Canyon 17 km north of the thesis area and in the Middle Fork of Little Sheep Creek about 4 km to the east. According to Moritz, the Sawtooth is made up of two distinct lithologies. The basal unit consists of 60-70 feet (18-21 m) of gray and gray green calcareous shales which weather into splintery, pencil-shaped fragments. The upper unit is a yellow brown argillaceous limestone, 70-80 feet (21-24 m) thick, which weathers into small blocky fragments.

The outcrops of the Sawtooth Formation in the thesis area are small and discontinuous, are not adequate for detailed descriptions, and do not even provide adequate appraisal for any lateral facies changes. The contacts with the overlying and underlying formations are covered. The Sawtooth Formation was mapped as that unit lying between the last outcrop of limestone of the Gypsum Spring Tongue of the Twin Creek Formation, to the first occurrence of oolitic limestone of the basal Rierdon.

A composite description taken from several localities agrees with Moritz's (1951) lithologic divisions even though thicknesses could not be determined. The most clearly defined contacts are located in a small saddle in the NE¼, NE¼, SE¼, Section 20, T. 15 S., R. 9 W. where the approximate thickness of the Sawtooth has been calculated to be 40 m. The lower part of the formation consists of very pale orange to grayish orange (10 YR 8/2 to 10 YR 7/4) and yellowish gray to dusky yellow (5 Y 7/2 to 5 Y 6/4) thinly laminated to very thinly bedded, micritic limestone. This lower unit weathers to acicular shards and contains rare pelecypod shell fragments.

The upper unit consists of massive, dense, argillaceous limestone which is pale yellowish brown (10 YR 6/2) when fresh and weathers to pinkish gray (5 YR 8/1) blocky centimeter-size fragments. Disseminated throughout the rock are clusters and stringers of fine sand-sized pyrite spherulites.

Fossils and Age

The upper limestone unit of the Sawtooth Formation contains numerous small, sharp, angular pelecypod shell fragments tentatively identified as Ostrea species. The wide range of this genus (Triassic to Recent; Shimer and Shrock, 1959, p. 395) is not sufficient to provide an age for this formation. Ostrea has been found in the Sawtooth Formation at other localities associated with faunal assemblages which have provided a more definitive age for the Sawtooth (Mann, 1954, p. 26; Moritz, 1951, p. 1808; Witkind, 1969, p. 34; Cobban, 1945, p. 1276).

Imlay (1952, p. 968) considered the Sawtooth Formation to be Middle Jurassic (Bajocian to Bathonian) in age, based on the presence of Defonticeras near the base, Arctocephalites and Procerites in the upper silty beds, and on the stratigraphic position beneath beds containing varied ammonite faunas that allow accurate correlation with the early Callovian ammonite zones of northwestern Europe. The Piper Formation has also yielded Defonticeras and Teloceras ammonites (Imlay et al., 1948). The second unit from the base of the Twin Creek Formation (the Sliderock Member) has furnished Bajocian ammonites while Callovian pelecypods have been collected from the upper members.

The Rierdon Formation

Cobban (1945, p. 1277) applied the name Rierdon to a sequence of alternating gray calcareous shales and limestones which overlie the Sawtooth Formation and disconformably underlie the Swift Formation. Cobban defined this formation strictly on its lithologic character and named it for exposures in Rierdon Gulch in the Sawtooth Mountains of

Montana. The Rierdon is 137 feet (42 m) thick at the type locality and it thickens to about 200 feet (61 m) 65 km in a westerly direction.

The distribution of the Rierdon Formation in southwestern Montana is similar to that of the other formations of the Ellis Group. It is restricted to the central and southern parts of the Tendoy Range (Scholten et al., 1955, p. 367); however, it does not crop out in the eastern part of the Centennial Mountains (Moran, 1971, p. 103; Murray, 1973, p. 74). Moritz (1951, p. 1809-1810) described 109 feet (33 m) of the Rierdon in Little Water Canyon and 104 feet (32 m) in the Middle Fork of Little Sheep Creek east of the thesis area. He noted that the basal unit of the formation in southwestern Montana is a prominent oolitic limestone which is overlain by argillaceous limestone.

Regional Distribution and Correlation

The thickness and distribution of the Rierdon Formation is similar to that of the Sawtooth. Distribution of sediments in the Jurassic seas continued to be controlled through the Callovian stage by the south arch of the Sweetgrass arch and the Belt arch (Cobban, 1945, p. 1277; McMannis, 1965, p. 1812) and a small positive area in southwestern Montana (Moritz, 1951, p. 1806) which was not emergent during Sawtooth time (McMannis, 1965, p. 1812). The Rierdon is absent in the vicinity of the Sweetgrass and Belt arches (Cobban, 1945). It generally thickens east- and westward from these positive areas to about 240 feet (73 m) in the subsurface (Imlay et al., 1948). Locally it is sandy suggesting possible emergence of some small positive elements (McMannis, 1965, p. 1813). The formation is distributed across central and eastern Montana where it overlies the Piper Formation (Imlay et al., 1948).

Regional correlation of the Rierdon is based on four distinct ammonite zones. They are, in ascending order, Arcticoceras, Gowericeras, and two varieties of Kepplerites. Peterson (1954, p. 475-477) stated that the Swift-Rierdon nomenclature could apply to Jurassic marine rocks of most of Wyoming and western South Dakota. For historical purposes, however, he preferred to retain the name Sundance for those units. The Rierdon Formation is correlative with the base of the Sundance Formation in northwestern and central Wyoming and the Black Hills, the "Lower" Sundance Formation of southeastern Wyoming, and the upper part of the Twin Creek Formation of eastern Idaho, northern Utah and western Wyoming (Imlay, 1952, p. 965, 968).

Distribution and Topographic Expression

The limestone of the Rierdon Formation forms distinct, low, white ridges which parallel the Sawtooth Formation in the south-central and southeastern parts of the thesis area in a line extending from the E½, Section 20 through Sections 21, 22, and 23, T. 15 S., R. 9 W. High angle reverse faults have juxtaposed partial sections of the Sawtooth and Rierdon Formations in the SE¼, Section 20, T. 15 S., R. 9 W. (Plate I). The outcrop pattern is disrupted by a series of faults in the SE¼, Section 22 and the SW¼, Section 23, T. 15 S., R. 9 W.

Thickness and Lithology

Moritz (1951, p. 1809-1810) described measured sections in Little Water Creek and on the east side of the Middle Fork of Little Sheep Creek 4 km east of the thesis area. In the Little Sheep Creek section Moritz described a basal gray to brownish gray, oolitic, thin- to medium-bedded limestone 26 feet (8 m) thick and an upper unit 78 feet (24 m) thick consisting of gray and thin-bedded limestone interbedded with gray, calcareous shale. According to Moritz (p. 1804), the Rierdon lies with apparent conformity above the Sawtooth and is conformably (?) overlain by the Swift Formation. Cobban (1945, p. 1290) noted an unconformity between the Rierdon and Swift Formations in western Montana. Witkind (1969, p. 35) described a very sharp contact between the Rierdon and Swift Formations near Yellowstone National Park and suggested that this unconformity extends into southwestern Montana as well. The contact in southwestern Montana was considered to represent a slight stratigraphic break by Scholten et al. (1955, p. 367). As will be discussed in a subsequent section, the Swift Formation is considered to be absent in the thesis area; the Rierdon Formation is, therefore, unconformably overlain by the red mudstones of the Morrison Formation.

In the thesis area the Rierdon Formation is approximately 31 m thick and is made up of two units. The basal unit consists of about 6 m of interbedded thin-bedded fossiliferous oolitic limestones and very thin-bedded fossiliferous micritic limestones that do not exhibit any sedimentary structures. The oolites are medium- to coarse-grained, and well sorted. The oolitic beds are yellowish gray (5 Y 7/2), pale yellowish gray (5 YR 8/1 to 5 Y 8/1), and moderate yellowish brown (10 YR 5/4). The basal limestone forms persistent low ridges and its presence was the basis for mapping the outcrop pattern of the Rierdon. The upper unit is generally covered but trenching in some strategically placed stream cuts revealed interbedded very thin-bedded to fissile argillaceous micrite and thinly bedded, sparse biomicrite. The argillaceous limestone is very pale orange (10 YR 8/2) and yellowish gray (5 Y 7/2) and weathers to platy and blocky shards 0.5 to 2.0 cm thick. The oolitic limestone is not restricted to the base of the formation but occurs as discontinuous lenses within the fissile argillaceous micrite. An outcrop of oolitic limestone occurs approximately 20 m above the base of the formation in its easternmost exposure. The oolite is thinly bedded and is intercalated with micritic limestone.

The contact with the Sawtooth Formation is considered to be the first occurrence upsection of the oolitic limestone. The contact with the Morrison Formation has been drawn at the marked break in slope upsection from the Rierdon ridges, or at the first appearance of the distinctive red soil of the Morrison.

Fossils and Age

The fossils in the Rierdon Formation are sparsely distributed throughout the lower unit, except for intercalated coquinoid beds up to 2 cm thick. The shell fragments in these layers are hydrodynamically oriented with convex surfaces up.

The fossils are generally too fragmented to allow positive identification; however, two pelecypod genera have been tentatively identified by me as *Camptonectes* and *Ostrea* (Shimer and Shrock, 1959). Other fossils included are gastropods, crinoids, especially the genus *Pentacrinus*, and other pelecypod fragments. *Camptonectes* and *Ostrea* species have been identified in the Rierdon in adjacent areas by Cobban (1945, p. 1281), Mann (1954, p. 26), and Imlay et al. (1948; 1957).

Imlay (1948; 1953b) places the Rierdon Formation as early Callovian in age (late Middle Jurassic) based on four ammonite zones corresponding to northwest European stages. They are, in ascending order, *Arcticoceras*, *Gowericeras*, a finely ribbed species of *Kepplerites*, and a coarsely ribbed species of *Kepplerites*.

The Swift Formation

Cobban (1945, p. 1281) applied the name Swift to the youngest marine Jurassic rocks found in western Montana. The type locality is in the vicinity of Swift reservoir in Pondera County, Montana. The formation ranges in thickness from 90 feet (27 m) to 166 feet (51 m). It consists of dark gray non-calcareous shale overlain by fine-grained glauconitic flaggy sandstone along the eastern flank of the Rocky Mountains and the Kevin-Sunburst dome. The unit is a thick-bedded fine-grained sandstone overlying a basal conglomerate, or a thick-bedded flaggy sandstone containing abundant dark non-calcareous shale partings where it covers the south arch of the Sweetgrass arch.

The Swift Formation is of widespread occurrence. It has been described in the Williston Basin (Storey, 1958) and the Swift-Rierdon nomenclature has been applied to formational units of the Sundance Group in eastern Wyoming and western South Dakota (Peterson, 1954). Witkind (1969, p. 35) described about 40 feet (12 m) of resistant, cross-bedded, oolitic, sandy limestone interbedded with thin claystones west of Yellowstone National Park. Moran (1971, p. 108) found 48 feet (15 m) of fossiliferous pebbly sandstone, fossiliferous glauconite limestone, and glauconitic sandstone in the eastern Centennial Mountains.

In southwestern Montana the Swift Formation is the thinnest and most widespread of the Ellis Group. It consists chiefly of dark greenish brown glauconitic sandstones and greenish gray shales and attains a maximum thickness of approximately 20 feet (6 m; Moritz, 1951, p. 1805). The formation is generally poorly exposed; a measured section was taken in Little Water Canyon only by trenching. No outcrops were described by Moritz (1951, p. 1810) in the Middle Fork of Little Sheep Creek section; he tentatively placed the unit in its proper stratigraphic position and assumed a thickness of about 20 feet (6 m). Scholten et al. (1955, Table 1) found the Swift Formation to be up to 10 feet (3 m) thick in the Tendoy Range. Klecker (1980, p. 136) found the Swift Formation to be generally nonresistant in Little Water Canyon with only one exposure discovered in a gully.

No outcrops, or any weathering products of the Swift Formation were exposed in the thesis area. One possible site for trenching in the formation is on a saddle in the SE¼, Section 21, T 15 S., R. 9 W. At this locality there is an abrupt change in the color of the thin soil cover from white and light gray typical of the Rierdon Formation to a moderate and dark reddish brown soil cover of the Morrison Formation.

Because the Swift Formation is generally thin and of variable thickness, and it could not be described in the thesis area, it is considered to be non-existent. It is a widespread formation of the Ellis Group and commonly the only representative of the Ellis (Moritz, 1951, p. 1804) where the underlying formations are absent. The absence of the Swift Formation in the thesis area is attributed to local pre-Morrison erosion (Suttner, 1969, p. 1393) and not to non-deposition.

The Morrison Formation

The Morrison Formation is one of the most extensive formations in the middle western states. A large body of literature exists concerning the Morrison, addressing itself not only to the facies and environments of deposition, but to very old controversies regarding the age and the lower and upper limits of the formation. Mook (1916) listed over 200 papers describing or discussing the formation published up to that time, excluding those of purely descriptive paleontology. Baker et al. (1936) reviewed the literature, particularly for the southern Rocky Mountains, and discussed the correlation of the Jurassic formations of parts of Utah, Arizona, New Mexico and Colorado. Reeside (in Yen, 1951b) provided a succinct summary of the stratigraphy of the Morrison Formation.

The name Morrison was first published by Cross (1894, p. 2) in his folio on the Pikes Peak Quandrangle, Colorado. Eldridge (in Emmons et al., 1896, p. 60-62) assigned the name formally to exposures north of the town of Morrison, Colorado. Eldridge's descriptions were based on incomplete sections and he did not designate a type locality. Lee (1920, p. 183-188; 1927, p. 28) redefined the upper and lower limits of the formation, but also did not designate a type section. Waldschmidt and LeRoy (1944, p. 1098) described a section of the Morrison Formation newly exposed in a roadcut two miles north of Morrison, and designated this locality as the type section. They subdivided the formation into six units with a total thickness of about 300 feet (92 m). They are, in ascending order, a basal sandstone, gray and red shale with charophyte oogonia, gray clay and limestone, gray shale and sandstone with dinosaur remains, red shale, and an upper unit of variegated shale and sandstone.

The depositional environment of the Morrison has been assumed to be one of an extensive alluvial plain with a low profile, of which no modern analogy exists. The plain possibly represents the distal end of a clastic wedge that extended east from Late Jurassicorogenic lands in Oregon and Washington (Suttner, 1969, p. 1393, 1406). The facies are, therefore, laterally discontinuous and correlation of individual units is impossible. The section described by Waldschmidt and LeRoy in 1944 serves as an acceptable standard for the formation.

In southwestern Montana the Morrision Formation is the most widespread Jurassic formation. Its thickness does not exceed 400 feet (122 m). The Morrison Formation is absent in southwestern Montana in the area of the Belt arch and it generally thickens eastward. The geometry of the formation was locally controlled by the Belt arch which was a pronounced positive element in Middle Jurassic time, but only of limited extent by the Late Jurassic. The Belt arch had subsided totally by the Early Cretaceous with subsequent development of a true clastic-wedge configuration of the Kootenai Formation (Suttner, 1966, p. 22). The outcrop pattern is generally one of gentle, covered slopes. Moritz (1951, p. 1810-1811) did find, however, relatively good exposures in Little Water Canyon and in the Middle Fork of Little Sheep Creek east of the thesis area. The formation thins north and east from 391 feet (119 m) in Little Sheep Creek south of Garfield Mountain to 130 feet (40 m) on the west slope of the Blacktail Range, 72 feet (22 m) in Little Water Canyon, and is absent in the northern part of the Tendoy Range (Moritz, 1951, p. 1810-1912; Scholten et al., 1955, p. 367).

Regional Distribution and Correlation

The Morrison Formation has been recognized in northern New Mexico, northeastern Arizona, central Utah, Wyoming, Montana, western South Dakota, western Nebraska, western Kansas, Colorado, and in the Oklahoma panhandle. Thickness and distribution of the strata recognized as belonging to the Morrison vary considerably. The range is from 300 feet (92 m) at the type section to a minimum of 21 feet (7 m) in the Black Hills (Waage, 1959, p. 38-40) and an apparent maximum of over 1500 feet (458 m) in northeastern Utah (Huddle and McCann, 1947). Reeside (in Yen, 1951b, p. 22-26) has provided a summary of the stratigraphy of the Morrison Formation over its entire distribution, including a historical approach to definitions of the upper and lower limits of the formation.

In Montana where the Morrison Formation lies on marine Jurassic strata, there is no physical evidence for interruption of sedimentation and the contact is considered to be conformable. There is no sharp change in lithologic character with the underlying formations and the boundary must be arbitrarily selected (Reeside, in Yen, 1951b, p. 25). In southwestern Montana the lower boundary is considered to be the top of the uppermost sparsely fossiliferous and glauconitic sandstone of the Swift. Local reworking of the glauconitic sandstone into the basal Morrison results in a boundary transitional over 20 feet (6 m) or less. The Morrison-Kootenai boundary is placed at the base of the Kootenai chert pebble conglomerate or the chert-rich "salt and pepper" sandstone (Suttner, 1969, p. 1393). The upper contact is unconformable and is observable, regionally, by a marked change in lithology and faunal content with the overlying Kootenai Formation (western Montana), Cloverly Formation (central Wyoming), Gannet Group (western Wyoming), the Lakota

Formation (western South Dakota) and the Dakota Group (Colorado) (Reeside, in Yen, 1951b, p. 25).

There is no difficulty with the regional correlation of the Morrison Formation because it is so extremely widespread and it is easily recognized in adjoining regions.

Distribution and Topographic Expression

No natural exposures of the Morrison Formation were encountered in the thesis area. Its presence is noted by a reddish brown soil cover which occurs in the break in slope upsection from the Rierdon ridges and below the basal chert pebble conglomerate and coarse-grained sandstone of the Kootenai Formation. The formation is located in a belt which parallels the outcrop pattern of the other Jurassic rocks from the SE¼, Section 21, eastward through Sections 22, 23, and 24, T. 15 S., R. 9 W. It has been displaced northward by faults in the SE¼, Section 22, and the S½, Section 23. The Morrison Formation is approximately 49 m thick.

Lithology

Moritz (1951, p. 1811, 1812) described a 391 foot (119 m) section of the Morrison Formation in the Middle Fork of the Little Sheep Creek about 4 km east of the thesis area. At that locality the formation is made up of a two-foot (0.6 m) bed of gray, fresh water limestone overlain by 37 feet (11 m) of poorly exposed, gray, soft, calcareous, "salt and pepper" sandstone. The top of the formation consists of 198 feet (60 m) of partly covered red mudstone, siltstone, and shale, with a few thin, gray, dense, fresh water limestones.

As stated in the previous section, no natural outcrops of the Morrison Formation were observed in the thesis area.

Age

The age of the formation is Late Jurassic. Yen (1951b, p. 26-35) provides a review of the literature pertaining to the paleontology of the Morrison Formation. He compared the published evidence and his own faunal collections with the collections made in the overlying Cloverly and Kootenai Formations. He observed that even though some fresh water molluscan fauna are cogeneric across the Jurassic-Cretaceous boundary, certain genera are present in the formations of one period and absent in the other. Even when cogeneric species are present, there is enough variation in the species and the assemblages to consider them as having lived during different periods of geologic time. The Morrison fauna is, therefore, not equivalent to Kootenai and Cloverly fauna, but is considered to be Kimmeridgian to early Portlandian (late Late Jurassic) in age.

THE CRETACEOUS SYSTEM

The Kootenai Formation

Sir William Dawson (1885, p. 1-2) is credited with publishing the name "Kootenie" by applying the name to a series of rocks in the Rocky Mountains of southeastern British Columbia which contained Jurassic-Cretaceous floras. Sir William used field descriptions by G. M. Dawson who had referred a succession of sandstones interbedded with shales, shaly sandstones, occasional conglomerates, and coal seams to the Kootenie Series. G. M. Dawson published his findings in 1886 and formally named Early Cretaceous rocks which crop out along the Rocky Mountain Front Range of Alberta as the Kootenie Series. Complete sections were not provided but the position of the Kootenie Series in the stratigraphic column was determined by its placement relative to overlying beds and by fossil plant remains (G. M. Dawson, 1886, p. 161B-164B).

Weed (1892) applied the name Kootenie to a sequence of coal-bearing interbedded sandstones and shales in the Great Falls area of Montana which overlie marine Jurassic rocks. He assigned an impure limestone containing brackish or fresh water gastropods to the top of the formation. In 1899 Weed discarded the name Kootenie for a sequence of alternating beds of sandstone and shale with a coal seam lying east of Great Falls, Montana; instead, he called the sequence the Cascade Formation. These rocks contained Early Cretaceous fossil leaves resembling those of the Kootenie Series of Canada. Fisher (1908, p. 79-81) called all the sedimentary rocks in the Great Falls-Judith Basin area of Montana which lie between the Jurassic Morrison Formation and the Late Cretaceous Colorado Group, the Kootenai Formation. Calkins and Emmons (1915,

p. 9) and Pardee (1917, p. 203) applied the name to a similar sequence of Early Cretaceous sandstones and variegated shales in the Drummond area of Montana. They used the new spelling proposed by Fisher; the Kootenai Formation is the accepted name for Early Cretaceous rocks of Montana. There is no type locality. The formation is named after the Kootenie Indians of south-central Alberta.

Up to about 1965 field investigations of the Kootenai in Montana were involved with the definition of the upper and lower boundaries. Cobban (1945, p. 1268) described about 1000 feet (305 m) of Early Cretaceous continental deposits which rest unconformably on Jurassic rocks in the area of the Sweetgrass arch of north-central Montana. He considered the rocks to belong to the Kootenai Formation and the lower contact to be at the base of the Cut Bank sandstone or the younger, Sunburst sandstone. Both of these sandstones are notable marker beds that lie either on the finer grained sedimentary rocks of the underlying Morrison Formation or on marine Jurassic rocks (Cobban, 1955, p. 107). Robinson (1963, p. 58) described the lower boundary of the Kootenai in the Three Forks area (Montana) to be a thick-bedded orange sandstone containing much chert. This sandstone has a distinctive "salt-and-pepper" appearance because of an admixture of gray and black chert, and jasper, with quartz; throughout southwestern Montana it is taken to be the base of the Kootenai and is locally called the Third Cat Creek sandstone, or where locally it is a chert pebble conglomerate, the Pryor Conglomerate (Suttner, 1969, p. 1393-1394). The upper boundary is considered to be at the top of the persistent and widespread gastropod-bearing limestone first described by Weed (1899, p. 310; also Cobban, 1955, p. 107; Suttner, 1969, p. 1394).

