AN ABSTRACT OF THE THESIS OF

Richard A. Klecker for the degree of Master of Science

in Geology presented on May 29, 1980

Title: Stratigraphy and Structure of the Dixon Mountain-Little Water Canyon Area, Beaverhead County, Montana

Abstract approved: Redacted for Privacy

Keith F. Oles

Approximately 23 square miles of the east-central part of the Tendoy Range are included in the thesis area which lies three miles west of Dell, Montana.

Rocks ranging in age from Late Mississippian to Recent are exposed in the study area. The upper Paleozoic section includes approximately 4,200 feet (1,281 m.) of strata which are divided into four formations. The nine formations of Mesozoic age total approximately 2,960 feet (903 m.) plus part of the Cretaceous-Tertiary Beaverhead Formation. Cenozoic strata in the thesis area include Beaverhead conglomerates, basaltic andesite flows, tuffs, and breccias, and unconsolidated gravels, alluvium, and alluvial fan gravels.

The thesis area was examined in detail to define formation distribution and structural complications. The Triassic and Jurassic (Ellis Group) rocks were examined specifically in an effort to interpret their depositional

environments. Sedimentary structures, fossils, and petrographic evidence indicate a shallow subtidal and intertidal setting for the deposition of Triassic rocks. Algal laminated strata within the Woodside Formation have not previously been described. Much of the Ellis Group was deposited under shallow subtidal and locally shoaling conditions. Glauconite calcarenites in the Swift Formation may represent deposition in sand waves.

Folding and faulting within the thesis area represent the overlapping influence of geosynclinal and cratonic tectonism in the hinge line setting of the Tendoy Range. The imposition of a southward-plunging anticline on a southwestward-plunging syncline has resulted in the overturning of the northern limb of the syncline. A low-angle thrust with eastward vergence has been developed in the Dry Canyon area. The thrust has been cut by a later high-angle reverse fault. White Knob strata (Mississippian) have been carried into the western margin of the thesis area on the Medicine Lodge thrust.

The economic possibilities are not encouraging. The phosphate deposits of the thesis area have been analyzed and deemed uneconomical. Distance from intrusions precludes the chance of significant mineralization in the area. The petroleum potential for the area is doubtful. Late Cretaceous folding and erosion may have allowed escape of hydrocarbons from the potential source and reservoir rocks of the region. However, the Medicine Lodge thrust sheet may conceal structural and stratigraphic configurations capable of trapping oil and gas.

Stratigraphy and Structure of the
Dixon Mountain-Little Water Canyon Area
Beaverhead County, Montana

by

Richard Alan Klecker

A THESIS

submitted to

Oregon State University

in partial fulfillment of
the requirements for the
degree of

Master of Science

Completed June 1980

Commencement June 1981

APPROVED:

Redacted for Privacy

Professor of Geology
in charge of major

Redacted for Privacy

Chairman of Department of Geology

Redacted for Privacy

Dean of Graduate School

Date thesis is presented May 29, 1980

Typed by Lynnette Klecker for Richard Alan Klecker

ACKNOWLEDGEMENTS

To my wife, Lynnette, I wish to express my gratitude for her support, encouragement, and assistance in bringing this thesis to completion. Without her help this all would have taken much longer.

To Dr. Keith F. Oles goes my gratitude for his critical reading of the manuscript. I would also like to thank him for his visit to the field area and the advise and encouragement he offered. To Drs. J.G. Johnson and A.R. Niem go my appreciation for their critical reading of the manuscript and their advise and guidance. To Dr. R.D. Lawrence goes my gratitude for his assistance in unravelling some of the structure of the thesis area.

The orientation and critiques offered by Dr. Robert Scholten, of Pennsylvania State University, are also greatly appreciated, as they aided in an early familiarization with the geology of the study area.

The many conversations with my fellow students, Gail Hildreth and Kumbe Sadler, prior to, during, and after the field season, provided reenforcement and criticisms for many of the fledgling hypothesis which rose and fell during the writing of this thesis.

The financial support provided by Union Oil Company of California is gratefully acknowledged.

TABLE OF CONTENTS

	Page
INTRODUCTION	
Location and Accessibility	1
Major Geographic Features	3
Investigative Methods	7
Previous Work	8
THESIS AREA STRATIGRAPHY	
White Knob Group	
Nomenclature	13
Middle Canyon Formation	
Nomenclature	14
Distribution and Topographic Expression	14
Lithology	16
Big Snowy Formation	
Nomenclature	18
Distribution and Topographic Expression	19
Fossils and Age	21
Lithology	22
Amsden Formation	
Nomenclature	25
Distribution and Topographic Expression	28
Fossils and Age	28
Lithology	30
Quadrant Formation	
Nomenclature	33
Distribution and Topographic Expression	35
Fossils and Age	37
Lithology	37
Phosphoria Formation	
Nomenclature	39
Distribution and Topographic Expression	41
Fossils and Age	44
Lithology	45
Depositional Environment	53
Dinwoody Formation	
Nomenclature	55
Distribution and Topographic Expression	56
Fossils and Age	59
Lithology	62
Depositional Environment	69
Woodside Formation	
Nomenclature	75
Distribution and Topographic Expression	75
Fossils and Age	77
Lithology	78

Page

Depositional Environment 86

Thaynes Formation

Nomenclature 92

Distribution and Topographic Expression 93

Fossils and Age 95

Lithology 97

Depositional Environment 108

Ellis Group

Nomenclature 114

Sawtooth Formation

Nomenclature 114

Distribution and Topographic Expression 115

Fossils and Age 117

Lithology 118

Depositional Environment 122

Rierdon Formation

Nomenclature 124

Distribution and Topographic Expression 125

Fossils and Age 125

Lithology 126

Depositional Environment 133

Swift Formation

Nomenclature 135

Distribution and Topographic Expression 136

Fossils and Age 136

Lithology 137

Depositional Environment 142

Morrison Formation

Nomenclature 146

Distribution and Topographic Expression 147

Fossils and Age 147

Lithology 148

Kootenai Formation

Nomenclature 149

Distribution and Topographic Expression 151

Fossils and Age 153

Lithology 154

Beaverhead Formation

Nomenclature 159

Distribution and Topographic Expression 160

Fossils and Age 162

Lithology 162

Medicine Lodge Volcanics

Nomenclature 165

Distribution and Topographic Expression 166

Lithology 166

Unconsolidated Gravels

Description 171

	Page
STRUCTURAL GEOLOGY	
Regional Structure	174
Thesis Area Structure	177
Folds	177
Faults	181
GEOLOGIC HISTORY	189
ECONOMIC GEOLOGY	204
BIBLIOGRAPHY	209

LIST OF FIGURES

Figure		Page
1	Location map of thesis area.	2
2	View of Dixon Mountain and western part of thesis area.	5
3	View of Dixon Mountain, Timber Butte, Little Water Canyon, and Dry Canyon.	6
4	Generalized stratigraphic column for thesis area.	12
5	Distribution of Middle Canyon Formation.	15
6	Distribution of Big Snowy Formation.	20
7	Rhythmically bedded limestone and black shale of Big Snowy Formation.	24
8	Distribution of Amsden Formation.	29
9	Limestone conglomerate at base of Amsden Formation.	32
10	Distribution of Quadrant Formation.	36
11	Distribution of Phosphoria Formation.	43
12	Thin-bedded chert of the Rex Chert Member of the Phosphoria Formation.	48
13	Chert breccia of Grandeur Tongue of the Park City Formation.	50
14	Distribution of Dinwoody Formation.	51
15	Exposures in Hidden Pasture Valley.	58
16	Photomicrograph of calcareous siltstone of Dinwoody Formation.	65
17	Photomicrograph of silty pelecypod-*Lingula* packstone of Dinwoody Formation.	68

Figure		Page
18	Distribution of Woodside Formation.	76
19	Photomicrograph of algal laminated, silty micrite of Woodside Formation.	84
20	Distribution of Thaynes Formation.	94
21	Thin-bedded limestones of Thaynes Formation.	98
22	Multiple erosionl surfaces of upper Thaynes Formation.	102
23	Photomicrograph of whole pelecypod packstone of Thaynes Formation.	104
24	Photomicrograph of pelecypod-gastropod grainstone of Thaynes Formation.	106
25	Feeding trails (snail trails ?) of Thaynes Formation.	109
26	Distribution of Jurassic age strata.	116
27	Photomicrograph of fossiliferous packstone of Sawtooth Formation.	127
28	Photomicrograph of fossiliferous oolitic grainstone of Rierdon Formation.	130
29	Photomicrograph of Bio-lithoclastic grainstone of Swift Formation.	140
30	Distribution of Kootenai Formation.	152
31	Photomicrograph of gastropod grain-stone of Kootenai Formation.	158
32	Distribution of Beaverhead Formation.	161
33	Distribution of Medicine Lodge volcanics.	167
34	Basaltic andesite flows of Medicine Lodge volcanics.	169

Figure		Page
35	Photomicrograph of basaltic andesite of Medicine Lodge volcanics.	170
36	Distribution of unconsolidated gravels.	172
37	Physiographic features and tectonic provinces of southwestern Montana.	175
38	Timber Butte anticline.	178
39	Axes of thesis area folds.	179
40	Surface trace of faults present in thesis area.	182
41	Planar jointing in Beaverhead Formation.	185
42	Triangular facets along trace of Red Rock Fault.	187

LIST OF TABLES

Table		Page
1	Summary of sedimentary structures and environmental indicators of Dinwoody Formation.	74
2	Summary of sedimentary structures and environmental indicators of Woodside Formation.	91
3	Summary of sedimentary structures and environmental indicators of Thaynes Formation.	112

LIST OF PLATES

Plate		Page
1	Geologic map of thesis area.	Pocket
2	Geologic cross-sections of thesis area.	Pocket
3	Correlation chart.	Pocket

STRATIGRAPHY AND STRUCTURE OF THE DIXON MOUNTAIN-LITTLE WATER CANYON AREA BEAVERHEAD COUNTY, MONTANA

INTRODUCTION

Location and Accessibility

The Dixon Mountain-Little Water Canyon area lies in the central part of the Tendoy Range of southwestern Montana. The Red Rock basin and the western Centennial basin compose the eastern margin of the Tendoy Range. The western limit of the range is defined by a series of basins; the Horse Prairie, North Medicine Lodge, Big Sheep Creek, and South Medicine Lodge. The Hap Hawkins Reservoir, occupying the site of the former town of Armstead, is at the northern limit of the range. The Continental Divide and the Idaho-Montana border follow the crest of the southwestern segment of the Tendoys.

The Dixon Mountain-Little Water Canyon area comprises approximately 23 square miles of the eastern flank of the Tendoy Range in Beaverhead County, Montana and includes much of the eastern two thirds of Township 13 South, Range 11 West. The area is approximately 42 miles south of Dillon, Montana and 12 miles north of Lima, Montana. The small town of Dell, Montana, on U.S. Highway 91 and Interstate 15, is three miles east of the thesis area. Approximately 11 square miles of the area fall within the

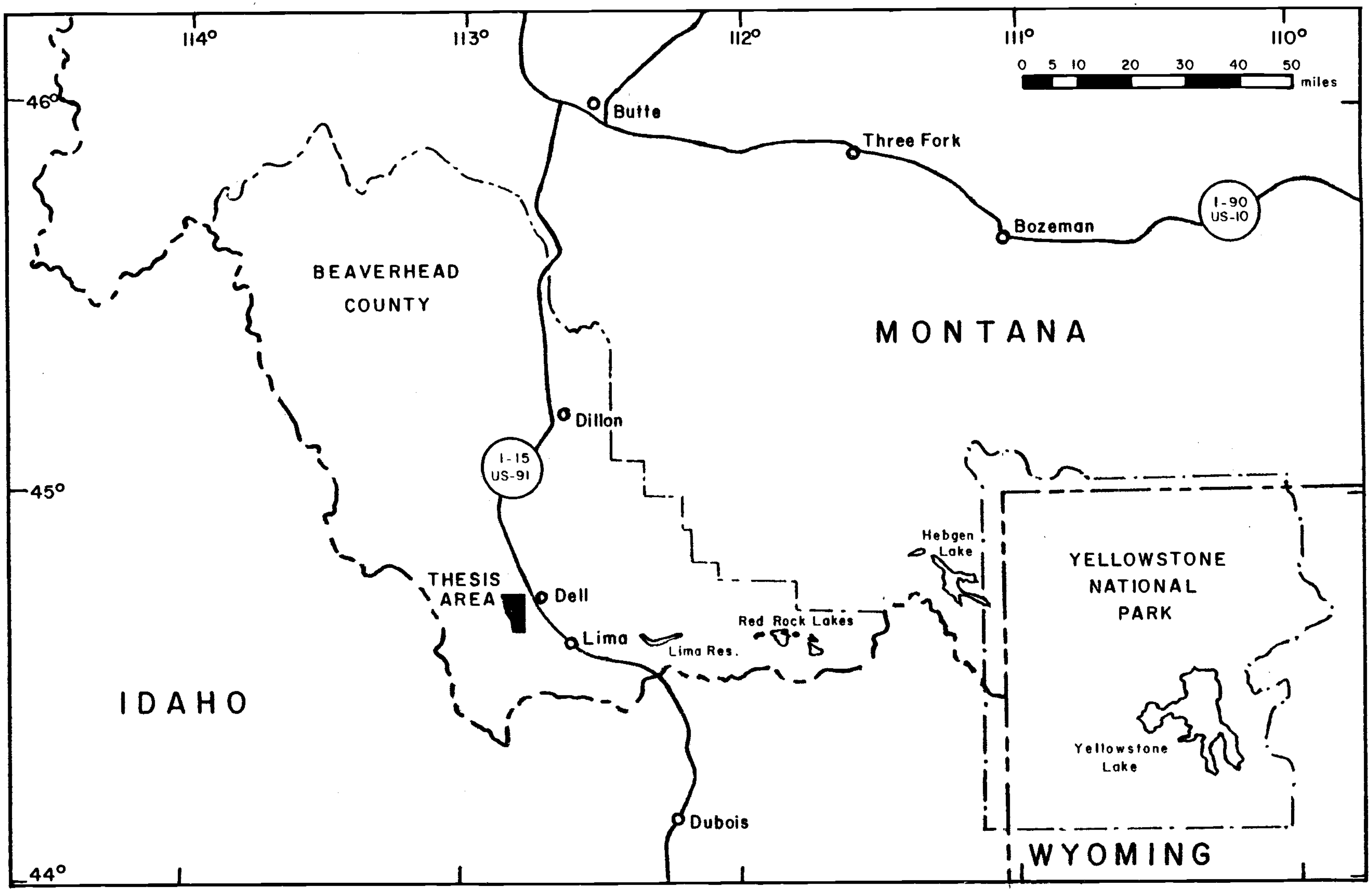

Figure 1. Location map of Dixon Mountain-Little Water Canyon thesis area.

boundaries of the Beaverhead National Forest.

A well maintained gravel road, following the course of Big Sheep Creek, traverses the southeasternmost corner of the area and provides access to the intermountane Muddy Creek basin. An unimproved road, originating at the Gene Hildreth ranch, due west of Dell, extends up the Little Water Canyon and bifurcates in open terrain to provide adequate access to a major part of the area.

Major Geographic Features

The lowest elevation within the thesis area (6,273 ft.; 1913 m.) lies upon the Quaternary fan deposits of the Red Rock Valley. The mountains and ridges of the Tendoy Range rise abruptly in elevation above the valley. The northwest-trending Red Rock fault zone separates the low-lying gravels of the valley from the rugged ridges and valleys of the Tendoys. Interfluvial ridges are truncated by the dip-slip displacement of the Red Rock fault zone, developing numerous triangular facets along the fault trace.

Mount Garfield (9,674 ft.; 2,950 m.) is the most prominent geographic feature within the thesis area. The southern flank of a second high ridge, Timber Butte (9,472 ft.; 2,889 m.) lies within the study area to the northwest of Mount Garfield. Both of these ridges are composed of highly resistant quartzose sandstone of the Pennsylvanian

Quadrant Formation. Both summits show evidence of the development of erosional surfaces prior to uplift.

The steep-walled canyons of Little Water Creek and Dry Creek lie between the prominent sandstone ridges, Little Water Creek is a perennial stream which drains a structural basin to the west and south of the canyon. Dry Creek canyon is the northern part of the Dry Creek-Hidden Pasture drainage system, a pair of strike valleys developed in Permian and Triassic limestones to the west of Mount Garfield. The Hidden Pasture area provides some excellent exposures of the Triassic strata along the antidip slope of an extensive (4.5 mi.) cuesta.

The western part of the study area is generally of low relief, though intermediate elevation, with a few minor drainages flowing westward into the Muddy Creek basin.

The highest discharge stream in the area is Big Sheep Creek. This stream, flowing through the southeastern corner of the study area, has incised a deep canyon through most rock units in the area, including the Quadrant sandstone. The lack of structural or lithologic control indicates that this stream is antecedent, having maintained its course as the Tendoy Range was block-faulted upward.

All other streams within the area of study display structural and lithologic control and are clearly not antecedent. Runoff within the area is low, therefore most streams show little evidence of stream bed erosion.

Figure 2. Dixon Mountain (left center) with Phosphoria strata on the dipslope (west of summit). Triassic rocks of Dry Canyon can be seen west of the notch in the center skyline. Looking south.

Figure 3. Dixon Mountain (left) and Timber Butte (right) as seen from eastern approach to thesis area from Dell, Montana. Little Water Canyon and Dry Canyon are in the low hills at center of photograph.

Investigative Methods

The field investigation commenced on July 3 and was concluded on August 23, 1979. During this eight week period, reconnaissance, mapping, and sample collection were completed.

Geologic findings were plotted on 1958 U.S. Department of Agriculture aerial photographs with a scale of 1:18,200.

The geology was subsequently transferred to a part of the Dixon Mountain quadrangle which has been enlarged to a scale of 1:12,000.

Bedding attitudes were measured with a Brunton compass. Stratigraphic thicknesses were measured directly with a Jacob's staff and Abney level. Paleocurrent indicators were measured directly and corrected for tectonic tilt using the technique described by Briggs and Cline (1967).

All sandstones are classified using the scheme outlined by Williams, Turner, and Gilbert (1954). A reference sand grain guage was used to judge the predominant grain size, as well as the degree of rounding and sorting of the clastic sediments.

Carbonate classifications are according to Folk (1962), or, where better characterization is possible, according to Dunham (1962).

Stratification and cross-stratification terminology is that of McKee and Weir (1953). The Geological Society of America Rock-Color Chart (1963) was employed to make field determinations of rock colors.

Previous Work

Previous work within the thesis area has been conducted by a number of investigators for a variety of purposes. The area provides one of a relatively few good exposures of the upper Paleozoic and Mesozoic section within the Tendoy Range.

The first geologic study of the area was conducted by the Hayden Survey. The results of this survey were published in 1872. During the first part of this century much of the stratigraphy of the area was completed on a regional scale. Douglass (1905) and Condit (1918) were responsible for much of this work in the southwestern Montana area and named some of the units present.

Few geologic investigations were undertaken following the work of Douglass and Condit until the late 1940's. During a few summers prior to 1948, a number of graduate students from the University of Michigan undertook a series of thesis problems within the Tendoy Range. The theses of Krusekopf (1948) and Wallace (1948) included parts of the present study area. These theses, as well as those of several others, were incorporated into a regional geologic study of the area produced by three of the University of Michigan students (Scholten, Keenmon, and Kupsch, 1955). Scholten subsequently was to spend many more summers in southwestern Montana and publish a great deal on the region.

The Cenozoic block faulting present in the area, including the Red Rock fault zone, was discussed by Pardee (1950). The Paleozoic stratigraphy of southwestern Montana was summarized by Sloss (1951). The Mississippian stratigraphy was considered by Holland (1952).

Interest in the economic potential of the Permian Phosphoria Formation resulted in a series of detailed studies of that unit. Good exposures of the Phosphoria within the thesis area led several workers to measure detailed sections. Klepper and others (1953) were the first to undertake such an investigation. Subsequently, Cressman (1955) produced a U.S.G.S. Professional Paper on the Phosphoria which was greatly expanded by Cressman and Swanson (1964) in a second U.S.G.S. Professional Paper.

Kummel's (1954) Professional Paper on the Triassic of southwestern Montana incorporated much of the material previously published on the subject.

Beginning, perhaps, with the regional study of the area published by Scholten and others (1955), a number of papers were written specifically on the Tendoy Range. Scholten's own interests were primarily those of regional tectonics but other contributions provided a better understanding of the stratigraphic and sedimentologic aspects.

Scholten (1961) and McMannis (1961) published papers outlining the depositional events represented in the southwestern Montana area. Scholten (1961) also summarized the tectonic events of the Tendoy and Beaverhead Ranges.

The regionally extensive, locally thick accumulations of syntectonic conglomerate, the Beaverhead Formation, were first named and described by Lowell and Klepper (1953). Ryder (1968), a student of Scholten, undertook a detailed study of the Beaverhead Formation. Ryder unravelled the complex depositional history of this conglomerate unit. Subsequently, Ryder and Ames (1970) and Ryder and Scholten (1973) were able to firmly date the upper two-thirds of the Beaverhead Formation and employ these dates to place time limits upon the multiple episodes of deformation that affected the region.

Maugham and Roberts (1967) produced a U.S.G.S. Professional Paper which added to the understanding of Big Snowy and Amsden Group stratigraphy and their relationship to the Mississippian-Pennsylvanian boundary. Additional valuable stratigraphic understanding was provided by Suttner (1968, 1969) for the Jurassic Morrison and Cretaceous Kootenai Formations. Detailed petrographic studies by Suttner provided an insight to the nature and proximity of the Laramide-age orogenic source to the west.

A brief paper by Scholten (1967) summarizes the tectonism of the region and gives a bleak appraisal of regional hydrocarbon potential.

A paper by Huh (1968), also a student of Scholten, based on a Ph.D. thesis, provides clarification of the Mississippian stratigraphy of the region. This paper detailed the nature of facies and age relationships between Mississippian carbonates deposited on the platform and the much thicker sequences deposited beyond the hinge zone in a miogeoclinal setting. The facies relationships that Huh documented are of importance to the tectonic interpretations of the region.

The most recently published papers are those of Ruppel (1978) and Skipp and Hait (1977). Both of these publications present tectonic interpretations of the area which differ from those of Scholten and his former students.

The theses of Hildreth and Sadler, as well as the present thesis, involve studies within the Tendoy Range. In addition to the three theses from Oregon State University, Timm Carr, a student of Dott at the University of Wisconsin, is preparing a Ph.D. thesis on the Triassic Thaynes Formation.

SYSTEM	ROCK UNIT	THICKNESS
TERTIARY	MEDICINE LODGE VOL.	
	BEAVERHEAD	15,000 ft. in region
--?--?---		
CRETACEOUS	KOOTENIA	approx. 1,000 ft.
JURASSIC	MORRISON	71 ft.
	ELLIS GRP.	230 ft.
TRIASSIC	THAYNES	680 ft.
	WOODSIDE	120 ft.
	DINWOODY	850 ft.
PERMIAN	PHOSPHORIA	833 ft.
PENSYLV-VANIAN	QUADRANT	2,662 ft.
	AMSDEN	approx. 200 ft.
--?--?---		
MISSISS-IPPIAN	BIG SNOWY	approx. 500 ft.
	WHITE KNOB GROUP ALLOCHTH.	unknown

Figure 4. Generalized stratigraphic column for thesis area.

THESIS AREA STRATIGRAPHY

White Knob Group

The name White Knob was first applied by Ross (1962) to rocks previously termed Brazer (Richardson, 1913) in south-central Idaho. At the type locality, on a ridge above Cabin Creek in the White Knob Mountains, Blaine County, Idaho, Ross described more than 7,300 feet (2227 m.) of predominantly pure limestone commonly containing chert nodules and laminae.

The White Knob was raised to group status by Huh (1967). Huh subdivided the group into four newly named formations. In ascending order they are: the Middle Canyon, Scott Peak, South Creek, and Surrett Canyon.

Huh (1967) described the Tendoy Range as lying along a Paleozoic transition zone between cratonic and miogeoclinal sedimentation regimes. Two Mississippian sequences, representing both sides of the transition, are present within the Tendoy Range. The White Knob Group is part of the miogeoclinal suite whereas the Madison Group is considered cratonic.

In the transition zone the White Knob Group is represented by the Middle Canyon and lower Scott Peak Formations and the Madison Group consists of the Lodgepole Formation and the thinned or absent Mission Canyon Formation.

Middle Canyon Formation
Late Mississippian

The Middle Canyon Formation was described and named by Huh (1967) at its type section along Middle Canyon, in the southwestern Lemhi Range, near Howe, Idaho. Huh characterized the 1,000 foot (305 m.) type section as being composed of two lithologies: thin-bedded, dark gray, chert-bearing, fine-grained limestone; and light-brown weathering, calcareous, quartz siltstone and fine-grained sandstone.

In Dixon Mountain-Little Water Canyon area the Middle Canyon is allochthonous, being part of the Medicine Lodge thrust sheet. The contacts with other strata are, therefore, entirely tectonic (Plate 1).

Distribution and Topographic Expression

The resistant limestones and sandstones of the Middle Canyon Formation support ledges, ridges, and isolated spines and points. The deformation this unit has undergone has left the strata dipping nearly vertically.

The topographically high ridges and hills extending approximately two miles from SW ¼, sec. 16, T.13 S., R.10 W. are entirely Middle Canyon strata. Two much smaller outcrops occur to the north of this ridge system, near the western boundary of the thesis area (Plate 1).

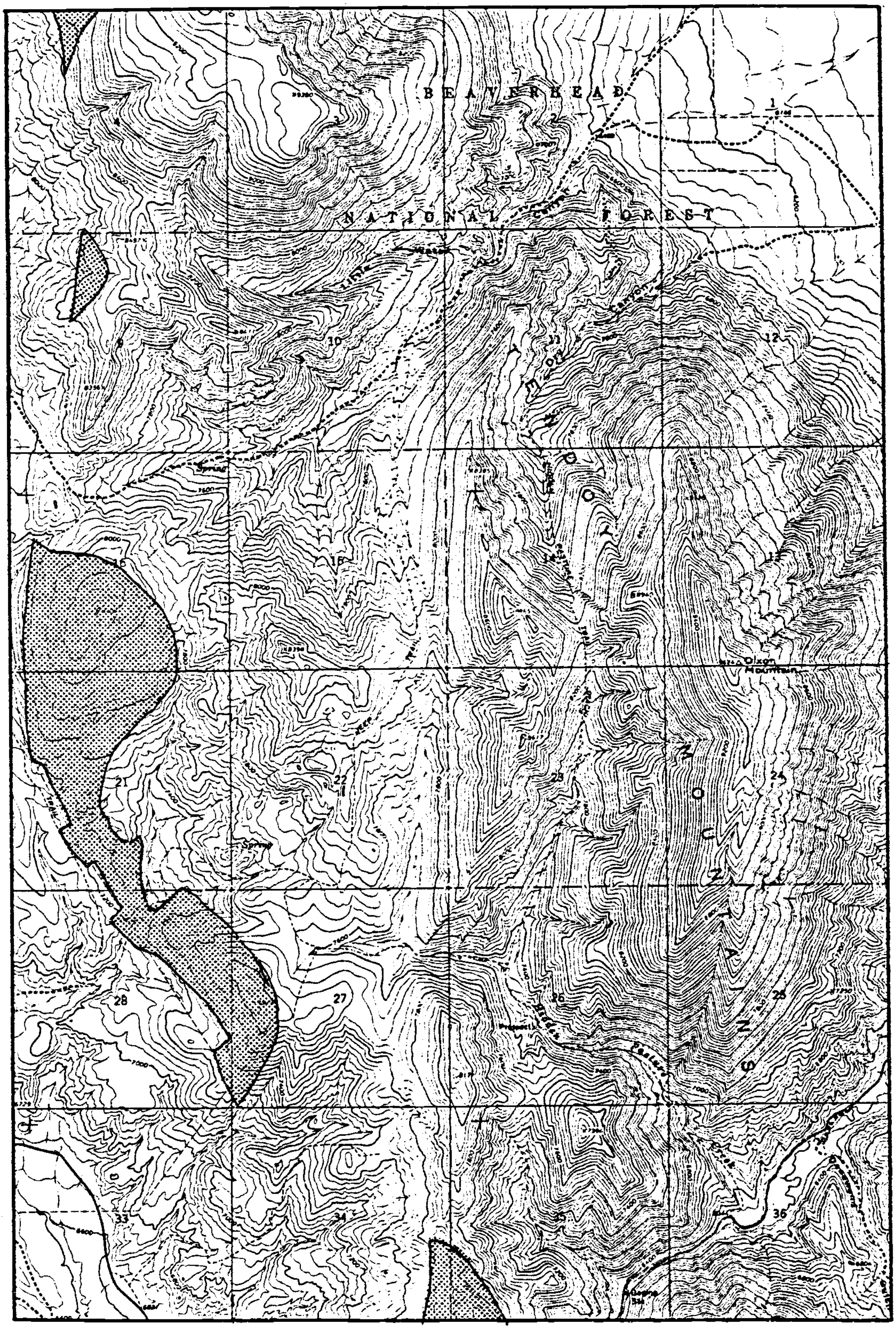

Figure 5. Distribution of Middle Canyon Formation.

Lithology

<u>Outcrop Character</u> The Middle Canyon Formation may be divided into limestone and calcareous quartz siltstone strata (Huh, 1967). The limestone strata are typically thin-bedded, unfossiliferous, dark, and micritic with isolated fossiliferous beds containing bryozoan and echinoderm fragments. These limestones are dark gray (N 3) to medium light gray (N 6) or yellowish gray (5 Y 8/1) and generally weather to slightly lighter colors. Chert occurs within these limestones as thin beds, laminations, and irregular nodules which are black (N 1) to dark gray (N 3).

The calcareous quartz siltstones are commonly a distinctive yellowish brown mottled color. Colors range from light yellowish brown (10 YR 6/4) to very pale orange (10 YR 8/2) on fresh surfaces and very pale orange (10 YR 8/2) to dark yellowish orange (10 YR 6/6) on weathered surfaces. These siltstones and sandstones are commonly very well indurated with siliceous and/or calcareous cement.

The entire extent of the outcrop exposures are part of an allochthonous block. The nature of the contacts with stratigraphically overlying and underlying units was not determinable because of tectonic complications.

<u>Thin-section Description</u> In light of the lack of stratigraphic evidence to support the assignment of the limestones and siltstones of this unit to the Middle Canyon Formation, it seems appropriate to examine these lithologies in more detail. The calcareous siltstones are composed of angular silt-sized quartz grains and silt-sized bioclastic carbonate grains with little else in evidence. Minor amounts of siliceous sandy siltstone are also present within this unit. These rocks are siltstones with an admixture of well rounded, very fine quartz sand grains which display evidence of authigenic overgrowths. Most (95%) of the silt- and sand-sized grains have an unstrained extinction. The accessory minerals zircon and biotite comprise approximately one percent of the rock.

Much of the limestone appears to be unfossiliferous in hand sample. However, when these rocks are examined petrographically they are found to be fossiliferous packstones, grainstones, and wackestones (Dunham, 1962) which are commonly silty. Fossil fragments within these rocks are most commonly medium to fine sand-sized grains of crinoid plates and columnals and fenestral bryozoan fragments. Minor amounts of echinoderm and brachiopod spines, brachiopod valve fragments, and foraminifera are also present in several thin-sections. The limestones show evidence of pervasive replacement of many of the biotic components, neomorphic aggradation of much of the matrix material, one or two episodes of calcite vein filling, and stylolitization. One sample shows partial chert replacement of calcite within some of the biotic fragments.

Huh (1968) presents a detailed analysis of the Middle Canyon Formation, as well as the other formations within the White Knob Group. While the limestones described above could be assigned to the Scott Peak Formation on the basis of lithology, their intimate association with the calcareous siltstones can only be taken as indicative of a Middle Canyon affinity.

Big Snowy Formation
Upper Mississippian

The name Big Snowy was first applied by Scott (1935) to strata exposed in the Big Snowy Mountains of south-central Montana. At the type section Scott defined the Big Snowy Group from strata that had previously been considered part of either the Quadrant or Amsden Formations. Scott defined three formations within the group. They are in ascending order: the Kibby Formation, a red shaly limestone; the Otter Formation, a black shale and clayey siltstone; and the Heath Formation, a shaly and clayey limestone. The Kibby and Otter Formations had been previously defined by Weed (1892, 1899). Scott determined the age of the Big Snowy to be late Meramecian to late Chester.

The Big Snowy Group was redefined at the type section by Gardner (1959). The revised group included the Kibby, Otter, and Heath Formations as well as the Cameron Creek (Scott, 1935), Alaska Bench (Scott, 1935), and Devils Pocket Formations. Gardner believed the Big Snowy to be of Mississippian and Pennsylvanian age.

In the Tendoy Range, Big Snowy stratigraphy is poorly understood. This cratonic sequence seems to lose some of its distinctiveness as it approaches the transition zone. The strata between the Mission Canyon Formation and the Amsden Formation are probably Big Snowy or a lateral correlative. Poor exposures and loss of distinctive character make identification of the member formations of the Big Snowy Group difficult. For this reason the Big Snowy will be dealt with as a formation, rather than a group, in the thesis area.

In the Dixon Mountain-Little Water Canyon area the Big Snowy Formation disconformably overlies the Mission Canyon Formation and underlies the Amsden Formation unconformably.

Distribution and Topographic Expression

The Big Snowy Formation is poorly exposed in the thesis area. The shaly nature of the unit has led to the development of soil and debris covered gentle slopes with isolated ledges formed by limestone beds.

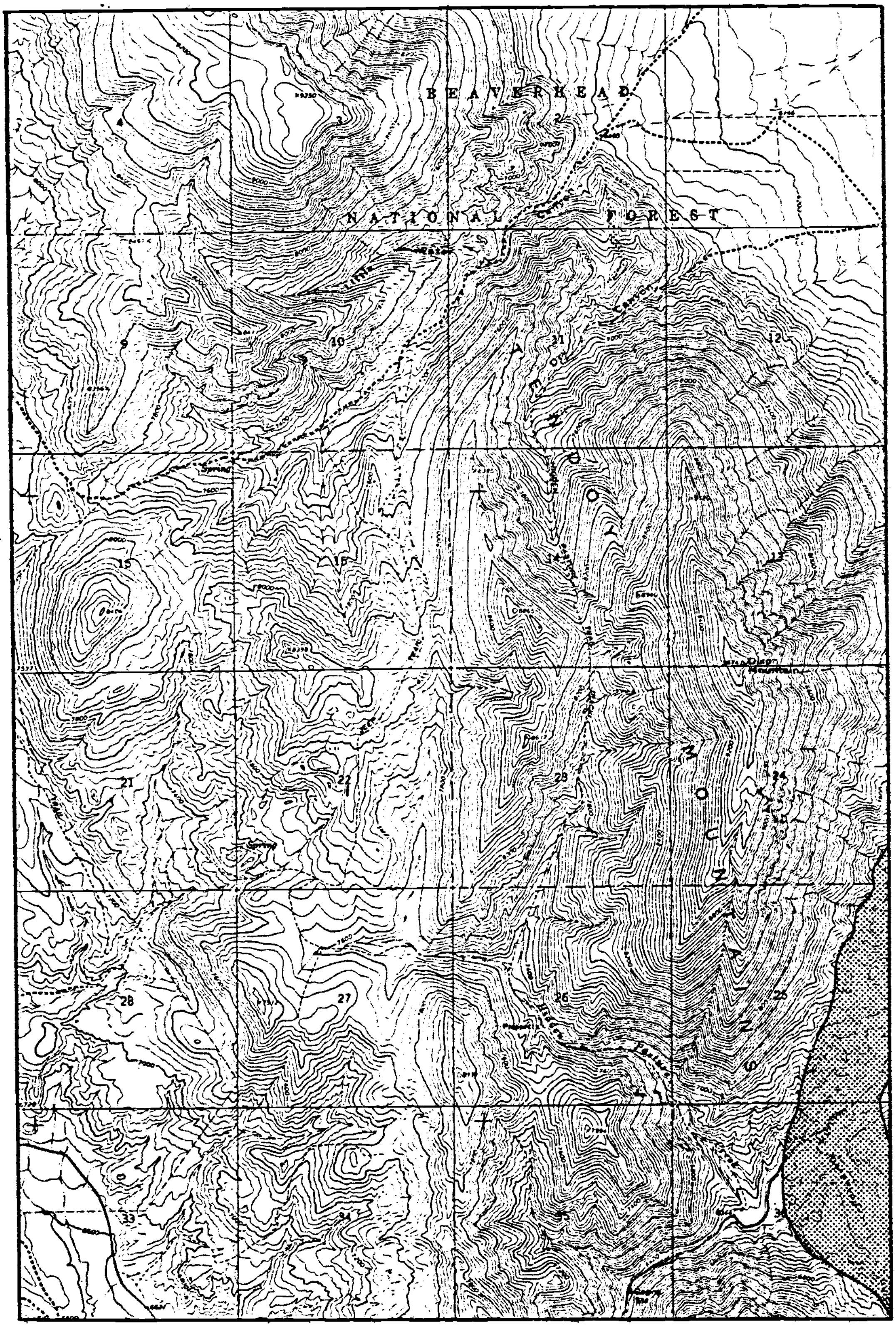

Figure 6. Distribution of Big Snowy Formation.

The Big Snowy occurs in the Big Sheep Creek area, in the eastern parts of sec. 25 and 36, T.13 S., R.10 W. Deadwood Gulch has developed as a strike valley within the Big Snowy. Proximity to a high-angle reverse fault, the Tendoy fault, has resulted in intensely deformed bedding within the Big Snowy. This deformation, primarily plastic in nature, is observed to wane in intensity as one moves away from the trace of the fault.

Some good exposures of the Big Snowy may be examined along the eastern side of Deadwood Gulch and along the road to the north of Big Sheep Creek.

Fossils and Age

Much of the limestone strata within the Big Snowy Formation was found to be fossiliferous. On the contrary, the shales and sandstones of the Big Snowy are, in general, unfossiliferous.

Within the thesis area many of the fossils are found in coquinoid beds; articulated and disarticulated brachiopod valves commonly are found in association with fragments of branching bryozoa and crinoidal debris. Brachiopods identified by the writer include a variety of strophomenids, *Dictyoclostus* and possible *Orthotetes*. The coquinoid nature of the fossiliferous strata suggests deposition in a shell bank setting under shallow, fairly high energy conditions.

Scott (1935) collected a variety of brachiopods from Big Snowy strata to the east of the thesis area and established the age as Late Mississippian, no younger than late Chesterian. Scholten et al., (1955) list a rather extensive fauna for the Big Snowy of southwestern Montana. A number of productids, spiriferids, and other families of brachiopods were listed. In addition, the pelecypod *Myalina*, the coral *Caninia*, and fenestral bryozoans occur.

Dutro and Sando (1963) determined the age of the Big Snowy Formation to be latest Meramecian to middle Chesterian. This age was based on the occurrence of the brachiopod *Straitifera brazerianus* and coral *Caninia* in the uppermost Big Snowy.

McKerrow (1978) suggests that *Caninia* is part of the upper reef slope community, a shallow water setting characterized by turbulent water conditions. Many of the faunal elements named by Scott (1935) and Scholten et al. (1955) are included in the reef-related communities described by McKerrow.

Lithology

Outcrop Character Little of the Big Snowy Formation is well exposed within the thesis area. The exact thickness of the formation in the Tendoy Range has never been determined, but it appears that several hundred feet are present (Scholten, et al., 1955).

The upper Big Snowy strata are composed of limestones and shales. The limestones are very thin- and thin-bedded fossiliferous packstones and grainstones. These rocks are brownish black (5 YR 2/1) to moderate brownish gray (5 YR 5/1) on fresh surfaces and weather medium dark gray (N 4), light brown (5 YR 7/4), and yellowish gray (5 Y 8/1). The shales of the upper Big Snowy are rhythmically interbedded with fine-grained limestones. The shales are very thin-bedded, partly fossiliferous and calcareous.
These rocks are black (N 1) when fresh and grayish black (N 2) when weathered. The interbedded limestones are fossiliferous silty micrites which are dark to medium gray (N 3 to N 5) on fresh surfaces and weather to light olive gray (5 Y 6/1) and yellowish gray (5 Y 8/1).

The lower part of the Big Snowy is almost entirely limestone. Fossiliferous grainstones (Dunham, 1962) within this part of the formation contain a diverse assemblage of crinoid columnals, echinoderm plates, bryozoa fragments, brachiopod valves, and locally, solitary corals. The grainstones are light to medium dark gray (N 7 to N 4) on fresh surfaces and weather to yellowish gray (5 Y 7/2). Much of the remainder of the lower Big Snowy is fossiliferous silty packstones and wackestones (Dunham, 1962). These rocks have the same type of fossil debris as the grainstones but generally are more fragmented and lack coral remains. They range

Figure 7. Rhythmically bedded dark micritic limestone interbedded with carbonaceous calcareous silty shale in the upper Big Snowy Formation. Exposed on the north side of Big Sheep Creek.

in fresh surface color from medium dark gray (N 4) to pale brown (5 YR 5/2) and brownish black (5 YR 2/1). The weathered surfaces of these packstones and wackestones range in color from gray (N 8) to yellowish gray (5 Y 7/2) and light brown (5 YR 7/4).

The upper contact between Big Snowy and Amsden strata is marked by a limestone pebble conglomerate within the lowermost Amsden which has a calcareous siltstone matrix. This conglomerate is overlain by a moderate reddish orange (10 R 6/6) soil covered slope. This reddish soil zone coincides with a slight break in slope to the more gently dipping overlying Amsden strata.

The lower contact with the underlying Mission Canyon Formation is recognized in the region as an erosional unconformity which is characterized as a reddish paleosol developed on a paleokarst surface (Scholten, et al., 1955). No such horizon was observed within the thesis area. This contact probably exists nearby, but outside the study area boundaries.

Amsden Formation
Mississippian-Pennsylvanian

The Amsden Formation, as presently defined, straddles the Mississippian-Pennsylvanian systemic boundary. This position has resulted in numerous attempts to rectify the time-stratigraphic situation by redefining the lithostratigraphic terminology.

The name Amsden was first applied by Darton (1904) to exposures of red shale, limestone, and cherty and sandy limestone along the Amsden Branch of the Tongue River, west of Dayton, Wyoming. The rocks described at this type section were considered to be correlative to strata Peale (1893) had assigned to the base of the Quadrant Formation, near Three Forks, Montana.

Branson and Gregor (1918) found that the Amsden Formation exposures in the Wind River Mountains of Wyoming were of Mississippian age rather than the Pennsylvanian age described elsewhere. This led Branson and Gregor to assign all the strata containing Mississippian fossils to the Sacajawea Formation.

Scott (1935) defined the Amsden Formation to be 109 feet (33 m.) of strata occurring between the base of the Quadrant Formation and the top of the Madison Group.

Branson (1939) and Branson and Branson (1941) suggested that the part of the Amsden Formation which is of Pennsylvanian age should be included in the Tensleep Formation (a correlative of the Quadrant Formation). It also was suggested that the lower, Mississippian age part of the Amsden be assigned to the Sacajawea Formation. Finally, the name Amsden was dropped from use. Thompson and Scott (1941) concurred in this reassignment and deletion proposal.

The stratigraphic difficulties remained however.

Berry (1943) suggested the reestablishment of the name Amsden for the strata between the Quadrant Formation and the Madison Group. It was Berry's contention that the lower part of the formation was a lateral equivalent of the Sacajawea Formation of Wyoming. This was followed by Perry and Sloss (1943) who noted that strata of latest Mississippian (Chester) age are present in the Williston Basin between the redefined Tensleep and Sacajawea formations of Branson (1939) and Thompson and Scott (1941). Perry and Sloss proposed that the name Amsden be reinstated for the Chester age strata.

Sloss (Gardner and others, 1946) concluded that the Mississippian-Pennsylvanian boundary was within the Amsden Formation. Sloss believed that the Amsden of northern Wyoming was younger than the Amsden strata of central Montana. Wilson (1962) followed this same line of thought by suggesting a strata-transgressive Mississippian-Pennsylvanian boundry within the Amsden-Quadrant sequence.

The continued use of the name Amsden was urged by Maugham and Roberts (1967). They suggested that the Amsden of central and eastern Montana be raised to a group status.

In the Dixon Mountain-Little Water Canyon area the Amsden Formation is about 200 feet (61 m.) thick (Scholten and others, 1955). The Amsden is conformably overlain by the Quadrant Formation. The presence of a thin limestone conglomerate at the base of the Amsden suggests a disconformable contact with the underlying Big Snowy Formation.

Distribution and Topographic Expression

Resistant limestone and sandstone beds within the Amsden Formation have weathered to moderately steep slopes. Outcrop quality is, however, poor because of numerous rock glaciers and talus cover derived from the overlying Quadrant quartzites.

A narrow band of Amsden strata occurs low on the eastern flank of Dixon Mountain, SE ¼, sec. 24, and southwestward through the center of sec. 25 and 36, T.13 S., R.10 W. Big Sheep Creek traverses Amsden strata obliquely in the vicinity of its confluence with Hidden Pasture Creek.

Fossils and Age

A number of beds within the Amsden Formation were found to be fossiliferous. However, most of the fossiliferous strata are composed of fragmented fossil debris which is, for the most part, unidentifiable beyond phyllum. Scott (1935) collected and described a variety of productids, spiriferids, rhynchonellids, and Composita from Amsden strata. Scott assigned a middle or late Chester age (latest Mississippian) to these strata based on the fossil collection.

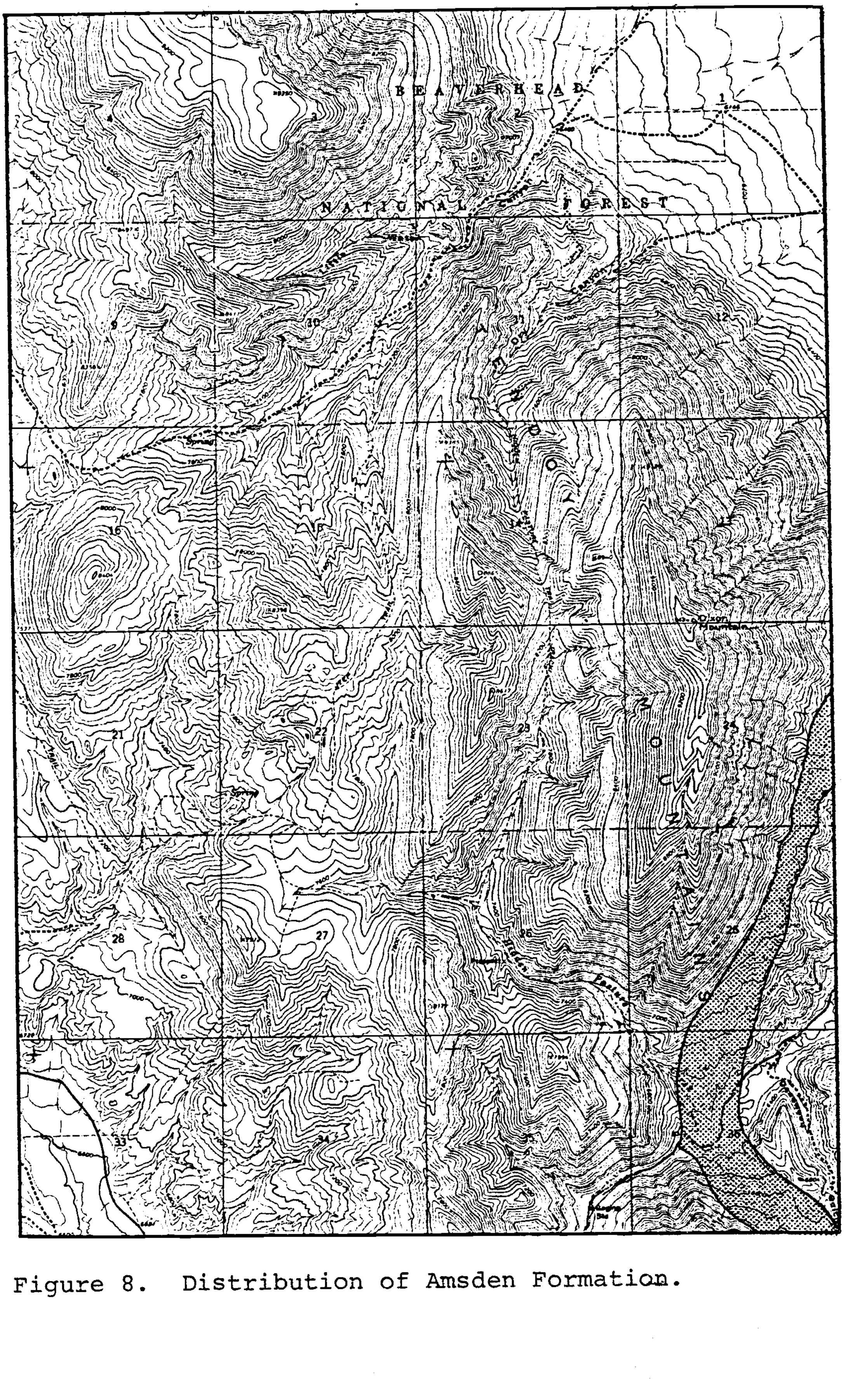

Figure 8. Distribution of Amsden Formation.

Subsequently, Scott (1945) found fusulinids of Morrowan age (earliest Pennsylvanian) within the uppermost Amsden. The straddling of the Mississippian-Pennsylvanian boundary by Amsden strata was to be the subject of much investigation and discussion for many years following Scott's work.

Additional studies by Gardner, et al. (1946) also described both Mississippian and Pennsylvanian fauna. Mundt (1956) found the Amsden strata he examined to range in age from Chesterian to Atokan.

An extensive examination of the Amsden by a number of investigators culminated in the publication of a U.S. Geological Survey Professional Paper (Sando, Dutro and Gordon, 1975) which catalogs the fauna of the formation and outlines a depositional history. This report identified and describes 210 taxa of mega fauna and 140 taxa of microfauna. This paleontological evidence indicates that the Amsden ranges in age from Late Mississippian to Middle Pennsylvanian.

Within the thesis area samples containing crinoidal debris and productids were collected by the author.

Lithology

Outcrop Character The Amsden Formation within the thesis area is, in general, poorly exposed. Slopes formed on Amsden strata are gentle, commonly soil covered, and generally obscured by talus streams of Quadrant sandstone. There is estimated to be approximately 200 feet (61 m.) of

Amsden exposed along Big Sheep Creek (Scholten, et al., 1955).

An upper limestone unit, a middle sandstone unit and lower limestone and red shale unit may be discerned from the available outcrops.

The upper limestone unit is composed of a thin-bedded petroliferous brachiopod-crinoidal sparry limestone which is light olive gray (5 Y 6/1) and weathers grayish orange (10 YR 7/4). The brachiopods are usually whole but disarticulated and locally are fragmented. Associated with this fossiliferous limestone is a thin-bedded limestone which is medium grey (N 5) and weathers to light olive gray (5 Y 6/1). This limestone is commonly interbedded with chert stringers which are medium dark gray (N 4) and weathers to grayish orange (10 YR 7/4).

The middle part of the Amsden is predominantly composed of thick- and thin-bedded, friable medium-grained sandstone which is light brownish gray (5 YR 6/1) with light brown (5 YR 5/6) mottling on a fresh surface; it weathers light brown (5 YR 6/4). Some of the sandstone beds are in part composed of chert pebbles up to two inches (5 cm.) in diameter and commonly grade upward from a pebbly sandstone to a coarse grained sandstone over approximately three feet (1 m.).

That part of the lower Amsden which is exposed is composed of very thin- and thin-bedded fossiliferous limestone which is light gray (N 7) and weathers to nearly the

Figure 9. Limestone conglomerate immediately up-section from the contact between the Big Snowy Formation and the overlying Amsden Formation. Exposed 0.5 miles north of Big Sheep Creek.

same color or light olive gray (5 Y 6/1). The fossils are disarticulated whole valves of a spiriferid brachiopod. The basal Amsden is marked by a soil-covered slope-forming zone of pale reddish brown (10 R 5/4) soil which appears to be laterally persistant.

The contact with the underlying Big Snowy Formation seldom can be distinguished in the field. Maughan and Roberts (1967) have noted that a limestone conglomerate marks the base of the Amsden in the southwest Montana region. In one poorly exposed outcrop, such a limestone conglomerate was observed. The conglomerate is composed of well rounded limestone pebbles and cobbles (1.5-4 in.; 4-10 cm.) with a calcareous siltstone matrix. The pebbles appear to be of a silty limestone which weathers to olive gray (5 Y 4/1).

The uppermost Big Snowy strata observed below the limestone conglomerate unit are very thin-bedded, fissile, petroliferous silty limestones which are dark gray (N 3) and weather to a light olive gray (5 Y 6/1). This limestone is interbedded with fossiliferous calcareous black shales (N 1) which weather to a grayish black (N 2).

Quadrant Formation
Pennsylvanian

The name Quadrant was first used by Peale (1893) for those rocks exposed in the Three Forks Quadrangle between the Mississippian Madison Limestone and the Jurassic Ellis

Group. Peale did not designate a type section, but the name is known to have been derived from Quadrant Mountain in the Gallatin Range of northwestern Yellowstone Park.

Strata exposed in bluffs encircling Quadrant Mountain were designated as the Quadrant Formation type section by Weed (1896). Weed measured and described 400 feet (122 m.) of light colored quartzite and interbedded dark saccharoidal limestone.

The type section was reexamined by Scott (1935). The lowermost 109 feet (33 m.), mainly limy shale, was assigned to the Amsden Formation. Scott suggested that the name Quadrant be dropped from use and that the remaining units be reassigned to the Big Snowy Group. Branson (1939) believed that the reassigned units of Scott (1935) should be considered as the attenuated edge of the Tensleep Sandstone.

Newly discovered fusulinids aided Thompson and Scott (1941) in their reappraisal of the section described by Scott (1935). The lowest beds were found to be of Mississippian age and were assigned to the Sacajawea Formation, a name later restricted to northern Wyoming (Scott and Wilson, (1953). The remaining beds (279 ft.; 85 m.) all of which were believed to be of Pennsylvanian age, were included in the Quadrant Formation. Thompson and Scott excluded the name Amsden from the section.

Sloss and Moritz (1951) measured 2,662 feet (812 m.)

of Quadrant sandstone along Big Sheep Creek, within the thesis area. The marked thickening of the section westward from the type section was taken as an indication of a post-Pennsylvanian erosion of a positive area to the east of Big Sheep Creek.

In the Dixon Mountain-Little Water Canyon area the Quadrant Formation is conformably overlain by the Permian Phosphoria Formation. The Mississippian-Pennsylvanian Amsden Formation underlies the Quadrant across a conformable contact.

Distribution and Topographic Expression

The highly resistant quartzose sandstones of the Quadrant Formation form the two highest topographic features in the area, Dixon Mountain and Timber Butte. The position and attitude of the Quadrant strata appear to control much of the surrounding drainage patterns and landforms.

In the vicinity of Dixon Mountain the antidip slope of Quadrant sandstone rises as a steep talus covered slope to the maximum elevation in the thesis area. The Quadrant dip slope of Dixon Mountain is in part covered by the flatirons of the overlying Phosphoria Formation. Big Sheep Creek passes through the Quadrant in a steep-walled canyon, SE ¼, sec. 35, and SW ¼, sec. 36, T.13 S., R.10 W.

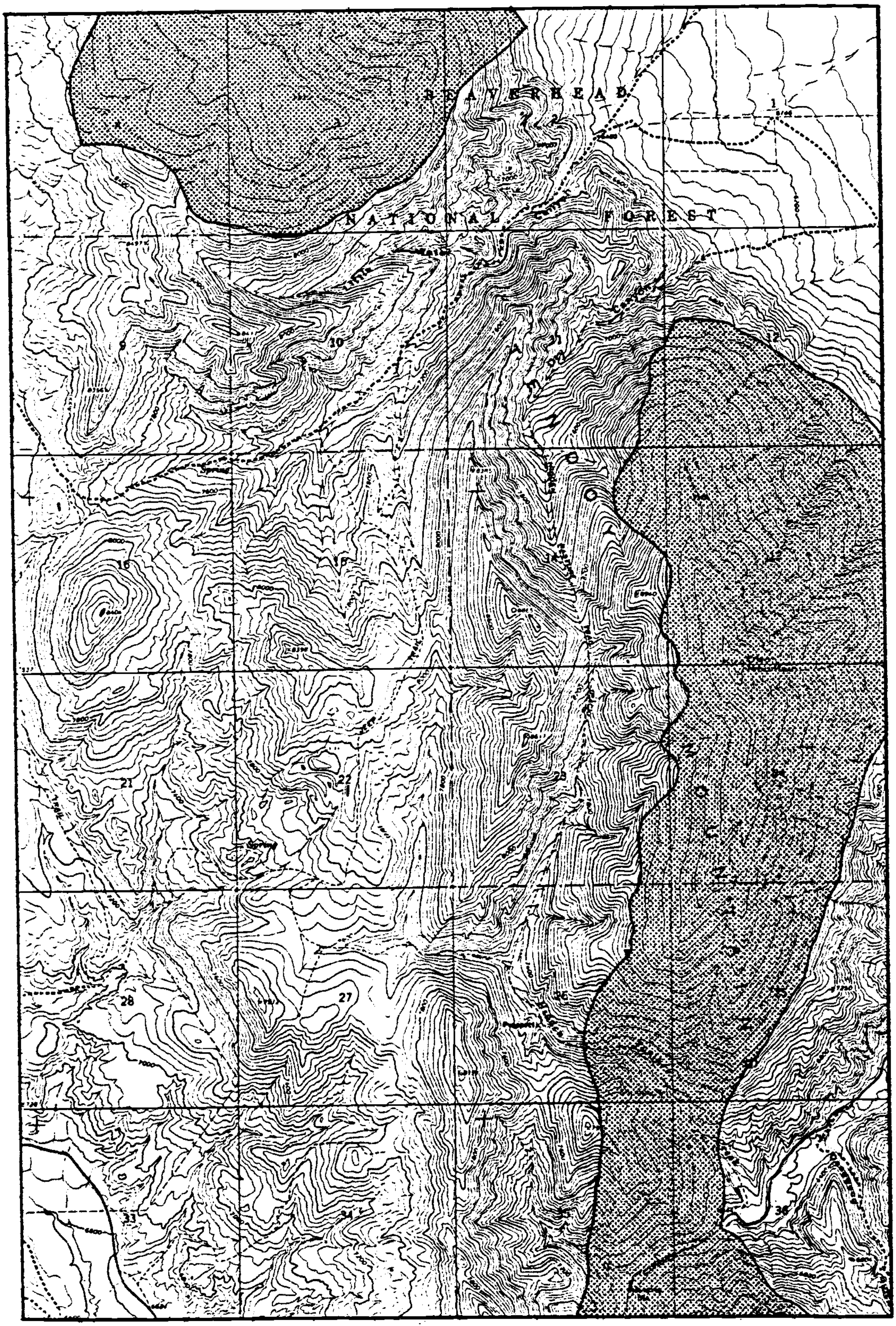

Figure 10. Distribution of Quadrant Formation.

In the vicinity of Timber Butte, blocky talus covered slopes rise steeply to form the second highest feature (9,472 ft.; 2,888 m.) in the thesis area.

Fossils and Age

No fossils were collected from the Quadrant Formation. The formation is characteristically unfossiliferous. Thompson and Scott (1941) found *Wedekindellina* sp. and *Fusulina* sp. at the type section of the Quadrant. These fusulinids indicate a middle Des Moinesian age (Middle Pennsylvanian) for the upper Quadrant. Siliceous sponge spicules from the Quadrant indicate a similar Middle Pennsylvanian Age (Scott, 1935).

Litholgy

Outcrop Character The highly resistant calcareous and siliceous arenites of the Quadrant Formation compose the high topographic features of the thesis area. This unit is 2,662 feet (812 m.) thick along Big Sheep Creek (Sloss & Moritz, 1951) and is most commonly observed as a few resistant ledges and cliffs surrounded by large talus streams and fields of subangular cobbles and boulders, some of which are larger than an automobile.

The Amsden-Quadrant contact is marked by a gradational change over approximately 20 inches (50 cm.) from limestones to interbedded limestone and sandstone to sandstone. The uppermost unit of the Amsden Formation is a petroliferous and fossiliferous limestone which is medium gray (N 5) on a fresh surface and weathers to a yellowish gray (5 Y 7/2). Many of the bedding surfaces display parting lineations. The lowest rocks assigned to the Quadrant Formation are medium-bedded, fine-grained, calcareous sandstones that are pinkish gray (5 YR 8/1) on a fresh break and weather to medium gray (N 5). These beds show shallow trough cross-bedding in sets (4-8 in.; 10-20 cm.) thick.

A great deal of the Quadrant is composed of well sorted, subangular, fine-grained siliceous sandstones that are dark yellowish brown (10 YR 6/6) with moderate reddish orange (10 R 6/6) bands defining cross-bedding on a fresh surface. These sandstones weather to grayish orange (10 YR 7/4). Many sandstones display well developed trough cross-beds, some of which are locally associated with evidence of penecontemporaneous slumping and bedding contortion. Near faults the sandstones are extremely well indurated (sedimentary quartzites?) with slickensided fractures having glassy surfaces. One zone within the Quadrant contains clasts of chert and limestone, one to four inches (3-10 cm.) in diameter, in a thin-bedded, fine-grained sandstone which displays low-angle planar cross-bedding.

The contact between the Quadrant and the overlying

Phosphoria Formation is gradational over 15 feet (5 m.) from the quartz sandstone to interbedded dolomites, sandstones and limestones. The uppermost strata assigned to the Quadrant Formation are thin- to thick-bedded, fine- and medium grained, calcareous and siliceous sandstones; these are very pale orange (10 YR 8/2), weather to yellowish gray (5 Y 8/1), and are commonly cross-bedded. The presence of the Phosphoria is marked by the first appearance of irregularly thin-bedded (2-30 cm.) dolomite which is very light gray when fresh and weathers to a pinkish gray (5 YR 8/1). Spheroid and ovoid chert nodules (2-20 cm. in diameter) are observed within many of the dolomite beds.

Detailed examination of the Quadrant Formation within the thesis area is impossible as little of the formation is seen in outcrop. A few ridges are exposed rather continuously but the presence of a great deal of talus makes bed by bed examination hopeless.

Phosphoria Formation
Permian

The name Phosphoria was introduced by Richards and Mansfield (1912) for rocks of Permian age exposed in Phosphoria Gulch, near Meade Park, Idaho. At the type section, 420 feet (128 m.) strata were measured and described. Two members, a lower phosphatic shale unit (the Meade Peak of McKelvey and others, 1956), and an upper bedded chert, the Rex chert, were defined.

The name Phosphoria was extended into Montana by Bonine (1914). The name was applied to rocks which had been previously assigned to either the upper Quadrant or the Teton Formation, a unit previously considered to be of Triassic age.

A reappraisal of the Permian stratigraphic nomenclature of the western United States was undertaken by McKelvey and others (1956). These workers more clearly defined the intertonguing relationship which existed between the Phosphoria Formation and the Park City Formation of Utah. The Shedhorn Sandstone Member was named for Permian quartzose sandstone exposed in the Madison Range. Three members were defined in the type section: the lowermost Meade Peak phosphatic shale, the Rex Chert, and an unnamed siliceous shale. Three additional member names were introduced for strata exposed outside the type locality: a lowermost chert member, believed correlative to the Meade Peak; the Retort phosphatic shale, described at Retort Mountain, south of Dillon, Montana; and the Tosi Chert, a correlative of the uppermost Retort and the siliceous shale member. McKelvey and others recognized the westward thickening of the formation to approximately 1,300 feet (397 m.) in south-central Idaho.

In the Dixon Mountain-Little Water Canyon area the Phosphoria Formation is believed to overlie the Quadrant

Formation conformably. The lack of adequate exposures makes positive field determination of the contact difficult. The Phosphoria Formation is unconformably overlain by the Triassic Dinwoody Formation (Moritz, 1951).

Distribution and Topographic Expression

The most extensive and best exposed outcrops of the Phosphoria Formation occur on the eastern side of the Hidden Pasture-Dry Creek drainage system (Plate 1). At these exposures the resistant chert and sandstone beds have been eroded into flatiron ridges on the dip slope of the Quadrant. The Hidden Pasture and Dry Creek drainages appear to have had their position controlled by the differential weathering of the Phosphoria and the overlying Dinwoody. Much of the course of these two streams is at or near the Phosphoria-Dinwoody contact.

Elsewhere in the thesis area, the Phosphoria is much less prominent. A part of the upper Phosphoria is present on the flat-topped interfluvial ridge between Dry Creek and Little Water Creek, SE ¼, sec. 2, and NE ¼, sec. 11, T.13 S., R.10 W. The Phosphoria is also present, but poorly exposed, low on the southern and southeastern flanks of Timber Butte. These outcrops are generally covered by talus from the Quadrant strata exposed upslope.

No Page 42

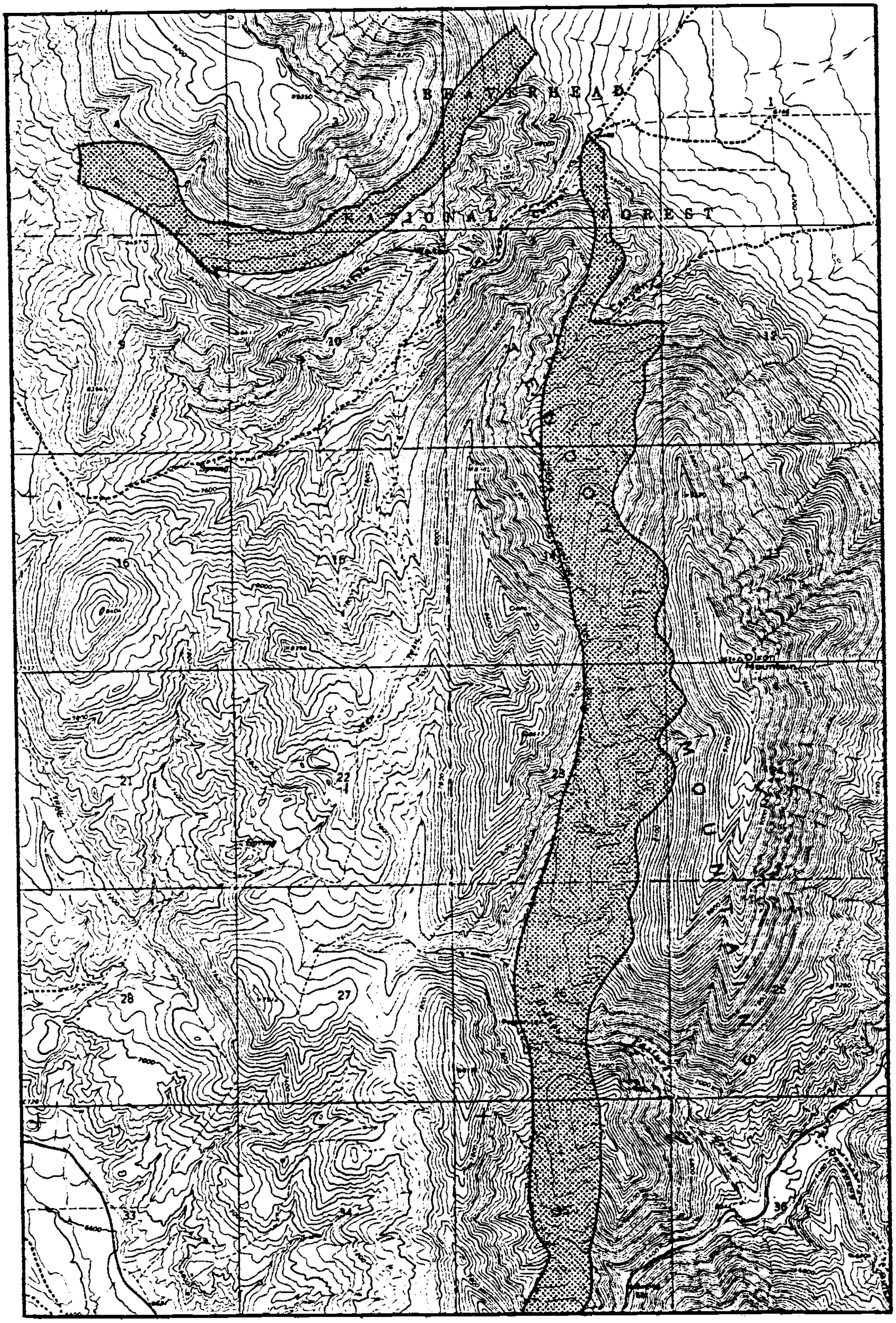

Figure 11. Distribution of Phosphoria Formation.

Fossils and Age

Many of the limestone beds and some of the shaly strata within the Phosphoria Formation are fossiliferous. Much of the fossil material is well preserved and readily identifiable as to genus. King (1930) examined the Phosphoria and collected a variety of specimens from the various lithologies present.

King found that the limestone and chert members of the Phosphoria are characterized by a different faunal assemblage than the phosphatic shale members. The brachiopods *Productus*, *Camarophoris*, *Composita*, and spiriferids characterize the limestone and chert facies (King, 1930). The occurrence of productids and spiriferids suggests a shallow normal marine environment. However, the presence of *Composita* suggests conditions were somewhat restricted (McKerrow, 1978).

The phosphatic shales are characterized by a variety of productids, rhynchonellids, *Orbiculoides*, and *Lingula* (King, 1930). This phosphatic facies is believed to have accumulated slowly under reducing conditions in quiet waters (Cressman and Swanson, 1964). However, the fauna associated with this facies suggests shallow, normal marine waters. The implications of this apparent contradiction will be investigated in the discussion of environments of deposition.

Within the thesis area, various productids and spiriferids were collected as well as bryozoans, crinoid columnals, Orbiculoides, and Composita.

The age of the Phosphoria Formation has been determined, based on macrofossils and microfossils. Nautiloid and ammonite genera collected from the upper phosphatic shale member of the Sublette Range of western Wyoming indicate an early Guadalupian (medial Permian) age (Miller and Cline, 1934). This age was also determined for the phosphatic shales of the Park City and Phosphoria or northern Utah (Clark, et al., 1977). Fusulinids collected from the base of the Phosphoria, near Three Forks, Montana, indicates a Wolfcampian (Early Permian) age (Frenzel and Mundorft, 1942). The age of the uppermost Phosphoria strata in southwestern Montana has not been determined and there is some question as to whether Phosphoria sedimentation continued to the end of the Permian in this region (Newell & Kummel, 1942, Moritz, 1951, Sloss and Moritz, 1951).

Lithology

Outcrop Character The Phosphoria Formation has been the subject of a number of studies. Economically significant deposits of phosphate within the formation have provided the impetus for several investigations of regional scale (Cressman, 1955; McKelvey, et al., 1956; and Cressman and Swanson, 1964). In particular, the Professional Paper prepared by Cressman and Swanson (1964) contains a detailed study of the distribution of the various Phosphoria and Park City members, petrographic and chemical analyses of the strata of the various members, and interpretations of the environments of deposition believed responsible for the various lithologies present. Included in the many detailed measured sections reported within the Cressman and Swanson paper are two composite sections constructed in part from data gathered within the thesis area. I refer those readers interested in greater detail concerning Phosphoria strata to this professional paper.