The gastropod limestone covers an area of about 16,000 square miles (40,960 km^2) in southwestern Montana. It has a wedge geometry with a maximum thickness of 200 feet (61 m) in its westernmost exposures. The unit thins systematically eastward to a north-south depositional zero edge east of the Bridger Range (Paine and Suttner, 1971). Where the gastropod limestone is absent, the top of the Kootenai is taken to be the medium- to fine-grained basal quartz arenite of the Colorado Group, or the first appearance of dark brown to black shale and siltstone of the lower Colorado Group (Suttner, 1969, p. 1394).

Other important contributions concerning the Kootenai Formation in southwestern Montana include a discussion of the petrography and stratigraphy of the unit in the Clark Fork Valley area near Drummond by Gwinn (1965). Robinson (1963) described the Kootenai as it occurs in the Three Forks area, and Mann (1954) described the formation in the Gravelly Range. The most comprehensive study of the Morrison-Kootenai petrography and stratigraphy with tectonic implications in southwestern Montana is by Suttner (1966, 1969), parts of which will be discussed in the section on the geologic history of the thesis area.

Regional Distribution and Correlation

The Kootenai Formation is a prominent unit that occurs throughout western and southwestern Montana. The Kootenai Formation forms a clastic wedge which thickens markedly in western Montana where the Morrison is thinnest. The greatest thickness is in excess of 1400 feet (427+ m) west of Anaconda, Montana and the formation thins eastward (Suttner, 1969, p. 1399-1400, figure 7). The basal sandstone is present throughout Montana (Suttner, 1969, p. 1394), whereas the gastropod limestone can be traced eastward to a north-south-trending depositional zero edge east of the Bridger Range (Paine and Suttner, 1971) where it grades laterally into sandstone and mudstone typical of the Blackleaf Formation of the Colorado Group (James, 1977).

The basal Kootenai sandstone has stratigraphic and lithologic equivalents in the basal Cloverly Formation of central, eastern, and northern Wyoming and central-southern Montana, the Lakota Sandstone of western South Dakota and eastern Wyoming, the Buckhorn Conglomerate of central-eastern Utah, and the Burro Canyon Conglomerate of the four-corners region of Colorado, New Mexico, Arizona and Utah (Suttner, 1969, p. 1394). The gastropod limestone can be correlated with the Peterson and Draney Limestones of the Gannett Group of eastern Idaho and western Wyoming (Holm et al., 1977, p. 259-260; and references therein) and with the lower Blairmore Formation of Canada (Brown, 1946; Peck, 1957, p. 9).

Distribution and Topographic Expression

The distribution of the Kootenai Formation parallels the belt of Mesozoic rocks in the southeastern corner of the thesis area. Except for some minor faulting the formation otherwise occurs as a continuous belt extending across the southeastern part of the map in the S½, Section 21, NE¼, Section 28, N½, Sections 27 and 26, T. 15 S., R. 9 W.

The basal salt-and-pepper cherty sandstone and chert pebble conglomerate forms a discontinuous low ridge up section from the Morrison swale. The gastropod limestone is a very distinct marker bed which forms a continuous outcrop either as a low ridge or ledge, or as a resistant unit underlying rounded hilltops with a very thin to nonexistent soil cover. The middle part of the formation forms low, soil-covered slopes underlain by variegated and non-resistant shales and mudstones with local isolated resistant outcrops of sandstone or limestone.

The Kootenai Formation overlies the Morrison Formation with an apparent unconformity. Where the Morrison is absent the Kootenai rests directly on the Ellis Group (McMannis, 1965, figure 10) or pre-Jurassic rocks north of Lima, Montana (Moritz, 1951, figure 18). The Kootenai is overlain by the Colorado Group east of a line which passes roughly north-south through Butte, Montana. In southwestern Montana the Kootenai is overlain by strata equivalent to the Aspen Formation of Scholten et al. (1955, p. 368) with a gradational and apparent conformable contact. Where pre-Tertiary erosion has removed the Aspen (?) Formation, the Beaverhead Formation rests with an angular unconformity on the Kootenai (Lowell and Klepper, 1953, p. 241).

Thickness and Lithology

Gwinn (1965, p. 36-37; and references therein) was able to distinguish four mappable units of the Kootenai Formation in the Clark Fork area of Montana. These are informally called the lower clastic member, the lower calcareous member, the upper clastic member, and the upper calcareous member. The lower clastic member is the regionally recognized basal chert pebble conglomerate or the salt-and-pepper medium- to coarse-grained sandstone. The lower calcareous member consists of dark gray and varicolored siltstones, shales, mudstones and minor shales ranging in thickness from less than 1 foot to 20 feet (0.3-7 m). The upper clastic member ranges in thickness from 400 to 500 feet (122 to 153 m) and consists of variegated shales, mudstones, and siltstones.

The upper calcareous member, better known as the gastropod limestone, is a ridge-forming gray biomicritic limestone attaining a maximum thickness of 232 feet (71 m) west of Drummond, Montana. Suttner (1969, p. 1394) recognized the four members elsewhere in southwestern Montana. He found that the two clastic members have a distinct mineralogy that persists farther east; the lower calcareous member is not persistent and the gastropod limestone can only be traced as far east as Bozeman, Montana (Suttner, 1969; Holm et al., 1977, figure 1). Scholten et al. (1955, Table 1) described the Kootenai Formation in the Lima region as consisting of brown and reddish shales interbedded with gray salt-and-pepper sandstones overlying a basal cherty conglomerate and underlying a white coquina. The lower calcareous member is not present in the area.

The Kootenai Formation in the thesis area is approximately 275 m thick. It disconformably overlies the Morrison Formation and underlies the Colorado Shale with an apparent conformable and gradational contact. The formation consists of a basal conglomerate and cross-bedded medium- to coarse-grained sandstones, with an upper, very distinctive gastropod and pelecypod coquinoid limestone. The middle part of the formation is made up of mudstones with discontinuous lenses of limestone and medium- to coarse-grained sandstone.

The basal unit consists of cross-laminated chert pebble conglomerate and medium- to coarse-grained cherty sandstone. The laminations show normal grading and the unit as a whole grades upward to a fine-grained sandstone. The conglomerate and coarse-grained sandstone are light olive gray (5 Y 6/1) and are composed of approximately 40% chert and 60% quartz. The pebbles are predominantly chert, ranging in size from 2 to 8 mm with an average median diameter of 5 mm, and rare rounded grayish orange (10 YR 7/4) non-calcareous siltstone clasts. The chert pebbles are light olive gray (5 Y 6/1), olive gray (5 Y 4/1), dark greenish gray (5 GY 4/1) and brownish gray (5 YR 4/1) in color. Eighty-five percent of the pebbles are angular and the rest are rounded. The matrix consists of silt-sized to coarse-grained sand and granule-sized angular to subangular chert and quartz grains. It is this admixture of quartz and chert which gives the rock its unique salt-and-pepper appearance. The basal unit is generally moderately to poorly indurated and forms low, discontinuous ridges. Thicknesses of the basal unit are variable but the ranges are approximately 6-10 m. The top of the unit is intercalated with variegated mudstones which make up the middle of the formation. An outcrop of fine-grained quartz arenite, located approximately 15 m above the base of the formation in S½, Section 23, is a good example of the intercalated sandstones of the upper part of the basal unit. This sandstone, approximately 2 m thick, is pinkish to yellowish gray (5 YR 8/1 to 5 Y 8/1) on a fresh surface, and weathers to light gray (N 8) slabs. The quartz grains are poorly sorted, angular, and are in framework support with calcite cement.

The middle part of the formation consists of about 200 m of variegated mudstones with lenses of sandstone and limestone. The mudstone is non-resistant but exposures in stream cuts reveal variegated calcareous siltstones that are moderate reddish brown (10 R 4/6) with light bluish gray (5 B 7/1) mottling, moderate medium light gray (N 6), and moderate reddish brown (10 R 4/6). About 180 m above the base of the formation is a laterally discontinuous ledge-forming unit of very light gray gastropod coquinoid limestone. Above the limestone is an exposure of a fossiliferous, medium-grained, calcareous, salt-and-pepper sandstone approximately 1.5 m thick. The sand particles are poorly sorted angular grains of chert (30%) and quartz (70%); there is an occasional moderately reddish brown (10 R 4/6) mudstone clast. The fossils include whole and fragmented gastropod and pelecypod shells. The entire outcrop is about 14 m thick.

The top of the formation consists of 35 to 65 m of a distinct molluscan limestone interbedded with fine- to medium-grained salt-and-pepper sandstone. The limestone consists of thin-bedded, packed, gastropod and pelecypod beds, and mixed pelecypod-gastropod coquinoid spar that is medium olive gray (5 Y 5/1) on a fresh surface, weathering to light olive gray (5 Y 6/1). The sandstone is calcareous and consists of moderately well sorted fine-grained particles of angular chert (25%) and quartz (75%) with calcite cement.

Fossils and Age

Stanton (1903) collected an invertebrate fauna in the upper Kootenai near Harlowtown, Montana and concluded that the forms were not older than Early Cretaceous. His assemblage is still accepted as the standard Early Cretaceous fresh water molluscan fauna in North America (Yen, 1951a, p. 1). Brown (1946) compared the flora as it existed above and below the unconformity between the Kootenai and Morrison Formations and concluded that the upper part of the Kootenai is Early Cretaceous in age. Yen (1951a) studied additional faunal collections from Harlowtown and further comfirmed an Early Cretaceous age for the Kootenai. Peck (1957) stated that the Kootenai is Aptian in age. He based his conclusion on the presence in the upper Kootenai of the fossil charophyte *Atopochara trivolis* and *Clavator harrisi*, both of which can be correlated with Aptian strata of Europe, Syria and North Africa. The widespread presence in the upper Kootenai of Eupera onestae (Katich, 1951), a characteristic Aptian pelecypod, supports the Yen and Peck analyses.

The accepted age for the Kootenai is Aptian. The contact between the Kootenai and Morrison Formations represents a hiatus which spans Portlandian and Neocomian time. Because the Aptian fossils were collected in the upper gastropod limestone, and there is as much as 1500 feet (458 m) of Kootenai strata beneath the limestone, Suttner (1969, p. 1395), suggests that the unconformity is suspect. He draws no specific conclusions, but suggests more physical and paleontological evidence is needed to confirm the disconformity.

Two genera of gastropods, Gyraulus and Circamelanie, were tentatively identified by me from the gastropod limestone.

Of perhaps some historical significance was the discovery in 1974 of a fossil moss, Diettertia montanensis, in shales of the Kootenai Formation by Brown and Robinson (1979, p. 170). The fossil moss possibly represents the oldest known gametophyte recorded from the Mesozoic of North America.

The Colorado Shale

Lying above the Kootenai Formation in the thesis area is approximately 200 m of variegated mudstones with discontinuous beds of limestone, sandstone, and siltstone which will herein be called the Colorado Shale. Further subdivision of the group is not possible. Exposures are poor and lithologic similarities with formations named in Wyoming, Idaho, and elsewhere in southwestern Montana make applying a name to this unit difficult.

Scholten et al. (1955, p. 368) tentatively mapped as the Aspen Formation all Cretaceous post-Kootenai beds in the Lima region. These beds consist of shale, siltstone, salt-and-pepper sandstone, some lithographic limestones, interbedded bentonite and porcellanite, and volcanic breccia in the upper part. The Aspen Shale is a late Early Cretaceous formation which is considered a member of the Mancos Shale in southwestern Wyoming, eastern Idaho, and northeastern Utah (Keroher, 1966, p. 154).

Moran (1971, p. 126) assigned rocks which occupy a similar stratigraphic position, approximately 80 km east in the Centennial Range, to the Thermopolis Formation. Moran described about 800 feet (244 m) of marine, yellowish brown, well sorted, thin-bedded, and thinly cross-bedded quartz arenites interbedded with fissile, olive black, and carbonaceous mudstones containing ferruginous concretions. Mann (1954, p. 32) described an estimated 2000 feet (610 m) of Late Cretaceous rocks in the Gravelly Range. He tentatively correlated the lower 200 feet (61 m) of black shale and reddish brown and yellowish brown fine-grained sandstone containing Inoceramus with the Thermopolis Formation. The upper 1800 feet (549 m) of alternating shale, sandstone, and claystone was correlated with the Mowry Shale of northeastern Wyoming.

In the Teepee Creek Quadrangle, which encompasses the northwestern part of Yellowstone National Park and adjoining parts of Montana, Witkind (1969, p. 37) tentatively correlated Early Cretaceous age rocks with the Thermopolis Shale. He recognized two lithologic units: a basal sandstone member, and an upper dark gray shale member. The lower sandstone unit was correlated with the Rusty Beds of the Thermopolis Shale of central, northern, and western Wyoming; the upper member apparently can be correlated with parts of the Thermopolis Shale in Wyoming and Montana and the Mowry Shale of south-central Montana. The Thermopolis Shale is time transgressive, and proper correlation appears to be difficult at best.

Peters (1970, p. 48-62) described over 6000 feet (1830 m) of Early Cretaceous rocks in the foothills of the Pioneer Range north of Dillon, Montana. He defined three members which could be correlated with the Colorado Group near Drummond, Montana, and on the Sweetgrass arch. The lower, or first member consists mostly of 1350 feet (412 m) of dull hued shale and argillite with subordinate amounts of sandstone and porcellanite. The second member is 1300 feet (397 m) thick and is composed of a basal pebbly sandstone which grades upward into interbedded variegated siltstone, shales, and argillites. The third member consists of interbedded siltstone, shale, and sandstone with local conglomerates, and pebbly sandstones.

Lowell (1965) tentatively assigned to the Early and Late Cretaceous Colorado Shale a single exposure of 150 feet (46 m) of yellowish brown sandstone, siltstone, and dark gray shale located approximately 50 km north of the thesis area along the Beaverhead River.

The original reference to the Colorado Group was made by Hayden in 1876 (p. 45) to black shales with a few laminated sandstones which occur along the Front Range of Colorado. The name was applied by early workers to correlative marine shales in central and northwestern Montana where they were ranked as a formation and called the Colorado Shale (Keroher, 1966, p. 891). Collier and Carthcart in 1922 (p. 172) raised the Colorado Shales to group status in the Little Rocky Mountains. The formations included were the Thermopolis, Mowry, and Warm

Creek Shales. In 1951 Cobban divided the Colorado Shale into lithologic units that could be correlated with the standard Cretaceous section of the Black Hills, one formation of which included the Mowry Shale. Cobban et al. (1959) later revised the Colorado Group on the Sweetgrass arch by dividing it into a lower unit, the Blackleaf Formation, and an upper unit called the Marias River Formation. The Blackleaf Formation is made up of four members of Early Cretaceous age (Albian), whereas four members described in the Marias River Formation are Late Cretaceous (Cenomanian) in age. The Colorado Group on the Sweetgrass arch lies unconformably above the Kootenai Formation and conformably underlies the Telegraph Creek Formation of the Montana Group.

Gwinn (1965) described the Colorado Group in the Clark Fork Valley of west-central Montana. He subdivided the Colorado Group into the Blackleaf Formation of Early Cretaceous age and the Coberly, Jens, and Carter Creek Formations of the Late Cretaceous.

The Colorado Group, with discrete correlative formations, is generally considered to be of Early to Late Cretaceous age and can be found in Colorado, Idaho, Iowa, Kansas, Nebraska, New Mexico, North and South Dakota, as well as in Montana and Wyoming (Keroher, 1966, p. 890). In a correlation chart, Suttner (1969, p. 1396-1397) included the Thermopolis and Mowry Shales, and the Frontier Formation in the Colorado Group in south-central Montana. Farther west in Montana he followed the nomenclature of Gwinn (1965) and Cobban et al. (1959); in the Lima, Dillon, and Melrose areas of Beaverhead County he considered any Early to Late Cretaceous rocks overlying the gastropod limestone of the Kootenai Formation to belong to the Colorado Group.

On an isophach map of western Montana, McMannis (1956, Figure

11) extended the lower Colorado into southwestern Montana. In extreme northwestern Wyoming and adjacent parts of Montana the lower Colorado units were mapped by McMannis as being non-marine in origin. Scholten et al. (1955), Moran (1971), Mann (1954), Witkind (1969), Gwinn (1965), and Peters (1970) described Colorado Group equivalent strata in southwestern Montana as having been deposited in both fluvial and paralic environments with minor strand line fluctuations.

Within any marginal marine-terrestrial environment the sedimentary rocks will reflect the narrow limits of depositional environments and the facies will be restricted in their vertical and horizontal continuity. Regional correlation of such rocks is difficult and is achieved by fossil and lithologic associations which may represent changes in depositional environments.

In continuity with the regional distribution of the Colorado Group and the lithologic similarities of the Lima region (Scholten et al., 1955) and known Colorado Group formations elsewhere, I propose that the Early Cretaceous rocks above the Kootenai Formation in the Tendoy Range be considered part of the Colorado Group. More work is required to further clarify and define the Colorado Group in the area. Until then, and for purposes of this thesis, the unit in the Little Sheep Creek area will be called the Colorado Shale.

Distribution and Topographic Expression

The formation forms a continuous belt of strata which parallels the Kootenai Formation in the southeastern part of the thesis area in the NE¼, Section 28 across the NE½, Section 26 and 27, T. 15 S., R. 9 W. It forms gentle slopes which are continuous with the Kootenai slopes that end upsection and uphill at the marked steepening in gradient at the base of the overlying Beaverhead Formation. Resistant lenses of sandstone, limestone, and siltstone underlie spurs of land which often rise above the surrounding slopes.

Thickness and Lithology

Scholten et al. (1955, p. 268) state that the contact with the Kootenai Formation is gradational and conformable in the Lima region. Evidence in the thesis area indicates that this is possible. Rocks of similar lithology occur in the lower half of the Colorado Shale and throughout the Kootenai Formation. It can be postulated that the sequence of Cretaceous rocks beneath the Beaverhead Formation represents continuous deposition and that the persistent bed of gastropod-bearing limestone is a convenient marker for the top of the Kootenai Formation.

The contact with the gastropod limestone is covered. The thickness of the formation is unknown because of pre-Tertiary erosion prior to deposition of the Beaverhead Formation. In the thesis area the Colorado Shale is at least 218 m thick.

The formation is made up mostly of non-resistant mudstones, and is generally covered. A sample of the mudstone was collected from a stream bank and consists of a medium bluish gray (5 B 5/1) and greenish gray (5 G 6/1) calcareous non-fissile silty claystone. Resistant outcrops of laterally discontinuous limestones, sandstones, and siltstones occur from place to place throughout the formation.

A low ridge underlain by a thin-bedded fine-grained sandstone unit 9 m thick occurs approximately 63 m above the gastropod limestone. The sandstone is a non-calcareous, thinly laminated mixture of well sorted angular grains of monocrystalline quartz (85%), chert (13%), and unidentifiable detrital grains (2%) that is grayish orange (10 YR 7/4) on a fresh surface and pale yellowish brown (10 YR 6/2) when weathered. The rock weathers to slabby talus.

Cropping out at different levels throughout the lower half of the formation is a calcareous medium- to coarse-grained sandstone that has the salt-and-pepper appearance caused by a mixture of black and brown chert, and quartz. The good exposures of the sandstone exhibit festooned trough cross-bedding, or planar cross-bedding with the cross-bed sets up to 0.7 m thick. The cross-bed sets range in thickness from 8 mm to 15 mm with the bedding plane angles averaging 18°. Paleocurrent directions could not be determined with confidence. The outcrops are approximately 3-4 m thick, are light gray (N 7) and light olive gray (5 Y 6/1) on fresh surface and weather to dusky yellow (5 Y 6/4) and light olive gray (5 Y 6/1).

The limestones are thin- to thick-bedded pelecypod-rich or gastropod-rich coquinas. The gastropod limestone is a well sorted packed biosparite which is yellowish gray (5 Y 8/1) on fresh surfaces and weathers to light olive gray (5 Y 6/1). The rock is dense, apparently thick-bedded, and weathers out to block debris. The pelecypod limestone is packed with shell fragments aligned parallel to bedding and contains minor amounts of quartz silt and ferruginous stain. The rock is pale yellowish brown (10 YR 6/2) on a fresh surface and moderate yellowish brown (10 YR 5/4) when weathered.

About 10 m along strike, with an outcrop of medium-grained salt-and-pepper sandstone, is a resistant spur underlain by a dark yellowish orange (10 YR 6/6) and grayish orange (10 YR 7/4), thin-bedded, fossiliferous, calcareous siltstone. On a fresh surface the rock is pale yellowish brown (10 YR 6/2). The fossils are mostly gastropods with some pelecypods and carbonized plant stems loosely and randomly arranged in the silty matrix. There are no internal structures in the rock. The outcrop is not exposed well enough to determine bedding characteristics and the relationship with the sandstone.

Fossils and Age

Cobban (1959) considered the Blackleaf Formation in central Montana to be Early Cretaceous (Albian) in age. His conclusions were based on an *Inoceramus comancheanus* faunal zone in the Taft Hill Member and a *Neogastroplites* species zone in the Bootlegger Member. There is a slight hiatus, representing part of the early Albian, between the Kootenai and the basal Colorado Group. There also is a disconformity between the Blackleaf Formation and the Marias River Shale which represents the early Cenomanian. The Marias River Shale is of Late Cretaceous age, having been deposited during late Cenomanian to early Santonian time. The contact with the overlying Telegraph Creek Formation does not represent a break in time.

Witkind (1969, Table 4 and p. 38) correlated the strata he tentatively called the Thermopolis Shale in the Teepee Creek Quadrangle of Montana with lithologically similar rocks in Wyoming and Montana which are now considered to be Early Cretaceous in age (Suttner, 1969, Figure 2; Cobban et al., 1959, Figure 3). Mann (1965, p. 34) considered the undifferentiated rocks in the Gravelly Range to be of Late Cretaceous age. He used fossil and lithologic similarities to correlate the strata with the Thermopolis and Mowry Shales of northwestern Wyoming, and the

Bear River Formation and part of the Aspen Shales in southwestern Montana. Moran (1971, p. 129) did not find any fossils in the Thermopolis Shale in the Centennial Range. He did correlate the Thermopolis Shale with similar Early Cretaceous strata in western Wyoming.

Scholten et al. (1955, p. 368) found Cissites (or Platanus) affinis, Betula beatriciana, and bone fragments of crocodiles or small dinosaurs within the strata they called the Aspen Formation in the Lima region. The fossils are apparently chronologically inconclusive but they do indicate lacustrine and swampy environments. Scholten et al. considered the Aspen Shale to be Late Cretaceous in age. Ryder and Ames (1970, p. 1161) found palynomorphs in the "Aspen" Formation stratigraphically below the Beverhead Formation near Monida, Montana that provide an Albian age for the formation. The palynomorphs also indicate swampy conditions of deposition.

The Beaverhead Formation

The Beaverhead Formation was first named by Lowell and Klepper in 1953. They applied the name to a thick sequence of conglomerate, sandstone, siltstone, and limestone that crops out over a 400 square mile (1037 km^2) area in Beaverhead County, Montana and adjacent parts of Idaho. They designated the type area to be at the mouth of McKnight Canyon six miles (10 km) west of Dell. At this locality the top of the formation has been eroded and the base of the formation has been faulted; however, approximately 9700 feet (2959 m) of the formation is well exposed and can be divided into three mappable members. They are an upper and a lower conglomerate member separated by a middle limestone and sandstone member. Elsewhere, the formation may be as much as

15,000 feet (4575 m) thick (Ryder and Scholten, 1973, p. 774).

Wilson (1967) mapped an area between Lima and Monida as part of a Ph.D. thesis (Ryder and Scholten, 1973, p. 774). At that time he had elevated the Beaverhead Formation to group status. Later, in 1970, Wilson dropped the group nomenclature and distinguished two Late Cretaceous formations in that area: the Beaverhead Formation and the Monida Sandstone. Ryder (1967) described eleven intertonguing "lithosomes" within the Beaverhead which are not laterally continous for more than 25 miles (40 km).

In 1973 Ryder and Scholten published a comprehensive study on the sedimentology of the Beaverhead Formation and its Laramide orogeny syntectonic significance. At that time they chose to drop Ryder's (1967) lithosome concept, and Wilson's (1970) Monida Sandstone as a formational unit and described, instead, five major lithofacies. Several of the lithofacies could be subdivided into additional compositional entities creating a total of ten mappable units. The subdivisions mutually intertongue and have a lateral continuity of between one and 20 miles (1.6 to 32 km). The ten lithofacies recognized are the:

Divide quartzite conglomerate unit
Kidd quartzite conglomerate unit
Divide limestone conglomerate unit
Lima limestone conglomerate unit
McKnight limestone conglomerate unit
Chute sandstone unit
Monida sandstone unit
Snowline sandstone unit
McKnight limestone-siltstone unit

The Beaverhead Formation is syntectonic with the Laramide orogeny. Ryder and Scholten (1973) reconstructed the paleogeography and the timing of the beginning of the Laramide orogeny in southwestern Montana by integrating knowledge involving the lithofacies, the timing of structural elements which involve the Beaverhead Formation, and the Albian (Early Cretaceous) to late Paleocene time bracket established for the formation by Yen (in Lowell and Klepper, 1953, p. 240; Ryder and Ames, 1970).

At most places the Beaverhead rests with distinct angular unconformity on rocks as young as the Colorado Shale (or equivalent strata). The contact between the Colorado Shale and Beaverhead Formation is, at some places, in obvious discordance (Ryder and Scholten, 1973, p. 781). At other localities, the upper part of the Colorado Shale forms a transitional boundary with the quartzite conglomerate of the Beaverhead. There is an apparently increasing hiatus between the two formations going from south to north in the direction of the Blacktail-Snowcrest uplift. The Beaverhead Formation is unconformably overlain by rocks no older than the late Eocene.

Age and Correlation

Studies by Yen (in Lowell and Klepper, 1953, p. 240) and Wilson (1970) assigned the Beaverhead Formation a Late Cretaceous to Paleocene age. Ryder and Ames (1970) used palynomorphs to assign a late Albian to late Cenomanian age to the basal Beaverhead. The middle part of the formation is assigned a middle Turonian to late Coniacian age. Floral associations in the upper third of the formation indicate a late Late Cretaceous to mid-Paleocene age for the upper part. A considerable thickness of quartzite conglomerate above the highest datable units indicates that Beaverhead deposition possibly continued into the early Eocene (Ryder and Scholten, 1973, p. 783).

The Beaverhead Formation is correlative, or partly correlative to clastic rocks in south-central Montana and northwestern Wyoming, including the Frontier Formation, Cody Shale, Eagle Sandstone, Livingstone Group, Fort Union Formation, Bacon Ridge Sandstone, Mesaverde Formation, Meeteetse Formation, the Harebell and Pinyon Conglomerates (Ryder and Scholten, 1973, p. 783), and the Willow Creek Formation in north-central Montana (Honkala, 1955, p. 127).

Ryder and Scholten (1973, p. 774) list other studies involving the Beaverhead Formation and other Laramide syntectonic deposits in east-central Idaho and northwestern Wyoming.

Distribution and Topographic Expression

The outcrop pattern of the Beaverhead Formation in the Little Sheep Creek area is one of distinct angular unconformity over rocks which range in age from the Pennsylvania to the Late Cretaceous. It forms a continuous pattern across the southern part of the area and a discontinuous belt of rock between Triassic rocks underneath and Mississippian rocks above in the northwestern part of the area (Plate I).

The formation is in depositional contact with the Colorado Shales in Sections 26, 27, and 28, and with the Quadrant Formation in Sections 19, 29, and 30, T. 15 S., R. 9 W. Part of the formation extends north from Sections 28 and 29 into Section 21, overlapping the sequence of Permian, Triassic, Jurassic and Cretaceous rocks with angular discordance.

In the northwest sector of the map the Beaverhead Formation lies on an erosional surface that describes a broad swale which cuts across the Thaynes Formation and down into the Woodside Formation. It is overlain by the Mississippian rocks involved in the Medicine Lodge thrust. In Sections 29 and 30, T. 15 S., R. 9 W. the Beaverhead is overlain by

Mississippian rocks emplaced along the Medicine Lodge thrust and by fresh water limestones of the Round Timber Member of the Medicine Lodge Beds.

The Beaverhead Formation is, for the most part, a cobble conglomerate in a poorly sorted calcareous sandy matrix. It is a poorly indurated rock and weathers to grass-covered slopes littered with rounded quartzite cobbles and pebbles, and angular quartzite debris on a soil cover that is pale reddish brown or moderate yellowish brown in color. The well-indurated and well-rounded quartzite cobbles which litter the eroded surface of the Beaverhead Formation have been reworked by Recent erosional processes so that the Beaverhead debris is ubiquitous not only on its erosional surface but also scattered over a large area. The outcrop pattern cannot be mapped by float alone.

Outcrops of the formation can be observed in the NE¼, Section 31, in a stream cut uphill from Two Springs in Section 32, T. 14 S., R. 9 W., under mountain mahogany bushes on the north slope of a drainage in the SE¼, Section 30, T. 15 S., R. 9 W., and in the NW¼, SW¼, NE¼, Section 26, T. 15 S., R. 9 W.

Thickness and Lithology

The Beaverhead Formation attains a maximum thickness of 245 m at Little Sheep Creek. The thickness is merely an approximation. The erosional surface upon which the Beaverhead was deposited provides too irregular a contact to be able to estimate or measure thickness. Furthermore, an indeterminate amount of the formation was removed or covered by the Medicine Lodge thrust, and Recent erosion has removed an indeterminable amount from the Beaverhead along the Continental Divide.

Two of the lithofacies described by Ryder and Scholten (1973) are present in the Little Sheep Creek area. They are the Divide Quartzite conglomerate and the Divide Limestone conglomerate. The outcrop pattern of the Beaverhead is very sporadic. For this reason the boundaries of the two lithologies have not been delineated and they have been mapped as one with the Beaverhead Formation.

The Divide Limestone Conglomerate is exposed along the scarp of a small slump in the NW¼, SW¼, SW¼, NE¼, Section 26, T. 15 S., R. 9 W. The overall color of the exposure is a yellowish gray (5 Y 8/1). The rock is a cobble conglomerate in which 95% of the clasts are limestone, and 5% are metaquartzite. The limestone cobbles are typical of the Mississippian rocks found in the region. Fifty percent of the granule-sized particles are quartzite, 50% are limestone. The matrix is a fine to medium sand cemented with calcite. There is no apparent bedding and the cobbles are randomly arranged without imbrication.

The Divide Quartzite Conglomerate is best exposed in a stream in the NE¼, SE¼, NW¼, Section 20, T. 15 S., R. 9 W. and on a cliff face in the NE¼, Section 31, T. 14 S. R. 9 W. The overall color of the unit is moderate yellowish brown (10 YR 5/4) and the unit is made up of quartzite cobbles and pebbles in a sandy matrix cemented with calcite.

An exotic block of limestone is imbedded in the Beaverhead in the NE¼, Section 30, T. 15 S., R. 9 W. The block is 335 m long and is a maximum of 61 m wide. It consists of a fractured and brecciated, very thick-bedded, light gray (N 7) limestone interbedded with thin-bedded, pale yellowish orange (10 YR 8/6), calcareous, shaly siltstone and thin-bedded, yellowish gray (5 Y 7/2), silty limestone. A brachiopod extracted from the silty limestone bed was identified by McKenzie

Gordon (pers. comm.) of the United States Geological Survey to be Pugnoides cf. P. ottumwa (White), a fauna restricted to the Mississippian. The occurrence of this limestone block in the Beaverhead is similar to observations made by Ryder and Scholten (1973, p. 779) elsewhere in the Lima region.

THE TERTIARY SYSTEM

Round Timber Limestone

The Round Timber limestone is an informal name applied to a thick unit of fresh water limestone which crops out in the southern part of the thesis area (Figure 35). Scholten et al. (1955, p. 369) described Tertiary rocks which are found in the intermontane basin between the Tendoy and Beaverhead Ranges and gave them the informal name, Medicine Lodge beds. The beds are made up of a series of varicolored shales, bentonitic clays, sandstones, pebble conglomerates, fanglomerates, some lignite, and fresh water limestone. Because the dominant lithology recognized in the Little Sheep Creek area is the limestone, with only very subordinate basal conglomerate, I am informally naming the unit the Round Timber limestone for exposures north of Round Timber Spring in the NW¼, Section 29, T. 15 S., R. 9 W.

The maximum thickness of the Medicine Lodge beds in the intermontane basin to the south was estimated by Scholten et al. (1955) to be about 5000 feet (1525 m); a partial section of 2600 feet (793 m) was measured in the Big Sheep Creek basin.

Distribution and Topographic Expression

The Round Timber limestone in the Little Sheep Creek area overlies with distinct angular unconformity the Triassic Thaynes Formation in the E½ and NE¼, Section 17 (Figure 36), and also in parts of Sections 19 and 20, T. 15 S., R. 9 W. Limited erosional remnants of the unit overlie Permian and Triassic rocks in the S½ of Section 19, and excellent outcrops of the unit occur in the E½, Section 30, T. 15 S., R. 9 W., where

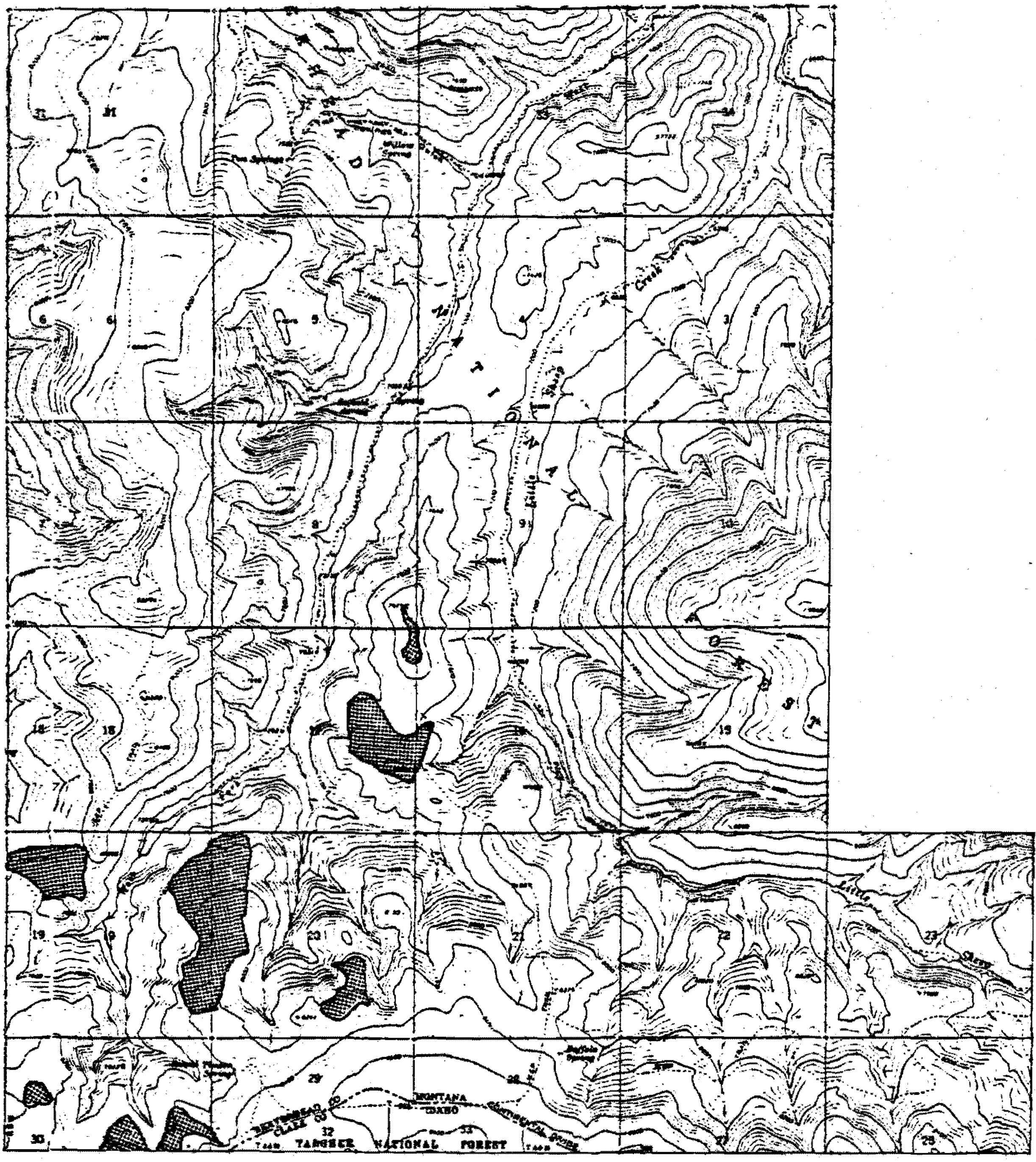


Figure 35. Distribution map of the Round Timber Limestone.

Figure 36. Round Timber limestone lying with angular unconformity on the Thaynes Formation. Here the Thaynes beds form part of the northern limb of a northward yielding overturned syncline. Picture taken looking NE from Section 30, T. 15 S., R. 9 W.

it overlies the Scott Peak and Beaverhead Formations, and is, in turn, overlain by the Edie School rhyolite.

The limestone is of variable thickness and is found at different elevations. The outcrops of the unit are on surfaces which have not been bisected by post-Laramide normal faults, so the distribution of the limestone most likely represents deposition on topographic irregularities (Figure 45). The limestone ranges in thickness up to 73 m, and the elevations at which it is found vary by as much as 185 m.

The Medicine Lodge limestone forms a very resistant, light colored, and nearly horizontal grass-covered surface littered with limestone slabs and blocks.

Lithology and Diagenesis

The Round Timber limestone consists of fenestral limestones underlain by pebble and cobble conglomerates.

The limestone is of a dense, hammer-ringing quality and is represented by two noticeable color variations. The more common variety is a light colored rock which is grayish pink (5 YR 8/2), very pale orange (10 YR 8/2), pale yellowish brown (10 YR 6/2), or grayish orange pink (10 YR 8/2) on fresh, broken surfaces. The rocks weather to shades of gray (N 6 to N 9) and pale yellowish brown (10 YR 6/2). The less common, darker variety occurs only as scattered erosional blocks overlying the lighter variety in the vicinity of the central half of the section line between Sections 19 and 20, and on the Continental Divide in the SE¼, Section 30, T. 15 S., R. 9 W. The darker variety is medium dark gray (N 4) on a fresh surface, and weathers to a medium gray (N 5) and pale yellowish brown (10 YR 6/2). It has a distinct petroleum odor on fresh break.

The limestone varies in outcrop pattern from being thickly bedded or massive, to locally thin-bedded and laminated. The best exposures of the unit occur in the SE¼, Section 19, T. 15 S., R. 9 W., where the outcrops are 6-7 m thick. The rocks are thick-bedded in the lower two-thirds of the outcrop (Figure 37), and thin-bedded to laminated in the upper third.

Peculiar to the Round Timber limestone is a fenestral texture which provides a high porosity to the rock. The fenestrae range in size from a fraction of a millimeter in diameter to as large as 10 cm. They are found randomly distributed in the thick- and thin-bedded limestone, or are confined to bedding planes in the laminated beds. The upper surfaces of some samples have clusters of fenestrae which resemble the molds of grass stems. The dark variety of limestone is crisscrossed on fresh, broken surfaces with the possible molds of matted grass roots (Figure 47).

The thick- to thin-bedded limestones are fenestral pseudosparites and sparse biopseudosparites with minor amounts of quartz silt and no internal sedimentary structures. The laminated limestone is an algal-mat boundstone. The limestones are generally fine- to medium-crystalline sparites. In thin section the rocks show various stages of aggrading neomorphism of micrite to pseudospar.

The components of the limestones are micrite, minor amounts of quartz silt (less than 1%), oncolites, micritic intraclasts, pelloids, rare gastropods, ostracods, the green algae <u>Chara</u>, and phosphorite (as pellets and fragments of <u>Lingula</u> shell fragments reworked from the underlying Triassic rocks). Generally the rocks have no internal sedi-

Figure 37. Light colored limestone, Round Timber limestone. Note homogeneous nature of the outcrop and the distribution of the fenestrae. Scale is 15 mm. SE¼, Section 19, T. 15 S., R. 9 W.

mentary structures; the components are thoroughly mixed as if bioturbated and the fenestrae are randomly distributed throughout. The laminated beds are the only limestones to exhibit a regular internal structure (Figure 38). In hand sample the laminae range in thickness from 1 to 4 mm and the thickness varies along strike within each lamina. The laminae are laterally discontinuous over a distance of 3 to 7 mm and intertongue with other laminae.

The laminae are made up of an accumulation of ooliths, small oncolites (2.5 mm median diameter), micrite intraclasts, and pelloids loosely packed in micrite. Each lamina forms a sequence of convex-up domes, 1-2 mm wide separated by narrower concave-up hollows. The dips and hollows are filled with allochems and some of the microlaminations are bifurcating. Immediately above the stromatolitic microlaminations are fenestrae 1 to 2 mm high and up to 2 cm long which separate one lamina from the next algally bound lamination (Figure 38). The laminations are algal-bound stromatolites with a microstructure of spaced laterally linked hemispheroids (Logan et al., 1964, p. 74).

The porosity in the limestones ranges from 20% to 60% with an average of 38%. Primary porosity of marine deposited micrite mudstones is about 50-70%. Assuming the original porosity of the freshwater micrites to have been about 50-70%, only one sample had a reduction of porosity to about 20%. The majority (70% to 90%) of the pores are primary and are fenestral; they are several magnitudes larger than normal interparticle spaces (Niem, 1979, unpublished lecture notes). A smaller proportion of the pores are secondary molds formed by the dissolution of oncolite laminations, or the decomposition of the vegetative part of encrusted Chara stems (Figure 44), or are well-preserved

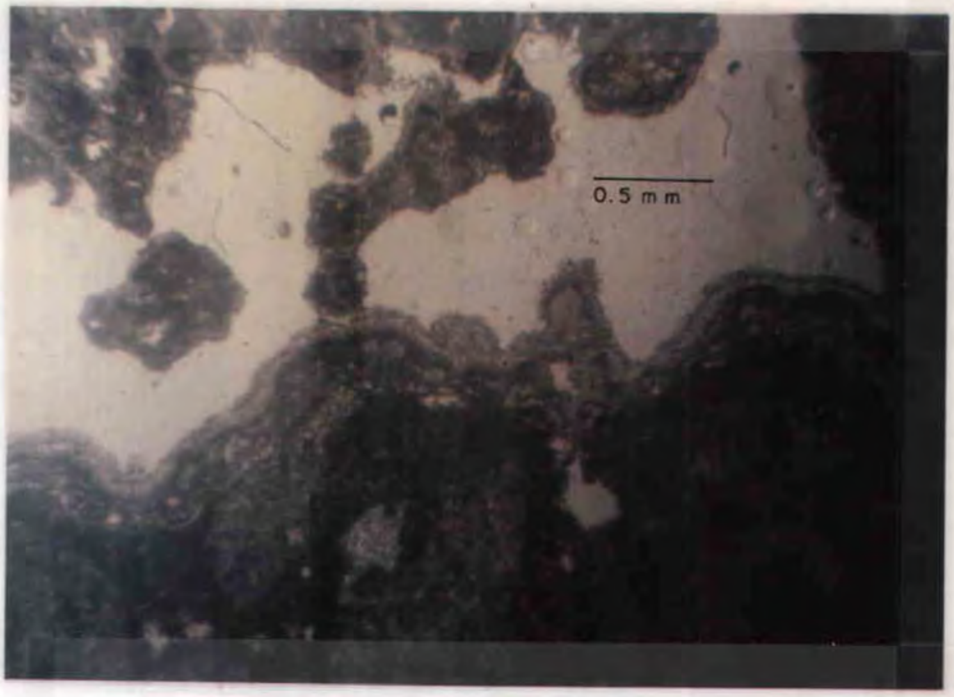


Figure 38. Photomicrograph showing relationship between stromatolitic microlaminations binding an accumulation of allochems loosely packed in micrite. Sample Kmb 79-73, SE¼, Section 19, T. 15 S., R. 9 W. Plane light.

burrows. Solution vugs occur in some samples.