Within the thesis area 833 feet (254 m.) of Phosphoria and Park City strata have been reported (Cressman and Swanson, 1964). This thickness has been divided into seven members as follows (in descending order):

Tosi Chert Member (Phosphoria Fm.) - 92 ft. (28 m.)

Cherty Shale Member (Phosphoria Fm.) - 57 ft. (17 m.)

Retort Phosphatic Shale Member (Phosphoria Fm.)
83 ft. (25 m.)

Franson Tongue (Park City Fm.) - 156 ft. (48 m.)

Rex Chert Member (Phosphoria Fm.) - 69 ft. (21 m.)

Meade Peak Phosphate Shale Member (Phosphoria Fm.)
29 ft. (9 m.)

Grandeau Tongue (Park City Fm.) - 347 ft. (106 m.)

The Park City Formation (Boutwell, 1907) is composed of two carbonate members (Franson and Grandeur) which are

500 feet (153 m.) in total thickness at the type locality. At the Park City District type section, in Utah, the Franson and Grandeur members are separated by a tongue of the Meade Peak Phosphatic Shale Member of the Phosphoria (McKelvey, et al., 1956) Tongues of the Park City members extend northward over much of the Western Phosphate Field (Honkala, 1967). To the south Permian strata becomes entirely that of the Park City Formation.

The Tosi Chert Member within the thesis area is rarely well exposed. Where encountered, the Tosi Chert consists of interbedded argillaceous dolomites or limestones and bedded cherts which are thin- to thick-bedded and contain minor carbonaceous mudstone intercalations.

Distinction between the cherty shale member and the underlying Retort Phosphatic Shale Member was never made in the field. These units are most commonly expressed as gentle slopes which are yellowish brown to olive gray in color. The most commonly exposed strata of the Retort member are light olive gray (5 Y 6/1), unevenly thin-bedded, burrowed, calcareous mudstones which weather to yellowish gray (5 Y 7/2) and medium gray (N 5).

The Franson Tongue of the Park City Formation is characteristically a series of thin- to thick-bedded cherty, phosphatic, or glauconitic calcareous sandstones within minor interbedded dolomites and limestones.

Figure 12. Irregular thin-bedded chert of the Rex Chert Member of the Phosphoria Formation. Exposed on the east side of the divide between Dry Canyon and Hidden Pasture Creek.

Typically these beds are brownish gray (5 YR 4/1). Within the thesis area the Franson strata most commonly exposed are medium- to very coarse-grained chert-bearing sandstones and pebbly sandstones.

The thin- to thick-bedded cherts of the Rex Chert Member are easily located within the thesis area and commonly from ledges and low cliffs. These light gray to grayish black (N 7-N 2) strata contain minor amounts of less resistant cherty and phosphatic mudstones.

The thin section (29 ft.; 9 m.) of argillaceous dolomites, phosphorites and dolomitic mudstones which comprise the Meade Peak Phosphatic Shale Member were not observed in outcrop within the thesis area. These olive gray to dark gray strata would be expected to be nonresistant slope-formers.

The Grandeur Tongue of the Park City Formation is composed of dolomite and limestone with chert nodules, dolomiic mudstones and several chert beds up to 10 feet (3 m.) in thickness. Two units noted by Cressman and Swanson (1964) as diagnostic of the Grandeur Tongue were located within the thesis area. An irregular bed (5-25 in.; 13-64 cm. thick) of angular chert fragments was found within a sequence of bedded chert. Downsection from this chert breccia, near the lower contact of the Phosphoria, is a series of thin- and very thin-bedded, poorly indurated,

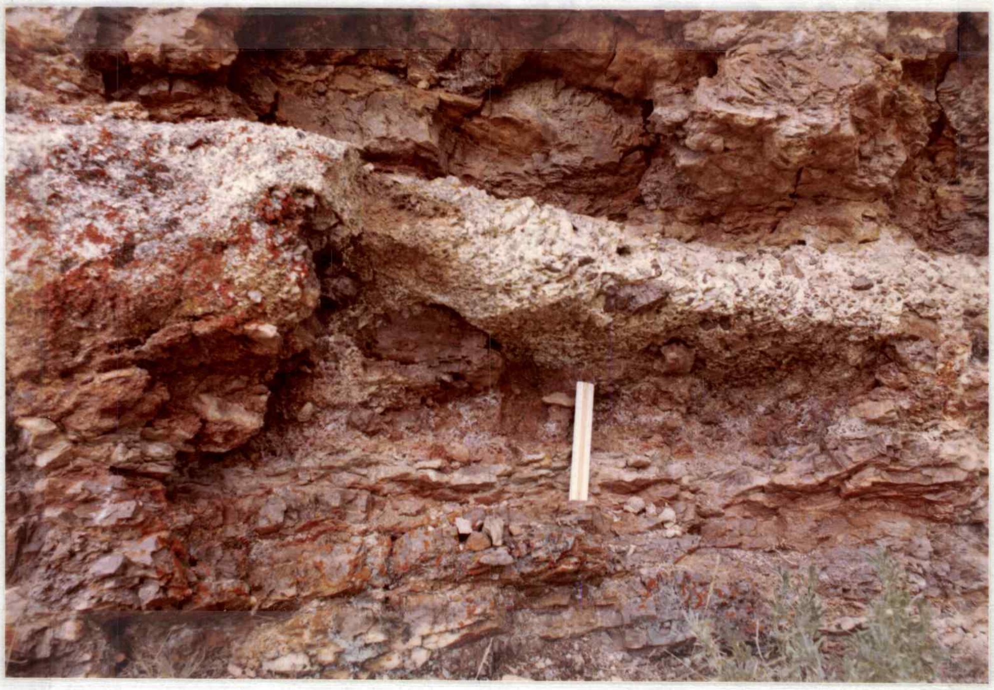

Figure 13. Chert breccia developed between bedded cherts of the Grandeur Tongue of the Park City Formation. Exposed S.E. of the divide between Dry Canyon and Hidden Pasture Creek.

very fine-grained quartz sandstones which are moderate red (5 Y 5/4) on the fresh and weathered surfaces. This unit is unique to the Grandeur Tongue and delineates the lower part of that member.

The contact between the Phosphoria and the underlying Quadrant Formation is gradational. The first appearance of irregularly thin-bedded, very light gray dolomite above yellowish gray fine- and medium-grained sandstones indicates the location of the lowest Phosphoria strata.

The contact between the Phosphoria and the overlying lower shale member of the Dinwoody Formation was not observed within the field area. The uppermost Phosphoria strata commonly observed in outcrop are yellowish olive gray (5 Y 7/1), glauconitic, calcareous, very fine-grained sandstones which weather light olive gray (5 Y 6/1), and pale yellowish brown (10 YR 6/2), commonly fissile, phosphatic siltstones or shale which weather grayish orange (10 YR 7/4). The argillaceous cherts of the Tosi Chert Member are rarely observed above these strata.

Regional studies by Newell and Kummel (1942) and Moritz (1951) suggest that a very slight angular unconformity, representing late Permian time, may exist between the Dinwoody and Phosphoria. Studies of the Phosphoria by Cressman and Swanson (1964) indicate a disconformity between the Permian and Triassic strata of Idaho, but they suggest deposition may have continued in southwestern Montana Regionally, the onlapping nature of Dinwoody strata upon the Phosphoria suggests the presence of a disconformity or unconformity on the basin margins. The thinness of strata between units positively dated as Middle Permian and the overlying Triassic rocks of the Dinwoody suggests to Wardlaw (personal commun.) that if deposition did continue into the Late Permian it was of minimal thickness. It may be that the Phosphoria basin was filled to near base level during the Middle Permian. If sedimentation did continue into the Late Permian it was of minor volume because of the establishment of a profile of equilibrium. Sediment bypassing or subsequent erosion would then leave little of this strata remaining as Dinwoody deposition commenced. It seems that the Dinwoody-Phosphoria contact within southwestern Montana would then be a disconformity.

The Dinwoody-Phosphoria contact was rarely seen in outcrop in the thesis area. Where observed the bedding was parallel with no evidence of the development of an erosional surface. This would suggest a conformable contact. However, if the lack of uppermost Permian fossils are considered, a paraconformity is suggested.

Depositional Environment

The great variety of lithologies present within the Phosphoria reflect a number of lithotopes. Environments ranging from sublittoral settings under the influence of wave activity to deep marine with very little circulation must have been present within the Phosphoria basin. Sediments ranging from coarse clastics to carbonates, phosphorites, and siliceous oozes accumulated.

Much of the sedimentary strata of the Phosphoria can be readily interpreted as to depositional environment. However, the phosphatic material and the abundant chert are more troublesome. Cressman and Swanson (1964) suggest two possible origins for the phosphatic mudstones: chemogenic and biogenic. The phosphatic material may have precipitated under anaerobic conditions as cold, deep marine waters, rich in dissolved phosphorus, upwelled onto the outer continental shelf. Kazakov (1937) believes this situation could occur below the photic zone where planktonic organisms would have no opportunity to use the phosphorus. It is equally possible that the phosphorus was extracted from the water by planktonic organisms. This material was then redissolved upon death of the organisms and precipitated from the water column inorganically. In either case deposition would have taken place in quiet water with slow sedimentation rates, below wave base.

The phosphorites of the Phosphoria probably formed under a low paleolatitudinal setting with paleogeographic and paleobathymetric conditions similar to the Miocene phosphorites of California (Monterey Shale) and Peru and the modern deposits off the coast of Baja California (Cook and McElhinny, 1979). Phosphorite metasomatism may have also contributed to the development of phosphatic strata but definitive evidence of replaced calcareous fossils is lacking.

An alternative hypothesis for phosphorite accumulation has been provided by Bushinski (1964). Phosphates may arrive at a basin through fluvial processes. Once delivered, the dissolved phosphorus would then be assimilated by plankton and ultimately deposited upon the bottom of nearshore waters, with little reaching the deeper parts of the basin. Bushinski supports this hypothesis by citing the intimate association of phosphatic deposits and clastic material, fragments of shallow water fauna, evidence of agitation, and the presence of cross-bedding Within the Phosphoria Formation all of these criteria are recognized with the exception of cross-bedding. The presence of glauconite within the Phosphoria strata also indicates a gently agitated, continental shelf setting. The absence of phosphorites in most fluvial-marine basins does, however, suggest that the depositional conditions necessary for phosphorite accumulation are more complex than simply supplying dissolved phosphorus to a shallow basin.

The many bedded chert sequences within the Phosphoria are attributable organic activity under conditions of gentle circulation and low terrigenous influx (Cressman and Swanson, 1964). One half to two thirds of the silica has been derived from the siliceous sponge spicules of Demospongia. This sponge could inhabit a variety of substrates but did not thrive below the photic layer. The remainder of the silica present in the cherts of the Phosphoria were probably derived from the siliceous skeletal remains of radiolaria and other silicoflagellates. Direct precipitation of amorphous silica does not appear to have been a major factor, for many of the sponge spicules display corrosion of axial channels indicating that surrounding waters were undersaturated with respect to silica (Cressman and Swanson, 1964).

Dinwoody Formation
Early Triassic

The name Dinwoody was first used by Condit (1916), but it was Blackwelder (1918) who formalized the name.

Blackwelder designated exposures along Dinwoody Canyon in the Wind River Mountains, near Dubois, Wyoming, as the type section. At that location, Blackwelder described 250 feet (76 m.) of interbedded platy calcareous sandstone and greenish-gray shale.

Color change criteria used by Blackwelder for recognition of the upper Dinwoody Formation contact were judged inadequate upon reexamination by Newell and Kummel (1942). It was suggested that the formation be restricted, at the type locality, to the limits of resistant siltstone exposures. Newell and Kummel recognized three subdivisions of the Dinwoody Formation: a basal siltstone; a Lingula zone of thin limestone and dark shale; and a Claria zone of calcareous siltstone and interbedded limestone.

In the Dixon Mountain-Little Water Canyon area the Dinwoody Formation overlies the Phosphoria Formation unconformably (Moritz, 1951). The Dinwoody Formation is conformably overlain by the Woodside Formation.

Distribution and Topographic Expression

The Dinwoody Formation is generally a nonresistant slope-former. However, its position downslope from the resistant beds of the Thaynes Formation has resulted in some good exposures.

The best exposures of the Dinwoody occur at the base

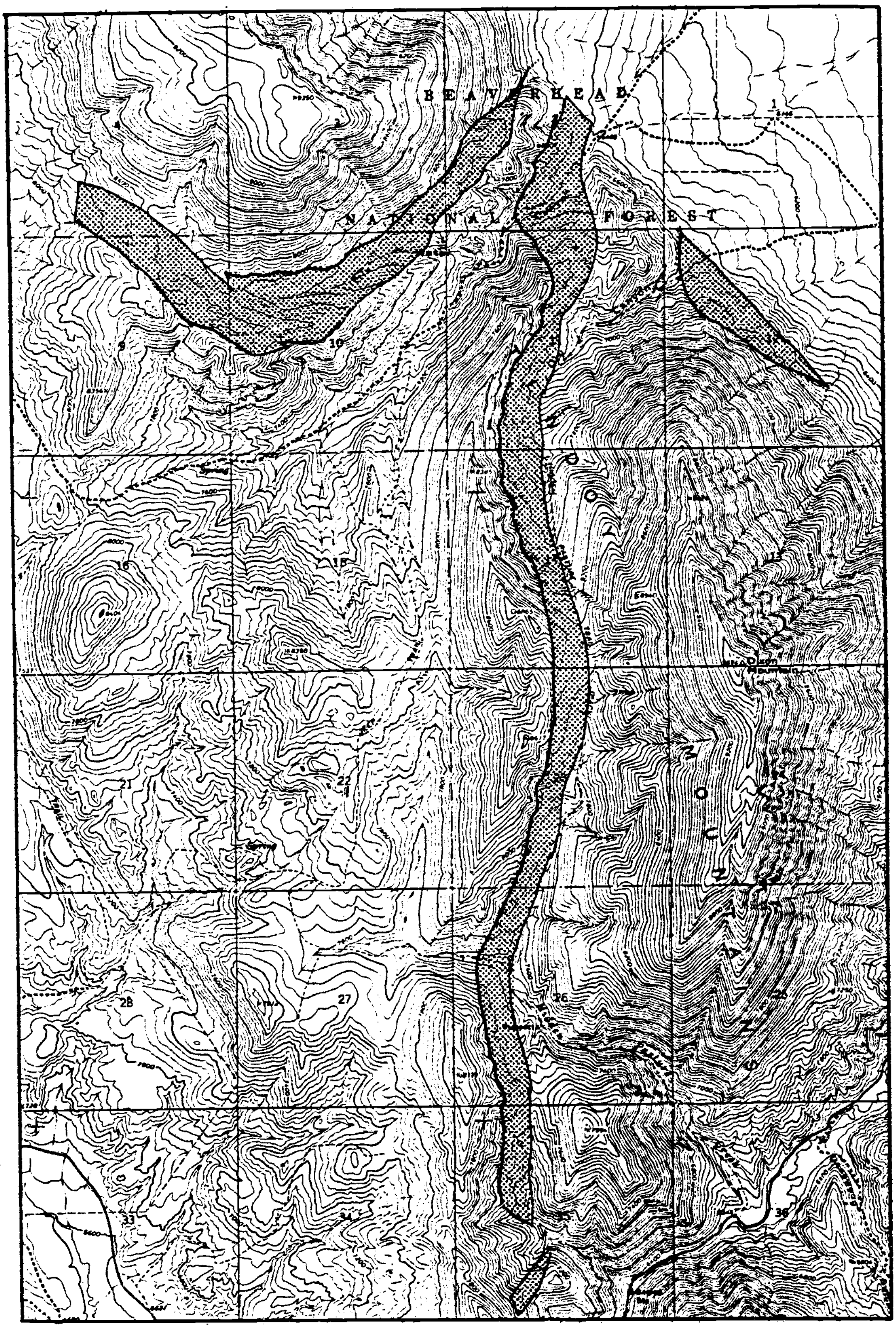

Figure 14. Distribution of Dinwoody Formation.

Figure 15. Exposures in Hidden Pasture Valley. Quadrant talus-covered slopes of Dixon Mountain at upper right skyline. Low triangular ridges of the Phosphoria to right (east) of valley. Triassic sequence exposed on slopes to left (west) of valley. Timber Butte in left distance. Looking north.

of a 4.5 mile long cuesta west of Dixon Mountain. Along this ridge, immediately west of Hidden Pasture Creek, relatively resistant silty limestone and calcareous siltstone form numerous ledges. To the north of the Hidden Pasture divide, SE ¼, sec. 14, T.13 S., R.10 W., outcrop quality decreases due to heavy forestation.

A second area of good Dinwoody exposures occurs on the north side of Little Water Canyon, near the mouth of the canyon. At this location, high cliffs have been developed.

Elsewhere in the thesis area the Dinwoody is poorly exposed. It is present, but generally soil covered, on the lowest part of the southern and southeastern slopes of Timber Butte, and along the northern slopes of the ridge immediately to the south.

Fossils and Age

The Dinwoody Formation contains a number of fossiliferous beds. The fossils are, however, rarely well preserved. Recrystallization, in the limestone has obliterated much of the detail of the fossils and molds, many of the fossils occur as fragmental debris. This poor preservation results in the recognition of numerous fossiliferous beds where identifications can not be refined beyond the phyllum or class.

Regional studies of Lower Triassic strata in the Montana-Idaho-Wyoming area by Kummel (1957) culminated in the development of a paleontologic and paleoecologic summary for the Dinwoody and Thaynes formations. Three lithologic-paleontologic facies are recognized within the Dinwoody.

The most common is the calcareous siltstone facies. The characteristic fauna of this facies includes: Lingula, Claria, Myalina and Anodontophora. This assemblage is characteristic of Early Triassic clastic-calcareous strata throughout the world. Less frequently, Mytilus and the brachiopods Terebratula, Rhynchonella, and Spiriferina are also found within the calcarous siltstone facies. This faunal assemblage suggests a shallow marine setting, not deeper than the uppermost neritic (Imlay, 1957, Kummel, 1957).

A terrigenous-free limestone facies is also present within the Dinwoody Formation. This facies commonly contains abundant Pentacrinus stem ossicles, as well as numerous valves of a small number of pelecypod and brachiopod genera. Rarely, small gastropods and echinoid spines can be found in this facies. Coquinoid banks composed of Eumorphotis, Pseudomonotis and Myalina occur within the limestone facies. These beds are interpreted as ancient shell banks (biocoenose). The fragmental nature of much of the shell debris within this facies indicates strong wave and bottom current activity in a shallow marine environment.

Infrequently, minor shaly layers are seen within the Dinwoody Formation. These beds commonly contain fragments of _Lingula_ and _Claria_. This, again, indicates a shallow water environment.

Within the thesis area the author has collected a great many _Pentacrinus_ ossicles from the terrigenous-free limestones. A great many articulated and disarticulated pelecypod valves were found within this facies as well, but recrystallization makes identification impossible. Molds of _Lingula_ and _Claria_ have been found in the calcareous siltstones. The _Lingula_ molds were found within beds that display snail trails (_Aulichnites_) along the bedding surfaces. This evidence supports the interpretation of a low energy shallow nearshore depositional environment. One bed of poorly preserved rough molds of an ammonite, possibly _Metophiceras_, was identified and sampled.

The ammonities _Discophiceras subkyokiticum_ and _Metophiceras subdemissum_ are noted within the calcareous siltstone facies. These ammonites are believed to be part of the _Otoceras_ (earliest Scythian) ammonite zone (Newell and Kummel, 1942). Investigations of Dinwoody strata in north-central Utah recovered the conodont _Anchignathodus isarcicus_. This is interpreted to be of Griesbachian (earliest Scythian) age (Clark, et al., 1977).

Lithology

<u>Outcrop Character</u> The Dinwoody Formation is commonly well exposed in outcrop throughout its occurrence within the thesis area. Thicknesses range from 870 feet (265 m.) at Little Water Canyon to 820 feet (250 m.) at the Hidden Pasture section (Moritz, 1951). The formation may be readily identified by its characteristic chocolate brown-weathering, thin-bedded limestones and siltstones.

The Dinwoody can be subdivided in the field into a lower unit of thin bedded siltstones and shales with minor interbedded limestones and a upper unit of silty limestone interbedded with shales. Moritz (1951) has suggested the informal names shale member and limestone member for the two units of the Dinwoody. Geographic names seem inappropriate, as these members do not represent easily mappable units.

The lower shale member lies upon Permian Phosphoria strata above a disconformable contact. Contrasting susceptibility to weathering commonly has resulted in the development of drainages at or near the Phosphoria-Dinwoody contact. The non-resistant strata of the uppermost Phosphoria generally form much gentler slopes than the more resistant ledge-forming Dinwoody. The lithologic evidence within the thesis area useful in locating the contact is easily recognized. The uppermost Phosphoria may be distinguished by a succession of two distinctive rock types. The lower unit of the sequence is a thin-bedded, (as thick as 24 in.; 60 cm.) sparsely fossiliferous, glauconitic, calcareous very fine-grained sandstone which is yellowish olive gray (5 Y 7/1) and weathers light olive gray (5 Y 6/1). This sandstone is overlain by several tens of meters of thin-bedded, usually fissile, fossiliferous, phosphatic siltstone or shale which is pale yellowish brown (10 YR 6/2) on a fresh surface and weathers grayish orange (10 YR 7/4). This sequence is rarely visible in outcrop but the soils derived from these rocks have a distinctive yellowish brown to olive gray color.

The lowest Dinwoody rocks above the inferred Dinwoody-Phosphoria contact are thin-bedded, in part trough cross-bedded, laminated siltstones which weather to a typical chocolate brown color (5 YR 5/4). These are interbedded with a silty pelecypod-bearing limestone which are light olive gray (5 Y 6/1) on fresh and weathered surfaces. These two rock types characterize the lower shale member of the Dinwoody Formation.

The uppermost part of the formation, the limestone member, contains a great deal (approx. 40%), of the thin-bedded and laminated calcareous siltstones described above, interbedded with thin- and very thin-bedded silty pelecypod limestones which are light brownish gray (5 Y 6/1) on the fresh surface and weather to medium gray (N 5) and pale brown (5 YR 5/2). The pelecypod fragments within these limestones are usually whole and commonly are articulated. The shell fragments and valves are commonly in framework support.

The uppermost Dinwoody in the thesis area consists of thin-bedded, very slightly calcareous siltstone which is very light gray (N 8) on fresh break and weathers to a distinctive bluish white (5 B 9/1). The nonresistant nature of the uppermost Dinwoody strata and the entire overlying Woodside rocks results in the contact between the two formations commonly being covered by talus, soil, and vegetation. Commonly there is a break in slope from the ledge-forming Dinwoody strata to the slope-forming rocks of the overlying Woodside. However, several small outcrop exposures of the contact were examined. The contact is entirely gradational and was determined in the field by the occurrence of fenestral laminae and changes in color. The uppermost Dinwoody rocks are thin- and very thin-bedded calcareous siltstones which are very light gray (N 8) fresh and weather yellowish gray (5 Y 8/1). These rocks are overlain by a thin-bedded and laminated calcareous siltstone which is moderate orange pink (10 R 7/4) on the fresh surface and weathers pale reddish brown (10 R 5/4). The laminae of this rock are crenulated and wavy and have the appearance of a fenestral limestone deposited in association with algal mats.

<u>Thin Section Description</u> In thin section the calcareous siltstones, which typify much of the Dinwoody, are composed

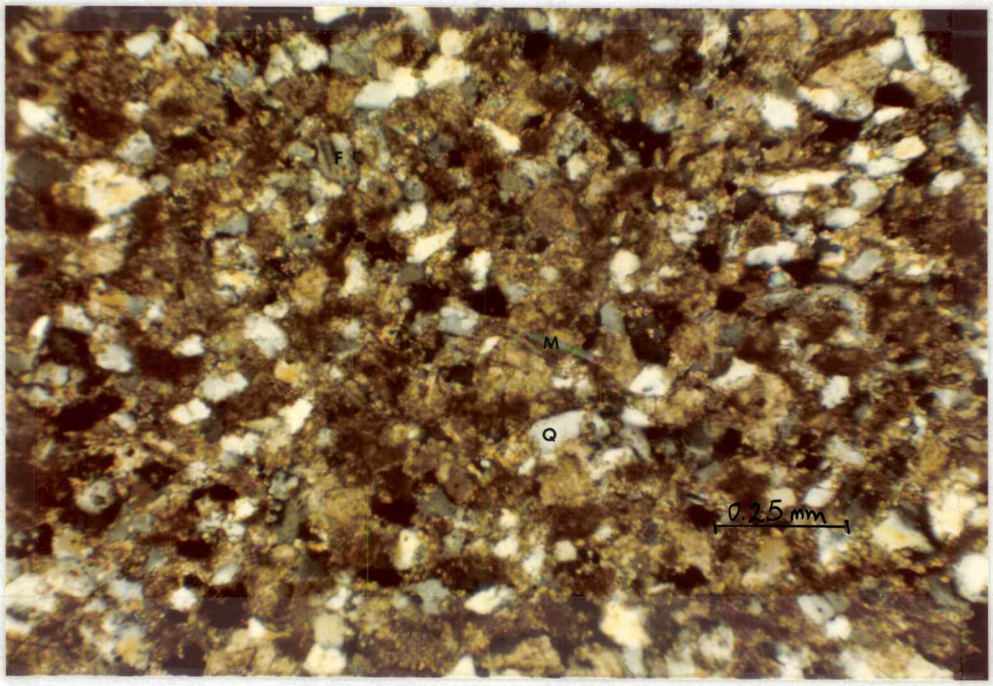

Figure 16. Photomicrograph of calcareous siltstone, composed of angular to subangular quartz (Q) with minor feldspar (F) and muscovite (M). Matrix is secondary calcite cement. Crossed nicols. Sample RAK-187, Dinwoody Formation, Hidden Pasture Valley.

of silt-sized quartz which is angular to subrounded and very uniform in size. The majority of the monocrystalline silt grains (99%) display nonundulatory extinction. A small number of polycrystalline quartz grains show undulatory extinction, suggesting a metamorphic origin. The silt grains tend to have the long axis of the grain aligned subparallel to the bedding plane. The matrix is approximately 30% of the siltstone and is microspar with a small (less than 10%) admixture of clay minerals.

The limestones within the Dinwoody display a much greater variety of lithologies. These variations will be useful in better understanding some of the conditions under which Dinwoody rocks were deposited.

Three thin sections representive of the major limestone types collected from the Dinwoody were examined. Rocks which in hand sample are seen to be silty limestones with ammonite molds are in thin section silty intraclastic biomicrosparites (Folk, 1962). Quartz silt grains within a microspar matrix occur in 0.1 inch (2.5 mm.) laminations. These silt laminae are separated by sharp planar surfaces from laminae of varing thicknesses containing molluscan fragments, silt grains, and isolated intraclasts in a microspar matrix. The molluscan fragments are angular to subangular and show no evidence of micrite coatings. The shell fragments have inverted to sparry calcite and show no remnant internal shell structure. The intraclasts are composed of micrite and subrounded quartz silt grains. The intraclasts are circular in outline, suggesting a spherical configuration, and show no evidence of micrite envelope development.

A second limestone contains much more shell debris. This rock is silty biomicrosparite (Folk, 1962) and is packstone according to Dunham's (1962) classification. The shell fragments are of pelecypods (inverted to sparry calcite) and the brachiopod Lingula. These fragments are in framework support and have their long axes oriented subparallel to bedding. A small number of the pelecypod fragments (5%) show very thin micrite envelope development. The interstices between allochems are filled with silty microspar or pseudospar. The silt grains are much like those of the previously described thin section, however many of the quartz grains with nonundulatory extinction contain black mineral inclusions.

A silty biopelmicrosparate (Folk, 1962) is composed of a variety of allochems useful for environmental interpretation. Pelecypod and Lingula valves are in framework support with voids filled with a silty, pelletal microspar and pseudospar. The shell structure of the pelecypod valves have inverted to sparry calcite but the phosphatic Lingula valves show distinctive internal microshell structure and sharp outlines. The pellets are composed of micrite and carbonaeous

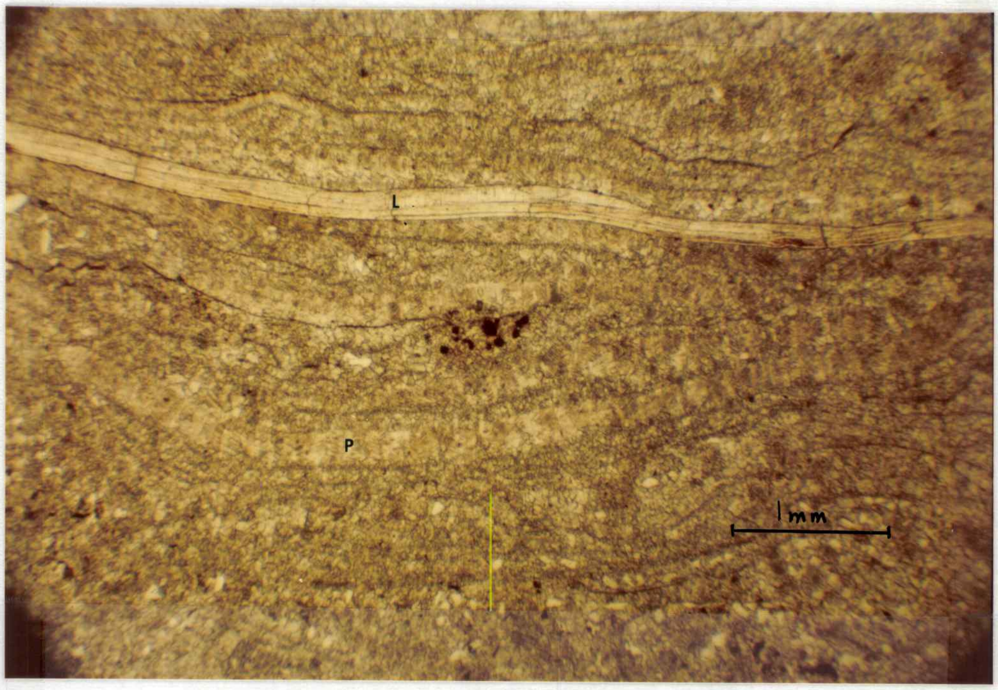

Figure 17. Photomicrograph of silty pelecypod-Lingula packstone composed of phosphatic Lingula valves (L) and pelecypod valves (P) neomorphosed to calcite spar. Note many valves are unbroken. Matrix is micrite neomorphosed to micro- and pseudospar. Sample RAK-101, Dinwoody Formation, north of Little Water Canyon.

material and are of a narrow range of shape and size. There is evidence of sheltering or "umbrella effect" (Wilson, 1975) in the nature of the matrix filling of the voids between bioclasts. Numerous interstices are filled with microspar or pseudospar, having little or no quartz silt admixture. One rounded, slightly elliptical intraclast of calcareous siltstone was also observed within the thin section.

There is no evidence of sparry calcite void fillings in any of the rocks examined in thin section. The presence of microspar and pseudospar suggest the depositional textures were entirely micritic (Bathurst, 1975) and that no pore spaces remained at the time of burial. The neomorphic processes of inversion of the aragonitic molluscan fragments to sparry calcite and aggradation of micrite matrix occurred subsequent to burial (Bathurst, 1975).

Depositional Environment

Four lines of evidence may be employed to interpret the conditions under which the Dinwoody Formation was deposited within the thesis area.

Fossils found within the Dinwoody are of a very limited number of genera though several beds contain large numbers of individuals. The lack of diversity may not be the result of environmental factors, but may be the result of the widespread extinction, at the close of the Permian, of many of the forms which could have inhabited this setting. Benthonic animals present indicate a shallow-water habitat. The presence of Lingula suggests water depth of less than 120 feet (37 m.) and most probably a very shallow nearshore or intertidal setting (Rudwick, 1965). Many fossils present in the Dinwoody are suspension feeders indicative of a fairly clear, generally well circulated environment. The presence of ammonites within the Dinwoody suggests adequate connection with open marine conditions.

A nearshore setting is compatible with the presence of silty dolomite containing subhedral dolomite rhombs in the Centennial Range (50 miles ESE of the thesis area). The presence of dolomite may be interpreted as caused by subaerial exposure in a supratidal setting (Deffeyes et al., 1965; and Shinn et al., 1965) or fresh water flushing of calcareous sediments (Folk and Land, 1975; and Dunham and Olson, 1978). The occurrence of dolomite interbedded with gypsum and anhydrite in one section examined by Heim (1962) suggests that the dolomite is, at least in part, of primary origin. This primary dolomite suggests a restricted, hot, arid setting with high evaporation rates.

Petrographic analysis and hand sample examination of Dinwoody rocks provide additional insights into the environment of deposition. The presence of a great deal of angular to subrounded silt-sized quartz grains indicates a nearshore setting, for much of this material could not remain in suspension beyond the influence of wave and current agitation. The pelecypod and Lingula valves indicate low or moderate energy conditions, as the shells are often found whole and articulated. In a few locations biostromal-like accumulations were observed. Those shells which are fragmented are angular and show little evidence of transportation. Most probably these valves were broken by predators or scavengers or possibly by storm waves.

Two of the three limestone thin sections examined contain fecal pellets in varying abundance. These indicate a relatively quiet water condition, possibly a protected lagoonal setting, for agitation would have disintegrated the pellets before lithification could have taken place (Newell et al., 1959, Bathurst, 1975). A few intraclasts of calcareous siltstone have also been seen in thin-sections. These may have been produced by either subtidal erosion, possibly by storms, or wave attack on semi-consolidated carbonate sediments.

A variety of sedimentary structures have been observed in association with many of the Dinwoody rocks. These sedimentary structures and their possible environmental significance are summarized in Table 1.

The Dinwoody was deposited on a broad cratonic platform margin which was 80-120 miles wide and may have had a slope as low as 1 ft./mi. (Newell and Kummel, 1942). In general, this setting fits into the theoretical epeiric sea sedimentation model developed by Irwin (1965). Within this model three zones of sedimentation are established, based on water depth and related energy levels of marine processes. The three zones—a low energy zone below wave base, a high energy zone under the influence of waves, and a low energy zone of dampened wave and tidal effects inland of the wave zone—provide a conceptual starting point for shallow slope platform sedimentation. Dinwoody sedimentation seems to best fit within the nearshore low energy zone where wave and tidal activity have little energy to transport sediments and calcareous muds and their contained biota are deposited in a relatively undisturbed state.

Evidence of current and wave activity within Dinwoody sediments may be attributed to wind-driven waves generated within the nearshore low energy zone and meager tidal effects which may have influenced the nearshore sediments. Recently Klein and Ryer (1978) have taken issue with Irwin (1965) and Shaw (1964) over the degree to which tidal effects traverse a broad shallow sea. Klein and Ryer have cited the presence of tidal currents in shallow Holocene seas and believe the same effects can be expected in the past. The presence of a moderate tidal range on the Dinwoody platform could explain the presence of various sedimentary structures attributable to emergent flow, falling water levels, intermittent subaerial exposure, and changes in current velocity (see Table 1).

The quartz silt within the Dinwoody may have been derived from either river-borne detritus, delivered to the shoreline and transport along shore by wave-driven longshore currents, or it may have been derived from eolian dunes which caused shoreline progradation as they advanced into the sea from the arid or semi-arid (Perry, 1962) land and were redistributed by shoreline processes. The angularity of the silt grains would indicate the former to be the case but eolian silt-sized particles need not display the typical rounding of sand-sized grains.

The nearshore deposits of the Gulf of Batabano, Cuba (Bathurst, 1975) may provide a modern analog for much of the Dinwoody deposition. In this setting a variety of molluscs live in a calcareous mud environment where salinity levels are near normal, circulation is good, wind driven currents provide necessary agitation for suspension feeding, and depths are less than 36 feet (11 m.). The shoreline is separated from the open ocean by as much as 90 miles and tidal ranges are progressively reduced shoreward.

Because of the low slope of the platform upon which deposition took place, the variations in lithology of the Dinwoody may be explained by variations in the relative

Table 1. Summary of Sedimentary Structures and Environmental Indicators

Dinwoody Formation

Characteristic Feature	Environmental Inference	Relative Abundance	Paleocurrent Direction	Reference
Intertidal Features				
Parallel laminations	Intermittency of deposition due to slight changes in current velocity	very common		Heckel, 1972
Festoon cross-bedding	Channel cutting of tidal flat by receding current	rare	N 70 W	Reineck & Singh, 1975
Symmetrical ripples	Small-scale wave activity	uncommon-rare	N 68 W	Reineck & Singh, 1975
Asymmetrical ripples	Late-stage sheet-like runoff	rare		Klein, 1971
Flat-topped asymmetrical ripples	Tidal scour, intermittent subaerial exposure	rare		Reineck & Singh, 1975; Klein & Ryer, 1978
Rill marks	Tidal scour	rare	N 33 W	Reineck & Singh, 1975; Klein & Ryer, 1978
Subtidal Features				
Feeding tracks on bedding planes	Invertebrate feeding in sub-tidal-intertidal transition	rare		Reineck & Singh, 1975; Howard, 1972
Silty fossiliferous laminae	Subtidal deposition influ-enced by current activity	common		Heckel, 1972
Parallel laminations	Fluctuations of tractive and slack water deposition	very common		Heckel, 1972; Reineck & Singh, 1975
Symmetrical ripples	Small-scale wave activity	uncommon-rare		Reineck & Singh, 1975
Asymmetrical ripples	Late-stage sheet-like runoff	rare		Klein, 1971

sealevel of the region. Variations in subsidence rates, rates of sediment accumulation, or eustatic sealevel change acting in concert or separately may have been responsible for slight variations in depositional setting.

Woodside Formation
Early Triassic

The name Woodside was first used by Boutwell (1907) for an 1,180 foot (360 m.) section of red and maroon, unfossiliferous, siltstone and shale exposed along Woodside Gulch in the Park City mining district of north-central Utah.

The stratigraphic limits and age of the Woodside Formation were defined by Newell and Kummel (1942). Those workers placed the age of the Woodside as Early Triassic (*Otoceras* zone) and defined the upper limit of the formation as the appearance of *Meekoceras*.

In the Dixon Mountain-Little Water Canyon area the Woodside Formation overlies the Dinwoody Formation conformably. The contact with the overlying Thaynes Formation is conformable as well.

Distribution and Topographic Expression

The Woodside Formation is a nonresistant slope- and valley-former in the thesis area, as it is throughout its extent. Outcrops of the Woodside are generally of poor

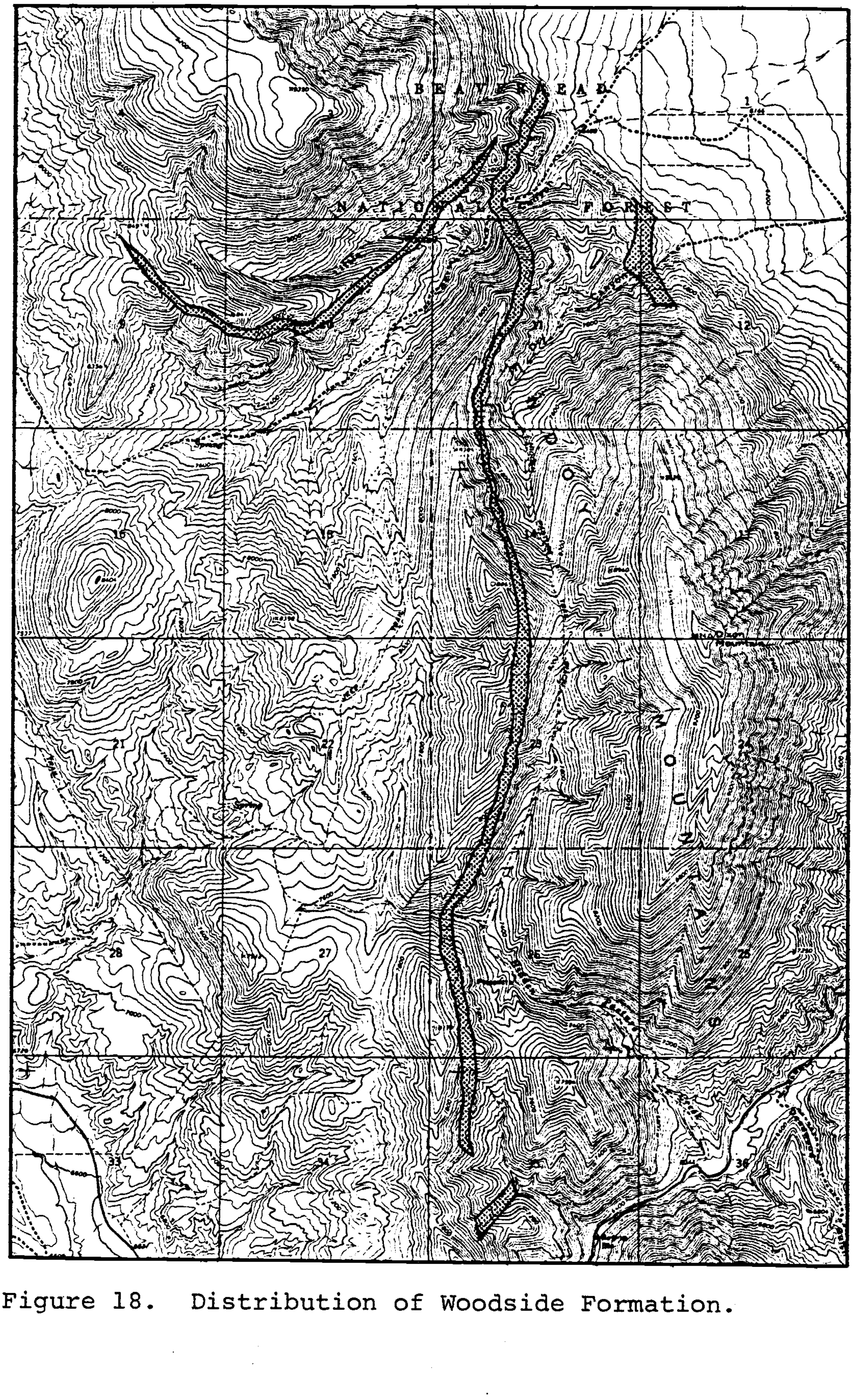

Figure 18. Distribution of Woodside Formation.

quality. Often its position is marked only by the presence of reddish soils.

The Woodside occurs most extensively midway up the scarp of the long cuesta to the west of Dixon Mountain (Plate 1). It also occurs as a trough near the top of the ridge to the south of Timber Butte and in limited exposures near the mouth of Dry Creek, NW ¼, sec. 12, T.13 S., R.10 W.

A few locations within the thesis area provide good exposures of the Woodside Formation. These occur where gullying has removed the soil cover to expose bedrock. The most notable of the exposures are: near the head of a dip slope drainage, NE ¼, NW ¼, sec. 23; the north side of a canyon, SW ¼, NW ¼, sec. 25; the north side of Dry Creek, NW ¼, NW ¼, sec. 12; and in a small drainage, NW ¼, SE ¼, sec. 2, T.13 S., R.10 W.

Fossils and Age

The Woodside Formation is very sparsely fossiliferous is the thesis area. Regional studies have yielded few fossil collections from Woodside strata. Newell and Kummel (1941) undertook an extensive regional examination of the Woodside and Dinwoody strata in an effort to accurately determine their ages. Specimens of *Discophiceras* were collected from Woodside strata near Melrose, Montana and *Claria* was collected at a number of locations. The biostratigraphic range of Discophiceras is within the Otoceras zone of Spath (1930) while Claria ranges through the Otoceras and Genodiscus zones. The Otoceras and Genodiscus zones are part of the earliest Triassic (Scythian). The Woodside was subsequently assigned to the Genodiscus zone (Newell and Kummel, 1942).

The only fossils collected from the Woodside within the thesis area were the fragmented remains of a number of juvenile ammonites in a coquina bed.

The interfingering nature of Woodside-Dinwoody contact suggests that the Woodside is time-trangressive. The presence of ammonites of the Meekoceras zone immediately above the Woodside indicates that the formation is entirely of Genodiscus (pre-Meekoceras) age (Newell and Kummel, 1941).

Litholgy

Outcrop Character The Woodside Formation is very poorly exposed within the thesis area. The formation can be most easily recognized as present where a zone of reddish soil occurs above the chocolate brown-weathering strata of the Dinwoody Formation. Little detailed work has been performed on the Woodside in the southwestern Montana and a number of previous workers have not recognized it within the region or more particularly, within the study area (Kummel, 1942, 1960). The poor outcrop quality has left the thickness of the unit in dispute. Moritz (1951) indicates a Woodside section of approximately 250 feet (76 m.) in the Hidden Pasture area while Scholten et al. (1955) assign a thickness of 122 feet (37 m.) in Little Water Canyon. Field determinations based on mapping and cross-sections indicates a thickness of approximately 200 feet (61 m.).

The Woodside Formation is generally characterized as a red bed sequence in the literature but close examination of the few well exposed outcrops reveals red beds to be of only minor importance within the study area. The limited availability of outcrops makes a bed by bed analysis of the Woodside impossible. The approximate stratigraphic position of sample locations often could be ascertained, but the construction of an accurate composite section was not possible. For this reason, lithologies will be discussed as general groups rather than as sequential variations. Lateral variations between outcrops suggest that lithologies change along depositional strike as well as vertically.

The Woodside can be characterized best as a series of silty, algal-laminated dolomites and dolomitic limestones interbedded with thin-bedded and laminated calcareous siltstones and very fine-grained sandstones, with subordinate fossiliferous limestones.

The silty, algal-laminated, ferrous dolomites and dolomitic limestones are the most distinctive and diagnostic rock type found within the Woodside. These beds are commonly more resistant than the associated siltstones. Because of differential weathering, the generally planar, minutely crenulated laminae are easily recognized. These rocks range in color from very light gray (N 8) through yellowish gray (5 Y 8/1) and pinkish gray (5 YR 8/1) to very pale orange (10 YR 8/2) on the fresh surface. Weathered surfaces range in color from pinkish gray (5 YR 8/1) and yellowish gray (5 Y 8/1) to yellowish brown (5 YR 6/6) and very pale orange (10 YR 8/2).

The algal-laminated dolomitic rocks are commonly interbedded with thin-bedded to thinly laminated, slightly calcareous siltstones and sandy siltstones. Bedding within these siltstones is generally parallel, slightly wavy and frequently discontinuous. Most of the siltstones show little evidence of bioturbation although a few samples display bedding disturbance and mottling. Laminations within the siltstones are commonly defined by variations in grain size and/or grain sorting. These rocks are nonresistant slope-formers which are rarely seen in outcrop and are presumably responsible for the characteristic reddish soils of the Woodside. Calcareous siltstones of a variety of colors were observed in outcrop. Commonly, fresh rock surfaces are either yellowish gray (5 Y 7/2) or light gray (N 7), but there is a range of other colors including pale reddish orange (10 R 6/4), pale red (5 R 6/2), moderate orange pink (10 R 7/4), grayish orange (10 YR 7/4), and very pale orange (10 YR 8/2). These rocks weather to a variety of reddish and grayish browns but, most weather to pale yellowish brown (10 YR 6/2) or yellowish gray (5 Y 8/1).

Minor beds of sandstone and limestone also were observed within the Woodside Formation. The sandstones are composed of very fine-grained quartz sand which commonly grades into overlying and underlying sandy siltstones or siltstones with intercalated sandstones. Bedding features within the sandstone strata are commonly defined by variations in sorting or concentrations of hemitite and limonite grains. The sandstones are most frequently grayish orange (10 YR 7/4) on a fresh surface and weather pale yellowish brown (10 YR 6/2).

Only two "non-algal-laminated" limestones were encountered in the Woodside of the thesis area. These beds are fossiliferous packstones (Dunham, 1962) which contained gastropod shells and are thin-bedded. The colors of these limestones are light olive gray (5 Y 6/1) and very light gray (N 8) which weather to olive gray (5 Y 7/1) and yellowish gray (5 Y 8/1).

The contact with the underlying Dinwoody Formation is gradational and has been defined in the section describing the outcrop character of that unit. This contact is usually accompanied by a break in slope from the ledge-forming

Dinwoody to the slope-forming Woodside. The contact with the overlying Thaynes Formation is sharp and planar. The uppermost rocks of the Woodside are very thin-bedded and laminated calcareous siltstones which are yellowish gray (5 Y 7/2) on the fresh and weathered surfaces. The overlying strata form a thick sequence (50-60 ft; 15-18 m.) of thin- and very thin-bedded, calcareous siltstones which contain poorly preserved molds of pelecypods and large ammonites (as large as 5 in.; 14 cm.) in diameter. The abundant ammonite impressions make this horizon very distinctive and easily recognized where outcrop quality permits. The strata are yellowish gray (5 Y 8/1) on a fresh surface and weathers dark yellowish brown (10 YR 4/2).

The contact with the overlying Thaynes Formation is commonly observed as a marked change from the slope-forming strata of the Woodside to the ledge-forming lower Thaynes beds. This contact has been described as conformable by all those investigators who have considered the Woodside-Thaynes boundary in southwestern Montana (Moritz, 1951; Scholten, et al., 1955). The contact is continuously traceable over a distance of approximately three quarters of a mile. In that distance the yellowish gray siltstone immediately below the contact thins from 24 inches (60 cm.) at its northernmost exposure to 8 inches (20 cm.) where it was last observed. This thinning may have been a function of depositional variation but, if considered in conjunction with the abrupt appearance of fossiliferous Thaynes strata across a sharp planar contact from the underlying typically unfossiliferous Woodside, this could suggest a disconformity or slight erosional unconformity.

Thin Section Description Petrographic examination of the algal-laminated dolomitic limestone, limestone, and calcareous siltstone lithologies of the Woodside Formation yield information useful in interpretation of the depositional setting.

Several of the algal-laminated dolomitic limestones were examined in an attempt to understand the conditions under which these rocks were deposited. The laminations are typically thin (0.01-0.02 in.; 0.25-0.50 mm.) and, where present, vugs (fenestrae) are confined to individual laminae. Less commonly larger fenestrae (0.01-0.03 in.; 0.25-0.75 mm.) cross one or more laminae. The large voids are commonly filled with spary calcite whereas small vugs are usually open. The algal laminations can be classified as type LLH-S or LLH-C microlaminae based on the stromatolite classification system devised by Logan et al. (1964). All of the algal laminations examined contained silt grains. Individual laminae are defined by variations in the amount of silt present. In general, laminae are developed by the alternating layers of micrite or microspar with 5 to 20% quartz silt

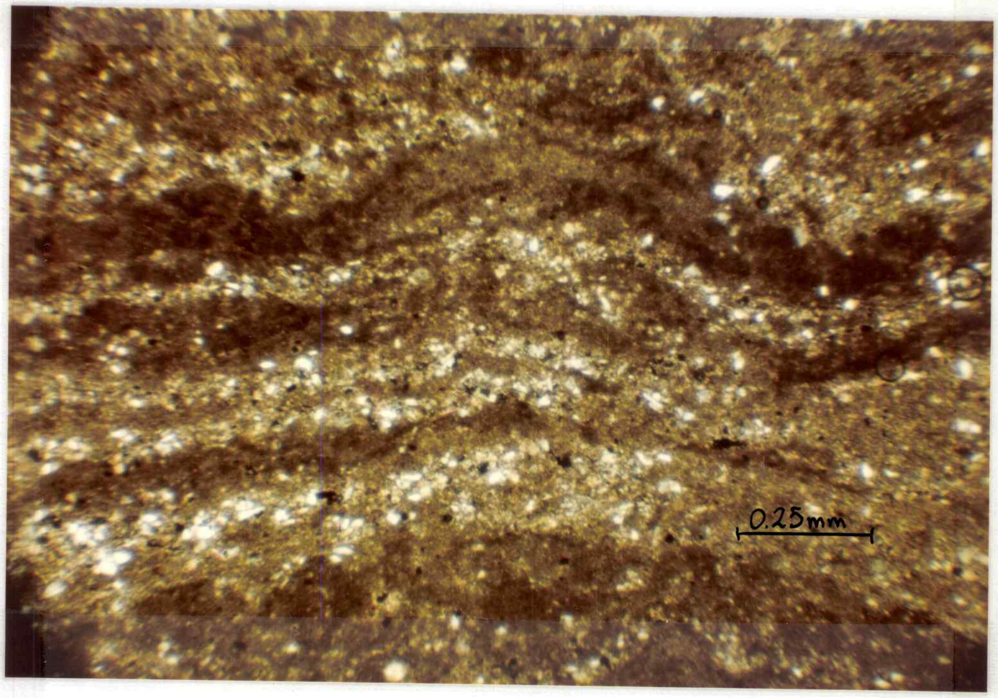

Figure 19. Photomicrograph of algal laminated, silty micrite. Note crenulation of laminations and alternation of micritic laminae with laminae containing angular quartz silt. Crossed nicols. Sample RAK-201, Woodside Formation, north side of Dry Canyon.

with layers containing 20 to 50% quartz silt. The monocrystalline silt grains are of a uniform size and are angular to subangular. Some samples contain approximately 1% silt-sized muscovite and feldspar.

Some caution must be employed in the interpretation of thinly laminated carbonate sediments as having been deposited in association with algal mats, for caliche laminates may have a very similar appearance. Read (1976) has outlined a series of criteria useful in making a distinction between these two types of laminated sediments. Of particular application to the sediments under consideration are the presence of laminae defined by grain size alterations. This is diagnostic of cryptalgal laminae. Calcrete laminates are generally defined by color changes. In addition cryptalgal laminates will display evidence of desiccation and convolutions in mat layers. The presence of abundant fenestrae and, less commonly, discontinuous, contorted, and domal laminae may be interpreted as desiccation features (Shinn, 1968). Lastly, calcrete is associated with weathering profiles and vadose diagenetic phenomena, none of which were observed within the Woodside.

The thin-bedded, vuggy and mottled, silty limestone examined in thin section is a silty pelmicrite (Folk, 1962) or a pelletal packstone (Dunham, 1962). The pellets all are of approximately the same size, but of two different shapes;

round to ovoid and cylindrical or rod-like. This rock contains about 10% angular to subrounded quartz silt but those parts which have a mottled appearance contain approximately 20% silt. The matrix of this rock is entirely micrite with no evidence of microspar development.

The calcareous siltstone examined in thin section is thinly laminated. This rock is composed of well sorted, angular and subangular quartz silt in framework support. Herringbone cross-laminations are defined by variations in abundance of a brown, clay-sized material (possibly iron-rich clays). The matrix material is at least in part calcareous and only partly fills intergranular voids. Both strained and unstrained quartz are present.

Depositional Environment

A variety of sedimentary structures observed in outcrop and hand sample, when combined with information derived from petrographic examination of thin section, provide useful indications of the setting responsible for Woodside deposition (see Table 2).

Perhaps the most diagnostic features are algal laminated ferrous dolomites. These rocks are composed of smooth or slightly undulating, continuous laminae which probably were generated by rather smooth, flat algal mats (Hoffman, 1976). Evidence of desiccation within these algal laminates is confined to abundant intralaminar fenestrae. This type of flat, fenestral mat is characteristic of the intertidal zone of a coastline protected from wave action (Logan, 1961; Kinsman and Park, 1976). Fenestral laminae are best developed in this intertidal zone where persistent flooding alternates with maximum exposure (Wilson, 1975). The alternation in abundance of silt within the algal laminae has been interpreted as caused by periodic or episodic influx of detrital particles (Monty, 1976). The detrital influx may have been cyclic (tidal effects) or erratic (storm effects).

It is possible that silt may have been brought into the nearshore environment by eolian processes. Investigations of modern algal belt and coastal sabkha sedimentation along the Trucial Coast of the Persian Gulf have shown that wind-blown detritus accumulates by adhesion on the moist algal and sabkha surface (Kinsman and Park, 1976). Kubal and Saadallah (1973) found an admixture of calcite, quartz (both rounded and angular), chert, muscovite, and plant fragments deposited by eolian processes in nearshore carbonate sediments of the northern Persian Gulf. Sedimentation rates as high as 1 inch/year (2.1 cm./yr.) were recorded.

Eolian contributions could have been a significant factor along the western margin of the North American continent during the Early Triassic. Climatic and paleomagnetic evidence indicate a lower latitude position of western North

America during the Triassic and Early Jurassic (Peterson, 1978). This near equatorial position would place the southwestern Montana area within the trade wind belt, a zone of almost constant northeasterly winds. This prevailing wind direction could have brought large amounts of airborne silt, derived from arid inland redbeds into the coast environment.

The dolomites and redbeds provide additional evidence of the climatic setting and depositional environment. The interbedding of dolomitic and nondolomitic strata within the Woodside indicates the dolomites present are not the result of hypersaline reflux conditions such as those described by Deffeyes et al. (1965) or diagenetic replacement in the subsurface by freshwater dilution of marine pore waters (Dunham and Olson, 1978). Both of these situations would have resulted in pervasive dolomitization of all Woodside strata. Algal laminated limestones forming in an arid climate are subject to considerable evaporation. The effects of evaporation on peritidal sediments in the Bahamas (Shinn et al., 1965) result in an increase in the concentration of dissolved salts just below the sediment surface. Magnesium ion concentration in interstitial waters can increase to the point that penecontemporaneous replacement of $CaCO_3$ by dolomite takes place. This is consistent with the findings of Folk and Land (1975). These findings show that under otherwise constant conditions, an increase in the Mg/Ca ratio will result in the precipitation of dolomite. In addition, the work of Folk and Land indicates that dolomitization can occur if the intertidal-supertidal zone were to experience a fresh water influx. This fresh water influx would lower the salinity of the interstitial waters and could lead to dolomite precipitation.

The reddish siltstones found within the Woodside are probably the westward extension of much greater thicknesses of red beds which lie east of the thesis area. These siltstones have a matrix composed of iron-rich clay material. Walker (1967) suggested that many red beds formed as the result of postdepositional diagenetic oxydation of ferromagnesium minerals in fluvial and fluvial-marine deposits of hot arid or semiarid climates. This may well be the case for the red beds of the Woodside, for the thick sequences of Woodside strata present throughout southwestern Montana and most of western Wyoming are interpreted as nonmarine. In addition paleoclimatic and paleomagnetic evidence indicate an arid setting.

These iron-stained sediments, carried by northeasterly wind to the shoreline, were probably responsible for the silt admixture of those peritidal sediments and may have been the source of iron in the ferrous dolomites.

The tidal flat subkha coastline of the southwestern coast of the Persian Gulf (Kendall & Skipwith, 1968; Kinsman and Park, 1976) and the tidal flats of Shark Bay, Australia (Hagen and Logan, 1975; Woods and Brown, 1975; Hoffman, 1976, Playford and Cockbain, 1976) offer modern analogs for the cryptalgal laminates of the Woodside Formation. Both of these locations are set in a arid climate where evaporation greatly exceeds precipitation. Sediment surfaces are often kept moist by the capillary rise of interstitial pore waters. Algal mat growth types have developed parallel to the shoreline at these two locations. Variations in tidal range and wave effect have resulted in slight differences between these two settings. In the Persian Gulf area, smooth algal mats are best developed in the upper tidal zone, whereas more polygonal and crenulated mats have developed seaward. This pattern has been attributed to the effects of tidal runoff (Kendall and Skipwith, 1968). In the Shark Bay setting, smooth algal mats are best developed in the lower intertidal zone (Woods and Brown, 1975). The effects of desiccation on the algal mats should result in the formation of flat pebble conglomerates as dried mats are eroded and redeposited in channels and potholes (Wilson, 1975). These features were not observed within the study area but were seen in Woodside outcrops 10 miles to the south (Sadler, personal commun.). Algal mats are observed in modern and ancient examples to form bulbous and columnar features.

Table 2. Summary of Sedimentary Structures and Environmental Indicators

Woodside Formation

Characteristic Feature	Environmental Inference	Relative Occurance	Reference
Undulatory small current ripples	low-energy to high-energy transition intertidal flat, tidal channel or inlet	rare	Reineck & Singh, 1975
Herringbone cross-bedding	rapid reversal of depositing rate current typical of tidal flat deposits	rare	Reineck & Singh, 1975; Ginsberg, 1975
Birdseye (fenestral) structure	entrapment of gas bubbles, desiccation of algal mats; diagnostic of supratidal and less commonly intertidal environment	common	Shinn, 1968; Wilson, 1970; Heckel, 1972
Cryptalgal laminates	cryptalgal mat growth and sediment adhesion; supratidal and high intertidal zone	abundant	Shinn, et al., 1969; Kinsman, et al., 1976 Heckel, 1972; Logan, 1961; Hoffman, 1976; Gebelein, 1976
Clay drapes (mud scum)	intermittent traction and suspension deposition; tidal flat deposits	rare	Ginsberg, 1975;
Mud-sand laminae couplets	alternation of eolian and tidal deposition or tidal flat	common	Wanless, 1975; Wilson, 1970
Flat pebble breccia of algal laminates	desiccation and erosion of algal mats; tidal flat environment	rare	Zamarrona, 1975
Red beds	Subaerial weathering of iron-rich soils or alluvium in warm humid to arid climate	uncommon	Walker, 1967; Van Houten, 1968
Dolomite	penecontemporaneous replacement of $CaCO_3$ where tidal or storm flooding alternates with subaerial exposure of supratidal flats	common	Shinn, et al., 1965

Hagan and Logan (1975) suggest the lack of such structures may reflect the combination of effects of low tidal range and low shore profile. The shallow cratonic sea of the Triassic of western North America, well isolated from the open ocean, may have had just such a low profile and low tidal range.

Thaynes Formation
Early Triassic

The name Thaynes was applied by Boutwell (1907) to 1,190 feet (363 m.) of strata exposed along Thaynes Canyon in the Park City mining district of north-central Utah. The type section was described as: a basal interbedded gray, tan, and black limestones, siltstones, and shales: and an upper tan sandstone. Boutwell noted some local tonguing of red shale in the middle part of the formation.

The base of the Thaynes Formation was defined by Newell and Kummel (1942) as the location of the *Meekoceras* zone. Kummel (1954) recognized the Thaynes over large areas of Utah, Idaho, Wyoming, and Montana.

In the Dixon Mountain-Little Water Canyon area the Thaynes Formation conformably overlies the Woodside Formation. The Thaynes Formation is overlain disconformably by the Jurassic Sawtooth Formation of the Ellis Group. This disconformity represents a major erosional episode in the region (Moritz, 1951). An undetermined part of the

Thaynes has been removed by erosion. No sedimentary record exists in southwestern Montana for the Middle Triassic through Early Jurassic.

Distribution and Topographic Expression

The two thick limestone units within the Thaynes Formation are predominant ledge- and ridge-formers throughout the thesis area.

The 4.5 mile long, north-trending cuesta to the west of Dixon Mountain has developed on the westward dipping resistant limestones of the Thaynes, shallow, cross-cutting canyons in NW ¼, sec. 23 and NE ¼, sec. 27, T.13 S., R.10 W., provide good exposures of much of the Thaynes strata.

The near vertically dipping flatirons developed on the southern flank of the ridge to the north of the south fork of Little Water Creek, SE ¼, sec. 9, and SW ¼, sec. 10, T.13 S., R.10 W., provide some spectacular exposures of the Thaynes Formation. These outcrops continue to the northeast for another mile as the strata becomes overturned.

Elsewhere in the thesis area, small low-lying outcrops are present near the eastern margins of the Medicine Lodge thrust sheet, NW ¼, sec. 16, and NW ¼, sec. 27, T.13 S., R.10 W. The location of these exposures appears to be controlled by faulting.

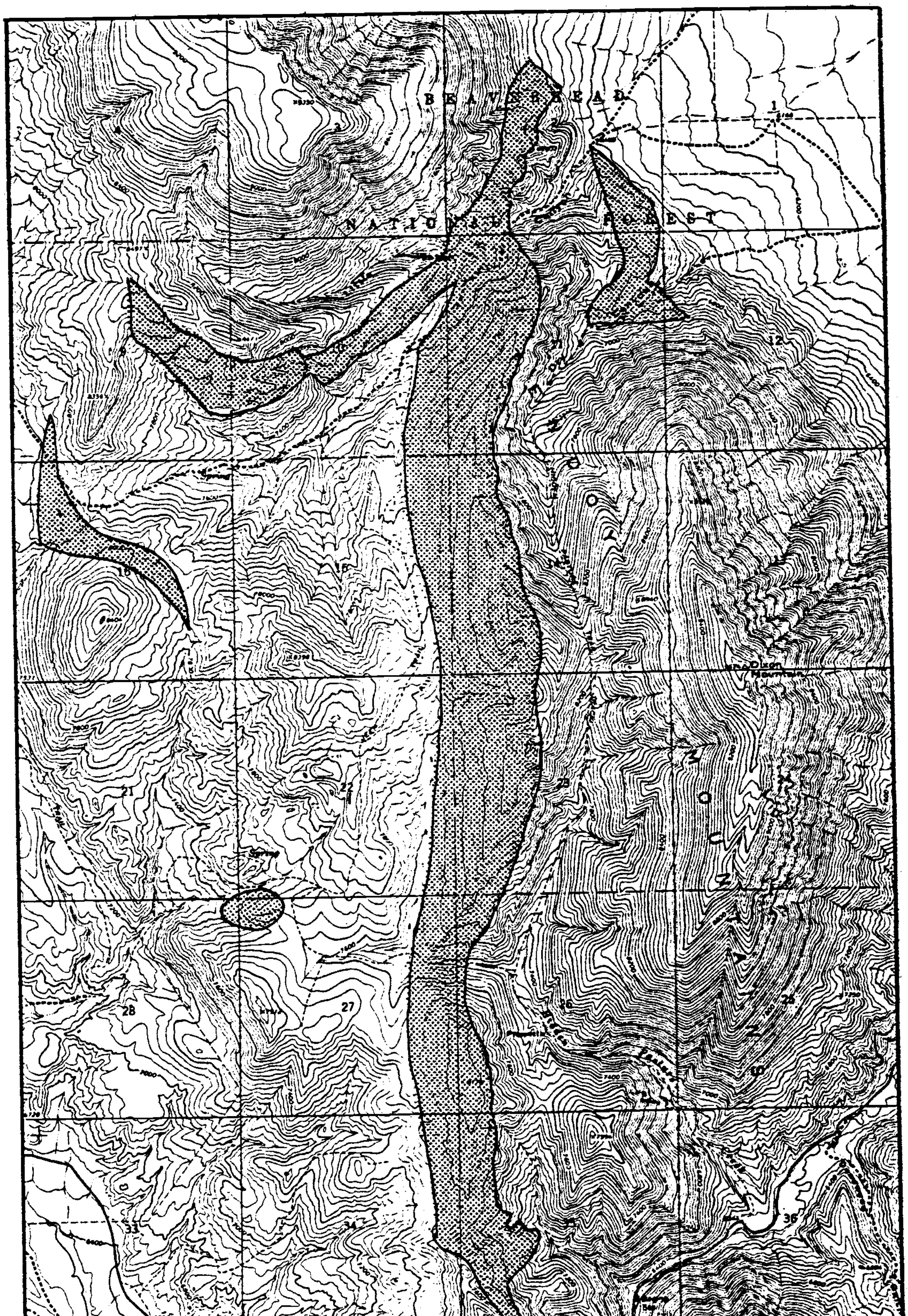

Figure 20. Distribution of Thaynes Formation.

Fossils and Age

Numerous beds within the Thaynes Formation are highly fossiliferous. In general, preservation is poor and much of the material is fragmental. None of the fossils collected were silicified and pervasive calcite recrystallization of shells has resulted in a few fossils breaking cleanly away from the enclosing rock. The most easily identifiable fossils were those which were exposed on weathered surfaces.

The basal limestone member of the Thaynes commonly contains *Meekoceras*. As many as fifty species of ammonites have been identified in basal Thaynes strata (Kummel, 1967). These ammonites are commonly assoicated with a variety of pelecypods and brachiopods, including *Lingula*. Shell banks containing a biocoenose assembly of *Eumorphotis*, *Pseudomonotis*, and myalinids are present at a number of levels in the lower Thaynes (Kummel, 1967). *Pentacrinus* is very common within the Thaynes limestones.

The limestone-sandstone facies of the Thaynes contains few well preserved fossils. Pelecypods including pseudomonotids, pectinoids, myalinids, and anodontophoras are those most frequently encountered. The brachiopod *Lingula* is also common within this facies.

The presence of the large ammonites *Pseudosageceras*, *Meekoceras*, *Flemingites*, and *Wyomingites* indicate an Early Triassic age for the Thaynes Formation (Kummel, 1967). A study of the conodont biostratigraphy of the Thaynes in Utah suggests a late Scythian (Smithian-Spathian) age (Solien, 1979).

A Spathian age has also been determined for conodont-brachiopod-bearing Thaynes strata examined in the Preuss Range of southeastern Idaho (Perry and Chatterton, 1979).

In the thesis area, abundant Pentacrinus ossicles are seen in much of the limestone strata. In addition, identifiable specimens of the pelecypod Eumorphotes and the brachiopod Rhynchonella were collected. Snail trail trace fossils (Aulichnites) were recognized in the calcareous sandstones of the Thaynes. These snail trails are believed to be indicative of the offshore-nearshore marine transition (Howard, 1972).

The fossil assemblage common to the Thaynes indicates a shallow marine environment. The presence of many fragmented shells suggests moderately active wave and current activity and a well oxygenated bottom (Kummel, 1967).

The fossiliferous horizons discussed by Perry and Chatterton (1979), in southeastern Idaho were interpreted to represent a shallow water, probably subtidal environment with moderate wave energy.

Lithology

Outcrop Character The Thaynes Formation is well exposed in the thesis area. The Thaynes is 681 feet (208 m.) thick in the Little Water Canyon area (Scholten, et al., 1955). Three members of the Thaynes are identifiable: a lower limestone member (109 ft.; 33 m.), a middle sandstone member (218 ft.; 67 m.), and an upper limestone member (354 ft.; 108 m.) (Kummel, 1960).

The lower limestone member consists of gray-weathering limestone interbedded with thin-bedded, gray, calcareous siltstone. The lower part of this member contains poorly preserved ammonite impressions, and pelecypod molds are conspicuous throughout much of the member. This unit is very similar in appearance and lithology to the upper limestone member of the Dinwoody Formation.

The limestones of the lower member of the Thaynes can be characterized as either fossiliferous micrites or fossiliferous sparites. The micritic limestones commonly contain *Pentacrinus* debris in matrix or grain support. These wackestones and packstones (Dunham, 1962) are pale yellowish brown (10 YR 6/2) or light olive gray (5 Y 6/1) on fresh surfaces and weather to very light or light gray (N 7 or N 8). The sparry limestones of the lower member are grainstones (Dunham, 1962) generally composed of either pelecypods or *Pentacrinus*,

Figure 21. Steeply dipping thin-bedded fossiliferous limestones of Thaynes Formation south of Timber Butte. Very thin intercalations of silty limestone.

locally containing calcareous siltstone intraclasts and pellets.