Calcite is the cementing material and occurs in three morphological types: thin micrite fringes lining some pores, scalenohedral spar arranged perpendicular to pore walls, and equant spar. The most common cement is a fringe of micrite over which has grown the scalenohedral spar (Figure 39). Equant calcite spar is very rare and was only observed filling *Chara* molds (Figure 40). Scalenohedral spar has occluded less than one percent of the fenestrae in the laminated beds.

Aggrading neomorphism (Folk, 1965; Bathurst, 1975) has occurred in all of the samples studied. It varies from a partial neomorphism resulting in a clotted appearance in thin section, to a pervasive medium-crystalline pseudospar which has obliterated any primary texture or structures. In all cases, the neomorphic calcite has a light gray or light brown cloudy appearance caused by organic inclusions, which contrasts sharply with the clear calcite cement (Figure 40).

The base of the Round Timber limestone is marked, at places, by a thin-bedded polymict conglomerate. The conglomerate can be observed in outcrop in the SW¼, SE¼, SW¼, NE¼, Section 17, the NE¼, SE¼, NW¼, SE¼, Section 20, and in a stream bed in the NE¼, NW¼, SE¼, Section 30, T. 15 S., R. 9 W. Elsewhere the conglomerate is covered but its position is marked by cobble- and pebble-sized gravels which litter slopes stratigraphically below most occurrences of the limestone. No Pleistocene summit surface gravels were mapped in the thesis area. The composition of any Quaternary colluvium reflects the local bedrock lithology mixed with the ubiquitous metaquartzite debris weathered from the Beaverhead Formation. Because of a possible mixing of weathered Round Timber conglomerate with detritus from the Beaverhead, a pebble count was not made

Figure 39. Photomicrograph of micrite cement (m) over which has developed scalenohedral spar (s) normal to the pore wall. The laminations (l) are dust inclusions in the spar which outline the bottom of the pore space. The laminations may represent times during cementation of the pore when ground water carried in small amounts of clay which settled to the bottom of the pore without interrupting the development of the calcite spar. Sample Kmb 79-71, center, NE¼, NE¼, Section 19, T. 15 S., R. 9 W. Plane light.

on the basal Round Timber limestone. However, the lithoclasts in the gravels associated with the Round Timber limestone are better sorted, represent a smaller range of clast sizes, and are lighter in color than those of the Beaverhead. There also is a greater variety of lithologies than found in the Beaverhead Formation, with calcareous siltstone, calcareous very fine-grained sandstone, and coquinoid limestones representative of the underlying Triassic formations. Such gravels, therefore, where associated with the fenestral limestones, were mapped as representing the approximate location of the base of the Round Timber limestone.

The outcrops of conglomerate are thin and discontinuous (Figure 41); the following is a composite description taken from several locations. The overall coloration of the outcrops is a very pale orange (10 YR 8/2) tempered with yellowish brown and grayish orange hues of the polymictic components. The conglomerates are thin-bedded, discontinuous, and lens-shaped bodies which are intercalated with the fenestral limestones (Figure 42). The lenses do not exceed 1 m in thickness and acquire a lateral extent of at least 17 m. The contacts with the limestones are gradational, and there is a grading upward to finer material within each bed. The dimensions of the largest lithoclast measured are 21 cm x 15 cm x 10 cm; however, a majority of the lithoclasts are in the pebble-size range. Cobble-sized clasts are broken but have rounded edges, whereas the pebbles are well rounded; the granule-sized lithoclasts are angular.

These conglomerates do not fit neatly into any one classification system. They are a polymictic conglomerate in a micrite matrix which has undergone various stages of aggrading neomorphism to pseudospar.

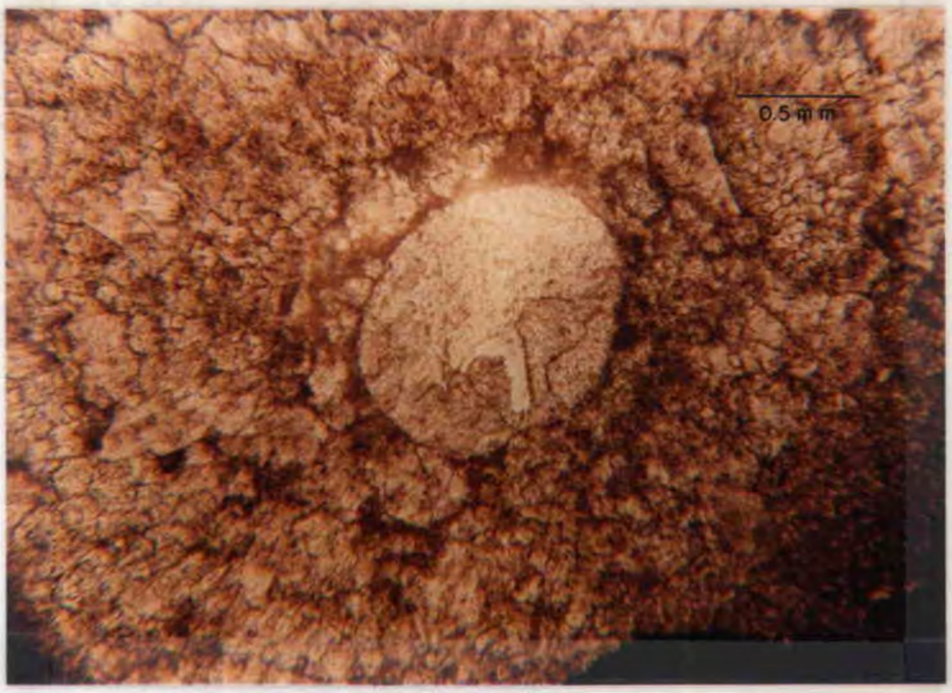


Figure 40. Photomicrograph. Equant spar filling Chara axis. Sample Kmb 79-71; center, NE¼, NE¼, Section 19, T. 15 S., R. 9 W. Plane light.

Figure 41. Basal Polymict conglomerate, Round Timber limestone. Clipboard is 38 mm long. SW¼, NE¼, Section 17, T. 15 S., R. 9 W.

Figure 42a. Basal conglomerate, Round Timber limestone. Lenses of conglomerate intercalated with fenestral limestones. SE¼, Section 20, T. 15 S., R. 9 W.

b. Close-up of Figure 42a. Scale is 15 mm.

Modal analyses of two samples are presented in Table 6. The volcaniclastic fragments, micritic mudstones, claystones, and oncolites are of particular interest because they are useful indicators of provenance and depositional environments.

The volcaniclastic fragments are a felted mass of plagioclase microlites (An_{25}) in a dark glassy (?) matrix. The An content is a mere approximation; the outlines of the microlites are not clear. The lithoclasts are well rounded, but the edges of the clasts are irregular and appear to have reacted with the calcite matrix. Scholten et al. (1955, p. 375-376) described basaltic and andesitic lavas and tuffs which underlie the Medicine Lodge beds in Big Sheep Creek basin about 5 km to the west. The volcanics are grouped under the heading of Medicine Lodge volcanics. The most common volcanics consist of microlites of plagioclase (andesite-labradorite) in a matrix made up of dust-sized particles of magnetite and augite. Another common variety is a porphyritic rock which contains phenocrysts of brown basaltic hornblende in a matrix of labradorite and hornblende microlites rimmed by alteration products, including calcite. It is conceivable that weathered products of the Medicine Lodge volcanics were incorporated into beach gravels of the ancient lake.

The limestones intercalated with the conglomerates are fenestral and have the texture reminiscent of petrified Swiss cheese (Figure 42b). The fenestrae are generally less than a millimeter in diameter and apparently follow bedding planes. Disrupting the primary texture of the rock are solution vugs with a maximum long dimension of 4 cm. The rock is light gray (N 7) in outcrop, but a refreshing pinking gray (5 YR 8/1) when broken. Microscopically the limestone is composed of

TABLE 6

Modal Analysis, Basal Conglomerate, Round Timber Limestone

Component	Sample Kmb 79-186	Sample Kmb 79-294
Calcite matrix (micrite, microspar, pseudospar)	70%	53%
Porosity	11	13
Clastic detritus (silt- and sand-size)		
quartz	4	--
collophane	1	--
Lithoclasts		
metaquartzite	6	14
chert	--	7
micrite intraclasts	5	5½
claystone intraclasts	2	2½
calcareous siltstone	1	3
oncolites and oolites	--	1
volcaniclastic	--	1
	100%	100%

Median diameter of lithoclasts range: 0.2 to 3.0 mm; average 1.06 mm.

Sample location:
79-186 - SW¼ SE¼ NW¼ SE¼ Section 20, T. 15 S., R. 9 W.
79-294 - SW¼ SE¼ SW¼ NE¼ Section 17, T. 15 S., R. 9 W.

about 90% micrite which has undergone partial neomorphism to microspar and pseudospar. The partial neomorphism has resulted in a clotted appearance in thin section. The remainder of the rock is composed of silt-sized detrital particles which consist of, in decreasing abundance, quartz, muscovite, zircon, collophane (pellets and Lingula shell fragments probably derived from underlying Triassic Formations), and plagioclase. The detrital material is dispersed throughout the rock, and is also concentrated in a thin lamina which transects one of the slides.

Porosity in the limestone is about 30%, whereas it is between 11% and 13% in the conglomerates. Calcite cement in the conglomerates occurs as micrite fringes, finely crystalline equant spar, and as isopachous scalenohedral spar. A rim of micrite is present on nearly all clasts and has two origins. The first is an algally induced adherence of micrite mud to the surface of clastic grains in the agitated waters which existed during deposition of the conglomerates as gravels. Several oncolitic structures were observed which had no visible nuclei. Many of the lithoclasts have a fringe of one to several concentric layers of micrite (Figure 43); a structure which Williamson and Picard (1974, p. 741) described in the Eocene Green River Formation as being oncolitic. The existence of algal structures elsewhere (Figure 38), and the association of carbonate mud with agitated conditions, make an organically induced deposition of the micrite rim a likely source. The second mechanism for the micrite rim is by algally induced diagenesis of the carbonate rock.

The drainage area for the lake was underlain by a great host of limestones and other calcareous sedimentary rocks. Fresh water entering

Figure 43. Photomicrograph. Oncolite, basal conglomerate, Round Timber limestone. Sample Kmb 79-294, SW¼, SE¼, SW¼, NE¼, Section 17, T. 15 S., R. 9 W. Plane light.

these rocks would be undersaturated with respect to calcium carbonate. Saturation would be reached relatively quickly (Bathurst, 1975, p. 445) depending on the chemical stability of the carbonate source and the chemistry of waters percolating through the system. Water entering the lake basin could be saturated with respect to calcium carbonate and thus would be the readily renewable source of calcite required for the cementation of the beach gravels.

An environment of fresh-water limestones and conglomerates undergoing diagenesis does not easily fit into some of the models provided which describe diagenesis of marine carbonates. Bathurst's (1975, p. 428) discussions on cementation of carbonates assume subaerial exposure of the rock and subsequent diagenesis in a fresh water phreatic environment. Longman (1980) provides a digest of diagenetic environments and textures of carbonates, but his discussion is biased toward petroleum exploration in marine carbonates.

The best fit model for diagenesis in the Miocene carbonates in the Little Sheep Creek area is one of an active fresh-water phreatic environment. The limestone was deposited in a fresh-water lake; however, "fresh water" only assumes a non-marine origin of the sediments, and does not reflect the chemical composition of the water. Fresh-water lakes are small when compared to most marine environments. The ecological and depositional environments of lakes are dynamic, short-lived, and are highly susceptible to fluctuations in the climate which ultimately controls water level, evaporation rates, rate of sediment influx, and the biological community (Surdam and Wray, 1976, p. 535).

The cementation of the conglomerate is in a more advanced state than the overlying limestone (average of 12% porosity compared to a range of 20%-60%). Cementation of the gravels probably commenced soon after deposition and continued after deposition of the overlying limestones. The conglomerates are a thin sheet of sediment which were deposited as beach gravels at the retreating shoreline as the lake basin filled. The thinness of the conglomerate, the paucity of clastic material in the overlying limestones, coupled with the observation that the composition of the conglomerate reflects the underlying rock type, indicate that the lake was slow in filling and that there was little detrital influx; the strand line was relatively stable. Waves progressing across a lake shelf and lapping against the shoreline would provide the pumping action required to drive water down into the sediment. This is one of the mechanisms proposed by Longman (1980, p. 464) and Bathurst (1975, p. 452) for pumping water into well-washed marine bioclastic debris, cementing them in what Longman calls the active marine phreatic diagenetic zone. The Round Timber conglomerates were sufficiently well washed to allow water, saturated with calcium carbonate, to be pumped into the sediment. Bathurst (1975, p. 453) states that the fabrics of submarine cements are more finely crystalline than fresh-water cements. Longman (1980) describes fibrous aragonite and Mg-calcite cement precipitated in a marine phreatic zone. He compares the fibrous cement with the more coarsely crystalline isopachous bladed calcite cement in a fresh-water phreatic diagenetic zone of marine rocks. When compared to the cement in the underlying Triassic rocks, that of the conglomerates is more finely crystalline. The crystallinity most likely reflects an intermediate composition of the water between the two end members of marine and fresh-water phreatic diagenesis.

The scalenohedral spar may have been precipitated in the active phreatic zone, or later, during a secondary stage of diagenesis, in a fresh-water phreatic environment. The high porosity and permeability of the Round Timber limestone creates a sharp permeability contrast with the underlying Paleozoic and Mesozoic rocks. Several springs were observed at the base of the limestone and I believe that the slump in the SE¼, Section 19, T. 15 S., R. 9 W. was a result of the permeability contrast. The base of the Round Timber limestone can be considered to be in a fresh-water phreatic zone of diagenesis, and cementation is continuing today. Some of the equant spar and the scalenohedral spar may be Recent and represent a continuing process of pore occlusion.

It would be a fallacy to leave the impression that the limestones, with porosities of up to 60%, were poorly cemented and friable. As already stated, all of the rocks have a very hard hammer-ringing quality. The well-indurated, but very porous limestone indicates that cementation of the lime mud occurred soon after deposition prior to any compaction of the sediment. This observation is consistent with the types of cement found in the limestone and the inferred chemistry of the lake water which was generally saturated with respect to calcium carbonate at the time of deposition of the sediment (see Environments of Deposition). Thin micrite fringes lining some pores, scalenohedral spar arranged perpendicular to pore walls, and equant spar are the types of cement observed in the limestone.

In the fresh-water phreatic zone of diagenesis, water saturated with calcium carbonate moves through the sediment. Calcite cement will start to precipitate as minute rhombs on grain surfaces. Continued precipitation of calcite reduces permeability and reduces the rate of nucleation with the result that crystals coarsen toward pore centers, producing an interlocking mosaic. Consequently, in fresh-water diagenetic zones the most common type of cement is a mosaic of equant spar (Longman, 1980, p. 475-477; Buchbinder and Friedman, 1980, p. 402). An intermediate stage exists between the calcite fringe and the occluded pore space. This can be a fringe of equant spar, small blades of spar, or a combination of the two which does not fill in the pore (Figure 40). The bladed calcite can form simultaneously with equant spar if nucleation occurs along a preferred crystallographic axis (Longman, 1980, p. 482). This type of diagenesis is consistent with the types of cement found in the limestone, and the inferred chemistry of the lake water.

Fossils and Age

No fossils diagnostic of the age of the Round Timber limestone were recovered in the Little Sheep Creek area. The calcified molds of the stems of the green algae, Chara, were observed in thin section (Figure 44; Scholle, 1978, p. 11). The close relationship between Recent Charophyta and fossilized counterparts make recognition of Chara species impossible from fragments in the limestone. Diagnostic specialized reproductive organs of the Charophytes, the gyrogonites (Peck, 1957), were not recovered. The only other fossils observed were two partially preserved gastropod fragments, and some ostracods. The fossils are useful in reconstructing the environment of deposition of the lacustrine rocks, but not the age.

Remains of Merychippus, a late Miocene horse, were recovered (Scholten et al., 1955, p. 369) in the northern part of the Beaverhead Range from rocks lithologically similar to the Medicine Lodge beds of the Lima region. The Medicine Lodge beds are underlain by basaltic and

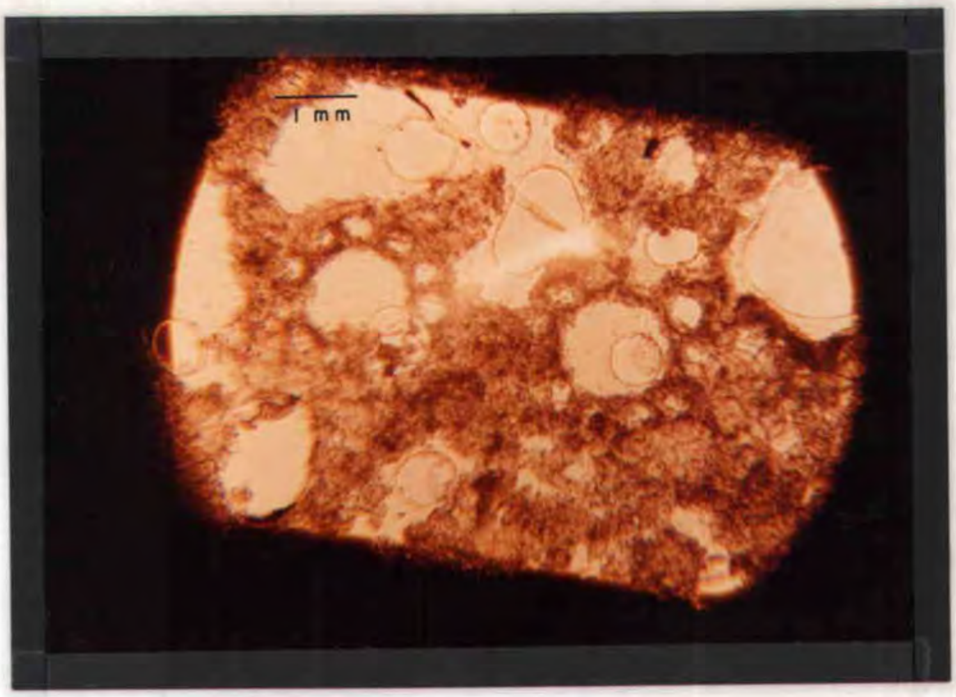


Figure 44. Photomicrograph. Round Timber limestone. Molds of Chara axes with small cortical tubes surrounding a larger central tube. Note irregular fenestrae. Sample Kmb 79-82, NE¼, SE¼, Section 19, T. 15 S., R. 9 W. Plane light.

andesitic lavas similar to those overlying the Muddy Creek beds of probable Oligocene age farther to the west (Scholten et al., 1955, p. 369-370). Locally, the Medicine Lodge strata are overlain by the Edie School volcanics of probable Pliocene age. For these reasons, the Round Timber limestone is tentatively assigned to the Miocene.

Environment of Deposition

The distribution and composition of the conglomerate, the limited occurrence of the laminated boundstones, the presence of *Chara* and other biotics, and the composition and texture of the fenestral limestones are four notable characteristics of the Round Timber limestone which aid in defining the environment of deposition of the unit.

The Conglomerate. The paleogeography of the lake bottom depicted in Figure 45 shows the variations in elevation at which the base of the Medicine Lodge limestone occurs. It is apparent that the limestone was not deposited as one continuous sheet. Instead it was deposited over a variable topography so that part was shallow, part was moderately deep, and at least one or two islands interrupted the sequence. The conglomerates are found on the slopes of the islands, as drawn on the map, and also along the shorelines of the ancient lake.