The siltstones of the lower limestone member of the Thaynes are very thin- or thin-bedded and are locally fossiliferous. These siltstones are composed of well sorted, angular to subangular quartz, with laminations and bedding defined by slight variations in grain size and sorting. Laminations are locally well defined, but numerous beds show disrupted laminae or a thoroughly homogenized texture indicative of varying degrees of bioturbation. Fossils present within the siltstone are almost exclusively pelecypods. The pelecypod debris is commonly whole and generally occurs in those beds which are bioturbated. Minor glauconite, hematite, and limonite occur in the siltstone strata.

The middle sandstone member of the Thaynes Formation is composed of very fine- and fine-grained, gray to tan sandstones with interbedded siltstones and minor limestones. These sandstones are commonly poorly exposed and generally form a gentle covered slope between the more resistant, ledge-forming, limestone members.

The sandstone of the middle member of the Thaynes may be characterized as thin-bedded, moderately sorted, fine-grained, rounded to subrounded quartz sandstone. These rocks are pale yellowish brown (10 YR 6/2) on a fresh surface and weather to yellowish gray (5 Y 8/1). Locally these sandstones contain coarse sand-sized chert grains or well rounded, very fine sand- to silt sized bioclastic material. The limestones of the middle sandstone member are mostly silty micrites, which are structureless. These rocks are light olive gray (5 Y 6/1) when fresh and very pale orange (10 YR 8/2) when weathered. Locally light olive gray (5 Y 6/1) weathering algal laminated limestones occur within the limestone strata.

The upper limestone member of the Thaynes is a series of thin-bedded, gray-weathering limestone with minor intercalations of calcareous sandstone and siltstone. A few chert beds and small chert nodules occur in the upper part of this member. This member resembles the lower limestone member in lithology and exposure, although limestone composes a much greater part of the total thickness of the member and *Pentacrinus* columnals are much more abundant.

Most of the limestones of the upper member of the Thaynes are fossiliferous, locally silty, packstones or grainstones (Dunham, 1962) containing either pelecypod or *Pentacrinus* fossils, but rarely both. These rocks are medium or very light gray (N 6 or N 7) or yellowish gray (5 Y 8/1) on fresh surfaces and weather to the same colors. The pelecypod valves present in these rocks are generally whole and locally are articulated. The *Pentacrinus* columnals are always disarticulated.

Locally, algal laminated limestone are present. These laminae display bedding contortion and zones of rip-ups.

The calcareous siltstones of the upper Thaynes are thin- and very thin-bedded, parallel laminated or structureless, and locally contain disarticulated, whole, pelecypod valves. The silt grains are well sorted, angular to subangular quartz. These rocks are yellowish gray (5 Y 8/1) to pale reddish brown (10 R 5/4) on fresh surfaces and weather to yellowish brown (5 Y 7/2) or moderate orange pink (10 R 7/4). A few thin, fine-grained calcareous sandstone beds also are present in the upper Thaynes. These rocks are composed of well sorted, subrounded quartz locally containing whole disarticulated pelecypod valves.

A number of locally persistent erosional surfaces were observed within the upper limestone member. These surfaces are sharp with approximately two inches (5 cm.) of relief. Siltstones immediately below these erosional surfaces commonly are pale red (10 R 6/2) with a noticeable hematite admixture to the calcareous cement. This coloration generally fades downsection from the erosional surfaces over four to eight inches (10-20 cm.).

The characteristics of the contact at the base of the Thaynes have been described in the outcrop description section of the Woodside Formation. Where exposed, the transi-

Figure 22. Multiple erosional surfaces developed in upper part of steeply dipping Thaynes Formation. Pencil points toward stratigraphic up. Exposed in flat-irons of the Thaynes north of Little Water Creek.

tion from the light colored, unfossiliferous strata of the Woodside to the dark yellowish brown (10 YR 4/2) fossiliferous, locally ammonite-rich Thaynes strata, is easily recognized. The upper contact with the overlying Sawtooth Formation was never seen in outcrop. However, the changes from gray, fossiliferous limestone of the upper member of the Thaynes to yellowish brown calcareous siltstones with no interbedded limestones may be taken as the position of the inferred contact. This contact is an erosional unconformity representing a period of post-Thaynes erosion which removed part of the Thaynes in southwestern Montana (Moritz, 1951). The Thaynes-Sawtooth contact is usually accompanied by a marked change in topographic expression. The uppermost Thaynes is a very resistant ledge- and ridge-former, whereas the overlying Sawtooth is very nonresistant. Where dipping strata is involved (commonly the case in the thesis area) dip slopes are developed at the top of the Thaynes with Sawtooth rocks forming low-lying soil-covered topography at the base of the dip slopes.

Thin-Section Description Those rocks of the Thaynes Formation examined petrographically include a calcareous siltstone and several fossiliferous grainstones, packstones, and wackestones (Dunham, 1962).

The fossiliferous packstones are the most common limestone encountered in the Thaynes. These rocks are generally

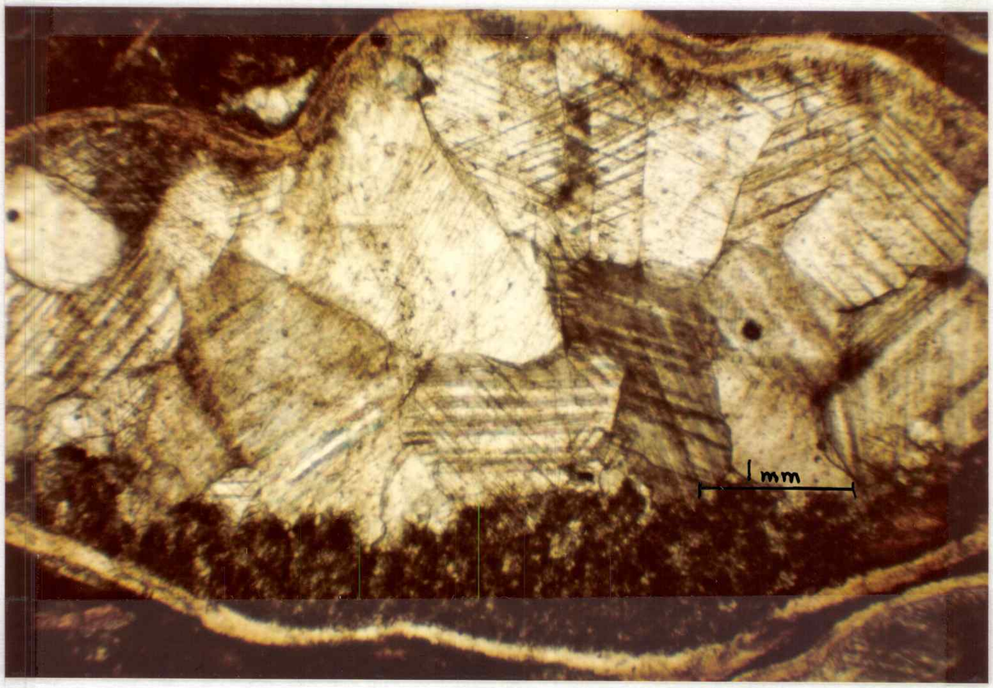

Figure 23. Photomicrograph of whole pelecypod packstone. Note well developed geopetal fabric within unbroken articulated pelecypod enclosed in micrite. Sample RAK-162-1, Thaynes Formation, NW¼, NW¼, sec. 27.

whole fossil packstones composed of disarticulated and, less commonly, articulated pelecypod valves. Where the pelecypods are articulated, geopetal fabric has been developed within shell interiors because of partial filling of void by carbonate mud and later filling of the remaining cavity by calcite spar. Pelletal or Pentacrinus-pelletal packstones also are common within the limestone members of the Thaynes. The packstones invariably show evidence of aggrading neomorphism of the micrite matrix to microspar and, less commonly, pseudospar. The molluscan fragments have all undergone neomorphic inversion of aragonite to calcite spar. Several rocks show partial silica replacement of calcite within the molluscan components.

A number of limestones are pelecypod grainstones. These rocks are composed of pelecypod, gastropod, and echinoderm remains with an admixture of Lingula fragments and silty micrite intraclasts. These rocks have undergone extensive diagenesis. One example displays the intensity of the diagenetic effects. This rock was deposited originally as a grain-supported accumulation of pelecypod and gastropod shells with some crinoidal debris. Calcium carbonate-rich waters, circulating through the interparticle porosity deposited calcite cement which filled much of the porosity. Syntaxial overgrowths of calcite developed on crinoidal debris during this period of diagenesis in an active fresh water phreatic environment, saturated with respect to $CaCO_3$.

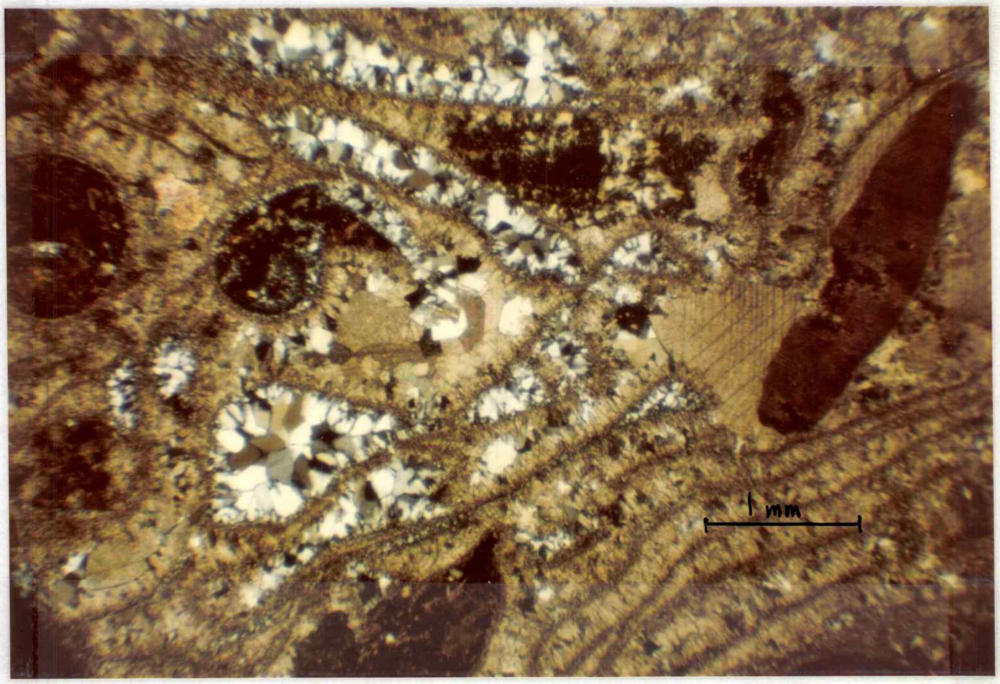

Figure 24. Photomicrograph of pelecypod-gastropod grainstone. Note moldic porosity (upper photo) subsequently filled by silica (lower photo, crossed nicols). Note syntaxial calcite overgrowth on echinoderm fragment (upper right) and remnant micrite rims around mollusc fragments. Sample RAK-112, upper limestone member of Thaynes Formation, west of Hidden Pasture.

The precipitation of calcite cement may have occurred synchronously with the dissolution of the aragonitic moll-uscan shells, for there is good evidence of the development of a moldic porosity (Longman, 1980). Subsequent to moldic porosity development, short euhedral calcite blades grew radially inward within moldic cavities as circulation of carbonate-rich fresh water decreased and pore fluids became stagnant. Most moldic pores were later filled with subhedral to euhedral quartz. The silica-rich fluids responsible for this cementation may have been derived from siliceous shales and cherts of the Phosphoria Formation. Those few pores still remaining after silica cementation have since been filled with calcite spar. Diagenesis within the limestones of the Thaynes is pervasive. Fossil mold boundaries are often blurred and neomorphism and recrystallization are commonplace. Rocks of very similar lithology within the Dinwoody do not show this degree of diagenetic alteration.

The uplift and erosion of the Thaynes in post-Early Triassic time may account for the very intensive diagenesis these rocks have undergone relative to underlying Triassic strata. This uplift, immediately following deposition, would have placed Thaynes strata in the active fresh water phreatic environment. Downward percolating meteoric waters would have quickly become saturated with respect to

$CaCO_3$. These saturated fresh waters, circulating through the Thaynes strata, could have produced the molluscan solution, sparry calcite cementation, and micrite neomorphism so pervasive in the Thaynes. Much of the siliceous cementation may have occurred subsequent to deeper burial later in the Mesozoic.

The calcareous siltstones and silty limestones of the Thaynes are gradational toward one another. These rocks are usually 40 to 60%, well sorted, angular to subangular quartz silt with a matrix of micrite neomorphosed to microspar. Many of the siltstones contain 1-2% muscovite, 5-10% hematite-stained quartz, and bioclastic fragments.

Depositional Environment

The fossils, sedimentary structures, outcrop character, and petrography of the rocks of the Thaynes Formation provide useful insights to the depositional environments responsible for Thaynes strata.

The molluscan fauna of the Thaynes suggests a shallow, normal marine environment with low to moderate wave activity. Wave action could not have been too severe for much of the pelecypod debris is whole and locally articulated.

The sedimentary structures observed within the Thaynes Formation (Table 3) provide evidence of subtidal and intertidal deposition. The presence of local erosional surfaces

Figure 25. Feeding trail (snail trail?) on bottom of rippled calcareous siltstone of the Thaynes Formation. Exposed S.W. of divide between Dry Canyon and Hidden Pasture Creek.

and solution cavities indicate occasional subaerial exposure and weathering.

The depositional environment of the upper and lower limestone members of the Thaynes was probably very similar to that described for the Dinwoody Formation. Limestones composed of worn and micritized grains indicate a shoaling environment within a zone of constant wave action. The lime mud was winnowed out of this setting. The majority of the limestones of the Thaynes are packstones of molluscan, crinoidal, or pelletal grains. These rocks indicate a quiet water setting where lime mud could accumulate. This may have been in a protected shallow lagoon setting with sufficient water circulation to maintain salinity levels within the toleration range of the abundant pelecypod fauna present. Those strata containing articualted bivalves probably rarely underwent any wave agitation or any intense bioturbation. Those beds containing disarticulated, whole pelecypods may have been deposited under very similar conditions but bioturbation was more intense. In addition, occasional storm waves may have "touched bottom" during early burial and a slight redistribution may have resulted. Where present, crinoidal debris is always disarticulated. These crinoids probably lived in a zone of stronger water circulation and were transported to quieter mud-rich environments after death.

The calcareous siltstones and silty micrites of the Thaynes may have been deposited in a setting similar to their counterparts in the Dinwoody. Silt-sized quartz may have been delivered to the marine environment by eolian and fluvio-deltaic processes. Longshore transport within the wave zone could have distributed this material along much of the Thaynes strandline.

The poorly sorted, lithic sandstones of the Thaynes have no counterpart in the Dinwoody Formation. These rocks are composed of a texturally submature mixture of detrital quartz, chert, and siltstone with an admixture of glauconite and apatite. The glauconite and apatite may have been brought into the sandstone facies by shoreward transport processes. These unstable minerals, as well as abundant chert, may have been derived from Phosphoria strata exposed to subaerial erosion a short distance inland (Hallam, 1975). The texturally submature and compositionally mixed sandstones may record the progradation or laterial spreading of fluvial sediments deposited along a deltaic margin. The submaturity of these rocks is characteristic of a variety of depositional settings but is most common in river channels, floodplains, and beach and bar deposits (Folk, 1975).

The nearshore deposits of the Thaynes of southwestern Montana were deposited on a shallow shelf or in a lagoon with good circulation protected from strong wave activity

Table 3. Summary of Sedimentary Structures and Environmental Indicators

Thaynes Formation

Subtidal Features

Characteristic Feature	Environmental Inference	Relative Abundance	Paleocurrent Direction	Reference
Parallel lamination	Fluctuation of tractive & slack water deposition	common		Heckel, 1972; Reineck & Singh, 1975
Primary current lineations	Backwash structure formed in upper flow regime	common		Reineck & Singh 1975
Current crescents	Depression caused by current flow around an obstacle	rare	S 35 E	Reineck & Singh, 1975
Symmetrical ripples	Small scale wave activity	uncommon	N 25 E bidirectional	Reineck & Singh, 1975
Asymmetrical ripples	Late-stage sheet-like runoff	uncommon	S 61 E S 04 E	Klein, 1971
Feeding tracks on bedding	Invertebrate feeding in subtidal-intertidal transition	rare		Howard, 1972 Reineck & Singh, 1975
Small erosional channels with channel fill layers which approximately conform to channel shape	Channel filling of tidal scour channel	rare		Reineck & Singh, 1975; Klein & Ryer, 1978
Oscillation ripples superimposed on current ripples	Late-stage emergent runoff changing to shallow slack water with wind generated oscillation waves	rare		

Table 3. Cont.

Characteristic Feature	Environmental Inference	Relative Abundance	Paleocurrent Direction	Reference
Intertidal Features				
Parallel laminations	Fluctuation of tractive and slack water deposition	common		Heckel, 1972; Reineck & Singh, 1975
Primary current linea - tions	Backwash structure formed in upper flow regime	common		Reineck & Singh, 1975
Current crescents	Depression caused by current flow around an obstacle	rare	S 35 E	Reineck & Singh, 1975
Symmetrical ripples	Small scale wave activity activity	uncommon	N 25 E bidirectional	Reineck & Singh, 1975
Asymmetrical ripples	Late-stage sheet-like runoff	uncommon	S 61 E S 04 E	Klein, 1971
Cryptalgal laminates	Cryptalgal mat growth and sediment adhesion	uncommon		Shinn, et al., 1969; Kinsman et al., 1971; Logan 1961; Hoffman, 1976; Gebelein, 1976
Reactivation surfaces	Alternation of tidal traction deposition and tidal scour	rare		Klein, 1970; Klein & Ryer, 1978
Prolonged Subaerial Exposure				
Solution cavities	Subaerial exposure at carbonate environ, possibly related to eustatic sea-level changes	rare		Purdy, 1974

by the attenuating effect of the broad, shallow cratonic platform. Detrital influx was generally low but the middle sandstone member records a marked increase in terrigenous sedimentation. Periodic shoaling and subaerial exposure occurred during deposition of the upper member of the Thaynes. Much of the Thaynes sedimentation took place at or near the subtidal-intertidal transition under conditions of low wave intensity and infrequent storm activity.

Ellis Group

The Ellis Formation was first described by Peale (1893) from exposures in the Livingstone-Three Forks area of Montana. The name was derived from nearby Fort Ellis.

A type section for the Ellis Formation was designated and described by Cobban and others (1945) in Rocky Creek Canyon, Gallatin County, Montana. Cobban (1945) subsequently raised the Ellis Formation to group status. Three formations were recognized. In ascending order they are: the Sawtooth, Rierdon, and Swift.

Sawtooth Formation
Middle Jurassic

The Sawtooth Formation was described by Cobban (1945) at its type section along Rierdon Gulch in the Sawtooth Range of Montana as a basal fine-grained sandstone, a middle dark gray shale with intercalated thin dark limestone layers, and an upper calcareous siltstone.

The Sawtooth Formation was dated as Middle Jurassic by Imlay (1952) based on the presence of Defonticeras at the base and Arctocephalites and Procerites in the upper part of the formation. Ammonite zones above and below the formation date the Sawtooth Formation as early Callovian.

In the Dixon Mountain-Little Water Canyon area the Sawtooth Formation overlies the Thaynes Formation at a major disconformity. The Rierdon Formation conformably overlies the Sawtooth Formation. Eastward from the thesis area the Sawtooth-Rierdon contact is an unconformity. This erosional contact has created difficulties in formation correlation within the Ellis Group (Storey, 1958).

Distribution and Topographic Expression

The Sawtooth Formation is a slope-former in the thesis area. Much of the formation is covered by soil and small chips of rock. Limited outcrops are present where gullying has removed the soil cover.

The Sawtooth occurs on the north side of the south fork of Little Water Creek and at the base of the dip slope of the north-trending cuesta west of Dixon Mountain (see map). A very poorly exposed Sawtooth sequence occurs on the ridge top north of the mouth of Dry Creek Canyon, NE ¼, sec. 11, T.13 S., R.10 W.

Figure 26. Distribution of Jurassic age strata.

Fossils and Age

The Sawtooth Formation is locally fossiliferous, but within the thesis area few identifiable fossils were found. Moritz (1951) found numerous Gryphaea and Camptonectes pelecypods in the Sawtooth strata of southwestern Montana. A few coquina beds of the bivalve Ostrea have also been described. Imlay (1967) has found numerous pelecypod genera to be associated with a small number of gastropod genera, rare ammonites, and locally abundant Pentacrinus ossicles.

The presence of the bivalves Ostrea, Trigonia, and Gryphaea indicates a shallow marine environment, possibly the uppermost neritic zone. The bivalves are believed to have preferred a firm calcareous mud bottom, as most were suspension feeders living on or in the substrate (Imlay, 1967).

Within the thesis area, no intact fossils were collected; fragments from one fossiliferous zone are tentatively identified as a pectinoid bivalve, possibly Camptonectes.

The presence of the ammonite genera Defonticeras, Arctocephalites, and Procerites in association with the above described benthonic fossils indicate late Bajocian (early Middle Jurassic) age (Strickland, 1960; Imlay, 1967).

Lithology

<u>Outcrop Character</u> The Sawtooth Formation is rarely seen in outcrop within the thesis area. Most frequently, the presence of Sawtooth strata is determined by the occurrence of splinters and chips of light gray and yellowish brown limestone and calcareous siltstone downsection from the light brown oolitic rocks of the Rierdon Formation. Only one good exposure of Sawtooth strata was encountered within the thesis area. This section has been examined by Moritz (1951) and 110 feet (34 m.) of limestone and siltstone were measured in a shallow man-made trench (NE¼, SW¼, sec. 10, T. 13 S., R. 10 W.). This trench has apparently not survived the ensueing twenty-eight years, for a search of the area revealed only poorly exposed and discontinuous ledges.

The lower two thirds of the Sawtooth is composed of thin-bedded calcareous siltstone. These rocks are grayish orange (10 YR 7/4) or yellowish brown (10 YR 4/2 and 10 YR 4/4) on fresh surfaces and they weather to gray orange (10 YR 7/4) or pale orange (10 YR 8/2). The siltstone strata are structureless or show evidence of a mottled texture which suggests thorough bioturbation.

The poor outcrop quality of the Sawtooth, and of the Ellis Group in general, made a determination of the nature of formational contacts impossible. However, field relationships may be established by the first appearance of diagnostic rock types. As one moves upsection, the change from the gray limestones of the upper limestone member of the Triassic Thaynes Formation to yellowish brown calcareous siltstones may be taken as indicative of the presence of lower Sawtooth strata. The position of the Sawtooth-Rierdon contact may be inferred by the stratigraphically lowest appearance of a light brown oolitic limestone. This oolitic limestone is of such a distinctive color and texture that it is easily recognizable and therefore serves as a good marker for the presence of lowermost Rierdon strata.

Thin Section Description The calcareous siltstones of the Sawtooth Formation are the most ubiquitous rock type within the formation. Those siltstones examined petrographically are composed of approximately 80% medium silt-sized, angular to subangular quartz grains. These quartz grains are mixed homogeneously with coarse silt-sized subangular calcite. The nature of intergranular boundaries, the presence of quartz silt grains partly engulfed by calcite, and the presence of dusty impurities indicates that the quartz silt was originally in framework support with a micrite matrix which has since undergone aggrading neomorphism to pseudospar. An admixture of approximately 1% muscovite and 1% well rounded medium silt-sized zircon was also observed. An examination of insoluble residues derived from siltstones of the Sawtooth reveal quartz, muscovite, zircon and magnetite (Moritz, 1951). These minerals can not be taken as diagnostic of any particular provenance, although the muscovite suggests an acid igneous or metamorphic source. Moritz suggests they may indicate derivation from a rising island arc terrane to the west.

The limestones of the Sawtooth are fossiliferous packstones (Dunham, 1962). They are composed of a variety of allochems including: pelecypod fragments, pellets, and silty micrite intraclasts, all displaying well developed micrite coating or envelopes; coarse silt-sized, subangular to rounded quartz; Pentacrinus columnals and echinoderm plates and spines which have been partly replaced by authigenic chert; and gastropod and ostracod fragments. There is evidence of a variety of environmental factors coming into play before ultimate deposition of these limestones. Worn and coated bioclasts indicate particle movement and boring by filamentous blue green algae in shallow sunlit water less than 30 feet (9 m.) deep (Wilson, 1975). Gastropod and ostracod tests show no such evidence of abrasion or micritization and were probably deposited under much quieter conditions. The coexistence of a great deal of micrite mud and well worn bioclastic debris indicates textural inversion, as abraded particles from a high energy zone were ultimately transported and deposited in quiet waters.

Some diagenetic effects are also in evidence. The

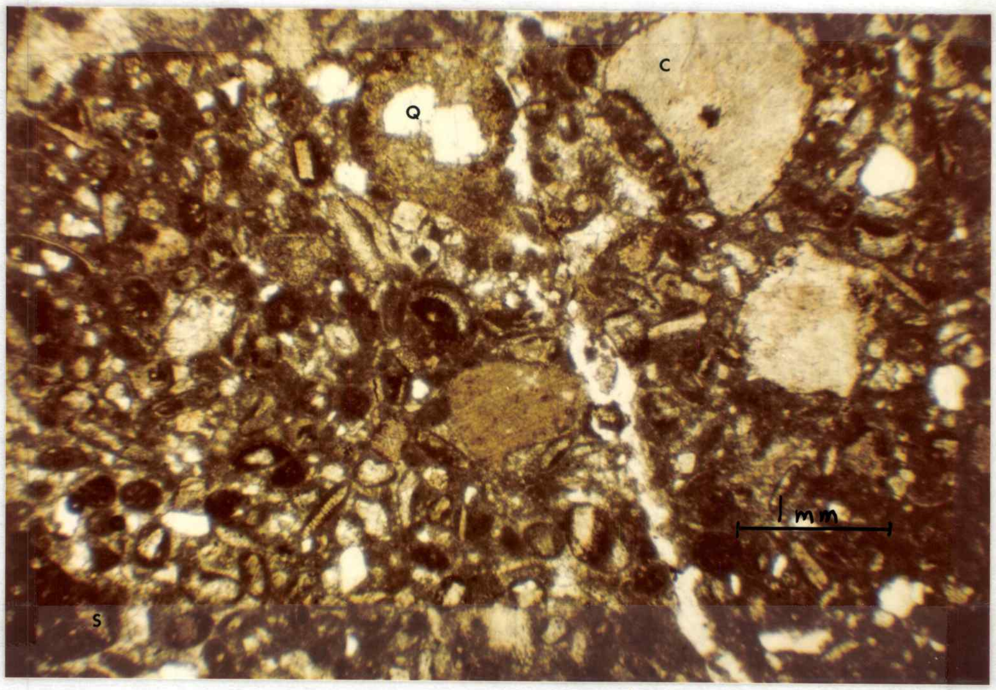

Figure 27. Photomicrograph of fossiliferous packstone, composed of crinoid oscicle (C), silty micrite, intraclasts (S), quartz silt (Q), and abraded molluscan and echinoderm debris with micrite coatings. Matrix is micrite. Sample RAK-335-11, Sawtooth Formation, north side of Little Water Creek.

molluscan fragments have undergone neomorphic inversion from the aragonite of the original shells to calcite spar. The micritization of some of the allochems has been discussed. Solution effects during early burial of the Sawtooth strata resulted in the development of moldic porosity within some of the echinoderm fragments. This porosity was subsequently filled by chert. The chert precipitation may have occurred during uplift of the strata and subaerial exposure. High formation temperatures would have dropped as a result of this uplift, resulting in an increase in calcite solubility and a decrease in silica solubility. Waters high in dissolved silica could then have precipitated chert in the porous echinoderm debris (Siever, 1959).

Depositional Environment

The fossiliferous packstones of the Sawtooth Formation offer the best evidence of the environment of deposition present in the Tendoy Range area during Bajocian-Bathonian time. Well rounded bioclastic grains with micrite rinds indicate a high energy setting within the upper photic zone (Wilson, 1975). A shallow shoal is the most probable situation where this would occur. Wave and tidal processes would keep bioclastic debris in constant agitation and continually bring additional biotic debris onto the shoal.

Occasional storm waves and tides would move this highly abraded material into quieter, deeper, protected waters behind the shoal. Once deposited in the lagoons, burrowing organisms would thoroughly mix this material with the micrite mud and remains of organisms indigenous to the lagoon (gastropods and ostracods). Bathurst (1975) notes the presence of a variety of micritized bioclastic grains within the modern lagoonal deposits of the Bahamas.

The processes responsible for packstone deposition provide some indication of the process of calcareous siltstone deposition. Whereas the fossiliferous packstones indicate shoreward transport of debris into a lagoonal environment, the quartz siltstones may represent seaward transport of minor terrigenous debris into a carbonate lagoonal setting. Fluvial processes may have moved considerable amounts of terrigenous sediment to the coastline. This material could have been reworked seaward during storm activity as the fine-grained clastic sediments of the shoreline and lagoon were put into suspension by strong waves and/or tidal action. This process could offer a partial explanation of the homogeneity of the silt-micrite mixture. Subsequent bioturbation could have completed the mixing of terrigenous and autochthonous carbonate muds.

Mixed clastic-carbonate environments are not common, however they are known to have existed. Selly (1978) describes a series of Miocene rocks in the Sirte basin of Libya which may offer an ancient analog. Terrigenous sediments were deposited at the mouths of low gradient rivers draining a low-lying hinterland. Onshore currents built carbonate bars seaward of the river mouths, thereby creating a mixed carbonate-clastic lagoon in the protected backbar setting. During the Middle Jurassic, western North America was within the subtropical high pressure zone and paleoclimatic conditions are believed to have been semiarid to subhumid (Peterson, 1978). Within this climatic belt, conditions would have been good for abundant carbonate production while terrigenous runoff would have been slow and stream loads would have been low.

Rierdon Formation
Middle Jurassic

Strata exposed along Rierdon Gulch in the Sawtooth Range of Montana were designated by Cobban (1945) as the type section of the Rierdon Formation. The 136 foot (41 m.) type section was described as: a basal calcareous shale with nodular limestone; a calcareous and noncalcareous shale with nodular limestone; and shale with a few interbedded limestone layers.

The use of the name Rierdon was extended from the type section into southwestern Montana by Moritz (1951). The basal unit of the Rierdon Formation in Montana was noted to be a prominent oolitic limestone overlain by argillaceous and sandy limestone.

In the Dixon Mountain-Little Water Canyon area the Rierdon Formation lies conformably upon the Sawtooth Formation. The Swift Formation Formation overlies the Rierdon Formation unconformably (Scholten and others, 1955).

Distribution and Topographic Expression

The Rierdon Formation is generally a slope-former, usually covered by soil or small rock fragments. The basal oolitic limestone unit locally supports ledges. Some good outcrops occur where gullying has removed the soil cover.

The Rierdon occurs on the north side of the south fork of Little Water Creek and immediately upslope and west of the drainage at the base of the dip slope of the north-trending cuesta west of Dixon Mountain (Plate 1). A limited, poorly exposed, lower Rierdon sequence occurs on the ridge top north of the mouth of Dry Creek Canyon, NE ¼, sec. 11, T.13 S., R.10 W.

Fossils and Age

A number of beds within the Rierdon Formation have been found to be fossiliferous in the thesis area. Imlay (1967) notes the presence of _Ostrea_, _Gryphaea_, _Mytilus_, _Echinotis_, and _Lingula_ as well as _Pentacrinus_ within Rierdon strata.

Imlay believes the community lived in water no deeper than the uppermost part of the neritic zone. The local occurrence of oyster banks at the base of the Rierdon reenforces this interpretation. The ammonite genera *Arcticoceras*, *Gowericeras*, *Kepperites*, *Gulielmiceras*, and *Procerites* are frequently found in association with the above described bottom community (Imlay, 1967). These ammonites indicate early to middle Callovian age (late Middle Jurassic).

Several pelecypod genera have been collected in the thesis area and tentively identified by the author as *Pleuromya*, *Meleagrinella*, *Trigonia* and *Gryphaea*. This assemblage suggests a littoral or shallow sublittoral environment (Wright, 1973).

Commonly the basal Rierdon is an oolitic limestone containing abundant oyster fragments. This facies is believed to have developed as offshore bars under conditions of relatively strong current and wave activity (Perry, 1979).

Lithology

Outcrop Character The presence of the Rierdon Formation may often be determined by the appearance of yellowish brown or light brown oolitic limestone. These oolitic strata are very easily recognized in outcrop exposures and, where present, as fragments on covered slopes. The Rierdon is commonly a slope-former, covered with a thin veneer of rock chips and soils. However, once the position of the Rierdon has been determined, close examination or shallow digging will reveal bedrock exposures.

The Rierdon has been measured and described by Moritz (1951) from exposures provided by a man-made trench (NE¼, SW¼, sec. 10, T. 13 S., R. 10 W.). This trench has not survived to the present time. A total of 109 feet (33 m.) of oolitic, argillaceous, or silty limestones and minor shales were found at this section.

An oolitic and, in part, fossiliferous limestone comprises much of the basal two-thirds of the Rierdon. These strata provide the most distinctive rocks of the formation. The rocks are pale yellowish brown (10 YR 6/2) when fresh and weather to the same color. The contained fossils are gastropods and pelecypods. A second oolitic unit lies within the upper third of the Rierdon. These rocks are light brown (5 YR 6/4) on a fresh surface and weather to very pale orange (10 YR 8/2) or pale yellowish brown (10 YR 6/2).

The remainder of the Rierdon Formation is composed of a variety of fossiliferous and unfossiliferous micritic limestones. These rocks range in fresh surface color from moderate and pale yellowish brown (10 YR 6/4 & 10 YR 6/2) to light brown (5 YR 6/6) and dark yellowish orange (5 YR 7/2). The weathered surfaces of these rocks are most commonly pale yellowish brown (10 YR 6/2), though some rocks are very pale orange (10 YR 8/2).

As noted above (Sawtooth Formation: Outcrop Character) the characteristics of formation boundaries are difficult to determine, because of poor outcrop quality. However, the readily recognized oolitic limestones do serve as an excellent marker for the Ellis Group in general and the Rierdon Formation in particular. The stratigraphically lowest occurrence of these oolitic limestones may be taken as the approximate location of the Rierdon-Sawtooth contact. The contact is considered to be conformable and marks a time-trangressive boundary developed in association with the onset of a late Bathonian-early Callovian transgression (Moritz, 1951; Hallam, 1975).

The Rierdon-Swift contact was not observed within the thesis area. The location of this contact may be inferred to be present by the first appearance of greenish gray glauconite sandstone and siltstone above the light brown to gray green limestones of the uppermost Rierdon. This contact is believed to be a disconformity representing a short regression during latest Callovian and earliest Oxfordian time (Hallam, 1975).

<u>Thin Section Description</u> Petrographic examination of Rierdon limestones reveal a number of fossiliferous oosparites and oomicrites, and fossiliferous and intraclastic pelmicrites (Folk, 1962).

The oolitic limestones of the Rierdon Formation are of two varities. The most common oolites are fossiliferous oosparites (Folk, 1962) or oolitic grainstones (Dunham, 1962). The ooids are composed of micrite accumulations around nuclei of either quartz silt or molluscan fragments. Concentric growth rings are indistinct and most ooids show evidence of algal boring. Bioclastic debris consists of bivalve and gastropod fragments, brachiopod fragments, echinoderm plates, and scaphopod tests. Several composite grains or grapestones of ooids also were observed. The ooids are moderately well sorted and the matrix is calcite spar.