Beginning at the northeastern edge of the outcrop of limestone in the NE¼, Section 19, T. 15 S., R. 9 W., and trending northwest for about 300 m are two terraces (Figure 35). They are not readily observed in the field but are clear features on aerial photos. The terraces follow bedding planes in the silty limestones of the Thaynes Formation. At this locality the Thaynes beds are dipping to the north and northeast at about 15°. It is possible that these terraces repre-

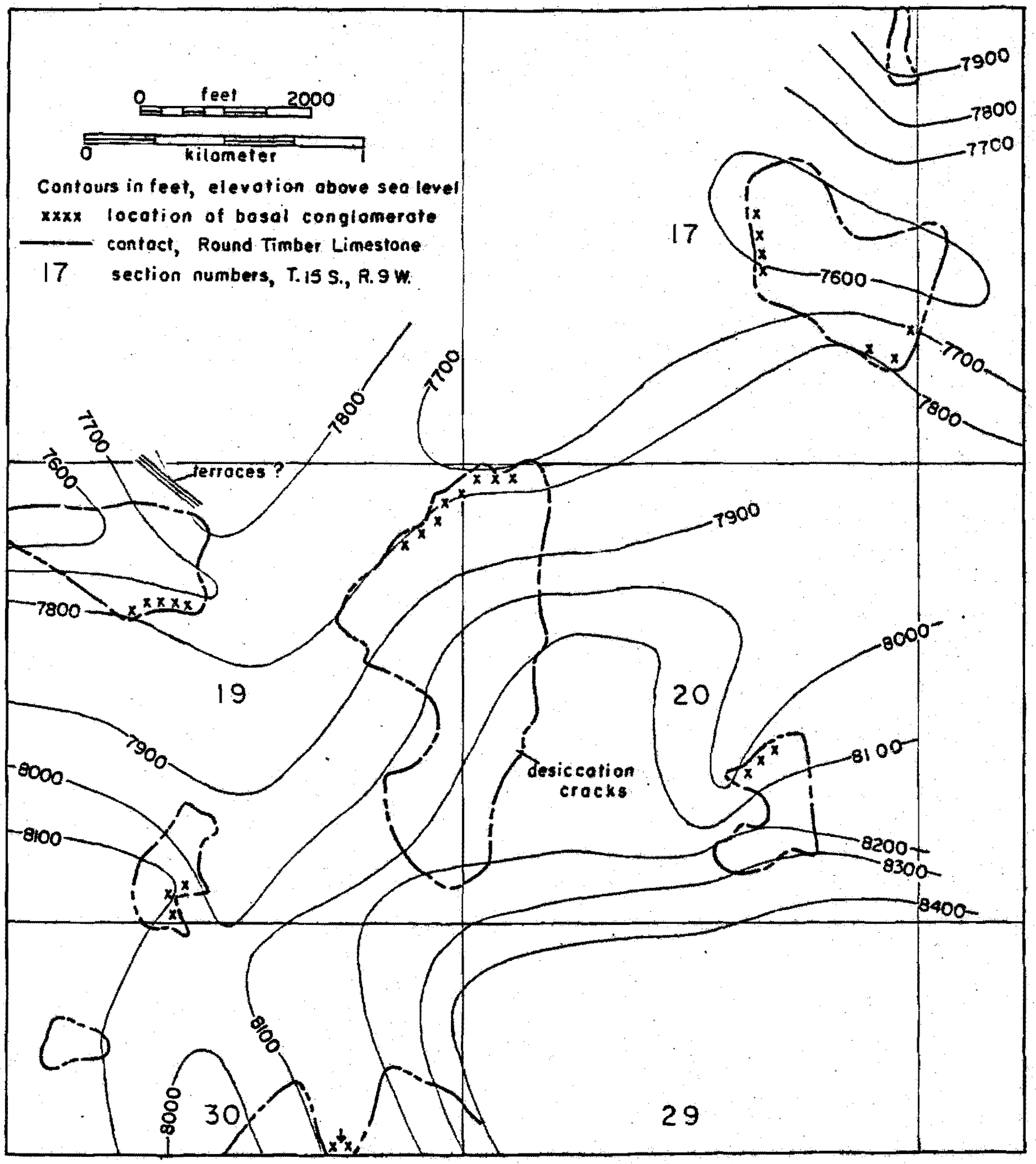


Figure 45. Paleogeographic map drawn at the base of the Round Timber Limestone.

sent a Recent erosional expression of the Thaynes Formation. However, the Thaynes elsewhere weathers to discontinuous ledges and cliffs up to 30 m high; such a step-like pattern was not observed elsewhere in the Little Sheep Creek area. Therefore, I believe the benches in the Thaynes are wave-cut terraces.

Volcaniclastic debris in the conglomerates is important for determining the age of the Round Timber limestone. It is also useful for confirming the fact that the conglomerate belongs in the Round Timber limestone, and is not a part of the Beaverhead Formation. The micrite intraclasts, claystone clasts, oolites and oncolites have some bearing on the deposition of the conglomerates.

Monty (1976, p. 194) defines oncolites as being "unattached subspherical, ovoid, lobate, or flattened structures otherwise similar to stromatolites". Golubic and Fisher (1975) described modern calcareous nodules forming in a creek near Lancaster, Pennsylvania. Carbonate encrustation on the nodules is concentrically arranged around a nucleus of a non-carbonate rock, and is associated with a light-dependent, carbonate-encrusted algal community which lives in the upper 2 mm of the nodule surface. Eggleston and Dean (1976, p. 487) describe oncolites as "stromatolites (which) are growing, . . .in yet a different form," in a Recent fresh-water lake in New York. The lake surface is sufficiently extensive to be affected by wind-generated waves. Oncolites occur on the soft, shallow (0-2 m) platform of the lake where enough agitation occurs to overturn and round them.

Oncolites are formed by algally induced encrustation of carbonate onto a suitable nucleus in shallow, agitated waters. Weiss (1969) used geographic, stratigraphic, petrologic, and peleontologic evidence to confirm a nearshore lacustrine origin for oncolites in the North Horn (Late Cretaceous-Paleocene) and Flagstaff (Paleocene-Eocene) Formations of central Utah. The oncolites occurred preferentially along positive tectonic lineaments associated with the late-Laramide orogeny. Weiss speculated that the oncolites were formed in agitated waters less than 4 m deep, and that they were associated with shore, or nearshore facies, specifically conglomeratic calcareous sandstone and sandy limestone.

The oncolites in the basal conglomerates of the Round Timber limestone were formed on shallow, carbonate banks in waters that ranged from 0-4 m. The shallower limit can be substantiated by the presence of micrite and claystone intraclasts. The intraclasts may have originated in nearshore environments or on banks which were occasionally subaerially exposed and desiccated (Figure 46). A maximum depth of about 4 m is established because the waters had to have been agitated enough, and the lake bottom to have been within the photic zone, to allow the oncolites to form. A zone within the limits of wave base is suggested because greater agitation would have destroyed the oncolites and pulverized the mud-chip intraclasts.

<u>Algal-mat Boundstone</u>. The algal-mat boundstone is a stratiform stromatolite whose presence in the Round Timber limestone is the exception rather than the norm. The single outcrop observed is significant in light of some observations made in Recent stromatolitic lakes.

Walter et al. (1973) described three morphological forms of algal stromatolites in two Recent ephemeral lakes in South Australia. The stromatolites represent similar biological communities, but the three different forms result from variations in the degree of subaerial exposure and desiccation. Of significance to this study are the stratiform

Figure 46. Desiccation polygons in the Round Timber limestone. NW¼, SW¼, Section 20, T. 15 S., R. 9 W.

stromatolites as described by Walter et al. (1973, p. 1023).

The maximum depth of the lakes is less than 1 m. The stratiform stromatolites are usually subaqueous except during a few months of the year. The stromatolites occur as almost flat, laterally continuous, sediment-rich mats which extend their margins in a lobate growth pattern. Storm waves suspend the sediment from the adjacent lake bottom, some of which settles onto the aglal mats. During the summer months, the algal mats are subaerially exposed and form desiccation polygons. The margins of the exposed polygons are eroded back by wind-generated wavelets; erosion is further enhanced by burrowing arthropods. The generation and erosion of the stromatolites is a continuing process with a cyclicity of about five years. The stromatolites are an agent of sedimentation; however, the algal mat is rarely preserved, only a laminated microcrystalline white carbonate mud. Van der Borch (1976, p. 418) studied cores taken from the ephemeral lakes and reported isolated stromatolites preserved at depth in the carbonate mud.

Johnson (1937, p. 1230) discussed two algal limestone deposits in an Oligocene lake in Colorado. The size and shape of the limestones were apparently governed by the profile of the lake shore and the depth of the lake. Fringing stromatolites occurred along steep, rocky shores, whereas on shallow lake bottoms laterally extensive flat stromatolites developed.

Eggleston and Dean (1976, p. 487) discovered that the morphology of stromatolites in three Recent fresh-water lakes in New York depended a great deal on surfaces which receive enough light for photosynthesis. They described fringing nearshore reefs, algal mounds on shallow platforms, and oncolites on shallow banks in agitated water.

The algal mat boundstone in the Round Timber limestone was probably formed on a shallow water bank in an area which was protected from extreme agitation by wind-generated waves. The water was so shallow that the bank was seasonally subaerially exposed. Eggleston and Dean (1976, p. 481) report an annual fluctation of Fayetteville Green Lake, New York of more than one meter. Seasonal depth variation in the Great Salt Lake is about 0.5 m (Halley, 1976, p. 435), and nearly total desiccation occurs in the ephemeral lakes of South Australia (Walter et al., 1973).

The stromatolites of the Round Timber limestone are made up of an accumulation of ooliths, oncolites, micrite intraclasts, and pelloids loosely packed in micrite, and bound with stromatolitic microlaminations. The components of the boundstone and its limited distribution suggest an environment of deposition similar to that described by Walter et al. (1973) for stratiform stromatolites in South Australia.

Charophyta and Other Biotics. The Charophyta is a class of nonmarine green algae that ranges from the Early Devonian to the Recent (Peck, 1957, p. 1). The occurrence of fresh-water limestones and associated strata in the geologic record is spotty and difficult to recognize (Picard and High, 1972). The structure, preservation, and occurrence of Charophyta is such that it will never be as useful age dating tool as marine organisms. However, they are found in non-marine rocks where other fossils are rare, or absent, and are consequently useful for stratigraphic correlations and paleoenvironmental interpretations.

Charophyta are restricted to nonmarine carbonates and associated sedimentary rocks. Similar restrictions of Recent Charophyta to nonmarine aquatic environments permit a direct correlation between ancient and Recent environments. Table 7 presents a summary of the conditions in which Charophyta, specifically the genus Chara, are living today. Of particular interest is the observation that the genus Chara thrives in waters which have a high concentration of calcium, are generally well aerated, clear, have a low salinity, and a high pH.

Charophytes are a major source of allochems in some Recent lakes (Eggleston and Dean, 1976, p. 485) and have been so in the geologic record (Peck, 1957, p. 2). They look very much like bushy higher plants and are distinctive among the algae becaue they have highly specialized reproductive organs, especially large female oogonia. The plant is of a simple structure with a rhizoidal system for anchorage, a main axis, branches, and branchlets (Peck, 1957, p. 4). Parts of the oognia become strongly calcified and usually are the only parts preserved as fossils (Johnson, 1961, p. 173; Peck, 1957, p. 5). Low Mg-calcite is secreted by the vegetative cells of the entire plant, forming a crust on the plant (Peck, 1957, p. 5; Eggleston and Dean, 1976, p. 484). The amount of encrustation is dependent upon the calcium content of the water, the pH, and light quality. Plants of the genera Chara and Tolypella are more often encrusted than other genera (Imahori, 1954, p. 22). Because the calcium carbonate is an encrustation rather than part of a skeletal system, the plant is rarely preserved. Instead, the calcite disintegrates into amorphous micrite upon the death and decomposition of the plant. If the morphology of the plant is retained, perhaps in a very quiet environment, subsequent bioturbation by arthropods, worms, root systems and the spreading rhizomes of living plants would destroy most of the delicate calcified plant molds. Some molds may be preserved, but their abundance would not be representative of

TABLE 7

Environmental Limits For The Growth of Chara and Other Charophyta

(Olsen, 1944; Imahori, 1954; Brunskill, 1969; Eggleston and Dean, 1976)

1. Areal extent of water.
   Ponds, stratified lakes, swamps, rivers, ditches, ephemeral bodies such as rice paddies.

2. Stratification.
   Meromictic lakes and shallow ponds, ephemeral lakes.

3. Salinity (S^{o}) and Chlorinity (Cl^{-}).

S^{o}	Cl^{o}	Approximate Percent Species
0-6 $^{o}/_{oo}$	0-3.5 g/1	75%
up to 12 $^{o}/_{oo}$	up to 6.0 g/1	25%
up to 18 $^{o}/_{oo}$	10.0 g/1	few, associated with marginal marine

4. Calcium.
   Most species can tolerate 15 mg/1 to 250 mg/1 CaO, and conditions of supersaturation with respect to $CaCO_3$.

5. Temperature.
   a. varies from Arctic ice water to hot springs
   b. seasonal variations of least 25°C.

6. Hydrogen ion concentration.

	pH	No. of Species
a.	constantly acid	None
b.	fluctuating acid-alkaline waters	Few
c.	constantly alkaline waters optimum pH is 7.0-9.0	Majority

7. Degree of Encrustation and Mineralogy.
   Chara species are heavily encrusted during alkaline conditions and CaO greater than 30 mg/1

8. Aeration, Light, and Depth.
   a. *Chara* species live in well-aerated waters, and they can grow where H_2S is abundant
   b. Require clear water and live to depth of photic zone
   c. *Chara* generally grow in waters 0.5-5 m deep with a maximum vertical range of 29 m.

9. Degree of agitation: very quiet to slow moving waters.

10. Elevation: Sea level to at least 1600 m.

the living population.

Golubic (1976, p. 127) writes,

> "Algal mats and Recent stromatolites are products of microbial communities composed of cyanophytes, eucaryotic algae, photosynthetic bacteria, and various heterotrophic bacteria. Cyanophytes, or blue-green algae, are the most significant single group engaged in the formation of stromatolites."

The reader is referred to the volume edited by M. R. Walter (1976) for a discussion of the morphogenesis, classification, and geologic significance of stromatolites. The contributions in Walter's tome represent a synthesis of the most recent analyses of stromatolites up until 1976. Other writers who have specifically linked calcite precipitation or entrapment by filamentous algae in modern and ancient lakes include Carozzi (1962), Walter et al. (1973), Johnson (1937), Golubic and Fisher (1975), and Weiss (1969).

Blue-green algae live under exceptionally diverse conditions and have a wide geographical distribution. They have been known to thrive in temperatures ranging from 0-73°C, pH 4 to $>$ 10.5, salinities of up to 250 $^{o}/_{oo}$, and even anaerobic conditions, but usually within the photic zone of most bodies of water. Self-induced light attenuation in the algal mats restricts the photosynthetic organisms to the upper 3 mm (Brock, 1976, p. 142, Table I, p. 146). The conditions in which cyanophytes live are compatible with the more restrictive conditions of the Charophyta.

Two partially preserved fragments of gastropod shells were the only macrofossils observed in the field. The fragments are too small to be identified, and no other mollusc shells were observed in thin section. A few disarticulated ostracod shells and shell fragments were found in thin sections of samples taken in the E½, Section 19, and the W½, Section 20, T. 15 S., R. 9W. Pelloids of micrite which could be either organic in origin or an inorganic accumulation of micrite (Pettijohn, 1975, p. 88, 333) were found in thin sections of the boundstone. If the pelloids are fecal pellets, they are the only remains of mud-ingesting fauna which may have contributed to the erosion of the stromatolites and the bioturbation of the micrite muds.

The environmental significance of gastropods and ostracods will not be attempted here. That their association with each other represents brackish or fresh-water environments is common knowledge (Niem, 1979, unpublished class notes; Majewski, 1969, p. 22, 72). The paucity of any invertebrate fossils in the Round Timber limestone, when compared to the apparent abundance of *Chara* and the blue-green algae, is not clearly understood. Perhaps it is related to the postulated hardness of the water and the high pH in which the algae lived.

Composition and Texture of the Fenestral Limestone. The limestone consists predominantly of micrite, which has undergone partial to complete neomorphism, and minor amounts of quartz silt. What is conspicuous is the lack of internal sedimentary structures (except for the laminated beds), the ubiquitous fenestral porosity, and the lack of terrigenous sediments.

Brunskill (1969) studied carbonate sedimentation in the stratified fresh-water Fayetteville Green Lake of New York. Most of the lake water is supplied by groundwater which drains an area underlain by limestone and gypsum. Seasonal measurements of pH, calcium, and alkalinity revealed that the lake is always supersaturated with respect to calcium carbonate. The degree of supersaturation ranges from one to two times in the winter, to eight times or more in the surface waters (epilimnion) during the spring and summer months. Ninety percent of the sedimentation occurs during the months May-October; 80% of the sediment is calcite directly precipitated from solution from the epilimnion. The sedimentation rate is 300 mg dry matter $m^{-2}yr^{-1}$. Eggleston and Dean (1976, p. 483) found that green algae, aquatic mosses, gastropods, and some fresh-water sponges inhabited stromatolitic bioherms developing along the shoreline. *Chara* populates the adjacent marl bench and provides some of the micrite carbonate sediment entrapped by the algal mats. A similar physical and chemical environment is postulated for the ancient Round Timber lake.

The lake may have had one or more tributary inlets, but the lack of terrigenous sediments indicates that the volume of water provided by run-off was probably minor. A steady source of water could have been supplied by groundwater. The drainage area is underlain predominantly by carbonate rocks. Groundwater passing through the carbonate rocks and entering the lake could have been saturated with respect to calcium carbonate. Seasonal variations in water supply, temperature, and possibly pH may have induced the direct precipitation of calcite from solution, and as encrustations on the vegetative parts of *Chara*.

The continued deposition of over 70 m of limestone indicates a steady state in the climate during Round Timber time. Lacustrine deposits of the Paleocene-Eocene Flagstone Limestone (Utah) and the Eocene Green River Formation (Wyoming-Utah) indicate major periodic fluctuations in the lake levels. The fluctuations produced the cyclic deposition of fluviatile, shallow water, and deep water sedimentary rocks found in the Green River Formation (Picard and High, 1968), and playa lake conditions in the deposition of the Flagstone Limestone (McCullough, 1977) and parts of the Green River Formation (Surdam and Wolfbauer, 1975).

Except for the basal conglomerate, there is no other evidence for an intermittent cessation of the deposition of the Round Timber limestone and an influx of terrigenous clastic debris. Several samples were stained for dolomite, but the rocks are made up almost entirely of calcite. The quartz silt probably was aeolian. Instead, the rather uniform composition and texture of the limestone indicates that the environment within the lake basin varied within narrow limits.

Two representative samples of the limestone were impregnated with a plastic compound (for method see Appendix A) and the calcite dissolved in hydrochloric acid. The porosity of the lighter colored variety of limestone is so high, and the impregnation with plastic so complete, that the calcite was only removed from the outside of the sample. The result is a rind of a dense, spongy, felt-like mass around the limestone core. The calcite in the darker colored variety of limestone dissolved completely leaving the cast of an entangled mass of root-like structures (Figure 47). Microscopically, the pores in the darker limestone are either perfectly circular when viewed in cross-section, or oval in oblique sections. Decomposition of the organic material was complete. There is no woody material; instead, there is a faint residue of black carbonaceous material lining some of the pores (Figure 48). The root-like stems do not have the node-internode structure found on modern Charophyta rhyzomes. The stems consist of a branching network with each succeeding branch smaller than its parent stem and oriented approximately normal to it. These structures have a

Figure 47. Cast of a matted and tangled mass of grass (?) roots. Sample of rock on left prior to impregnation with plastic and dissolution of the calcite matrix, Round Timber limestone.

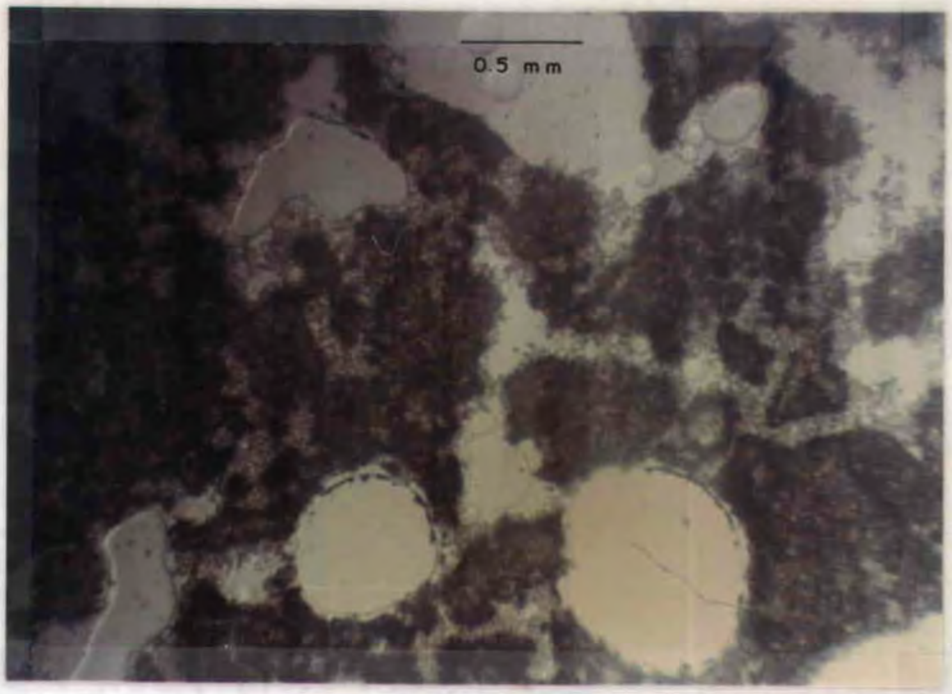


Figure 48. Photomicrograph of root (?) mold showing regular outline of the pore as compared with irregular fenestrae, and thin residue of carbonaceous (?) material. Sample Kmb 79-75, NE¼, NW¼, SE¼, Section 20, T. 15 S., R. 9 W. Plane light.

striking resemblance to modern root systems of dicotymous plants (Frank Smith, 1980, personal communication). It is possible that the mold of the root-like mass may represent the tangled root system of marsh grasses.

The most probable source for the fenestrae was the entrapment of gas bubbles in the lime mud. Under similar conditions in many Recent shallow water environments, where there is a high organic content in a muddy substrate, decomposition of the organic material by bacteria results in the production of carbon dioxide or methane gas. The gas bubbles were probably held in the sediment by the felt-like mass of rhizomes until lithification of the lime mud entombed them permanently in the geologic record.

Bioturbation of any primary structure in the lime mud was accomplished by the spreading rhyzomes of growing <u>Chara</u> plants, by burrowing organisms and by spreading grass root systems. The great porosity of the rock has two sources. The first is the entrapment of gas bubbles in the sediment, a biproduct of organic decomposition. The second is by the preservation of rhyzome and root molds, or the preservation of burrows.