Fossiliferous oolitic wackestones and packstones (Dunham, 1962) also occur within the Rierdon. The oolitic wackestone is composed of ooids which have undergone extensive micritization and abrasion. Echinoid spines and plates show evidence of varying degrees of micritization whereas gastropod tests are unworn and show little evidence of micrite coatings. A small number of brachiopod valves also were observed. These valves are much larger than the other allochems and have no micrite rims. The oolitic packstone examined in thin section is composed of micritized ooids which developed around quartz silt and biotic nuclei. These ooids are very closely packed. Intergranular boundaries are not tangential, as is normally the case, but rather, are concavo-convex, interfitting, or planar. It may be that micritization

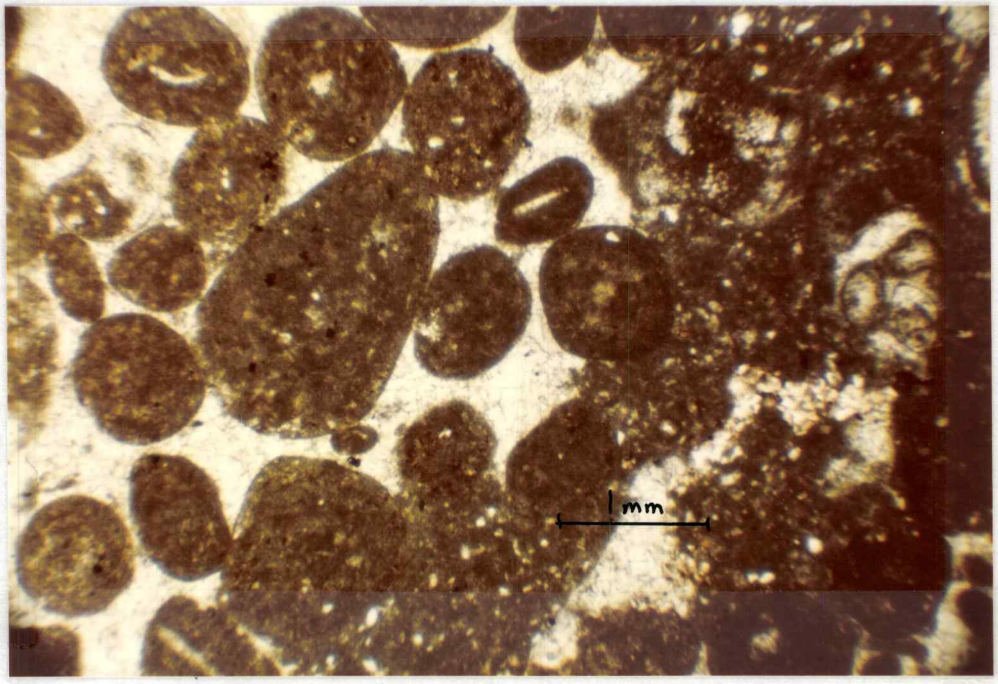

Figure 28. Photomicrograph of fossiliferous oolitic grainstone composed of silty micrite layers around a quartz silt or mollusc nucleus. Matrix is calcite spar. Sample RAK-152, Rierdon Formation, NE¼, SE¼, sec. 15.

has left the outer rim of these ooids so weakened that upon deposition, neighboring grain boundaries were deformed as grain stacking progressed. Alternatively, the ooids may have undergone intermittent solution during periodic sub-aerial exposure shortly after deposition. A small number of echinoid fragments also are present within this rock.

The question of ooid formation has had considerable attention. Chemical and physical conditions of modern ooid formation in the Bahamas have been investigated by Davies, et al. (1978) and Newell, et al. (1960). In general, oolites are associated with very shallow water conditions where evaporation reduces the CO_2 partial pressure and the water becomes supersaturated with respect to $CaCO_3$. Davies, et. al (1978) have described two types of ooids: those with a radial structure and those with tangential arrangment of aragonite needles. The lack of distinct concentric or radial layers within Rierdon ooids may be taken as indicative of shallow agitated conditions where aragonite needles were accumulated tangentially around a nucleus by Sorby's (1879) snowball effect.

Diagenesis within the oolitic limestones is confined to the neomorphic inversion of the aragonite of molluscan fragments to calcite spar. The circulation of $CaCO_3$-rich waters within the porous oolitic grainstones has resulted in the eogenetic cementation of the ooids. The presence of composite grains within the oolitic rocks indicates that cementation was occurring at or very near the sediment-water interface. The combined effects of depositional over-packing and eogenetic cementation of grainstones has left the oolitic limestones with very little porosity.

The pelletal packstone (Dunham, 1962) examined is composed of pellets of a uniform size range, angular to sub-angular quartz silt, and echinoderm and pelecypod fragments. These grains do not show any evidence of micrite rims. The matrix of this rock is micrite which has undergone neomorphic aggradation to microspar. The fossiliferous pelletal mud-stone (Dunham, 1962) examined is composed of a pelletal micrite matrix in which a variety of allochems are seen to be in matrix support. The allochems include well rounded, coarse and very coarse sand-sized, micrite intraclasts composed of quartz silt in a pelletal micrite matrix, gastropod and pelecypod fragments, echinoid plates and spines, and one disarticualted but entire pseudopunctate brachiopod valve.

The occurrence of pellets within these limestones is most commonly taken to imply an origin as fecal pellets derived from mud-ingesting organisms (Illing, 1954; Beales, 1958). However, Illing suggests that some pellet-like grains may form by progressive aggradation of lime mud by inorganic processes. The uniform shape and size of the Rierdon pellets, as well as the lack of composite grains of these pellets, suggest that these grains are probably of a fecal nature.

The pelletal limestones have undergone post-depositional neomorphic inversion of aragonitic molluscan fragments to sparry calcite neomorphic aggradation of micrite to microspar and, less commonly, pseudospar.

Depositional Environment

The lack of adequate outcrop exposures of Rierdon strata leaves only the evidence provided by fossils and petrographic and hand sample examination available for making paleoenvironmental interpretations.

The collection of bivalves taken from the Rierdon indicates a littoral or shallow sublittoral setting. The highly abraded nature of much of the fossil debris seen in thin section suggests that high energy wave or current conditions were prevalent.

The fossiliferous oolitic limestones of the Rierdon Formation indicate warm, clear, shallow, and agitated water conditions associated with local shoaling. This shoaling may have been either oolitic shoals, such as those developed in the Bahamas, or tidal bars subject to the constant motion of ebb and flood tides. Newell, et al. (1960) suggest that the optimal depth of ooid formation in the Bahamas is less than six feet (2 m.). Much of the oolitic and bioclastic material within this facies has well developed micrite rims and shows evidence of extensive abrasion. These worn and coated grains are indicative of extended residence in high energy setting within the upper photic zone (Wilson, 1975).

The pelletal and fossiliferous mudstones indicates deposition under much quieter conditions. The presence of abundant lime mud and well formed pellets indicates deposition within a zone below wave-base, with very slight water movement.

The association of winnowed, coated and worn bioclastic debris and oolites suggests high-energy shoals which were continually supplied with normal marine biotics by wave and current activity. The presence of abundant echinoderm material suggests neighboring waters must have been well oxygenated, with normal salinity and good circulation.

The oolitic and fossiliferous packstones and wackestones indicate the effects of two contrasting environments which were in close proximity. Worn and coated material within these rocks show evidence of wave action and abrasion in shallow agitated waters. These grains were then subsequently redeposited, possibly by storm activity, in much quieter waters. This type of textural inversion may have taken place in the bank or shoal interior facies or in a protected lagoon shoreward of a series of very shallow shoals or banks. The mixing of these two environments is also indicated by the mixture of worn and unworn bioclastic grains in a matrix support. These grains may have been deposited together and ultimately thoroughly homogenized by bioturbation.

The studies of modern carbonate bank environments have shown several well developed examples of oolitic and skeletal sand shoals developed at the edge of a platform with a lime mud or pelletal lime mud lithotope developed immediately shoreward. Two good examples of such settings would be the western margin of the Great Bahama Bank (Purdy, 1963; Newell, et al., 1959) and the Gulf of Batabano (Daetwyler and Kendall, 1959; Hoskins, 1964).

The rocks of the Rierdon may represent the lateral shifting and aggradation of wave, current, and/or tidally washed shoals or bars, quieter back-bar lagoons and the transitional facies between these two environments.

Swift Formation
Late Jurassic

The name Swift was applied by Cobban (1945) to 135 feet (41 m.) of dark noncalcareous shale and flaggy, glauconitic sandstone exposed near the Swift Resevoir, west of Dupuyer, Montana. Cobban noted lithologic and thickness changes along the strike of the formation. The Swift Formation in southern Montana is predominantly a calcareous glauconitic sandstone.

The use of the term Swift was extended eastward into Wyoming and South Dakota by Peterson (1954). Peterson noted an eastward thickening of the formation.

In the Dixon Mountain-Little Water Canyon area the Swift Formation overlies the Rierdon Formation disconformably. The overlying Morrison Formation is conformable upon the Swift Formation.

Distribution and Topographic Expression

The Swift Formation is a nonresistant slope-former in the thesis area. The formation is commonly covered by soil. Only one exposure of the Swift was encountered during field work: this outcrop is on the northeast side of a deep gully, north of the south fork of Little Water Creek, in the NE ¼, sec. 10, T.13 S., R.10 W.

The presence of glauconitic sandy weathering products was used in the field as an indicator of the location of the Swift. The Swift Formation is inferred to occur along the western side of the drainages developed at the base of the dip slope of the north-trending cuesta, west of Dixon Mountain (Plate 1). The occurrence of the Swift is also inferred to the north of the south fork of Little Water Creek.

Fossils and Age

The Swift Formation within the thesis area yielded no fossils. Regional studies of the Swift have yielded the ammonite genera _Cardioceras_, _Goliathiceras_, and _Pachycardioceras_ in association with as many as twelve bivalve genera. The presence of _Cardioceras_ within the Swift indicates an early Oxfordian (early Late Jurassic) age (Imlay, 1952).

The pelecypods found within the Swift suggest a shallow marine environment. The presence of _Mytilus_ within this assemblage restricts the setting to the littoral zone (Imlay, 1952). Hallam (1975) has examined sandstones of Oxfordian age in the western interior of the United States and found the trace fossils _Rhizocorallium_, _Teichichnus_, and _Thalassinoides_ to be common. These trace fossils are believed to be characteristic of the lagoonal to nearshore (infralittoral) setting (Chamberlaian, 1978).

Lithology

Outcrop Character The Swift Formation is seen in outcrop at only one location within the thesis area (NE¼, SW¼, sec. 10, T. 13 S, R. 10 W). This is approximately the same location at which Moritz (1951) measured and described 17 feet (5 m.) of Swift strata. Moritz characterized the Swift as a series of glauconitic sandstones and shales. Considerable emphasis has been placed upon the presence of glauconite as diagnostic of Swift strata (Moritz, 1951; Imlay, 1967).

Two glauconitic rocks are found within the Jurassic strata. Both of these rocks are highly fossiliferous and medium to coarse sand-sized glauconite grains are conspicuous. One of these rocks is a lithoclastic and bioclastic, moderately sorted, medium- to very coarse-grained sandstone which is yellowish gray (5 Y 8/1) on both the fresh and weathered surfaces. The second rock type is a coquina of thick shelled pelecypod valves (as large as 1 in.; 25 mm.) in a medium- to coarse-grained lithic sandstone matrix. This rock is yellowish gray (5 Y 8/1) on the fresh and weathered surfaces.

The determination of the nature of the upper and lower contacts of the Swift was never accomplished in the study area. The presence of Swift strata could occasionally be inferred where the yellowish brown and light brown rocks and rock-covered slopes of the Rierdon passed upsection to darker, gray green soils. The thinness of the formation has, in general, resulted in no surface expression of the strata. Regional studies indicate that the Rierdon-Swift contact is either disconformable or unconformable (Hallam, 1975).

The upper Swift contact also was never observed. This contact has been described as gradational (Moritz, 1951). The gray green and greenish brown sandstones and siltstones of the Swift are said to grade into overlying varicolored shales of the Morrison Formation (Moritz, 1951). This gradational contact represents the transition of sedimentation in southwestern Montana from marine to nonmarine (Hallam, 1975).

The location of Swift and Morrison strata may be inferred as those rocks which are immediately downsection from the basal pebble conglomerate of the Cretaceous Kootenai Formation.

Thin-Section Description The glauconitic rock of Jurassic age which was examined petrographically is a lithic, fossiliferous grainstone (Dunham, 1962). This rock is composed of a variety of lithoclasts, bioclasts, intraclasts, oolites and grapestones in framework support, with calcite spar filling intergranular pores.

The framework grains within this rock consist of very fine sand-sized chert, well rounded fine sand-sized, strained and unstrained quartz, medium sand-sized metamorphic rock fragments (phyllites?), coarse sand-sized calcareous siltstone grains, and rounded very fine sand-sized glauconite. The bioclastic material within this rock is assorted molluscan debris which has undergone neomorphic inversion to calcite, crinoid columnals, echinoderm plates and spines, brachiopod valve fragments and rounded fragments of belemnite rostra. A small number (5%) of the echinoderm plates have all or part of the reticulate pore pattern filled or partly replaced by glauconite. There are also present a few ooids and grapestone grains composed of cemented ooids.

Though this rock type can not be considered typical of the Swift Formation, two components of this rock are considered to be unique to the early Oxfordian of the Western Interior

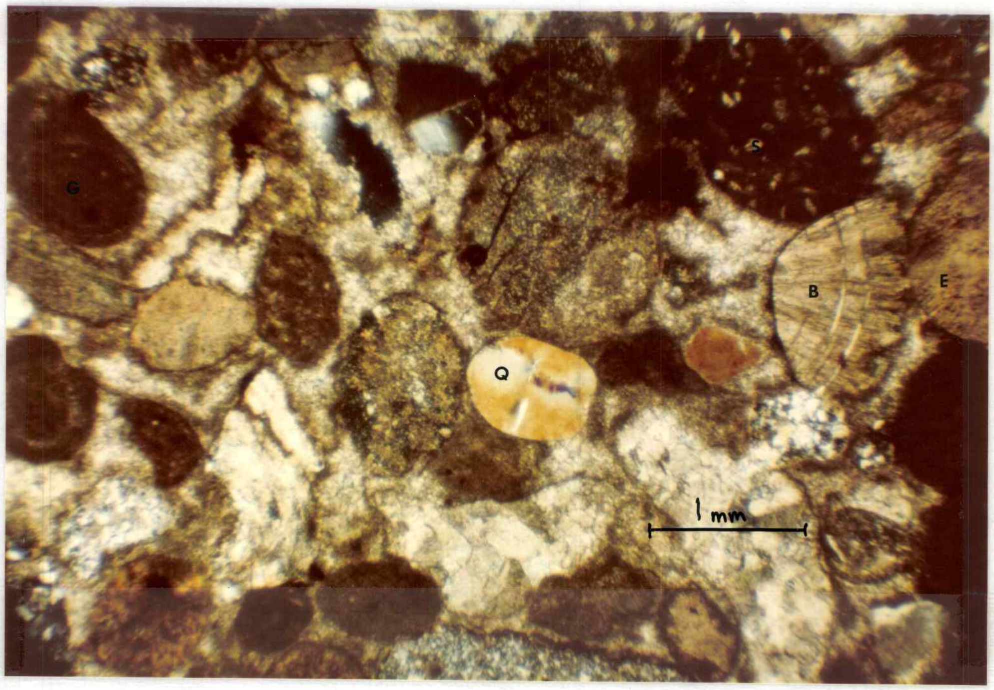

Figure 29. Photomicrograph of bio-lithoclastic grainstone. Well-rounded grains of quartz (Q), calcareous siltstone (S) and belemnite rostrum (B). Note rounded grapestone grain (G) and echinoderm fragment with pores partially filled with glauconite (E). Crossed nicols. Sample RAK-335-1, Swift Formation, north side of Little Water Creek.

seas. The abundance of glauconite is considered by Hallam (1975) to be characteristic of early Oxfordian strata and is known to be widespread throughout much of Montana, Wyoming, Idaho, South Dakota, Utah, and New Mexico. In addition, Hallam (1975) and Imlay (1967) have noted a marked increase in the abundance and diversity of early Oxfordian belemnites over earlier marine Jurassic Strata. The appearance of glauconite and belemnites in combination may be taken as a reliable indication that the rock under consideration is of the Swift Formation.

The glauconite present within the Swift indicates very early diagenesis of iron-rich micaceous muds on a shallow marine continental shelf where there was minor agitation, normal salinity, and slightly reducing conditions (Pettijohn, 1975; Reineck and Singh, 1975). The formation of glauconite is facilitated by the presence of organic material. This organic material may be instrumental in the generation of a reducing microenvironment near the sediment-water interface within a generally well circulated, oxidizing environment. Commonly, glauconite authigenesis takes place where there is little or no detrital influx (Heckel, 1972). The presence of glauconite in the pores of worn and abraded echinoderm fragments is considered to be indicative of relict sediment alteration (Milliman, 1974).

Depositional Environment

The compelling evidence of transportation of the grains of the Swift limestones precludes the use of fossils as an accurate paleoenvironmental indicator. The presence of such a great variety of lithologies, degrees of weathering, and sizes of grains suggest that the components of these rocks have originated from several different depositional environments.

The bioclastic grains are those or shallow water pelecypods and gastropods, brachiopods, echinoderms, and crinoids. Some of the large fragments are those of thick-shelled animals believed to be typical of high energy shoreline or shoal environments. The presence of ooids, grapestones composed of ooids, and intraclasts of calcareous siltstone provide additional evidence of a shallow, well agitated setting. The nektonic belemnites could have entered the depositional process at any point and provide no specific environmental indications.

The variety of lithoclasts present within the Swift indicate derivation from a number of sources. Chert grains are the most prominent component of the lithoclastic fraction and represent the erosional products of subaerially exposed sedimentary strata. These source strata may have been the Permian Phosphoria which is known to have been exposed during the Late Jurassic and contains a number of chert layers (Hallam, 1975). The well rounded quartz of these rocks is presumably the recycled quartz sand from exposed Paleozoic sandstone strata. This reworking of sedimentary rocks also is indicated by a variety of well rounded siltstone clasts. A distinctly different source of sediment supply is suggested by the presence of a small number of metamorphic rock grains. These may have been derived from uplifted Precambrian basement to the west. A number of fluvial dominated deltas are known to have existed along the western margin of the Oxfordian sea of the western interior (Brenner & Davis, 1974). These deltas were presumably responsible for the delivery of the siliciclastic sediments to the Swift basin.

The presence of glauconite indicates that a third sedimentary environment contributed material to the Swift limestones. The conditions for glauconite authigenesis have been discussed in the previous section. It is noteworthy that glauconite forms under agitated conditions where sediment accumulation rates are slow or nonexistent. It may be that, though a number of sedimentary environments were contributing sediments to the Swift basin, the accumulation rates were slow and sedimentary particles spent prolonged periods of time at the sediment-water interface.

Two contrasting explanations can be provided for the mechanism of aggregating sedimentary particles from a number of different environments into a high energy setting where fine-grained material was winnowed away.

The abundance of well worn bioclastic debris associated with ooids, intraclasts, and grapestones suggest a shoal or bar setting. Longshore currents may have brought in a variety of lithic grains from the reworking of nearby fluvio-deltaic deposits. Onshore wave and tidal processes could have moved glauconitic sediments toward the shoreline and contributed to the piling up of sediments on a shoaling nearshore platform. These offshore bars may have migrated shoreward as they developed, aggraded upward, or remained stationary. If they did migrate shoreward or aggrade upward it is possible that they eventually become emergent as barrier islands (Otros, 1970; Hoyt, 1967).

Alternately, the coquinoid sandstones of the Swift may have accumulated as sand waves on the shallow floor of the Oxfordian sea. These sand waves may have developed in response to tidal currents, storm wave activity, or oceanic currents. Tidal activity in the interior seas of the Jurassic is considered to have been minimal (Wilson, 1975). It may be that oceanic currents provided much of the energy to accumulate and winnow the coquinoid sands of the sand waves. Longshore currents may have combined with nearshore wind wave processes to generate a net seaward transport of sediments derived from fluvio-deltaic siliciclastic and interdeltaic bioclastic sources.

Selley (1976) has described tidal current sand bodies in the North Sea which are composed of winnowed, fine- to medium-grained, fairly well sorted, glauconitic, bioclastic sand. These deposits may provide a reasonable modern analog for the coquinoid sands of the Swift Formation.

The fossiliferous shales, orthoquartzitic, and sandy biosparites of the Oxfordian of the western interior of the United States have been investigated in some detail by Brenner and Davies (1973, 1974). These workers characterize the regressive phase (upper Swift) of the Oxfordian of Montana as a series of marine bar sands composed of sandy glauconitic coquina interfingering with siliciclastic sandstones and shales. Brenner and Davies (1973) believe that storm wave turbulence was responsible for the transportation of much of the biotic material (which ranges up to 60 mm.). Much of the coquinoid sands is interpreted as storm lag deposits produced as storm activity moving obliquely to sand bar trends incised channels across these bars. The coarse bioclastic and siliciclastic material would have been deposited as thin sheets and fans leeward of the sand ridges. Constant bottom current activity would have kept these deposits well winnowed as the debris lay at the sediment-water interface for extended periods of time.

The work of Brenner and Davis in better exposed and somewhat thicker Oxfordian sequences provides a workable explanation for the deposition of clean, mixed clastic-carbonate sediments. A determination of the conditions responsible for the deposition of the Swift limestones of the thesis area may never be possible because of the lack of adequate exposure. More information about facies relationships will be necessary before a well documented interpretation can be made.

Morrison Formation
Late Jurassic

The Morrison Formation was first described by Eldridge (1896) at an incomplete section exposed north of Morrison, Colorado. However, Cross (1894) was the first to publish the name in a folio of the Pikes Peak Quadrangle. No type section was designated by either Eldridge or Cross.

The Morrison Formation was redefined by Lee (1920). Some of the originally defined Morrison was reassigned to the Dakota and Sundance Formations. Lee (1927) reexamined his 1920 work and revised some of the thicknesses.

A type section was designated at Morrison and the type description was revised by Waldschmidt and LeRoy (1944). The 276 foot (84 m.) type section was subdivided into six lithologic units. In this new description the Morrison Formation was defined as strata lying between the Jurassic Sundance Formation, as defined by Lee (1920), and the Cretaceous Dakota Formation.

Cobban, Imlay, and Reeside (1945) restricted the Morrison Formation at the type section to 110 feet (34 m.) and assigned that part below this new section to the Ellis Group.

In the Dixon Mountain-Little Water Canyon area the Morrison Formation is poorly exposed, and the contacts cannot be observed in outcrop. Scholten and others (1955) indicate that the Morrison Formation is 71 feet (22 m.) thick in the Little Water Canyon. Greenish soils stratigraphically above the Ellis Group have been used as an indicator of the formation's presence.

In the Tendoy Range the Morrison Formation lies conformably upon the Swift Formation (Moritz, 1951). The Cretaceous Kootenai Formation overlies the Morrison Formation disconformably (Suttner, 1969).

Distribution and Topographic Expression

The nonresistant shales and siltstones of the Morrison Formation are not observed in outcrop in the thesis area. Regional studies, however, suggest its presence and locally, greenish soils have been taken as marking its occurrence.

The Morrison is inferred to be present stratigraphically below the highly resistant "salt and pepper" sandstones and conglomerates of the overlying Kootenai Formation.

Fossils and Age

The Morrison Formation within the thesis area is unfossiliferous. In a regional study of the Morrison in Montana and Wyoming, Yen (1951) collected specimens of the bivalves _Unio_, _Vetulonaria_, and _Hadridon_ as well as the gastropods _Valuata_, _Viviparus_, _Physa_, _Gyraulus_, and _Lymnaea_. In a subsequent study of Morrison microfossils, Peck (1957) collected a variety of ostracods and the charophyte oogonia _Clavator_ and _Latoehara_ from lacustrine limestone lenses. These charophytes indicate Kimmeridgian and Purbeckian ages (Late Jurassic), respectively, for the Morrison Formation. These age determinations are in agreement with those reached by Yen based on his macrofossil collection.

Lithology

Outcrop Character The Morrison Formation is poorly exposed in the thesis area. No outcrops of Morrison strata were encountered during the field investigation. Moritz (1951) describes "relatively good exposures" of the Morrison in the Little Water Canyon area and indicates that 72 feet (22 m.) are present in that area. These rocks are described as gray green, greenish brown and yellow siltstones and minor sandstones.

The presence of Morrison strata was rarely detectable in the field. Generally the unit is a non-resistant slope-former, of low relief. Locally, greenish soils, immediately downsection from the basal pebble conglomerate of the Kootenai

Formation were useful in locating the Morrison.

The Morrison conformably overlies the Swift Formation. The contact between these two units is gradational (Moritz, 1951). The contact with the overlying Cretaceous Kootenai Formation is an unconformity which span Purbeckian and Neocomian (latest Jurassic and earliest Cretaceous) time (Suttner, 1969). The upper Morrison contact can be reliably placed at the base of the pebble conglomerates, pebbly sandstones, and sandstones of the basal Kootenai.

The Morrison has been well studied by a number of workers. In the southwestern Montana area, Suttner (1966, 1969) presents a study of the regional distribution, petrography and paleotectonic implications of the nonmarine Morrison and Kootenai strata in the southwestern Montana region.

Kootenai Formation
Cretaceous

It is rather difficult to determine to whom the naming of the Kootenie Group should be attributed. The Lexicon of Geologic Names indicates that Sir William Dawson first publicly used the name. However, his source of information appears to have been field work performed by G.M. Dawson (Dawson, 1886). Furthermore, the Lexicon citation is from a paper written by J.W. Dawson (Dawson, 1885). No type section was designated by any of the Dawsons, but the name Kootenie was applied to a sequence of sandstone, shale, conglomerate, and coal beds exposed along the Old Man River and Martin and Coal Creeks in the Bow Valley area of Alberta.

The name Kootenie was extended into the Great Falls, Montana area by Weed (1892). At this location the name was applied to a series of coal-bearing rocks which are underlain by marine Jurassic strata. Subsequently, Weed (1899) discarded the name Kootenie for the strata described in 1892 in favor of the term Cascade Formation.

Fisher (1908) applied the name Kootenai, in its presently accepted spelling for central and southern Montana, to Early Cretaceous coal-bearing rocks in the Belt, Montana area. This new name was to replace the name Cascade suggested by Weed (1899) and apply to the same rocks. Subsequent work by Douglas (1909), Calkins and Emmons (1915), and Pardee (1917), extended use of the term Kootenai to Early Cretaceous rocks of southwestern Montana.

Plant fossils collections taken by Brown (1946) contain no specimens of Purbeckian and Neocomian age. This evidence supports the hypothesis that an unconformity exists at the base of the Kootenai Formation in Montana.

The "salt and pepper" sandstone (Cut Bank sand) was identified by Cobban (1955) as the basal unit of the Kootenai Formation. This basal sandstone was observed to be lying unconformably upon Jurassic Ellis and Morrison strata in the Little Belt Mountains-Sweetwater Arch area of Montana. Cobban noted that the Kootenai Formation thickens to the southwest.

In the Dixon Mountain-Little Water Canyon area the Kootenai Formation overlies the Morrison Formation unconformably (Moritz, 1951; Scholten, et al., 1955). The contact with the overlying Aspen Formation is not observed in the thesis area but is described as gradational where present. Within the study area the Kootenai is unconformably overlain by a variety of Late Cretaceous and Tertiary deposits.

Distribution and Topographic Expression

The variety of rock types within the Kootenai Formation and their varied responses to erosion have resulted in a diversity of topographic forms. The basal "salt and pepper" sandstone and chert pebble conglomerates are a persistent ledge- and outcrop-former wherever present. The sandstone and limestone beds and lenses within the formation, as well as a gastropod coquina at the top, support ledges and ridges wherever present. However, most of the Kootenai consists of nonresistant mudstones which are generally soil covered slope-formers.

The Kootenai occurs throughout much of the upper reaches of the Little Water drainage system. Much of sec.

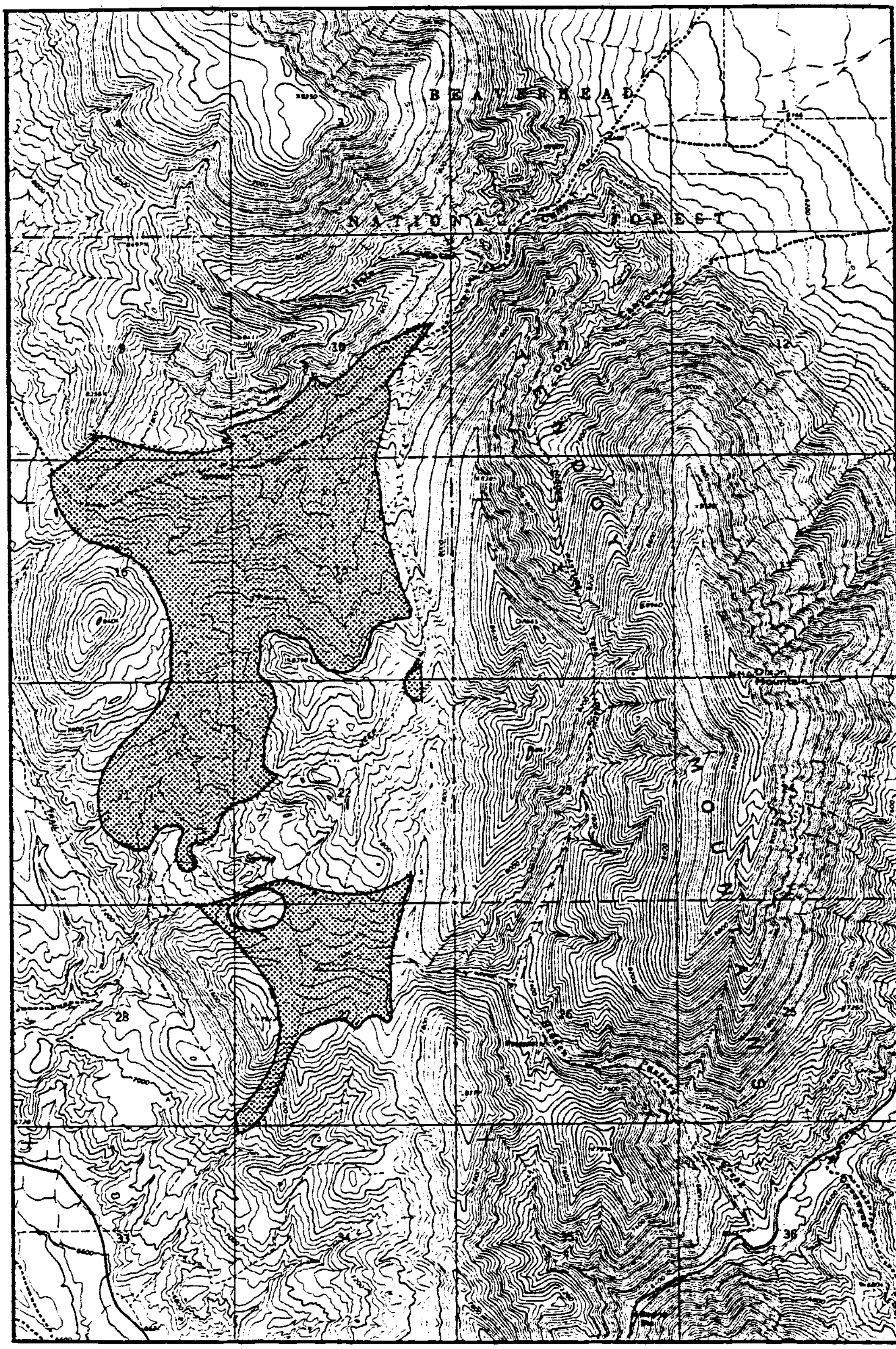

Figure 30. Distribution of Kootenai Formation.

15 and 16, and parts of sec. 21 and 27, T.13 S., R.10 W., are Kootenai exposures and soils. The reddish coloration of much of the mudstones and shales is a useful indicator for the presence of the formation.

Fossils and Age

The Kootenai Formation has two fossiliferous members: a lower calcareous member and an upper gastropod limestone member. The dominantly non-marine nature of the Kootenai has left much of the formation unfossiliferous.

The lower calcareous member of the Kootenai is predominantly a biomicrite with abundant articulated and disarticulated ostracods and numerous gastropods, charophytes, and possible pelecypods (Holm, et al., 1977).

The upper gastropod limestone is composed of biomicrites containing a variety of spiraled nonmarine gastropods or ostracods and oomicrites and algal limestones (Holm et al., 1977). The pelecypods *Eupera onestae*, *Protelleptio douglassi*, and *Unio farri* are known to occur throughout the upper Kootenai limestone (Suttner, 1969).

Floral remains from the lower Kootenai and uppermost Morrison have been collected and examined by Brown (1946). The lack of Purbeckian and Neocomian age plant remains suggests an interruption in deposition and supports the presence of a disconformity between the Kootenai and the underlying

Morrison Formation.

The lower calcareous member of the Kootenai has been interpreted by Holm, et al. (1977) to represent lacustrine deposition within a series of foreland basins. The gastropod limestones have been interpreted by the same workers as representing deposition within lakes that were intermittently connected with marine waters.

The age of the Kootenai has been difficult to determine accurately. The presence of Eupera onestae within the gastropod limestone firmly establishes an Aptian age for these strata (Katich, 1951). However, the remainder of the formation lies below that uppermost member and no reliable ages have been assigned to the older units. It is therefore possible, as Suttner (1969) conjectures, that the Kootenai may span the Neocomian and possibly the latest Jurassic.

Within the thesis area high-spiraled gastropods within the gastropod limestone have been tentatively identified as Circamelania.

Lithology

Outcrop Character Parts of the Kootenai Formation are well exposed within the thesis area. The resistant sandstone, conglomerate, and limestone units within the Kootenai generally support low-lying ledges. Much of the intervening strata is non-resistant calcareous siltstones and mudstones which are rarely seen in outcrop but impart a pale reddish brown color to much of the covering soils.

Approximately 1,000 feet (305 m.) of Kootenai strata are believed to be present in the thesis area (Suttner, 1969). This sequence may be subdivided into— a basal unit of chert pebbles conglomerates, pebbly sandstones, and coarse-grained sandstones, a middle section of interbedded calcareous mudstones and siltstones with local lenses of limestones, and a thin upper gastropod limestone unit.

The Kootenai overlies the Morrison Formation unconformably. This unconformity represents a period of non-deposition and erosion during the Purbeckian and Neocomian time (Suttner, 1969). The highly resistant, ledge-forming conglomerates at the base of the Kootenai overlie non-resistant Morrison strata.

The basal conglomerate of the Kootenai is a persistent ledge-former and is accepted as indicative of lowermost Kootenai strata wherever present. In the thesis area this unit is approximately 20 feet (6 m.) thick. The conglomerates of this unit are very thick-bedded sandy pebble conglomerates with the pebbles either well rounded chert or, much less commonly, subangular calcareous mudstone and siltstone. The matrix of these rocks is a medium- to coarse-grained sandstone which is light brown (5 YR 6/4) on fresh and weathered surfaces. The conglomeratic units grade upsection over a three foot (1 m.) interval to trough cross-bedded, medium-grained pebbly sandstone. These sandstones commonly have a "salt-and-pepper" appearance. The sandstones consist of well rounded black chert pebbles and sand grains, in a light gray (N 7) matrix. These rocks weather to grayish red (5 R 4/2).