The darker colored variety of limestone has an apparent higher organic content than the lighter colored variety. It is also less porous. It is possible that the darker limestones were deposited in a deeper part of the lake, below the thermocline, where the water temperature was cooler and bacterial activity reduced. Eggleston and Dean (1976, p. 480) report that the Fayetteville Green Lake, New York is stratified with the summer thermocline at approximately 5 m. Below the thermocline the temperature remains constant at about 7°C. This is consistent with lower bacterial metabolism, yet within the growth zone of Chara (Table 7).

The fenestrae aligned with the laminations in the algal boundstone have their origins either as gas bubbles, or as voids created by the desiccation of the algal mat during subaerial exposure (Shinn, 1968). When subjected to desiccation, an algal mat will sometimes shrink, or shrivel up, creating void spaces below, and an irregular surface above, upon which the next colony will establish itself (Pettijohn, 1975, p. 336; Niem, 1979, unpublished class notes). As the surface of the stromatolite is re-colonized, the substrate beneath becomes cemented and rigid with a laminated sponge-like texture (Eggleston and Dean, 1976, p. 484; Bradley, 1929, p. 205). The stromatolites can be easily eroded, or the primary structures destroyed, by Charophytes or other aquatic plants, or by grazing and scavenger fauna.

Conclusion. Brunskill (1969, p. 845) determined the sedimentation rate for calcium carbonate in the Fayetteville Green Lake to be 0.7 mm yr^{-1}. Using this figure as an approximate rate of sedimentation, the ancient Round Timber lake may have existed for at least 100,000 years during the Miocene. The areal extent of the Round Timber limestone is inadequate for determining the maximum or minimum extent of the ancient lake. The outcrops are insufficient for determining the inlets and outlets for the lake, or why a temporary depositional basin was ever created. Post-Laramide faulting, or the volcanic activity which produced the Medicine Lodge volcanics may have created a temporary basin.

The lake level rose slowly, producing basal sheets of beach gravels overlain by relatively pure calcite mud which lithified to a fenestral micritic limestone. Subsequent neomorphism has altered part of the limestone to pseudospar, and a clotted mixture of micrite and microspar. The lake was never very deep and its surface was broken by one or more islands which provided shallow, protected banks on which developed algal stromatolites. The rest of the lake was populated with the green algae _Chara_ anchored firmly by rhizomes to the lime muds of the lake bottom, and probably marsh grasses. _Chara_ plants were a source of low Mg-calcite mud and along with spreading grass roots their rhyzome systems served to destroy any primary sedimentary structures and entrap gas bubbles in the sediment. The drainage area for the lake was underlain by limestone and calcareous sediments which were a source of calcium carbonate in groundwater flowing into the lake. The lake waters were seasonally, or perpetually, supersaturated with respect to calcium carbonate. The primary source of sediment was by direct precipitation of calcite from solution, with an indeterminable amount provided by self-induced low Mg-calcite encrustations on _Chara_ plants.

The Round Timber limestone is overlain by a rhyolite of the Edie School volcanics. The volcanic rock may have filled the lake, causing its demise, or it may have covered parts of the limestone long after cessation of sedimentation in the lake basin.

Edie School Rhyolites

Scholten et al. (1955, p. 376-377) applied the name, Edie School Rhyolites. to slightly porous pale lavender rhyolitic rocks which overlie mid-Tertiary strata in the South Medicine Lodge basin an intermontane basin immediately south of the Little Sheep Creek area. Skipp et al. (1979) described similar rocks in the Edie Ranch Quadrangle of Idaho and Montana. The rhyolites in the Edie Ranch Quadrangle are densely welded ash flow tuffs. The unit ranges in thickness up to about 100 m in the South Medicine Lodge basin (Skipp et al., 1979) and it wedges out where it overlaps the flanks of adjoining ranges (Scholten et al., 1955, p. 377).

Scholten et al. (1955, p. 377) speculated on the source for the rhyolite tuffs. They suggest that the source may have been small acidic lava and tuff cones found at various localities in central Idaho. The cones are mostly covered by the Snake River basalts, but one cone located about 17 km south of the Tendoy Range near the village of Small, Idaho, may have been the source of the rhyolites in the South Medicine Lodge basin.

The Edie School Rhyolites unconformably overlie the Round Timber limestone of Miocene age and the Mississippian Scott Peak Formation in the thesis area. South of the Tendoy Range the rhyolites are disconformably overlain by the Snake River basalts of late Pliocene to Pleistocene age. The Edie School Rhyolites are, therefore, assigned to the Pliocene.

Distribution and Topographic Expression

The Edie School Rhyolites disconformably overlie the Round Timber limestone and the Scott Peak Formation in the E½, Section 30, T. 15 S., R. 9 W. Slabby erosional remnants can be found on the plateau underlain by the Round Timber limestone in the NW¼, SW¼, Section 20, T. 15 S., R. 9 W. (Plate I). Natural exposures of the rhyolites occur on a fault scarp about 800 m south of the thesis area in Section 30, T. 15 S., R. 9 W. No natural exposures of the rhyolite could be found in the thesis area; however, the unit has an apparent sub-horizontal attitude.

The surface is covered by a thin layer of soil which yields a liberal amount of rhyolite chips. The perimeter of the unit is marked in places by talus of slabby rhyolite.

Lithology and Chemistry

The rhyolite is a well-indurated aphanitic rock with phenocrysts of quartz and magnetite, and small (up to 2 cm) rounded xenoliths of calcareous sandstone. The rock is pale yellowish brown (10 YR 6/2) when weathered, and light brownish gray (5 YR 6/1) on freshly broken surfaces.

Microscopically the rock is composed of a micro- to cryptocrystalline groundmass which makes up about 95% of the rock mass. The phenocrysts, which make up about 5% of the rock, consist of 35% quartz, 33% orthoclase, 15% plagioclase (An_{27-30}), 13% magnetite and 3% zircon, apatite and biotite. The groundmass is made up of orthoclase (?) and quartz (?) microlites which form a trachytic texture around the phenocrysts. Small stretched-out laths of glass have been divitrified to form a pilotaxitic texture; suggesting this is a rhyolite from the densely welded zone of an ash flow tuff. Alteration is slight with halos of limonite around magnetite phenocrysts.

Chemical analyses of three samples of the Edie School Rhyolites are presented in Table 8. Noteworthy is the uniform composition of the samples, and the high silica and potassium values. The rocks can be classified as rhyolites according to Taylor's (1978, p. 11-12) classification of calc-alkaline rocks of the High Cascades of Oregon.

TABLE 8

Chemical Analysis of Edie School Rhyolites

Oxide	Weight Percent Kmb 79-76	 Kmb 79-74a	 Kmb 79-74b
FeO	0.2	0.2	0.2
TiO_2	0.24	0.23	0.23
CaO	0.5	0.9	1.2
K_2O	5.46	5.46	5.44
SiO_2	78.0	77.9	77.5
Al_2O_3	11.9	11.7	11.7
Na_2O	3.5	3.5	3.5
TOTAL	99.8%	99.89%	99.77%

Sample locations:

Kmb 79-74a and b, NW¼, NW¼, SW¼, Section 20, T. 15 S., R. 9 W.

Kmb 79-76, E½, Section 30, T. 15 S. R. 9 W.

Analysis done by Karl Wozniak, Oregon State University; samples prepared according to methods outlined by E. M. Taylor.

STRUCTURE

Regional Structure

In 1909 Douglass published a reconnaissance study of the geology of North Dakota, Montana, and Idaho. Upon reaching western Montana he observed that, "The country has the appearance of having been for ages the battleground of geological forces." (Douglass, 1909, p. 267). This observation is particularly applicable to that part of southwestern Montana and central Idaho north of the Snake River Plain. This region encompasses two distinct structural and sedimentological provinces that have undergone differential rates of subsidence and sedimentation since the Late Precambrian: stable cratonic shelf sedimentation to the east and sedimentation in the more rapidly subsiding Cordilleran miogeosyncline to the west (Scholten, 1967, p. 8; Huh, 1968, p. 31; Stokes, 1976, p. 17). Connecting the two provinces is a transition zone across which marked changes in the thicknesses of the sedimentary rocks occur; the transition zone was also the area of the most intense Laramide deformation.

The transition zone is coincident with the Wasatch Line of Kay (1951, p. 14) or the hinge line as outlined by Scholten (1957) that extends in a curvilinear belt from northern Mexico through the middle western states and into Canada. The line was drawn by Kay at the zero edge of the Early Cambrian strata; it roughly divides Cambrian strata more than 5,000 feet (1,525 m) thick on the west from lesser thicknesses to the east. In southwestern Montana the Wasatch Line is a linear zone trending northwest that includes the Tendoy Range and the eastern flank of the Beaverhead Range (Figure 49).

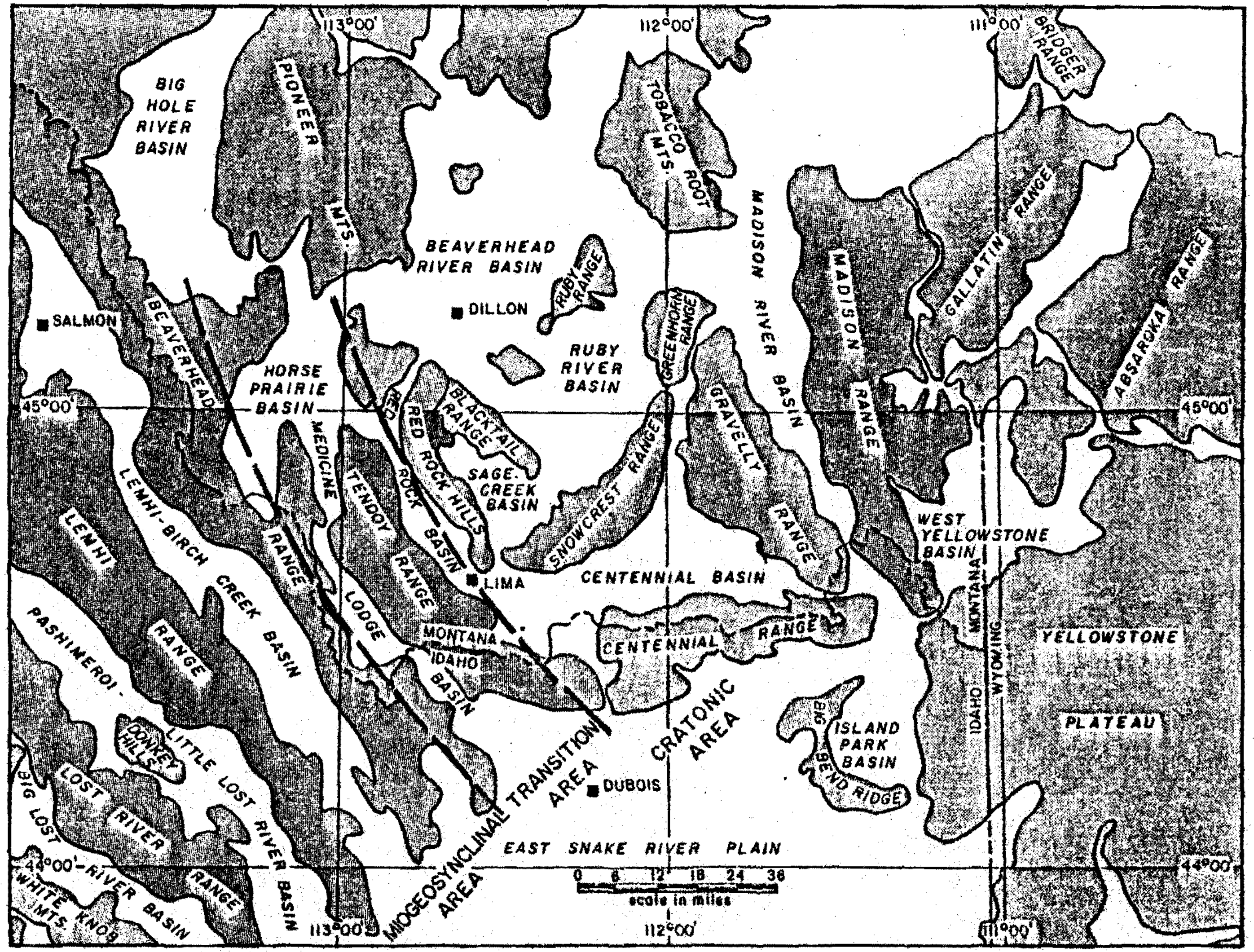


Figure 49. Physiographic features and tectonic provinces of southwestern Montana and adjacent areas (from Klecker, 1980).

Paleozoic strata on the craton east of the Wasatch Line have a maximum thickness of 7,500 feet (2,288 m) and overlie Precambrian metamorphic basement (Huh, 1968, p. 31). Laramide structures consist of broadly folded and faulted mountain ranges that involve the Precambrian basement. Radiating south away from the domal uplift of the Tobacco Root Mountains are several major anticlinoria which have gently dipping western limbs and more steeply dipping eastern limbs that are bounded by high angle reverse faults. Precambrian basement rocks are exposed along the axes of the anticlinoria and form the resistant crests of several mountain ranges.

The anticlinoria bifurcate into smaller less-extensive folds that plunge gently to the south. The western anticlinorium, the Blacktail-Snowcrest arch, curves to the southwest and divides into two anticlines separated by the Little Water syncline. The syncline and the easternmost anticline, the Garfield anticline, plunge gently into the Tendoy Range with a northeast-southwest aspect. The Garfield anticline continues into the Little Sheep Creek area and has acted as a buttress against which subsequent Laramide deformation has occurred. The eastern arch, the Madison-Gravelly arch, bifurcates into three smaller anticlines and subordinate synclines which may have been a controlling factor in the east-west trend of the Centennial Range (Witkind, 1977, p. 534-536).

Precambrian basement rocks are unknown in the miogeosyncline west of the Wasatch Line. Instead, the Paleozoic systems have a minimum thickness of 12,000-15,000 feet (3,660-4,575 m) and overlie the Precambrian Belt sedimentary strata (Huh, 1968, p. 31). The Laramide structures involve the entire stratigraphic sequence and are very complex, consisting of faulted asymmetric to recumbent folds, and low-angle thrust faults that have a north to northwest structural trend (Scholten, 1967, p. 9).

The transition zone, which corresponds to the miogeosynclinal-cratonic hinge line and the Wasatch Line, is a meeting ground of these two structural provinces. Precambrian basement rocks have been exposed along northwest-trending high-angle reverse faults in the Beaverhead and Tendoy Ranges and northeast structural trends occur in the Tendoy Range. The overall structural trend in the transition zone, however, is north to northwest. The northwest-trending structures in the Tendoy Range include low- to high-angle reverse faults such as the Tendoy fault, flat thrust sheets such as the Medicine Lodge thrust, and asymmetric to recumbent folds. Except for the high-angle reverse faults, the structures are confined to Paleozoic and Mesozoic strata (Scholten, 1967, p. 11). North- to northwest striking basin-and-range type post-Laramide block faulting also has occurred. These normal faults have created the Red Rock basin on the east front of the Tendoys and the North and South Medicine Lodge and Big Sheep Creek basins between the Tendoy and Beaverhead Ranges.

Structure at Little Sheep Creek

The structure in the Little Sheep Creek area reflects the general pattern of deformation described for the transition zone in southwestern Montana. Northeast-trending folds, northwest trending overturned and asymmetric folds, north to northwest-trending high-angle reverse faults, a flat thrust sheet, and normal faults representative of Laramide tectonism across the Wasatch Line are found in the 60 km^2 mapped in the

Little Sheep Creek area. The structure described herein reflects a merging of tectonic styles described for both the miogeosyncline and the craton.

Folds

Folds occur in three orders of magnitude. The primary folds are those which are easily distinguishable from outcrop pattern and are the controlling structural features of the area (Figure 50). The second-order folds are mapped on the outcrop pattern of the basal Thaynes and are minor folds involving the Thaynes and Woodside Formations and, possibly, the Dinwoody Formation. The third-order folds are intrafolial folds involving incompetent beds within the Thaynes Formation (Figure 51).

The primary folds include the Garfield anticline and a subordinate syncline located to the north, herein called the Middle Fork syncline. These are open folds that plunge gently approximately S65°W and are representative of the bifurcating nose of the Blacktail-Snowcrest anticlinorium exemplary of cratonic style tectonism. The Garfield anticline involves the thick sandstone of the Quadrant Formation and the overlying strata up to, and including, strata as young as the Colorado Shale. The expression of the Garfield anticline is masked in the southwestern part of the area by asymmetric northwest-southeast-trending folds (cross-section A-A'). These are also primary folds that are doubly plunging; the plunges of which are nearly perpendicular to the plunge direction of the Garfield anticline.

The third-order folds are small intrafolial folds within the thin-bedded carbonates of the Thaynes Formation. Except for two third-order

Figure 50. Axes of thesis area folds.

Figure 51. Intrafolial folds in the Thaynes Formation; west side, West Fork of Little Sheep Creek, NW¼, SW¼, NE¼, Section 19, T. 15 S., R. 9 W.

folds in the NE¼, Section 8, T. 15 S., R. 9 W., the second- and third-order folds are all essentially horizontal or plunge gently toward the northwest quadrant; the two exceptions plunge gently to the northeast.

The northwest-southeast-trending primary folds, specifically the Round Timber syncline, the Buffalo Spring anticline, and the Seybold syncline are northeast-yielding overturned asymmetric folds. The folding is tight but becomes open and dies out within a short distance in the direction of vergence (Cross section A-A", Plate II). These folds are exemplary of the miogeosynclinal style of deformation.

A question arises regarding the order of folding at Little Sheep Creek. The structures are Laramide and the area represents a merging of structural styles from the miogeosynclinal and cratonic provinces. Northeast of the thesis area the southeastern limb of the Blacktail-Snowcrest arch was refolded into the northwest-trending Lima anticline between Lima and Monida, and the Red Rock syncline north of the Lima reservoir (Scholten, 1967, p. 12). Scholten also discusses the creation of a triangular basin in the northeast-trending Little Water syncline where the geosynclinal deformation impinges upon the syncline.

Scholten (1967, p. 12) also cites, "In T. 15 S., R. 9 W. small-scale northwest folds in Triassic rocks were superposed on the main northeast structure in that area." He is most likely referring to the Round Timber syncline, the Buffalo Spring anticline, and the Seybold syncline. I agree with Scholten's interpretation of the sequence of events; however, I have interpreted the structure not as small-scale folds limited to the Triassic rocks, but as folds within the axis of a large scale syncline that has involved rocks representative of the entire stratigraphic column up to, and including, the Early Cretaceous rocks (see cross-section A-A', Plate II). These secondary axial folds apparently become recumbent east of the thesis area in Four Eyes Canyon (Scholten, 1979, personal communication). This larger syncline, unnamed in the thesis area, represents a refolding of the nose of the Garfield anticline. The structure contour map drawn at the base of the Phosphoria Formation (Plate IV, Figure 1) indicates that the major synclinal axis bifurcates into two synclines (Round Timber and Seybold synclines) with an intervening anticline (Buffalo Spring anticline) in a northwesterly direction as it approaches the southeastern limb of the Garfield anticline. That the Garfield anticline controlled the axial folding within the large-scale syncline is possible and consistent with the inferred sequence of events.

Further evidence for control of subsequent folding by the northeast-southwest folds is exemplified by a comparison of Figures 1 and 2, Plate V. The gently plunging Middle Fork syncline and Garfield anticline are easily distinguished in Figure 1. Figure 2, Plate V, is a structure contour map drawn at the base of the Triassic Thaynes Formation. This map augments what can be concluded from the geological map (Plate I): the primary folds evident in the curving pattern of the Quadrant and Phosphoria Formations are reflected in the overlying strata but their expression has been greatly subdued. The structure contours drawn at the base of the Thaynes Formation indicate that the expression of the Garfield anticline has been subdued upsection and confused by the refolding of the strata into the northwest-southeast folds.

The conclusion can be made that the northeast-southwest-trending folds, specifically the Garfield anticline, existed prior to the compressional forces that created the northwest-southeast folds. The Garfield anticline acted as a buttress against which subsequent deformation occurred, limiting the northward migration of northwest-southeast folds. Compressional forces were transmitted beyond these primary folds to create the second and third order folds.

Faults

Except for the Medicine Lodge thrust, faulting in the thesis area consists of northwest-trending and north- to northeast-trending normal and high-angle reverse faults that represent only tens of meters of displacement. These faults are not deep-rooted and die out within the ductile and less competent limestones and mudstones of the Mesozoic rocks (Figure 52; Plate II).

The Middle Fork fault, so named because it extends north into the Middle Fork of Little Sheep Creek where it dies out, is a normal fault with approximately 240 m of displacement and a right lateral strike-slip component juxtaposing Pennsylvania Quadrant strata with the Triassic Dinwoody. The fault trace is curvilinear and cuts across gullies without deviation indicating that the fault plane is vertical or subvertical. The Middle Fork fault extends southward beneath the Beaverhead Formation; there is no field evidence to suggest that the Beaverhead Formation was involved in the faulting. Immediately west of the southern extension of the Middle Fork fault is a second north-south normal fault with a displacement reverse to that of the Middle Fork fault; this has created a small graben between the two. Field evidence suggests that the western side of the graben continues beneath the Beaverhead Formation but does not involve the latter.