The middle part of the Kootenai is composed of a variety of rock types which are only locally persistent and commonly are lensoid. The interbedded sandstones and limestones of this unit are the only ledge-formers. These strata are very thin-bedded, medium-grained calcareous sandstones composed of subangular quartz and chert grains. The sandstones are grayish orange pink (5 YR 7/2) on the fresh and weathered surface. The limestones associated with these sandstones are thin-bedded, silty, sparsely fossiliferous micritic rocks that are light gray (N 7) on a fresh surface and weather to medium light gray (N 6) and light brownish gray (5 YR 6/1). Where present these limestones are never more than three feet (1 m.) thick but are easily traceable in outcrop and invariably pinch out along strike. Most (70%) of the middle part of the Kootenai is composed of non-resistant calcareous siltstones and mudstones. The siltstones are thin-bedded and unfossiliferous. These rocks are a variety of colors including moderate pink (5 R 7/4) strata which weather to pale reddish brown (10 R 5/4), grayish red purple

(5 RP 5/2) rocks which weather to pale red (5 R 6/2) and rocks which are very light gray (N 8) on fresh and weathered surfaces.

The mudstones are only seen in outcrop where stream gullying has removed the soil cover. These rocks are very thin- and thin-bedded and generally silty. Here again, a variety of colors were seen including moderate red orange (10 R 6/6), yellowish gray (5 Y 7/2), grayish red purple (5 RP 4/2) and light bluish gray (5 B 7/1). These rocks commonly weather to moderate reddish orange (10 R 6/6) and olive gray (5 Y 4/1). One small gully in the study area exposed dipping mudstone strata with well developed asymmetrical ripple marks on several of the bedding planes. These ripples indicate paleocurrent directions of S60W, S30W, and S40E.

The uppermost part of the Kootenai is an easily recognized series of thin-bedded gastropod limestones. Many of these beds are composed entirely of high-spiraled gastropods in framework support. A small number of interbedded limestones contain an admixture of fragmented pelecypod debris. The gastropod limestones are light gray (N 7) to light brownish gray (5 YR 6/1) on fresh surfaces and weather to yellowish gray (5 Y 8/1) and light brownish gray (5 YR 6/1). These fossiliferous limestones are underlain by light brown (5 YR 6/4) weathering, thin-bedded, fine- to coarse-grained

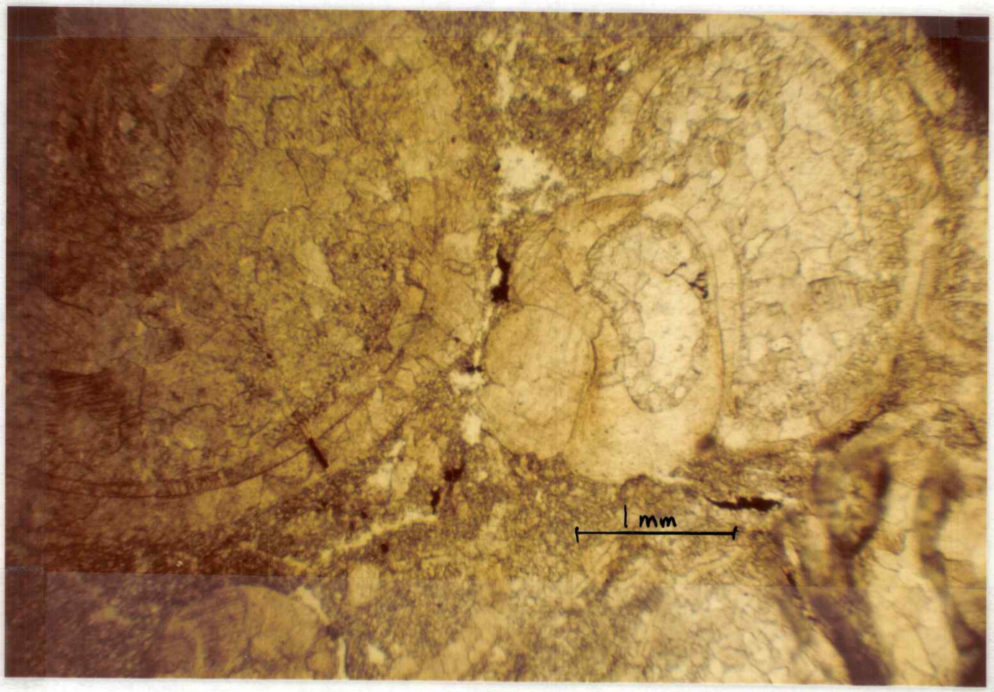

Figure 31. Photomicrograph of gastropod grainstone. Whole gastropod tests in framework support. Tests have inverted to calcite spar, intraparticle porosity is filled by calcite spar. Matrix is micrite neomorphosed to microspar. Sample RAK-108, Kootenai Formation, south side of Little Water Creek.

cherty sandstones which have numerous beds displaying planar cross-stratification and channel scour-and-fill structures.

The gastropod limestones of the uppermost Kootenai are the youngest strata involved in extensive folding in the study area. All younger strata lie upon strata of Kootenai age and older with marked angular discordance.

The Kootenai and underlying Morrison have been examined in considerable detail by Suttner in a Ph.D. thesis (1966) and in subsequent publications (1968; 1969). The age, distribution, petrography, provenance, clay mineralogy, and paleotectonic implications of the clastic rocks of these two formations are well described. It is Suttner's thesis that these two formations represent the retreat of Jurassic seas from the western interior and the eastward advance of the clastic wedge derived from initial Laramide pulses in central Idaho. Readers interested in further information on this subject should consult Suttner's work.

Beaverhead Formation
Cretaceous-Tertiary

The Beaverhead Formation was formally named by Lowell and Klepper (1953) for a sequence of conglomerates, sandstones, siltstones, and freshwater limestones first described by Eardley (1950). At the 9,700 foot (2,959 m.) type section in McKnight Canyon, west of Dell, Montana, an incomplete sequence (the upper and lower contacts are faulted out) was measured and described. The age of this sequence was believed to be Late Cretaceous to Early Eocene.

Ryder (1967) undertook a regional study of the stratigraphy of the Beaverhead Formation as part of his doctoral work. Ryder was able to show the intertonguing nature of the thick deposits of conglomerate. Palynological data suggested a Late Cretaceous to Paleocene age.

In subsequent studies, Ryder and Ames (1970) and Ryder and Scholten (1973) employed the age and stratigraphy of the Beaverhead Formation to interpret and date many of the tectonic events that affected the southwestern Montana region.

In the Dixon Mountain-Little Water Canyon area the Beaverhead Formation overlies Mesozoic strata unconformably. Elsewhere in the region Paleozoic and Mesozoic strata lie below an unconformity with discordance as great as 25°. The Beaverhead Formation is not overlain by any other rocks in the thesis area. Elsewhere in the region, Eocene and Oligocene sedimentary rocks and volcanics overlie the formation unconformably.

Distribution and Topographic Expression

The cobble and pebble conglomerates of the Beaverhead Formation are prominent cliff-, ledge-, and hill-formers throughout much of the region.

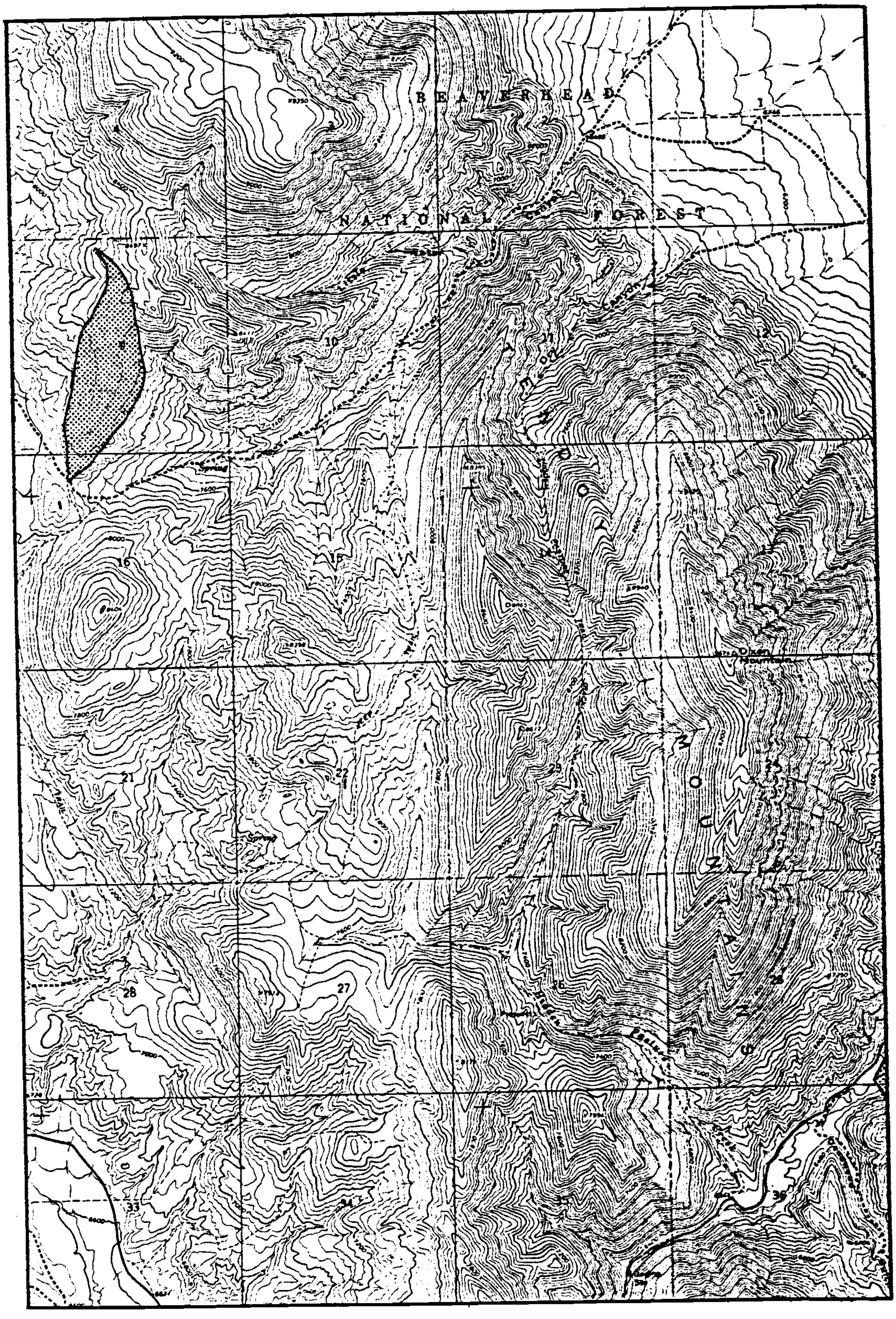

Figure 32. Distribution of Beaverhead Formation.

Only one large outcrop of the Beaverhead occurs within the thesis area. This cuesta, on the western margin of the area, sec. 9, T.13 S., R.10 W., stands above much of the surrounding terrain. Quartzite and limestone conglomerates form the ridge crest with abundant talus on a gentle slope below.

A second exposure of the Beaverhead is present to the east of the Tendoy fault, NE ¼, sec. 36, T.13 S., R.10 W. This exposure is part of an areally extensive occurrence mainly lying east of the thesis area.

Fossils and Age

No fossils were collected from the Beaverhead conglomerates within the thesis area. Ryder (1967, Ryder and Scholten, 1973) examined the Beaverhead in considerable detail and collected a variety of palynomorphs from three localities. These palynomorphs indicate ages of late Aptian through Latest Cretaceous. A considerable thickness of the Beaverhead overlies the highest dated beds and it has been speculated that Beaverhead deposition may have continued into the Paleocene or perhaps the early Eocene.

Lithology

Outcrop Character The Beaverhead conglomerates are very widespread throughout the region surrounding the thesis area. Stratigraphic relationships and thicknesses of the ten lithologies present within the formation are uncertain because of post-depositional tectonic events (Ryder, 1967; Ryder and Scholten, 1973). It is believed that at least 15,000 ft. (4,575 m.) of Beaverhead conglomerate is present within the region.

Within the thesis area only two exposures of the Beaverhead were found. Of these, only one was well enough exposed to permit adequate examination (sec. 9, T.13 S., R.10 W.). Two different conglomerates are present. The lower unit is a poorly sorted pebble-boulder conglomerate (clasts 0.4 in to 5 ft.; 1 cm. to 1.5 m.) with a poorly sorted coarse-grained sandstone matrix. This conglomerate displays evidence of rather crude, large scale, foreset cross-bedding. Clasts within this unit are predominantly rounded quartzite boulders associated with a variety of sandstone, shale, and limestone cobbles and pebbles.

Overlying the quartzite conglomerate is a very thick-bedded (16 to 23 ft.; 5 to 7 m.), poorly sorted, moderately rounded, limestone conglomerate. Clasts are composed almost exclusively of fossiliferous Paleozoic and Mesozoic limestones similar to those seen elsewhere in the thesis area within their respective formations.

These two contrasting conglomerates are angularly discordant with respect to one another and are separated by a very fine-grained pebbly sandstone which is moderate reddish orange (10 R 6/6), and a friable calcareous, medium- to fine-grained sandstone which is reddish brown (10 R 5/6). The conglomerates lie unconformably upon underlying Jurassic and Cretaceous strata.

This particular outcrop was identified by Ryder (1967) as part of the Divide Limestone Conglomerate unit. The Divide Limestone Conglomerate is characterized by a predominance of Triassic and Jurassic limestone clasts. Subsequently, Ryder and Scholten (1973) determined this outcrop to be part of the McKnight Limestone Conglomerate unit which typically is composed of Mississippian and Triassic limestone with some Belt quartzite clasts.

It may be that parts of the two units defined by Ryder (1967) are present in the outcrop under consideration. The multicolored quartzite clasts of the lower conglomerate unit may be part of the Divide Quartzite Conglomerate or the Kidd Quartzite Conglomerate. The overlying limestone conglomerate is, probably, part of the McKnight Limestone Conglomerate unit. These two lithologic units are known to intertonque elsewhere in the region and both are known to be present in McKnight Canyon, about 3 miles to the northeast (Ryder & Scholten, 1973).

Medicine Lodge volcanics
Tertiary

Scholten and others (1955) applied the informal name, Medicine Lodge volcanics, to basaltic and andesitic lavas, tuffs, and agglomerates in the Tendoy Range. These rocks have been correlated to rocks of similar composition and age distributed as far westward as the northern slopes of the Beaverhead Range. Sedimentary rocks associated with these volcanics suggest a late Oligocene or early Miocene age.

In the Dixon Mountain-Little Water Canyon area the Medicine Lodge volcanics lie unconformably upon Mesozoic strata. The volcanics lie unconformably upon Mesozoic strata. The volcanics are overlain by unconsolidated late Tertiary and Quaternary gravels and sediments.

Distribution and Topographic Expression

The flows, breccias, tuffs, and agglomerates of the Medicine Lodge volcanics are, in general, soil covered hill-formers with isolated outcrops of resistant flow rocks.

The best exposures and the areally most extensive occurrences of the Medicine Lodge volcanics are in the western half of sec. 22, T.13 S., R.10 W. At this location, banded flow rocks and agglomerates have been eroded into ledges, ridges, and isolated pinnacles. In this same area some tuffs have been exposed.

A second, less well exposed, occurrence of the volcanics is located in the SE ¼, sec. 27, T.13 S., R.10 W. At this location, flow rocks form a hill with some limited outcrops on the flanks.

Lithology

Description The informal term Medicine Lodge volcanics has been applied to Scholten et al. (1955) to a variety of andesitic and basaltic flows, tuffs, and volcanic breccias found in the Tendoy and Beaverhead Ranges and the intervening Muddy Creek, Big Sheep Creek, and Medicine Lodge Valleys. Within the region, these volcanics, or volcanics believed to be correlative in age, are underlain by sedi-

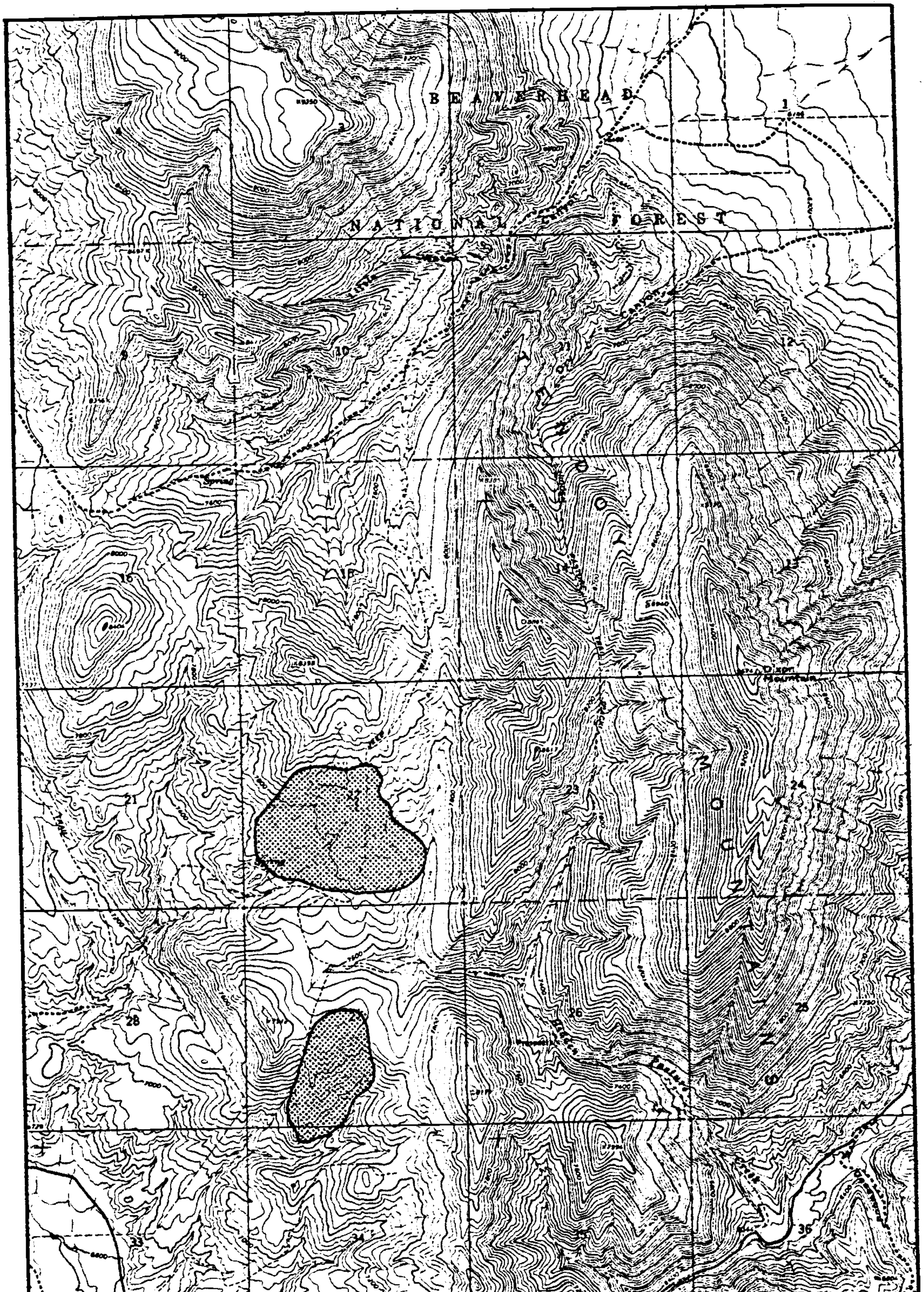

Figure 33. Distribution of Medicine Lodge Volcanics.

mentary strata and tuffs thought to be Oligocene in age and are overlain by sediments of probable Miocene age (Scholten, et al., 1955).

The Medicine Lodge volcanics are exposed at two locations within the thesis area. The similarity of rock type and proximity suggest these two exposures are genetically and temporally related. The best exposed and most easily examined series of volcanics includes a central semi-circular area of volcanic breccia and flows surrounded by additional flows and minor tuffs.

In hand sample the flow rocks are platy, aphanitic, locally vescicular and scoriaceous andesites which are dark to medium gray (N 3 to N 5) or olive gray (5 Y 4/1) on fresh surfaces. These rocks commonly weather to grayish red (10 R 4/2) or less commonly, pale red (5 R 6/2) and olive gray (5 Y 4/1).

In thin section the volcanic flow rocks are composed of 15% phenocrysts in a matrix of plagioclase microlites and devitrified glass. The phenocrysts are predominantly subhedral to anhedral clinopyroxene with a minor amount (10%) of orthopyroxene. The groundmass is laths of plagioclase arranged in a pilotaxitic texture. The plagioclase microlites are either andesine or labradorite with an An of 45 to 52. This range suggests these volcanics may best be termed basaltic andesites.

Figure 34. Curviplanar parting developed in basaltic andesite flows of the Medicine Lodge volcanics. Ragged pinnacles at the sky-line are composed of volcanic breccia. Exposed near the center sec. 22, T. 13 W., R. 10 S.

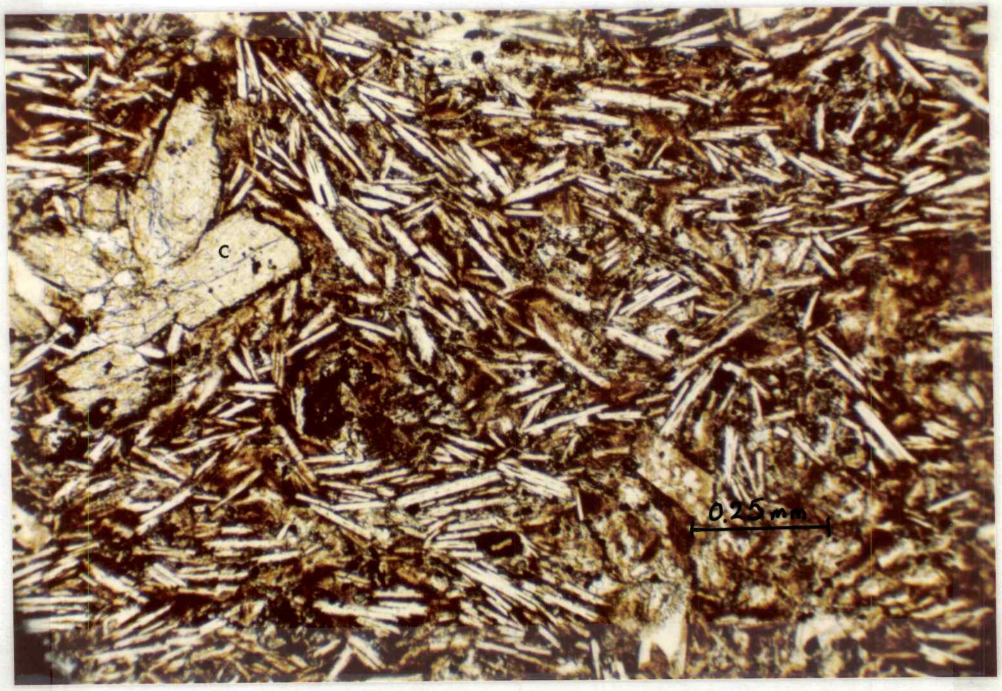

Figure 35. Photomicrograph of basaltic andesite. Note clinopyroxene phenocryst (C) and pilotaxitic arrangement of plagioclase AN_{45-52}) microlites. Groundmass is devitrified glass. Sample RAK-158, Medicine Lodge volcanics, center of sec. 22.

The volcanic breccias in the area are composed of angular to subangular fragments of the above mentioned basaltic andesite in a matrix of similar color and composition. Fragments within the breccia range in size from lapilli to blocks up to 31 inches (77 cm.) across.

The minor tuffs are very light gray (N 8) on fresh and weathered surfaces. These rocks are commonly crystal tuffs, though a few crystal lithic tuffs were also examined. The lithic fragments are lapilli and blocks up to 16 inches (40 cm.) of vesicular lava. In thin section these andesitic tuffs are seen to be composed of ash- and lapilli-sized anhedral plagioclase and clinopyroxene in a groundmass of brown divitrified glass.

Unconsolidated Gravels
Tertiary-Quaternary

Description The origin of the unconsolidated gravel cover on small plateaus and flat topped ridges of intermediate relief (7,500 to 8,500 ft.; 2,288 to 2,593 m.) within the thesis area is somewhat problematical. These gravels are composed of a very poorly sorted accumulation of boulders up to five feet (1.5 m.), cobbles, pebbles, sand, and silt. The boulders and cobbles are of a variety of rock types which are recognizable as having been derived from numerous locally exposed Paleozoic and Mesozoic strata. Subangular calcareous and noncalcareous sandstone and siltstone, sub-

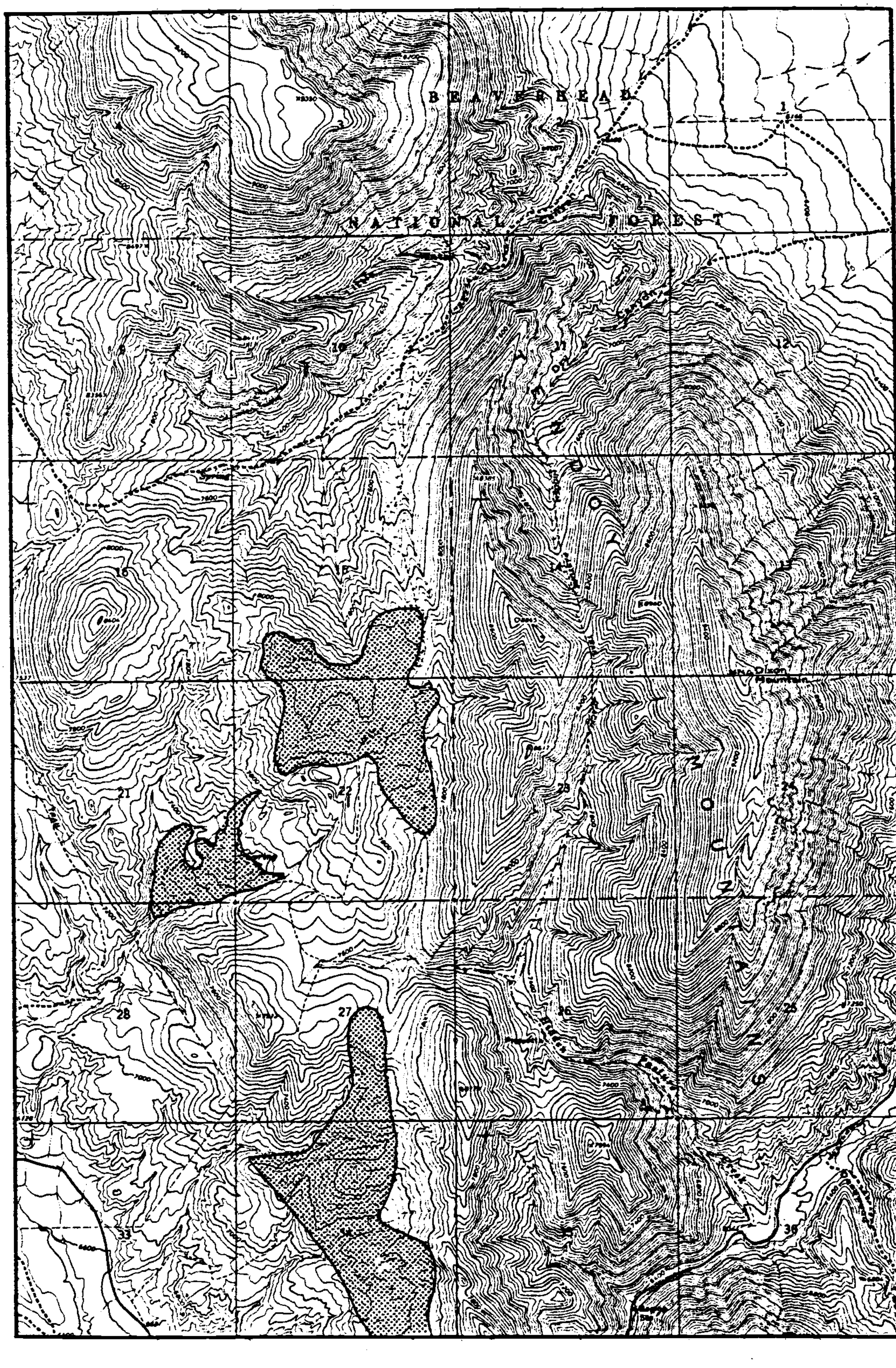

Figure 36. Distribution of unconsolidated gravels.

rounded pelecypod and crinoidal limestone, and well rounded red and purple quartzite and chert compose the bulk of the larger clasts. This material has the appearance of the Beaverhead conglomerates though it is unconsolidated and rests upon rocks younger than the Beaverhead.

Scholten, et al. (1955) believe this material represents an intermediate erosional surface developed during the late Tertiary, previous to block faulting in the area. The lack of sorting and the variety of rounding of the cobbles and boulders of the gravel deposits suggests a glacial or fanglomerate origin. Glaciation within the Tendoy Range is believed to have been of very limited extent. One morainal deposit has been reported along McNinch Creek, five miles west of the thesis area (Scholten, et al., 1955). The unlithified gravels within the thesis area may have had a glacial origin as well. Parts of the gravel layer have a topography suggestive of glacial deposition. Small enclosed depressions and semicircular ridges have the appearance of a knob-and-kettle topogrpahy developed on a terminal moraine.

These gravels could have formed as glacio-fluvial deposits on outwash fans at the front of a small mountain glacier. The small depressions may be the remains of kettles developed along the morianal front. The framework support manifested in much of this material would support this interpretation.

STRUCTURAL GEOLOGY

Regional Structure

The southwestern Montana area and the adjacent Idaho-Montana border lie within a region of abrupt and distinctive stratigraphic and structural changes. The Tendoy and nearby Beaverhead Ranges fall within a transitional zone between a cratonic and a geosynclinal sedimentary province (Scholten, 1967). This transition zone was an area of differential subsidence throughout Paleozoic and early Mesozoic time. Greater subsidence west of the transition zone or hinge line, resulted in the deposition of approximately 15,000 feet (4,575 m.) of Paleozoic strata. By comparison, only 7,500 (2,288 m.) of rock of similar age were deposited east of the hinge line (Huh, 1968).

The differences in subsidence rates and subsequent sedimentary accumulation between the cratonic platform and the miogeosynclinal trough had marked effects on later Laramide orogenic events. The structural framework that developed in the geosyncline, west of the hinge line, is characterized by low-angle thrust faulting and overturned and recumbent folds which do not involve crystalline basement rocks. These folds and faults are oriented in a subparallel manner and trend N35-40W (Huh, 1968).

Structural development in the cratonic province, east of the Tendoy Range, is much simpler than that of the geo-

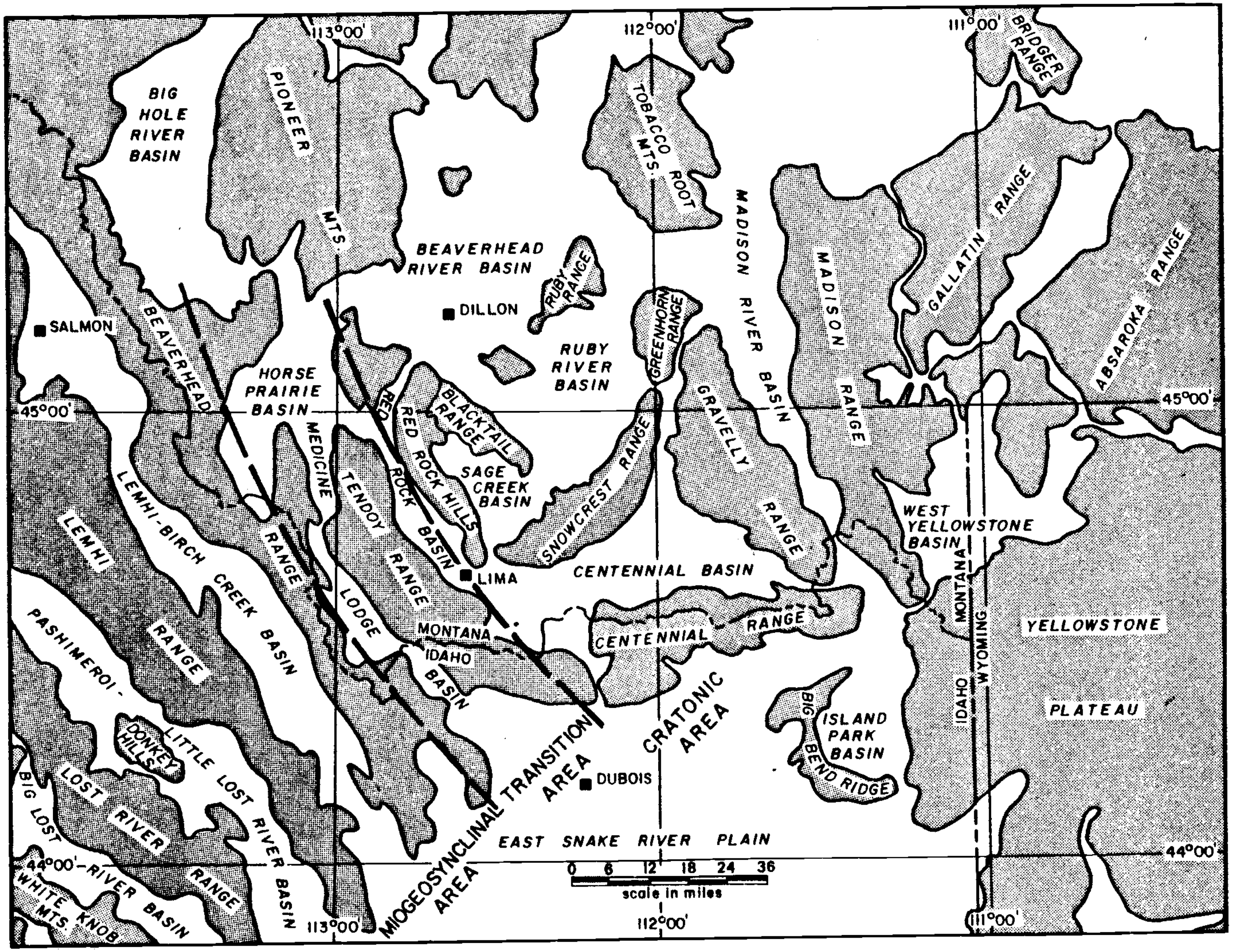

Figure 37. Physiographic features and tectonic provinces of southwestern Montana and adjacent areas (Modified from Scholten, 1967; and Huh, 1968).

synclinal province. Broad anticlinoria, commonly bounded by steeply dipping reverse faults, characterize much of the region. A series of uplifts radiating to the southwest, south, and southeast from the Tobacco Root Mountains dominates much of the area. These broadly folded and steeply faulted ranges commonly include crystalline basement rocks in their cores and are more randomly oriented than nearby structures of the geosyncline.

The Tendoy Range lies in the meeting ground between these contrasting structural provinces. The southwest-plunging folds related to cratonic-style uplifting of the Blacktail-Snowcrest arch have been highly deformed by subsequent "geosynclinal" deformation. As this deformation advanced into the transition area during the Paleocene a series of northwest-trending folds, high angle reverse faults, and low-angle thrust faults were superimposed on the cratonic structures.

The overlap of structural styles has resulted in a very complex structural framework within the Tendoy Range. Here, upthrusted basement blocks are interspersed with intricately refolded anticlines and synclines, downward-steepening reverse faults, and low-angle thrusts.

Late Tertiary block faulting is responsible for much of the present physiography of the region. Here again a change in structural pattern is observed across the transition zone.

The north- to northwest-trending basin-and-range style block faulting of the Lost River, Lemhi, and Beaverhead Ranges extends only to the eastern flank of the Tendoy Range. Block faulting continues to the east of the transition zone; however, the pattern of faulting becomes much more random.

Thesis Area Structure

The structural patterns within the thesis area are representative of the tectonic patterns of southwestern Montana. The variety of deformational trends and styles within the study area accurately depict a regional tectonic history which has been influenced by both geosynclinal and cratonic diastrophism.