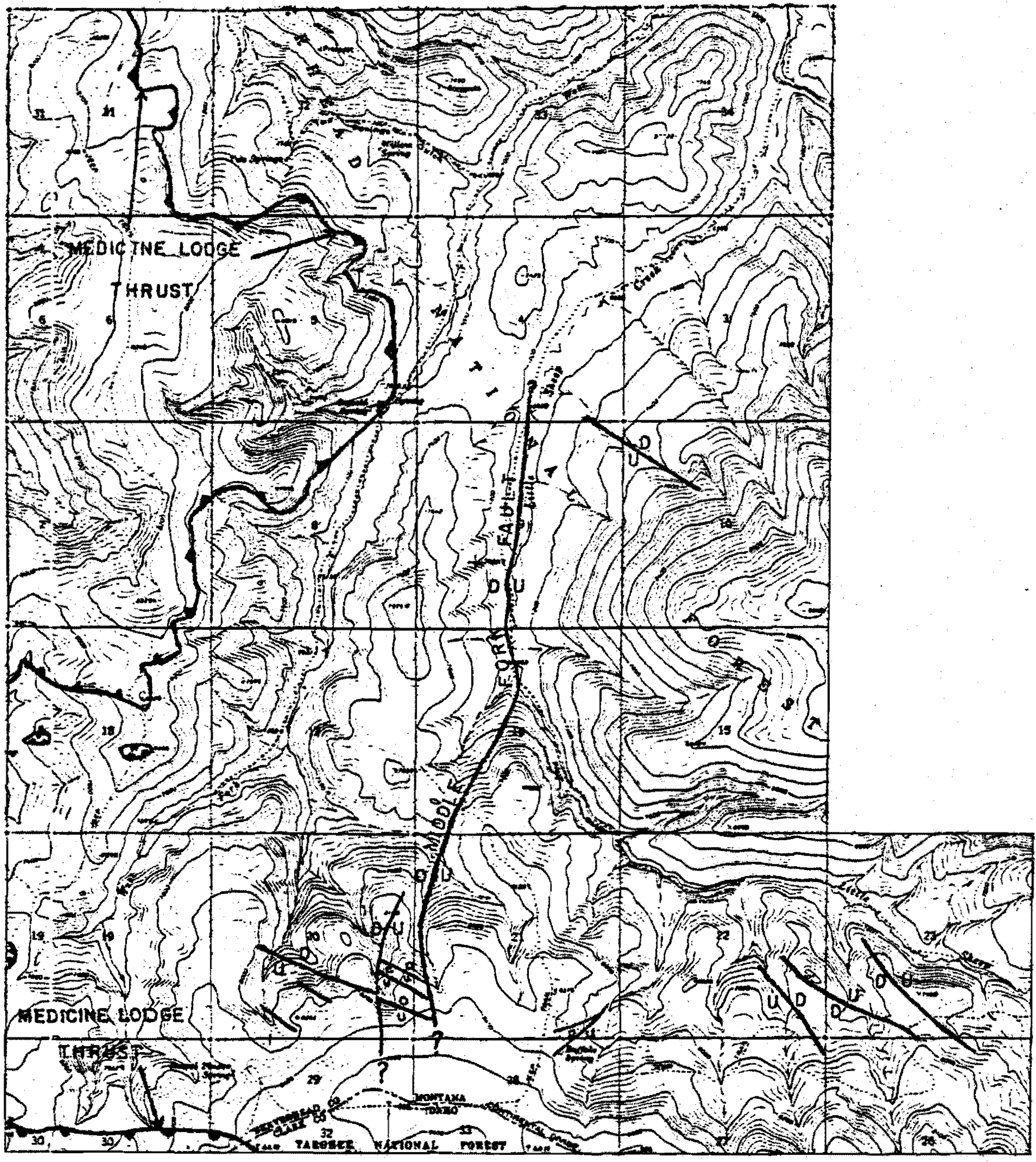


Figure 52. Surface traces of thesis area faults.

Other normal faults include the northwest-trending fault in the NW¼, Section 10, T. 15 S., R. 9 W. which has displaced the Phosphoria Formation and which dies out in the adjacent Quadrant and Dinwoody Formations; a northwest-trending fault in the SE¼, Section 22, T. 15 S., R. 9 W. which involves the Triassic, Jurassic, and Early Cretaceous strata; northwest-trending faults involving only the Thaynes Formation in the S½, Section 20, T. 15 S., R. 9 W.; and a bifurcating normal fault in the SE¼, Section 21, T. 15 S., R. 9 W. that involves displacement restricted to the Jurassic and Early Cretaceous rocks.

High-angle reverse faults trending northwest are found within the graben in the SE¼, Section 20, T. 15 S., R. 9 W., and a high-angle reverse fault is found in the up-thrown side of the fault immediately west of the graben. These faults are imbricate faults that developed in the axis of the Seybold syncline, and possibly become listric with depth and die out in the Quadrant Formation. Displacement along these faults is small, perhaps on the order of 30-40 m.

Two high-angle reverse faults of limited displacement are found in Sections 22 and 23, T. 15 S., R. 9 W. These faults have a northwest trend, are confined to the Mesozoic rocks and have a displacement of only about 210 m (cross-section C-C', Plate II). The temporal relationship of these two high-angle reverse faults is difficult to assess. Reconnaissance of the structure and stratigraphy farther up Little Sheep Creek immediately east of the thesis area reveals more northwest-trending faults that represent a cummulative horizontal right lateral displacement of approximately 850 m. These faults may be coincident with the folding of the Garfield anticline and the southwest plunge of its axis, folding and tearing of the rocks having taken place in the less competent strata upsection from the Quadrant sandstone.

The Beaverhead Formation was syntectonic with the onset of Laramide tectonism, but the formation itself was also folded as the deformation progressed. A good example of the folding is the Lima anticline north of Lima, Montana. Exposures of the Beaverhead Formation are limited within the thesis area and no attitudes of the bedding could be taken. The lower contact of the formation has been inferred from the distribution of float and weathering characteristics contrasting with the underlying formations. A structure contour map drawn at the base of the Beaverhead was inconclusive; it was not possible to differentiate between folds and the erosional contact of the unconformity. The only evidence within the thesis area that the Beaverhead has been a subject of some deformation are fractures within the formation found in outcrop in the SE¼, NW¼, Section 30, T. 15 S., R. 9 W. The timing of this fracturing cannot be determined but could be related to post-Laramide uplift of the Tendoy Range.

The sequence of folding and faulting in the thesis area first involved the formation of the northeast-southwest-trending folds, specifically the Garfield anticline and the Middle Fork syncline, and the subordinate folds evident in the structure contour map of Figure 1, Plate V. The high-angle reverse faults in Sections 22 and 23, T. 15 S., R. 9 W. may have been formed during this initial stage of folding. The second event involved the refolding of the nose of Garfield anticline into a major northwest-southeast syncline with folding of the Mesozoic strata into doubly plunging overturned folds within the axis of the syncline. Vergence is towards the northeast with the major syncline overturned. The Round Timber and Seybold synclines and the

Buffalo Spring anticline are the axial folds; the folding becomes less intense and dies out in the direction of vergence immediately beyond the Seybold syncline. High-angle reverse faults within the axis of Seybold syncline occurred at this time. Relaxing of compressional forces towards the close of Laramide tectonism created the Middle Fork fault and the other normal faults described in the thesis area.

The Medicine Lodge thrust is a nearly horizontal fault that is part of a late Paleocene-early Eocene thrust system that extends from the north margin of the Snake River Plain to the west margin of the Big Hole Basin, Montana (Ruppel, 1978, p. 3). Within the thesis area the Medicine Lodge thrust has juxtaposed strata of the Scott Peak Formation onto Jurassic and Cretaceous-Paleocene strata (Figure 53). Lithologic similarities between strata in the thrust sheet and the Scott Peak Formation on the eastern flank of the Beaverhead Range indicate that the source of the thrust was as far away as the Beaverhead Range or from the area in between that is now part of the South Medicine Lodge basin. The fault was therefore an eastward-yielding structure with a maximum total displacement of approximately 20 km. The limestones in the thrust sheet are in places remarkably undeformed, but elsewhere they are folded into disharmonic, recumbent, and detached folds.

Figure 53. Medicine Lodge thrust with limestones of the Scott Peak Formation (Msp) lying on the Beaverhead Formation (KTbh) which disconformably overlies the Woodside Formation (ꞦRw). Calcareous siltstones of the Dinwoody Formation (ꞦRd) in the foreground.

ECONOMIC GEOLOGY

The economic potential of the Little Sheep Creek area exists in four possibly exploitable realms: phosphates, hydrocarbons, gypsum, and building stone.

The color, texture, and hardness of the Round Timber limestone make it a desirable building material. When cut into slabs and polished the peculiar fenestral porosity provides a unique decorative panel that can be used as an ornamental facade on exterior or interior walls. Local reports in Lima indicate that the "travertine" limestone deposit is the largest of its kind in the United States. The limestone has been actively mined on the Idaho side of the Continental Divide south of the thesis area where it has been exposed by the Deadman fault.

The outcrop of the limestone that lies on the section line between Sections 19 and 20, T. 15 S., R. 9 W. is of particularly exploitable potential. In the summer of 1979, the Hallet Mining Company of Duluth, Minnesota actively pursued the possibility of mining the Round Timber limestone. Because of the numerous landslides and general slope instability in Little Sheep Creek, construction of access roads to the limestone would have had to follow stringent guidelines established by Forest Service engineers. The inaccessibility of the limestone, coupled with the current recession and the potential environmental hazards imposed on the residents in Little Sheep Creek made the limestone an economically infeasible venture (Lietch, 1979, personal communication).

Gypsum was mined between 1940 and 1942 from the shaly beds of the Big Snowy Formation along the East Fork of Little Sheep Creek approximately 2.5 km east of the thesis area (Scholten et al., 1955, p. 400).

Further investigation is required, but the gypsum in the gypsiferous shales of the Big Snowy Formation in the thesis area may not be concentrated sufficiently to make them an economically potential resource.

Kerogen shales in the Phospharia Formation in Little Sheep Creek can yield 16 to 25 gallons of oil per ton (Condit, 1919, in Scholten et al., 1955, p. 401). I did not observe these shales in the Phosphoria during my field studies and it is conceivable that their recovery is not economically feasible.

Swetland, Clayton, and Sable (1978, p. 82) found that samples of the Big Snowy Formation in Big Sheep Creek, Tendoy Range, and Blacktail Deer Creek, Blacktail Range, contained enough organic compounds to make this formation a potential source for hydrocarbons. The Quadrant sandstone and the basal sandstone of the Kootenai Formation may make likely stratigraphic hydrocarbon traps. However, exposure of the source beds and the reservoir sands by folding and faulting in the thesis area may have permitted the escape of any economically potential hydrocarbons. Any structural traps that may exist would be small and difficult to locate.

Phosphorite is treated either with sulfuric acid to produce fertilizer or in an electric furnace to produce elemental phosphorus. The sulfuric acid and electric furnace treatments require a minimum phosphate content of 31% and 24% respectively (Cressman and Swanson, 1964, p. 338). The average phosphate content of the Meade Peak and Retort Phosphatic Shale Members of the Phospharia Formation in the Little Sheep Creek Section measured by Cressman and Swanson is 16.7% and 8.5%, respectively. These figures do not make the phosphorites economically feasible within the thesis area.

GEOLOGIC HISTORY

The summary of the stratigraphy and structure of the Little Sheep Creek area can only be discussed in light of the sedimentological and tectonic events which make up the geological history of the area. That the geological history is related to the changes in sedimentation across the Wasatch Line and ultimately the style of deformation during the Laramide orogeny is an important concept to remember. The Wasatch Line did not control sedimentation, but was instead a transition zone between shallow marine deposition on a relatively stable cratonic shelf to the east and deeper lithotopes in the Cordilleran miogeosyncline to the west. This difference in sedimentation lasted throughout most of the Paleozoic and into the Triassic. The slope of the cratonic shelf was low throughout most of that time so that regional diastrophism or eustatic changes in sea level had far-reaching consequences on the craton; specifically, the creation of regional unconformities on the craton as the result of nondeposition or uplift and erosion. Generally these unconformities do not exist in the miogeosyncline, and commonly are not represented in the transition zone across the Wasatch Line.

Discussion of the geological history of the thesis area begins with deposition of the upper part of the Mission Canyon Limestone. Older Paleozoic strata are present in the Lima region (Scholten et al., 1955), but are not exposed in the thesis area.

The Mission Canyon Limestone at Little Sheep Creek is light in color and consists of thick beds of a coarsely crystalline mixture of fossil debris overlain by laminated micrite with interbeds of fossiliferous sparite. These carbonates are probably representative of open marine platform and platform edge sedimentation (Wilson, 1970, p. 231). This interpretation is consistent with the model presented by Rose (1976) and later modified by Gutschick, Sandberg, and Sando (1980). The Mission Canyon Limestone is a widespread carbonate extending across most of Montana and Wyoming with lateral equivalents in Colorado, Utah, Arizona, and New Mexico. The Mission Canyon represents the highest level of inundation on the craton during the Mississippian after the initial transgressive phase represented by the Lodgepole Limestone. The Mission Canyon Limestone was deposited on a shallow continental shelf with a slope as low as 0.5° (Gutschick et al., 1980, p. 124) so that deposition of the carbonate was rapid, creating a sedimentological regression.

Sedimentation of detritus derived from the Antler orogenic highland was continuous during the Early Mississippian as far east as South Dakota. The western limits of the Mission Canyon Limestone mark the edge of a stratigraphic reef and the margin of the epeiric shelf. The position of the reef is coincident with, and parallel to, the Wasatch Line. Sedimentation was continuous west of the Wasatch Line with deposition of calcareous shales of the upper McGowan Creek Formation in the miogeosyncline, and flysch deposits farther west at the base of the Antler highlands.

A eustatic drop in sea level (Gutschick et al., 1980) during the early Meramecian caused the cratonic shelf to become emergent. The drop in sea level and development of the karstic surface on the Mission Canyon during Meramecian time marked the end of what Rose (1976) called the lower depositional complex of the Mississippian System. During Meramecian time the karst surface on the Mission Canyon acted as a ramp across which terrigenous clastic sediments were transported and deposited in the miogeosyncline as the Middle Canyon Formation of the White Knob Group.

Rising sea level during the early Chesterian occupied two embayments along the continental margin. The northern embayment expanded during the Chesterian to create a trough that extended east from southwestern Montana across central Montana and into the western part of the Dakotas (Easton, 1962). This trough was the site of Big Snowy sedimentation. The southern embayment extended into Wyoming and was the site of Amsden sedimentation.

The carbonate shelf margin and the upper depositional complex of Rose (1976, p. 459-465) shifted west at this time. Bioherm-like masses of fossiliferous calcarenites were deposited along the shelf margin to form the limestones of the White Knob Group. The upper two members of the Scott Peak Formation, the South Creek Formation, and the Surrett Canyon Formation represent the shelf edge bioherm (Rose, 1976, p. 469). These formations are fully developed in the Lemhi Range of central Idaho. The source for the Scott Peak Formation in the Medicine Lodge thrust in Little Sheep Creek is postulated to be from the eastern flanks of the Beaverhead Range. In the Beaverhead Range the Scott Peak Formation consists of only the lower cyclic member and the lower part of the middle massive member. The Scott Peak is overlain by the black shales of the Big Snowy Formation. That the bioherms developing west of the Beaverhead Range controlled circulation of the back-reef area becomes apparent in the Big Snowy shales.

Rose (1976, p. 464) discovered that the transgressive-regressive couplet apparent in the lower depositional complex is not evident in the upper depositional complex. Instead, sedimentation during Late Mississippian time involved several changes in sea level. In the Little Sheep Creek area the Big Snowy Formation consists of a lower limestone unit that grades upward from a thin-bedded micrite to a thick-bedded fossiliferous spar; the upper part of the formation consists of a dark gypsiferous mudstone. The lower unit is indicative of a shallowing-upward sequence representing a lowering of sea level or a slowed subsidence. The dark gypsiferous mudstones were probably deposited in anoxic, hypersaline conditions in which normal marine circulation was restricted by the local development of shoals and bioherms.

The Big Snowy trough was drained during the latest Chesterian by the Big Snowy uplift of central Montana. This area became a source of sediments for the Horseshoe Shale Member of the Amsden Formation in the Wyoming sea which had persisted and expanded during the Chesterian. During the Morrowan, the Wyoming sea expanded north into Montana and became the site of Early Pennsylvanian Amsden Group sedimentation.

The foregoing discussion has been elaborated upon in greater detail in the section earlier in the text on the Mississippian-Pennsylvanian boundary in Montana and Wyoming. What is of significance here is that the retreat of the sea from central Montana resulted in an erosional unconformity at the top of the Big Snowy Formation that was only developed locally in the vicinity of the Wasatch Line in southwestern Montana. Hildreth (1980) cites a shallowing-upward carbonate sequence in the Big Snowy Formation on the Armstead anticline that is gradational and conformable with the overlying Amsden Formation. Klecker (1980, p. 25) cites a limestone conglomerate at the base of the Amsden Formation in Little Water Canyon and interprets this as representing an erosional unconformity. Scholten et al. (1955, p. 366) also noted local limestone conglomerates at the base of the Amsden in the Lima region but they considered the contact between the Big Snowy and the Amsden generally to be gradational and conformable.

After draining of the Big Snowy trough, the shoreline was apparently stabilized near the Idaho-Montana border so that sedimentation continued uninterrupted from the Mississippian into the Pennsylvanian in this region.

The lower part of the Amsden Formation in the thesis area is covered. The Amsden Formation in the Beaverhead Mountains is mostly sandstone (Maughan, 1975, p. 284); the Amsden Formation on the Armstead anticline consists entirely of calcareous, fine-grained sandstone that becomes less calcareous upsection and grades into the quartz arenites of the Quadrant Formation (Hildreth, 1980). Farther south, in Little Water Canyon, the Amsden consists of lower and upper limestones separated by sandstone (Klecker, 1980, p. 31). In the thesis area the upper Amsden consists of thin-bedded limestone with local concentrations of fossil debris. The terrigenous clastic content of the limestone increases upsection and the contact with the overlying quartz arenites of the Quadrant Formation is gradational and conformable. The Quadrant Formation is a thick wedge-shaped sequence of fine- to medium-grained, cross-bedded quartz sandstone. This wedge of sandstone is thickest in the west and thins markedly eastward where it becomes transitional with sandy dolomite, dolomite and limestone (Maughan, 1975). A western or northwestern source for the sandstone is therefore implied.

Up to Amsden time, sedimentation within the thesis area had been on a carbonate shelf; the deposition of carbonate sediments had been controlled by water depth and any terrigenous sediments were derived from the craton to the east. The Antler highland existed farther west in Idaho during Amsden time, but clastic sediments were deposited as flysch along the foreland basin and were not transported east onto the shelf (Rosé, 1976; Gutschick et al., 1980). This situation apparently started to change during the Early Pennsylvanian when west-derived terrigenous clastics became dominant.

Beginning in Early Pennsylvanian time and continuing into the Permian, a shallow sea transgressed across the craton. Marine, or marginal marine, sedimentation occurred over much of Montana, Wyoming, and parts of the Dakotas, and spread southeast to the Transcontinental arch. Strong positive areas existed in central-western Idaho during the Early Pennsylvanian exposing Early Paleozoic and Late Precambrian sandstones that became the source of clastic sediments in the Amsden and Quadrant Formations. The highlands persisted well into Late Pennsylvanian and Early Permian time on the site of the earlier Antler highlands, or were an extension of the Pennsylvanian-Permian Humboldt highlands proposed by Ketner (1977, p. 363). The thick accumulation of Quadrant sandstone can be attributed to the proximity of the source area and a more rapid subsidence rate in southwestern Montana. Maughan (1975, p. 287) describes this trough as a branch of the miogeosyncline that extended northeast from Idaho. The flexibility of this basin persisted into the Permian and was a contributing factor in the distribution of sediments in the Permian sea.

The Permian system has a remarkable variety of facies that represent continuous deposition in shifting environments in a relatively shallow epeiric sea. The carbonate facies of the Park City and Gerster Formations dominated sedimentation south in eastern Nevada, central and northern Utah, and southwestern Wyoming; the chert-mudstone phosphorite facies of the Phosphoria Formation dominated in eastern Idaho, northern Utah, western Wyoming and southwestern Montana; the Shedhorne Sandstone facies is best developed in northwestern Wyoming (McKelvey et al., 1956, p. 2834).

The Humboldt highland belt appeared to have been most influential during Early to Middle Permian time in central Nevada, and the position of the highlands in central Idaho is difficult to determine. The greatest influence of the northern highlands as a western source of sediment occurred during Desmoinesian and Missourian (Middle Pennsylvanian) time and diminished by the Early Permian (Maughan, 1975, p. 288-289). The northern highlands became insular at this time, and were less influential as a source of sediment. They may have channeled cold arctic waters south along the craton causing upwelling along the coast that contributed to the deposition of phosphorites in the Phosphoria Formation.

The Phosphoria basin extended north from Idaho through Lima, Montana north northeast past Dillon. The source of silt and sand was intermittently from the west and northwest, and from the east. The trough was deepest during Meade Peak and Retort time when the eastern and western sources of sediment were of little influence. Cold arctic waters were channeled into the area and deposition of the phosphorites ensued. The western source also supplied very little sediment during deposition of the spiculate cherts of the Rex and Tosi Members of the Phosphoria Formation. The water was shallower than during deposition of the phosphorites and the cherts accumulated as siliceous oozes in depths of less than 50 m. The Permian sea in southwestern Montana during these periods of time was warm and was the site of deposition of carbonates in the Grandeur and Franson Tongues of the Park City Formation (Cressman and Swanson, 1964, p. 378-383). The western highland apparently controlled the circulation of the sea water and became a source of silt and sand during Grandeur and Franson time.

Much of the western part of the North American continent was covered by a shallow sea during Permian time. Paleoclimate reconstruction indicates that a strong global thermal gradient existed with the paleoequator occupying the Texas-New Mexico-Coahuila area. Permian biotic provinces reflect a paleoclimatic control on the distribution of the biota; it becomes possible to recognize high-latitude, mid-latitude, and low-latitude provinces (Yancey, 1975, p. 759). The region of Phosphoria deposition occurs within the mid-latitude province. The boreal (Arctic) province was located in northern Canada and in northern Europe and Asia. This broad geographical reconstruction is compatible with a source for the cold, nutrient-rich waters that flowed into the Phosphoria basin causing phosphorite deposition. Southwestern Montana was located in the region of the paleo-northeasterly trade wind belt, a similar condition that contributes to the upwelling of the cold California current along the west coast today (Cressman and Swanson, 1964, p. 380). Brachiopod and conodont faunas in the Phosphoria Formation indicate that cool water existed in the area of maximum phosphorite deposition in eastern Idaho and southwestern Montana (Wardlaw, 1980, p. 360).

A hiatus between Permian and Triassic strata across the miogeosyncline is indicated by biostratigraphic evidence discussed above, under Permian System. The distribution of Permian and Early Triassic strata indicates that sedimentation in the miogeosyncline during the Triassic followed a pattern similar to that of the Permian. In general, a broad, shallow carbonate platform covered the area during the Triassic with the thickest accumulation in an embayment in extreme southeastern Idaho. In west-central Nevada and southeastern Idaho Early Triassic rocks disconformably overlie medial to Late Permian strata. Farther north in southwestern Montana lithologic similarity between the upper Phosphoria and lower Dinwoody Formations indicates that deposition may have been continuous; however, biostratigraphic evidence does exist (Kummel, 1954; Wardlaw and Collinson, 1979, Figure 8) that indicates a hiatus between the Phosphoria and Dinwoody Formations in southwestern Montana that spanned late Guadalupian and Dzfulfian time. The widespread Permian-Triassic unconformity was possibly created by a regional uplift associated with the Sonoman orogeny of central Nevada.

The Early Triassic stratigraphy of southwestern Montana, western Wyoming, eastern Idaho, and northeastern Utah indicates two transgressive stages across the craton separated by a regression. The slope on the submarine platform was very low so that slight changes in sea level or changes in the rate of terrigenous sediment supply were reflected in broad migrations of the facies. The intertonguing of the Dinwoody, Woodside, and the Thaynes Formations with the Ankareh Formation illustrates this observation. Generally, transgression was accompanied by deposition of carbonates on the shallow platforms in southwestern Montana, western Wyoming and northern Utah, and shale in the miogeosynclinal basin in eastern Idaho. Regression was accompanied by a migration of marginal marine and terrigenous sediments off of the craton towards the miogeosyncline.

This sequence of events is evident in the Triassic strata of the thesis area. The initial transgressive stage must have been rapid and was possibly accompanied by a rapid subsidence in the Dinwoody basin. The lower fissile siltstone member of the Dinwoody may reflect deposition during the maximum transgression when the Dinwoody sea extended as far east as central Wyoming. The upper siltstone-limestone member was deposited on a tide-dominated continental shelf within or just below wave base subjected to intermittent storm swells. This shallowing up sequence is indicative of the regressive stage that culminated with deposition of the intertidal redbeds and carbonates of the Woodside Formation. The regression indicated by the deposition of the Woodside Formation could be sedimentological and may have been the result of an increase in clastic terrigenous sediment influx coupled with a decline in the rate of subsidence of the Dinwoody basin. These conditions were reversed during Thaynes time, for the sea transgressed once more over southwestern Montana, western Wyoming, northern Utah, and spread as far south along the edge of the craton as northern Arizona (Collinson and Hasenmueller, 1978, Figure 6b, p. 182). The limestones, siltstones, and very fine-grained sandstones of the Thaynes Formation at Little Sheep Creek were deposited in a shallow shelf environment similar to the Dinwoody Formation.

Sedimentation continued at least into the Middle Triassic in Nevada (Collinson and Hasenmueller, 1978, Figure 2) and it can be assumed to have continued elsewhere in the Cordilleran miogeosyncline. However, Middle and Late Triassic and Early Jurassic strata are missing from the northern part of the miogeosyncline where Early Triassic rocks are overlain disconformably by Middle Jurassic rocks.

Emergence of the miogeosyncline during the Middle and Late Triassic and the Early Jurassic marked the end of miogeosynclinal sedimentation. The foreland basins that had existed during the Late Paleozoic and Early Triassic would no longer exist.

The emergence of the western interior prior to the Middle Jurassic transgression created a marked erosional unconformity; Middle and Late Jurassic marine and marginal marine strata overlie rocks in Montana, Wyoming, and Idaho that become progressively younger to the south and range in age from Mississippian to Early Jurassic. Transgression of the Jurassic seas occurred in four pulses that spread farther south with each pulse (Hallam, 1975, p. 119-123). Late Mesozoic and Tertiary uplift, folding, and faulting has limited the number of outcrops of Jurassic rocks west of the overthrust belt; the western limit of the Jurassic seas cannot be determined accurately, but a paleogeographic reconstruction of Jurassic North America (Hallam, 1975, Figures 8.2-8.6) indicates a positive area in part of the western states that limited the invasion of the seas during Bajocian-Oxfordian time. The Jurassic sea spread south across Canada into the western interior of the United States as far as northern Arizona and northern New Mexico. The gradient of the sea floor must have been very low because, except for the Sweetgrass-Belt line of uplift in central Montana, the sedimentation was uniform over a very broad region. The Sweetgrass-Belt uplift became emergent during the Bajocian and locally controlled sedimentation to the end of the Callovian. The emergent areas were low-lying and were not major sources of sediment, but they apparently did restrict the circulation of sea water south and east in southeastern Montana and Wyoming, as indicated by the distribution of gypsiferous red beds of the Piper Formation, Gypsum Spring Formation, and the lower Twin Creek Formation. The sandstone, shale, and limestone of the Sawtooth Formation are more indicative of normal marine conditions west and north of the islands.

The Bajocian sea extended south from the Arctic region into Montana, around the Belt and Sweetgrass arches into western Wyoming, eastern Idaho, and central Utah. The initial phase of the transgression resulted in the deposition of limestone, sandstone, and shale of the Sawtooth Formation in western Montana and limestone of the Piper Formation in northern Montana. Deposition of the intertidal deposits of the Gypsum Spring Tongue of the Twin Creek Formation in the Little Sheep Creek area and elsewhere in southwestern Montana and its correlatives in eastern Idaho, Wyoming and southern Montana was occurring simultaneously.

As transgression continued into the Bathonian the area encompassed by the Twin Creek Formation continued to subside faster than elsewhere on the craton. A north-south-trending trough was formed along the Idaho-Wyoming border in which was deposited the thickest accumulation of marine Jurassic rocks. The eastern margin of the basin is marked by the eastern boundary of the Twin Creek Formation, or the transition from normal marine open shelf sediments on the west to hypersaline red beds to the east (Wright, 1973, Figure 8).

The sea transgressed over southwestern Montana during late Bajocian and Bathonian time with deposition of the Sawtooth Formation. The lime muds and argillaceous lime muds of this formation were deposited rapidly in open shelf environments below wave base. Ostrea, tentatively identified in the formation, was an epifaunal suspension feeder entirely conditioned to living on soft muddy substrates (Wright, 1973, Figure 4). That the shells are fragmental indicates that storm-generated waves may have occasionally reached the bottom.

The oolites at the base of the Rierdon Formation indicate an initial shoaling of the Jurassic sea at the onset of Callovian time. The sea deepened across the entire area of Callovian sedimentation as the shoreline transgressed farther east into the Dakotas and south into Wyoming. The fine-grained, sparsely fossiliferous limestones in the upper Rierdon were deposited in open marine conditions below normal wave base. Local lenses of oolitic limestone in the micritic limestone in the Little Sheep Creek area indicate local areas of shoaling in the Rierdon sea.

A pronounced emergence of the Sweetgrass-Belt region of uplifts during the middle and upper Callovian created an erosional unconformity between the Rierdon and overlying Swift Formations.

The Oxfordian saw the submergence of the Sweetgrass arch, a nearly complete submergence of the Belt arch, and the maximum transgression of the sea east into the Dakotas and Nebraska, and as far south as New Mexico and Arizona. Sedimentation in Montana and Wyoming consisted dominantly of sandstones to the west and shales to the east. The sandstones of the Oxfordian Swift Formation were deposited as a nearly uniform blanket of clastics over most of Montana. The formation is the most widespread of the Ellis Group of formations.

Slow epeirogenic uplift of the area, possibly associated with the Nevadan orogeny farther west, caused a draining of seas from the area during late Oxfordian time. Deposition of the variegated mudstones, sandstones and freshwater limestones of the Morrison Formation on a broad alluvial floodplain ensued. Eastward transport of these sediments is indicated regionally from a source an unknown distance to the west. The Belt arch persisted into the latest Jurassic and locally controlled sedimentation of the Morrison Formation. Reworking of Swift sandstone into the base of the Morrison Formation indicates that the change from marine to fluvial sedimentation was transitional and that the contact is conformable. The Swift Formation ranges in thickness up to 6 m in the Tendoy Range. This very thinness and the reworking of the Swift sands into the Morrison may have created local unconformities. Such an unconformity exists in the thesis area where the Morrison Formation lies unconformably on the Rierdon Formation.

The Belt arch had completely subsided by the Early Cretaceous and became the site of the thickest accumulation of Kootenai sediments (Suttner, 1969, p. 1,406). Early Cretaceous uplift became more intense in Oregon and its position migrated east into Idaho at this time. The proximity of these highlands was closer in Kootenai than in Morrison time and is reflected in the coarser sediments found in the lower Kootenai. The coarse sediments and the variegated mudstones of the Kootenai were deposited in a broad alluvial plain similar to that of the Morrison Formation. The chert-pebble conglomerates and salt-and-pepper sandstones in the basal Kootenai were probably deposited as a wide sheet of sediment by laterally migrating rivers flowing east from the highlands. Locally, the streams may have cut channels into the underlying Morrison Formation. The freshwater gastropod limestone marking the top of the Kootenai Formation and its correlatives, the

Peterson and Draney Limestones, was deposited in temporary basins that developed in the alluvial floodplain marginal to the Early Cretaceous sea farther east in Montana (Suttner, 1969; Holm et al., 1977). The Colorado Shale (or, the Aspen Formation of Scholten et al., 1955) also consists of variegated mudstones with lenses of quartz sandstone, salt-and-pepper sandstone, and fresh water gastropod-bearing limestone deposited on a swampy alluvial plain.

The Morrison and Kootenai Formations and the Colorado Shale in the thesis area form a sequence of terrigenous clastic sediments that was deposited as a clastic wedge east of an orogenic highland in eastern Oregon. The highlands were probably responses to the Nevadan orogeny and their position migrated eastward during the Early Cretaceous from Oregon to Idaho. The proximity of the highlands is reflected in the coarser sediments found in the Kootenai and Colorado Shales. The late Albian to mid-Paleocene Beaverhead Formation lies unconformably on the Colorado Shale in parts of southwestern Montana. However, a conformable contact occurs near Monida, Montana where shales and sandstones typical of the Colorado Shale are found intercalated in the lower Beaverhead Formation (Ryder and Scholten, 1973, p. 181). This conformable-unconformable relationship indicates that the source of clastic sediments had migrated even farther east and was proximal to Beaverhead deposition. The propagation of these highlands eastward had begun to deform the strata on which the Beaverhead conglomerates were deposited.

Scholten et al. (1955), Ryder and Ames (1970), and Ryder and Scholten (1973) have presented stratigraphic and structural evidence indicating that the Beaverhead Formation represents syntectonic sedimentation in a low sedimentation basin proximal to mountainous regions.

The source of the cobbles and pebbles was from as far west as the Idaho batholith, from the "Ancestral Beaverhead Mountains", and locally from the emerging Blacktail-Snowcrest anticlinorium and other local highland areas. Regionally, the Beaverhead Formation had been deformed into northeast-trending structures coincident with the Blacktail-Snowcrest uplift and later faulted by northwest-trending high-angle reverse faults and folded into northwest-trending folds.

The sequence of events described above for the region is not as clear in the thesis area. The Beaverhead Formation lies with a distinct unconformity over older strata that were faulted and folded prior to deposition of the formation. The outcrop pattern of the formation indicates that it is relatively undeformed. Fracture patterns observed in the Beaverhead Formation may be related to post-Laramide basin-and-range faulting, and not to Laramide deformation.

The source for the quartzite and limestone cobbles of the Beaverhead Formation in Little Sheep Creek was from the southwest (Ryder and Scholten, 1973, p. 789). The sediments were deposited after deformation of the older strata into northeast- and northwest-trending structures. The sequence of structural events in the thesis area, therefore, does not precisely follow the regional trend.

The structures within the thesis area are "Laramide" in the broadest sense of the word as used by Eardley (1951, p. 287) and others. This definition incorporates the folding and thrusting of a miogeosynclinal sequence of strata along the overthrust belt, and the broad asymmetric uplifts typical of the eastern Rocky Mountains. The Laramide orogeny was a response to compressional forces that abated after the Paleocene. The onset of the deformation can be argued to have begun in the Late Jurassic, or more specifically, at least with the beginning of Beaverhead deposition in the late Albian (mid-Cretaceous).

Hamilton (1978, p. 54-58) provided a temporal distinction between the two styles of deformation. The thrusting and folding of the miogeosynclinal wedge of sediments began in the Late Jurassic and continued into the Paleocene, but was essentially a Cretaceous phenomenon. He considered the asymmetric uplifts on the craton to be Laramide and limited the orogeny to the time spanning the latest Cretaceous to the Eocene. The two events do overlap, but essentially the thrusting ceased soon after the onset of the Laramide orogeny. The overlapping of the structures is documented by the diversion of imbricate thrusting around the Uinta Mountains (Laramide) and the eastward progression of thrust faults south of the Teton-Gros Ventre uplift (Laramide) into western Wyoming. Hamilton (1978, p. 55) also suggests this relationship may exist in southwestern Montana "where Paleocene structures bulge into the cratonic Laramide Crazy Mountains Basin."

The structures in the thesis area represent overlapping styles of deformation, and it is difficult to establish the timing of the events. The southwestern source of the Divide Quartzite and the Divide Limestone Conglomerates of the Beaverhead Formation in the thesis area indicates that a highland source lay in that direction. The Beaverhead lies with distinct angular unconformity on older strata, including the Colorado Shale. The age of the Beaverhead Formation in the Blacktail-Snowcrest anticlinorium indicates that the arch emerged in Albian-Cenomanian time and contributed sediments to the Beaverhead until Late Cretaceous or Paleocene time. The Beaverhead Formation is truncated by major northwest-trending, high-angle reverse faults such as the Tendoy fault. The faults and the associated folds are, therefore, mid- to late Paleocene phenomena. The Medicine Lodge thrust truncates the youngest Beaverhead strata and may have developed after the northwest-trending structures and is possibly an early to mid-Eocene structure (Ryder and Ames, 1970, p. 1169).

The timing of the regional structural events and the angular unconformity beneath the Beaverhead Formation in the thesis area, indicate that eastward-directed compressional forces persisted into the Tertiary and were superimposed upon the northeast-trending Blacktail-Snowcrest anticlinorium. In the thesis area, the southwest-plunging Garfield anticline was, therefore, probably refolded into the doubly plunging series of folds in the southwest corner of the map. This sequence of events is made more reasonable because the enormous thickness of the Quadrant Formation and its near-homogeneous composition of quartz sandstone within the axis of the Garfield anticline would have acted as a competent buttress against east- and northeast-directed compressional forces.

The middle and late Tertiary was a period of relative tectonic quiescence disrupted occasionally by extrusive igneous activity. The Challis volcanics of central Idaho roughly parallel the fault trace of the Medicine Lodge thrust system (Ruppel, 1978, p. 18) and are approximately equivalent to the Oligocene Muddy Creek volcanics and the Miocene Medicine Lodge volcanics (Scholten et al., 1955, Table 3) in the Tendoy and Beaverhead Ranges. The spatial relationship between the fault trace and the volcanic rocks suggests that the fault could have served as a conduit for local eruptive centers satellite to the main Challis volcanic field (Ruppel, 1978, p. 18).

Local uplifts during the late Oligocene and Eocene contributed clastic material to the Muddy Creek and Medicine Lodge beds, respectively, of the Tendoy Range (Scholten et al., 1955, p. 398). Basin-and-range type normal faulting may have commenced during the Miocene in this region creating basins in which the Miocene sediments collected. The Round Timber limestone probably accumulated in one of these late Miocene basins. The lack of a significant detrital component in the Round Timber limestone indicates that the region may have been low-lying, a condition characteristic of the area by the late Tertiary. The Pliocene Edie School rhyolites covered the low-lying surface over most of the area of the South Medicine Lodge basin and lapped onto the Round Timber limestone.

Normal faulting along the Deadman and Kissick Faults on the west front of the Tendoys and the Red Rock fault zone on the eastern front started in the late Pliocene-early Pleistocene time, eventually creating the present Tendoy Range. Similar basin-and-range type faulting was creating the Beaverhead, Lemhi and other ranges in central Idaho with the intervening intermontane basins, and the system extended northeast to create the uplift of the modern Blacktail Range along the Blacktail fault. Triangular facets cutting across alluvial fans along the eastern front of the Tendoy Range indicate that uplift of the Tendoy Range along the Red Rock fault has continued into the Recent. Related Recent movement along normal faults has been documented along the Centennial fault (Witkind, 1975) and by the Hebgen Lake earthquake in 1959 in the Madison Range.

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APPENDIX A

Location of Slumps and Debris Slides

1. NE¼ of the NE¼ of the SE¼ of Section 19, T. 15 S., R. 9 W.: a slump located 150 m south of a spring occurring at the approximate contact between the Thaynes Formation and the overlying limestone of the Round Timber limestone.

2. E½ of the SW¼ of Section 17, T. 15 S., R. 9 W.: a slump involving the Woodside and Thaynes Formations.

3. W½ of the SE¼ of the NW¼ of the SE¼ of Section 7, T. 15 S., R. 9 W.: a slump involving Mississippian limestone.

4. SE¼ of the SE¼ of the SE¼ of Section 5, T. 15 S., R. 9 W: east across the West Fork of Little Sheep Creek from Seybold Spring and immediately south of the Leitch farm is a slump in which a segment of the Thaynes Formation has gravitationally overridden the Woodside Formation.

5. SW¼ of the NW¼ of the SE¼ of Section 9, T. 15 S., R. 9 W: a slide on the eastern side of the valley drained by the Middle Fork of Little Sheep Creek involves the Phosphoria Formation.

6. N½ of the SE¼ of the SE¼ of Section 31, T. 14 S., R. 9 W.: a slump and debris slide noticeable from the White Pine Ridge road involves Mississippian limestone and the underlying (because of thrusting) Beaverhead Formation.

7. S½ of the SW¼ of Section 16, T. 15 S., R. 9 W., and the N½ of the NW¼ of Section 21, T. 15 S., R. 9 W.: a slump involving the Dinwoody Formation.

References for Plate III
(Numbers correspond to numbered columns)

1. Mamet, 1975.
2. Sando, Mamet, and Dutro, 1969.
3. Gutschick, Sandberg, and Sando, 1980; Hildreth, 1980.
4. Douglass, 1977.
5. Sando, Mamet, and Dutro, 1969.
6. Huh, 1968; Mamet, Skipp, Sando, and Mapel, 1971; Sando, Gordon, and Dutro, 1975; Sando, Dutro, Sandberg, and Mamet, 1976.
7. Huh, 1968; Sando, Mamet, and Dutro, 1969; Sando, Gordon, and Dutro, 1975; Sando, Dutro, Sandberg, and Mamet, 1976; Skipp and Hall, 1980.
8. Hildreth, 1980.
9. Maughan and Roberts, 1967; Sando, Gordon, and Dutro, 1975; Skipp and Hall, 1980.
10. Sando, Mamet, and Dutro, 1969; Sando, Gordon, and Dutro, 1975; Sando, Dutro, Sandberg, and Mamet, 1976; Yancey, Ishibashi, and Bingman, 1980.
11. Maughan and Roberts, 1967; Skipp and Hall, 1980.

CALCITE - DOLOMITE STAIN PROCEDURE (after Niem, 1979)

Solutions Needed:

1) HCl - 10% soln.

2) Staining soln. - prepared as follows (under the hood):
   a) Put 1 ml. of concentrated HCl in a 500 ml. beaker
   b) Add about 50 ml. of distilled water
   c) Add 2.5 gm. of potassium ferricynide
   d) Add 0.5 gm. of Alizarin red S
   e) Mix with a stirring rod, add distilled water until the beaker contains 500 ml. and stir again.

Procedure:

1) Slab the rock to be examined; it is not necessary to polish the surface in order to obtain satisfactory results.

2) Etch the slabbed surface by holding it in a shallow dish containing the 10% HCl soln. for only 3-5 seconds.

3) Rinse the surface gently with distilled water.

4) Warm the staining soln. to approx. 40 deg. C. under the hood and keeping it at about this temp. hold the etched surface in this soln. for about 45 seconds. Rinse in distilled water, dry and spray with plastic coating.

Results:

Calcite - stains a dark red or purplish red

Dolomite - does not take the stain

Ferrous dolomite$_1$ - stains blue

Ferrous calcite - stains bluish red

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Method for Impregnating
Round Timber Limestone

Materials needed:

1. Suitable container in which rock sample can be completely immersed in a polyester resin, i.e., paper cup, or a cup molded from aluminum foil.

2. Polyester Resin manufactured by Fiberlay, Inc., 1158 Fairview North, Seattle, Wash., 98109.
   - a) Fiberlay resin #4116
   - b) Styrene monomer (thinner) #P-205
   - c) Fiberlay catalyst #P-102

3. Vacuum oven with temperature control.

4. Suitably sized sample.

Procedure:

1. Place sample in cup. Support the sample on several small pieces of foil or rock to allow resin to flow underneath the sample.

2. In a throw-away container mix Fiberlay resin #4116 with Styrene monomer (thinner) #P-205 in the proportions of 2 parts resin to 1 part thinner. Use more thinner for less porous samples.

3. Add Fiberlay catalyst #P-102 in the proportions of 10 drops catalyst to 1 pint of thinner/resin mix. Stir well. (For less porous rocks use fewer drops of catalyst. This will reduce the "set-up" time and allow the resin to penetrate the rock more completely.)

4. Pour mixture over sample in cup, completely covering the sample.

5. Place sample in vacuum oven and turn on vacuum. Leave under vacuum for about one hour.

6. Turn off vacuum and turn on heat; set temperature to about 60°C. Leave for 24 hours.

7. Remove from vacuum oven and break, or cut away, a part of the encrusting resin exposing the rock.

8. Immerse in hydrochloric acid until the limestone has dissolved completely.