Folds Folding in the study area is a cylindrical style, varying in degree of intensity. The dominant structure in the area is a southwestward-plunging broad syncline (Little Water syncline) with the axial trace passing through Little Water Canyon. A related anticlinal feature is developed to the south of the syncline with its axis plunging westward through the southernmost part of the study area (Big Sheep Creek anticline). This folding involves all strata older than late Early Cretaceous (Kootenai) and is probably related to the Late Cretaceous Blacktail-Snowcrest arch.

A number of northwest-trending structures have been superimposed on the gently plunging Little Water syncline.

Figure 38. Steeply southward plunging Timber Butte anticline. Quadrant talus-covered slopes of Timber Butte in center skyline. Steeply-dipping flat irons of Thaynes limestone define nose of anticline (center distance). Reddish beds at right are overturned Kootenai strata. Cliffs to extreme left center are of Beaverhead Formation. Looking N35E.

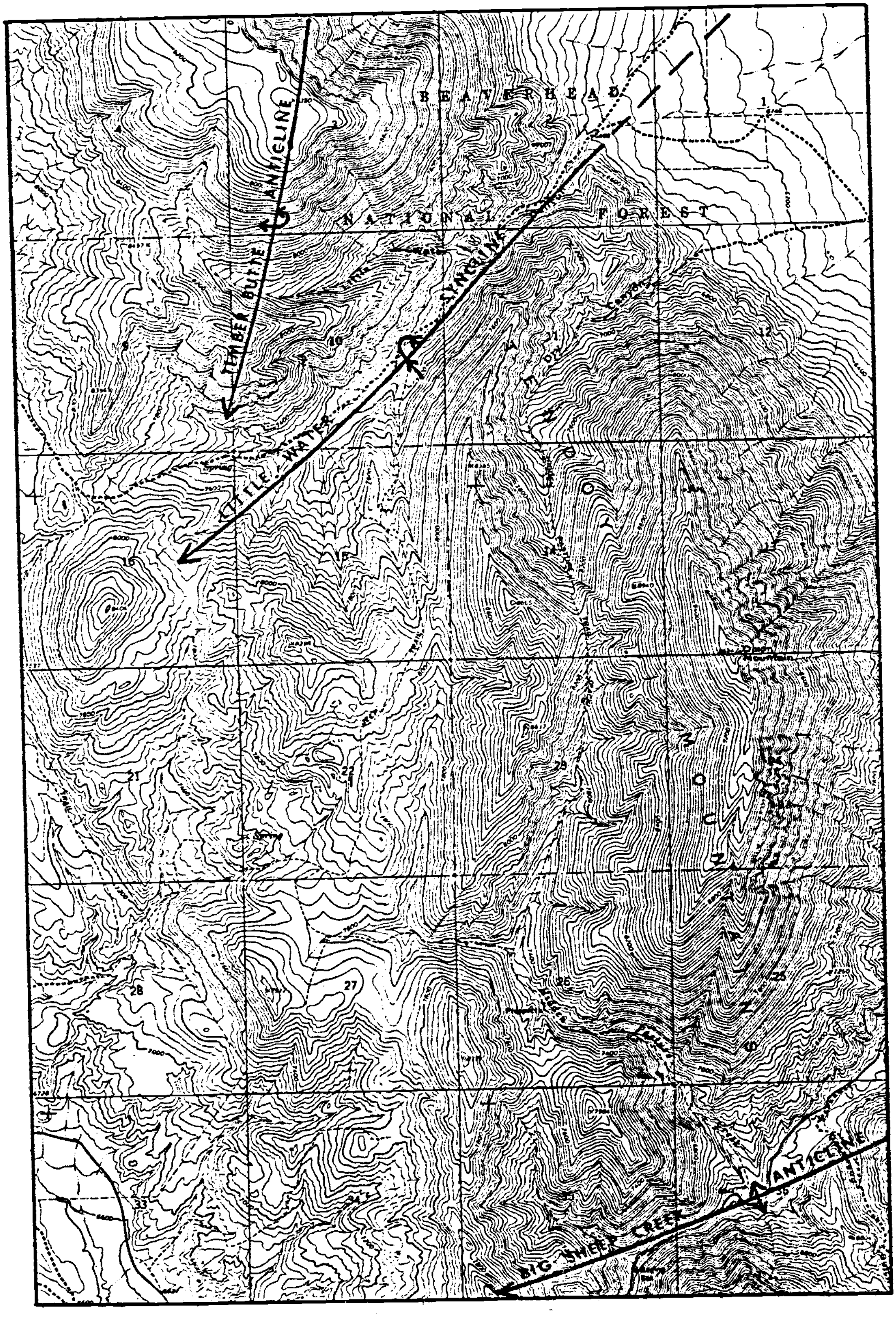

Figure 39. Axes of thesis area folds.

Northeastward-directed compression related to the eastward migration of "geosynclinal" deformation resulted in the refolding and faulting of the Little Water syncline. This refolding is manifested by the overprint of a steeply southward-plunging anticline (Timber Butte anticline). This anticlinal feature is most pronounced in the steeply dipping flat-irons and overturned Triassic strata north of Little Water Creek. This folding probably occurred subsequent to the development of a Late Cretaceous erosional unconformity and deposition of the Beaverhead conglomerates. The Beaverhead has been tilted to the west and the underlying unconformity is now dipping westward at approximately thirty degrees. The anticlinal folding overturned much of the northern flank of the Little Water syncline. The axial plane of the overturned fold flattens upward in a manner similar to the convex-up geometry of the major reverse faults in the area. This upward-flattening geometry may suggest an approach to an early Tertiary erosional surface. The potential for horizontal yielding would increase as structural depth decreased.

The folding of the Timber Butte anticline led to the development of a tightly closed syncline between Timber Butte and Dixon Mountain. Continued compression and folding of this syncline probably led to the high- and low-angle reverse faulting present in the Little Water Canyon-Dry Canyon area.

<u>Faults</u> Low-angle thrust faulting involving Mississippian limestone (Medicine Lodge Thrust) is the most dramatic faulting in the thesis area. This thrust involves implacement of strata markedly different in facies from coeval autochthonous rocks. Displacement of at least 20 miles is indicated.

The Medicine Lodge thrust is, however, of only minor importance in understanding the tectonic history of the area. Northwest-trending low-angle and high-angle reverse faults related to compressional stresses and intense folding are illustrative of the progression of deformational events in the area.

The Tendoy thrust is an upward-flattening reverse fault which has placed Pennsylvanian and Mississippian strata over the Beaverhead conglomerates (Late Cretaceous-early Tertiary). This strike fault involves a stratigraphic separation of at least 7,000 feet (2,135 m.) and perhaps as great at 15,000 feet (4,575 m.). The surface trace of the Tendoy thrust lies immediately east of the thesis area. The structural elevation of the upper Paleozoic and Mesozoic strata in the Tendoy Range is, in part, a result of displacement along the Tendoy thrust.

As compressional stresses continued into the early Tertiary the Timber Butte anticline was superimposed on the northern flank of the Little Water syncline. The intense folding that followed ultimately overturned much of the north-

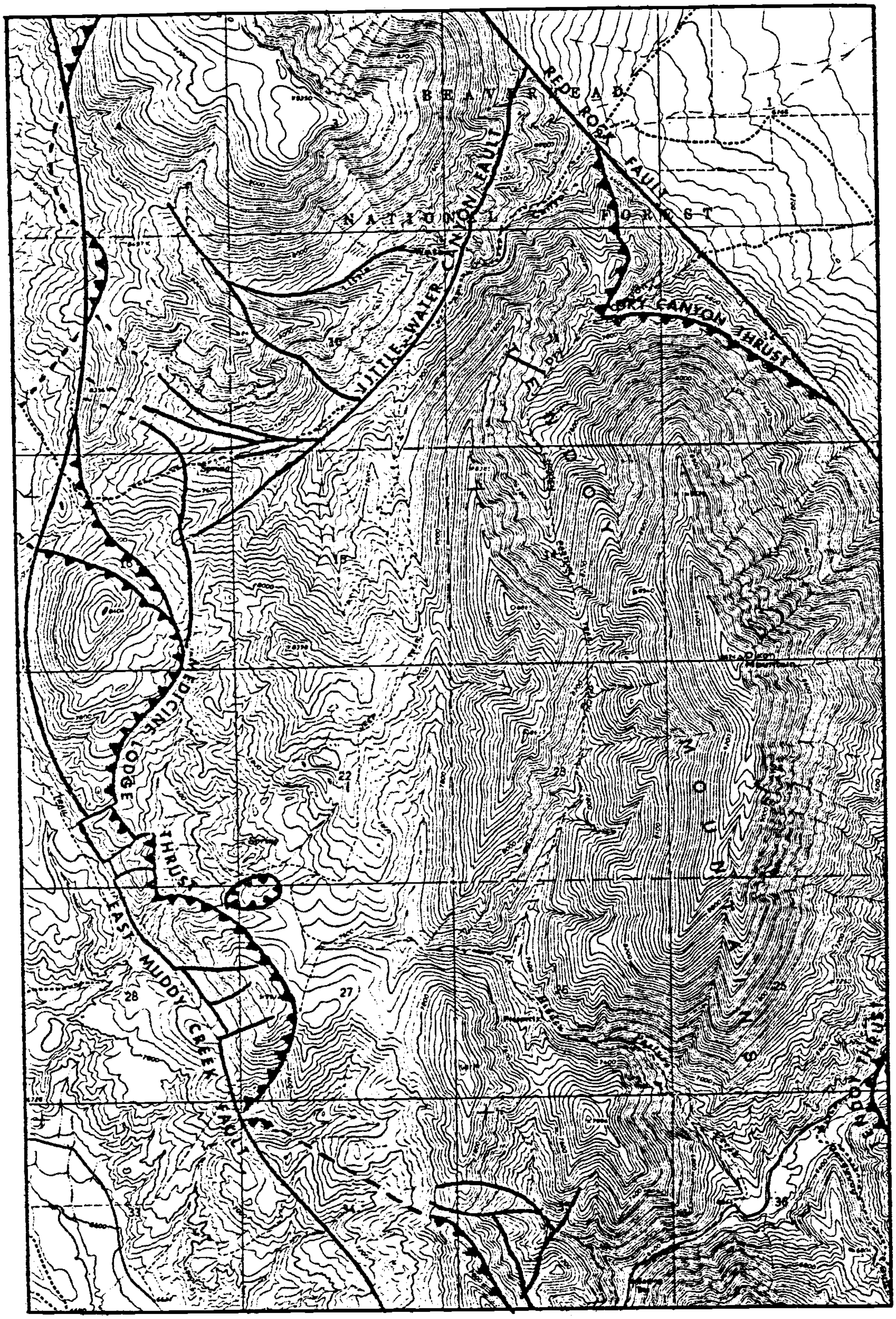

Figure 40. Surface trace of faults present in thesis area.

ern flank of the syncline. Structural readjustments near the nose of the steeply plunging Timber Butte anticline resulted in the development of northwest-trending reverse faulting of limited stratigraphic separation. These faults, of only local extent, have left parts of the Jurassic sequence and the Dinwoody Formation tectonically thickened (Plate 1).

The low-angle thrust exposed in Dry Canyon (Dry Canyon thrust) may have been developed during the folding of the Timber Butte anticline. Extension of fractures formed during Tendoy thrust motion may have led to a second upward flattening reverse fault forming subparallel to the first. At the mouth of Dry Canyon upper Phosphoria strata rest upon Jurassic rocks across the Dry Canyon thrust. This indicates a stratigraphic separation of approximately 2,000 feet (610 m.).

Extension of this fault into the subsurface suggests a horizontal displacement of at least one mile. Lithologic similarities of Thaynes (Triassic) strata above and below the thrust indicate horizontal movement could not have exceeded two or three miles.

Overturning along the southeast flank of the Timber Butte anticline must have continued as movement along the Dry Canyon thrust took place. This continued folding led to development of the northeast-trending high-angle reverse fault which can be traced through the western part of Little

Water Canyon and the northern slope of the canyon (Little Water Canyon fault). Displacement along the Little Water Canyon fault increases markedly eastward. Where last observed, north of the mouth of Little Water Canyon, overturned Dinwoody (Triassic) and Phosphoria (Permian) strata lie across the fault from a normal sequence of Triassic strata. Examination of this fault in cross-section suggests a stratigraphic separation of approximately 3,000 feet (915 m.). This displacement decreases westward over three miles to virtually zero.

The Medicine Lodge thrust sheet was implaced during early to mid-Eocene subsequent to most of the folding and faulting within the study area (Ryder and Ames, 1970). The thrust sheet shows no evidence of having been affected by the Late Cretaceous and early Tertiary tectonic events in the area. Implacement of this thrust involved local deformation of underlying strata at the thrust front. The incompetent mudstones and thin limestones and sandstones of the Kootenai Formation show evidence of faulting and folding along the thrust contact. Small slivers of Thaynes limestones are present along the base of the thrust. These limestone slivers must have been caught up in the thrusting as the Medicine Lodge thrust progressed upsection through the northwestern flank of the Little Water syncline.

In the northwestern corner of the thesis area, the Medicine Lodge thrust rests upon the folded dipslope of the

Figure 41. Parallel planar jointing pattern developed in thick-bedded limestone conglomerates of the Beaverhead Formation. Exposed in conglomerate cliffs on western boundary of thesis area.

Beaverhead conglomerate. As the thrust moved over the Beaverhead the conglomerates were pushed into the underlying Cretaceous (Kootenai) and Jurassic rocks. This caused local folding and minor reverse faulting in these strata. The conglomerates also bear witness to the period of stress. A series of well defined, subparallel fractures, passing through both matrix and clasts, are very conspicuous in the Beaverhead cliffs and ledges.

Normal faulting of regional and local significance occurs within the study area. Block faulting in the south-central part of the area appears to be the result of structure readjustments related to tensional stresses across the axis of the westward-plunging Big Sheep Creek anticline. A number of small normal faults also occur within the allochthonous limestones and sandstones of the Medicine lodge thrust sheet. This faulting may be related to differential motion during thrust implacement or a response to northwest-trending block faulting along the adjacent East Muddy Creek fault.

The East Muddy Creek fault on the western side of the study area and the subparallel Red Rock fault on the eastern side of the area are the most recent tectonic elements. These normal faults are the easternmost of a series of basin-and-range-type block faults which extend westward into the Beaverhead, Lemhi, and Lost River Ranges of Idaho. The

Figure 42. Triangular facets and fault scarplet developed along trace of Red Rock Fault between Lima and Dell, Montana. Five miles SE of thesis area.

present physiography of the Tendoy Range is defined by the Red Rock fault and its western counterpart the Deadman fault. The intermontane Muddy Creek Basin is bounded by the East and West Muddy Creek faults.

Displacement across the Red Rock fault is difficult to determine, for the downthrown block is covered by Quaternary fan and stream gravels. Truncated ridge spurs along the eastern front of the Tendoy Range indicate at least 300 feet (92 m.) of displacement. This is probably a very conservative estimate of fault motion. It is likely that the elevation of the Tendoy Range above the adjacent Red Rock Valley is the result of dip slip motion along the Red Rock fault. If this is the case, displacement of at least 3,000 feet (915 m.) is indicated.

Motion along the Red Rock fault continues to the present. Low scarps developed across fluvial gravels range from ten to 20 feet (3-6 m.) in height (Scholten, et al., 1955).

Stratigraphic separation across the East Muddy Creek fault has not been accurately determined. The downdropping of 700-800 feet (214-244 m.) of strata involved in the Medicine Lodge thrust and the presence of at least 600 feet (183 m.) of Tertiary deposits in the Muddy Creek Basin indicate dip slip displacement may exceed 1,500 feet (458 m.).

GEOLOGIC HISTORY

The oldest rocks exposed within the thesis area are of Mississippian age. These rocks, and younger Paleozoic and Mesozoic strata, record a sequence of depositional events common to the craton and miogeosyncline of the Cordillera of western North America. The southwestern Montana area was part of a transitional zone or hinge line throughout the Paleozoic. To the west of this region, thick basinal deposits accumulated in the deeply subsiding Cordilleran trough. To the east of the region a thin veneer of cratonic sediments accumulated, interrupted by numerous regional unconformities and diastems. The recurrently active positive elements of the transition zone migrated laterally between the Tendoy Range the Lemhi Range to the west.

This geosynclinal sedimentation pattern continued uninterrupted into the Early Triassic. Uplift, beginning in the Middle Triassic, marked the beginnings of the breakup of the Paleozoic geosyncline. The eastward shifting of tectonic belts and depositional basins is recorded in the sedimentary events of the Jurassic and Cretaceous and the tectonic events of the late Mesozoic and early Tertiary.

Mississippian sedimentation in southwestern Montana and adjacent parts of Idaho can be characterized as a lower transgressive sequence and an upper regressive sequence (Rose, 1977). During the transgressive phase (Kinderhook-

Osage), the open marine sediments of the Lodgepole Formation were deposited on the cratonic margin and in the Williston Basin while argillaceous sediments derived from the Antler highland to the west were deposited in the deeper parts of the miogeosyncline (Huh, 1968). As the seas regressed during Meramec and Chester times, deposition patterns across the hinge line changed markedly. The thickly bedded Mission Canyon limestones were deposited under conditions of shallow water depth and high sedimentation rates. Rapid sediment accumulation resulted in the westward progradation of the platform margin and the deposition of allodapic deposits along the platform edge (Huh, 1968).

As regression continued into the Chesteran time, evaporite sequences and ultimately an erosional surface developed on the Mission Canyon limestones. Terrigenous clastics derived from newly exposed cratonic areas were carried across the eroding platform and deposited in the deeper parts of the basin in what is now Idaho (Rose, 1977). These deep marine detrital deposits are typical of much of the Middle Canyon Formation of the White Knob Group. The cratonic sediments of the Middle Canyon interfinger with flysch deposits derived from the Antler highland. This intertonquing of deposits in the narrow basin between the highlands to the west and the cratonic margin precluded the formation of a starved basin within the geosyncline (Rose, 1977).

During the time of erosion of the Mission Canyon and deposition of the Middle Canyon Formation the sandstones, silty limestones and black shales of the Big Snowy Formation were deposited. These rocks accumulated in the Williston Basin and a narrow trough (Big Snowy trough) which extended southwestward to join the open marine waters of the Cordilleran geosynclinal sea (Maughan and Roberts, 1967). Regional uplift in middle Chesterian time, which centered around southern Montana, brought an end to deposition in the Big Snowy trough and much of the Williston Basin. An erosional surface was developed on the subaerially exposed Big Snowy strata which is manifested in the Tendoy Range as a thin deposit of limestone conglomerates. During Chesterian time the remainder of the White Knob Group was being deposied in the deeper parts of the geosyncline beyond the cratonic edge.

A late Chesterian transgression resulted in marine sedimentation returning to much of the area previously involved in Big Snowy sedimentation. The deposition of the limestones and sandstones of the Amsden Formation record the uplifting and deep erosion of Paleozoic strata in southern Canada, Idaho, and western Montana (Maughan and Roberts, 1967). As uplift and erosion continued into the Early Pennsylvanian, extensive areas of Ordovician sandstones (Kinnikinic) were exposed. These eroded sandstones are probably the source for much of the sand of the Quadrant and Tensleep Formations. Sandstone deposition must have kept pace with subsidence for all of the Quadrant was deposited under shallow shelf conditions. Similarly uplifted Ordovician sandstones in west-central Utah are believed to have contributed considerable amounts of sand to the Tensleep, Wells, Oquirrh, and Weber formations south of Montana (Maughan and Roberts, 1967). Deposition of the Amsden and Quadrant formations in southwestern Montana was continuous and records the increased supply of sand to the basin as limestones and dolomite become increasingly less common upsection until the Quadrant becomes almost entirely sandstones.

Quadrant deposition may have continued into the earliest Permian (Wolfcampian); however, increased subsidence rates or decreased sediment supply lowered the seafloor below wave base in the area west of the Ruby Range during the Early Permian. This lowering of the basin floor was accompanied by differential tectonic movement within and around the Phosphoria basin (Cressman and Swanson, 1964). An area including the present Tendoy and Snowcrest ranges subsided more rapidly than did the Beaverhead Range area to the west. The many lithologic changes present in the Phosphoria Formation represent variations in subsidence rates, detrital sediment supply, tectonic movement in and around the basin, and changes in circulation patterns within the basin (Cressman &

Swanson, 1964). These variations in basin sedimentation processes had a strong influence on the localization of areas of maximum carbonate and siliceous sponge and microfossil production and the deposition of limestone and chert strata. Phosphatic sediments are believed to have been deposited during times of maximum transgression within the basin (Boyd and Maughan, 1973).

Subsidence within the Phosphoria embayment, which at times extended as far east as central Wyoming, kept pace with or surpassed a eustatic sea level fall that began in the pre-middle Leonardian and continued into the Guadalupian (Vail et al., 1977). During the Late Permian, sea level fall surpassed subsidence rates and parts of the basin margin became emergent. Although parts of the Phosphoria strata became subaerially exposed, low relief and arid climatic conditions resulted in little erosion or soil development.

Tectonic and sedimentary conditions remained much the same in the Early Triassic of southwestern Montana as they had been in the Middle and Late Permian. This, however, was not the case to the west of the study area. Orogenic events similar in style and location to the Late Devonian Early Mississippian Antler orogeny developed in northern Nevada and adjacent Idaho. Compressional events during the Late Permian, possibly related to marginal basin closing, resulted in late Paleozoic continental rise and slope sediments (Havallah Sequence) being thrust eastward over coeval continental shelf deposits along the Golconda thrust (Silberling, 1973). This deformational event had little effect on Early Triassic sedimentation in southwestern Montana. However, the subsequent eastward migration of related epeirogenic movements led to the post-Early Triassic erosion which affected much of western Montana.

Although physical conditions of sedimentation changed little across the Permo-Triassic boundary, biological changes were significant. The diminution of skeletal-producing and mound-building organisms during the Late Permian extinctions resulted in the deposition of fewer bioclastic strata. The accompanying reduction in planktonic organisms may explain the lack of organic-rich deep basinal deposits (Boyd and Maughan, 1973).

As Early Triassic seas transgressed northward into Utah, eastern Idaho, western Wyoming and southwestern Montana, the sediments of the Dinwoody and Woodside Formations were deposited. The Dinwoody strata accumulated in an elongate miogeosynclinal basin and cratonic platform between the Sonoma highland of north-central Nevada and central Idaho to the west and the low lying craton to the east (Collinson and Hasenmueller, 1978). Detritus within the Dinwoody was derived from cratonic sources. The Dinwoody shoreline was a zone of intertonguing with the continental Woodside deposits. The basin margin can be characterized as broad mudflats and estuaries of very low relief upon which algal mats were commonly developed. These strandline deposits pass basinward to low offshore bars and back bar lagoons, gray shales and limestones of the platform, and the black shales of the deep basin beyond the platform edge. Climatic conditions in the region during the Early Triassic were hot and arid. Warm waters saturated with calcium carbonate supplied from eroded Paleozic carbonates of the craton, were very conducive to the growth of calcareous skeleton-producing organisms. Seasonal storm activity may have been influential in the resedimentation and distribution of skeletal and detrital material.

The Dinwoody-Woodside transition zone migrated across the basin margin as the result of a slowly rising sea level and changes in detritial sediment supply. As terrigenous influx exceeded sea level rise the shoreline regressed basinward. The very low relief of the shoreline led to very slight sea level changes affecting wide areas. An extensive transgression during the Early Triassic brought the deposition of the Thaynes Formation. Tectonism and sedimentation conditions during Thaynes time were very similar to those that existed during deposition of the Dinwoody-Woodside sequence. However, the Thaynes shoreline was extended farther onto the craton than was the case earlier in the Triassic.

Sea level fall and emergence of the Dinwoody-Thaynes basin during the Middle Triassic exposed Thaynes strata to extensive erosion. As much as 1000 feet (305 m.) of Triassic and late Paleozoic rock were removed from much of western Montana (Moritz, 1951). This period of uplift and subaerial exposure marked a change from the geosynclinal sedimentation which had characterized the Paleozoic and Early Triassic to a foreland basin setting which persisted throughout the remainder of the Mesozoic (Collinson & Hasenmueller, 1978).

During the Middle and Late Jurassic the southwestern Montana region was affected by three transgressive episodes, each spreading farther southward than the previous into the interior seaway of western North America. This seaway, spreading out of northern Canada, affected much of Montana and Wyoming and parts of Idaho, Utah, Colorado, New Mexico and North and South Dakota. The western margin of this seaway was the site of maximum sediment accumulation (Imlay, 1957). Jurassic sedimentation in western Montana was influenced by the Belt arch, an intermittently positive tectonic element located in west-central Montana that commonly produced shoaling, influenced marine circulation patterns, and was infrequently emergent. In addition, pulses of orogenic activity to the west contributed to changing patterns of sediment supply.

The first Jurassic marine incursion to affect southwestern Montana occurred during latest Bajocian time (Hallam, 1975). This transgression brought deposition of the Sawtooth Formation carbonates and shales along the western margin and fine-grained terrigenous clastics of the Piper Formation in the Wyoming shelf area to the east (Wright, et al., 1977).

Rising sea levels during the early Callovian coincide with deposition of the Rierdon Formation. These shallow marine limestones and oolitic limestones show evidence of effects of the nearby Belt arch. Northward flowing warm water influenced sedimentation in southwestern Montana and much of the Williston Basin area of eastern Montana. Belt arch uplift in late Rierdon time subjected Rierdon strata near the uplift to erosion and altered basinwide circulation patterns (Peterson, 1957). Subsequently, sedimentation continued in the eastern part of the interior seaway but altered current flow brought cooler waters into the area and ended carbonate sedimentation (Peterson, 1957). Following the cessation of Rierdon-Twin Creek deposition, carbonate sedimentation would never again be significant in the western interior seaway of North America.

The final and most areally extensive transgression came in the early Oxfordian (Hallam, 1975). During this time uplifting of the orogenic areas to the west began to supply major amounts of terrigenous clasts to the interior seaway. Much of the sediment on the western margin of the seaway was deposited as laterally migrating and locally prograding deltaic deposits which were reworked by marine processes (Brenner and Davies, 1974). These sediments become finer eastward with sandy shallow water deposits surrounding the Belt arch (Peterson, 1957). As tectonic activity west of the seaway increased in intensity and migrated eastward, detrital influx into the interior seaway increased. This increased sediment supply first brought a halt to the regional westward transgression and ultimately initiated a regression (Brenner and Davies, 1974). This regression continued until the western part of the seaway was filled and sedimentation was entirely non-marine.

By Kimmeridgian time the non-marine Morrison Formation was accumulating in southwestern Montana. The distribution and thickness of the Morrison in Montana were controlled by the position of the Belt arch positive element and subsidence within the reactivated, northeast-trending, Big Snowy trough (Suttner, 1969). Relatively large rivers meandered over the low-lying floodplain deposits of western Montana. Locally, ephemeral lakes developed and fresh water limestones accumulated (Moritz, 1951). Suttner (1969) believes that the Morrison represents the distal end of a clastic wedge extending eastward from latest Jurassic orogenic highlands in eastern Oregon and Washington. The aggradation and uplift of the Morrison floodplain must have occurred at rather high rates for a significant worldwide sea level rise was underway (Vail et al., 1977). Following a brief sea level fall at the close of the Jurassic this eustatic rise continued through much of the Cretaceous but marine sedimentation never returned to western Montana.

An erosional unconformity was developed on Morrison strata during much of the Early Cretaceous. During Aptian time diastrophic pulses in central Idaho exposed chert-rich Pennsylvanian and Permian strata to erosion. This tectonic activity is manifested in southwest Montana by an abrupt influx of high energy coarse detrital material in the basal conglomerate of the Kootenai Formation (Suttner, 1966). As older Paleozoic strata was exposed in the orogenic belt, sediment size decreased and sandstone and siltstone deposition prevailed. These fine-grained terrigenous deposits contain thin lenses of fresh water limestone, deposited in small basins. As sedimentation continued in the foreland areas east of the orogenic belt, the Belt arch became a site of rapid subsidence and ultimately was an area of very thick accumulations of Early Cretaceous sediments.

Near the close of the Aptian a series of foreland basin lakes developed between uplands of moderate relief in Idaho and coastal plains along the Cretaceous seaway to the east. These lake deposits are believed to have been the result of the fortuitous combination of local downwarping to form the lake basins, carbonate-rich streams entering the lakes from Paleozoic source rocks, and a warm, semi-arid climate conducive to carbonate precipitation with little terrigenous influx (Paine and Suttner, 1971). These lacustrine deposits are almost entirely gastropod tests. These gastropod limestones cover an area of nearly 16,000 square miles in western Montana (Paine and Suttner, 1971).

Deposition of the early Late Cretaceous Aspen Formation followed conformably upon the Kootenai. Much of these floodplain, lacustrine and swamp deposits have been removed by Late Cretaceous erosion in southwestern Montana.

Beginning in the Albian a crystalline basement-cored anticlinorium began to rise in southwestern Montana. This southwest-trending arch, the Blacktail-Snowcrest arch, was part of a series of south-trending uplifts related to the basement uplifts of the Tobacco Root Mountains. The Gravelly, Madison and Gallatin anticlinoria are all related to the same sequence of cratonic uplifts (Scholten, 1967). These uplifts appear to have been temporally related to the intrusion of the Idaho batholith and the doming up of a great thickness of Paleozoic and Belt age strata in Idaho (ancestral Beaverhead Range). As the Blacktail-Snowcrest arch continued to rise, the Mesozoic and Paleozoic strata on its western and southern flanks were folded and eroded. The Little Water syncline developed in the thesis area between two southwestward plunging anticlines at the nose of the Blacktail-Snowcrest arch

Scholten, 1967). As uplift continued, clasts of Mesozoic and Paleozoic limestones were shed onto the developing angular unconformity. Concurrently, highly resistant Belt Quartzite cobbles and boulders were being shed eastward by torrential braided streams that flowed from the mountains onto an actively downwarping depression (Scholten, 1968).

The 15,000 feet (4,575 m.) of the Beaverhead Formation are composed almost entirely of these intertonquing quartzite and limestone conglomerates. Uplifting and rapid erosion of the Blacktail-Snowcrest arch and the ancestral Beaverhead Range continued well into the Paleocene and perhaps into the early Eocene (Ryder, 1968). As the Beaverhead conglomerates were being deposited, northwest-trending folds were superposed on the southwest-trending Late Cretaceous structures (Scholten, 1968). This resulted in the refolding of the Little Water Syncline.

Deposition of the Beaverhead conglomerates ceased in the late Paleocene or early Eocene as deformation migrated eastward and began affecting the Tendoy Range (Ryder and Scholten, 1973). Northeastward directed compressional forces resulted in a series of northwest-trending, downward-steepening reverse faults cutting through the Beaverhead strata and juxtaposing rocks as old as Mississippian. In addition northwest-trending folds continued to develop throughout the region.

The same compressional events which generated the reverse faulting in southwestern Montana were probably responsible for the initiation of the Medicine Lodge Thrust. Movement of this thrust began as early as Late Cretaceous or early Paleocene (Ruppel, 1978). The Medicine Lodge thrust sheet moved northeastward and may have been the source of uppermost Beaverhead conglomerates. These same conglomerates were ultimately overridden by the thrust in the Tendoy Range. Thrust motion appears to have stopped by the early Eocene (Ruppel, 1978).

Subsequent to the termination of thrust movement, volcanic activity was initiated in areas nearby to the west of the Tendoy Range (Challis volcanics). This activity began in the Oligocene and continued into the Miocene (Scholten, et al., 1955). The Medicine Lodge volcanics commonly are found near the trace of the Medicine Lodge Thrust. This proximity suggests that the fault may have served as a conduit for local eruptive activity related to the main Challis volcanics of the Beaverhead and Lemhi Ranges to the west (Ruppel, 1978).

The relaxation of compressional stresses during the middle Tertiary led to crustal readjustments in the Pliocene and block faulting. A series of normal faults developed approximately parallel to the strike of Sevier-Laramide structures. In the Tendoy Range, the Muddy Creek and Red Rock faults developed in response to the regional extension. Block faulting in the Tendoy Range shows maximum displacements along the eastern side of the range and a resultant southwestward tilting of late Tertiary erosional surfaces (Scholten, 1968). Maximum displacement of as great as 3,000 feet (915 m.) is indicated on the Red Rock fault along the eastern edge of the Tendoy Range (Pardee, 1950).

Late Tertiary and Pleistocene drainage systems have eroded terraces and deposited sands and gravels throughout much of the Tendoy Range. These deposits and terraces have been uplifted as block faulting continues in the region. The Red Rock fault has been historically active. A recent scarp 20 feet (6 m.) high is present at Big Sheep Creek. This continued faulting activity may be responsible for much of the landslide activity observed in the region. Related normal faulting in the Madison Range to the east was responsible for the disastrous Hebgen Lake earthquake and Madison landslide near West Yellowstone in August of 1959.

ECONOMIC GEOLOGY

Few deposits of economic significance are present within the thesis area or in the Tendoy Range and adjacent Red Rock Valley.

Known metallic mineral deposits in the Tendoy Range are confined to a single lead-zinc mine near Ellis Peak, about eight miles west of the study area. The potential for mineralization exists in the brecciated, crushed, and mylonitized zone along the sole of the Medicine Lodge thrust (Ruppel, 1978). Hydrothermal fluids related to Tertiary volcanism in the area may have migrated along this zone and produced mineralized zones of economic significance. A graphite deposit along Kate Creek, five miles northwest of the thesis area, appears to be just such a deposit (Scholten, et al., 1955).

The presence of extensive exposures of the Phosphoria Formation within the study area and southward through the Tendoy Range suggest economical deposits of phosphates may be present. Regional studies of the western phosphate field undertaken by the U.S. Geological Survey (McKelvey, 1949) indicate that the phosphatic shales of the study area are not of high enough concentration to be economical. This report and subsequent reports (Cressman and Swanson, 1964) indicate low concentration of fluorine, nickel, molybdenum, uranium, vanadium, and solid hydrocarbons are also present within the

Phosphoria. Kerogen shales of the Phosphoria may be distilled to yield eight to 26 gallons of oil per ton (Scholten, et al., 1955). These shales were quarried near the mouth of Dry Creek Canyon during the 1920's for use as heating fuel.

Within the Tendoy Range, the Mission Canyon Formation was quarried and dried for lime but this operation has long since been abandoned. Near the southern end of the Tendoy Range gypsum was mined from the Big Snowy Formation from 1940 to 1942 (Scholten, et al., 1955).

Gravels in the Red Rock Valley have proved to be the most utilized and abundant economic sedimentary deposits of the region. A number of quarries have been established to provide aggregate for the past road construction and the continuing work on the Interstate Highway system.

The sedimentary and tectonic history of the region surrounding the thesis area suggests the possibility of hydrocarbon generation and accumulation. The black shales of the Phosphoria Formation and the black shales and dark carbonates of the Mississippian strata are excellent source beds in the region (Cannon, 1971). The juxtaposition of reservoir quality sandstones in the Quadrant and Kootenai formations provides good possibilities for migration and accumulation.

A number of wells have been drilled in the region with no success. A well drilled by Cities Service in 1952 and 1953 southeast of Lima was abandoned at a depth of 11,212 feet (3,420 m.) in Morrison strata. A well drilled in the

Centennial Valley by Shell Oil in 1964 was abandoned in Madison rocks at a depth of 10,244 feet (3,124 m.). This well encountered oil shows and dead oil in Triassic, Permian, and Pennsylvanian strata. Since those early attempts little drilling has been undertaken in the region. Recently, stratigraphic tests have been made southeast of Lima and in the Centennial Valley. The test six-and-one-half miles southeast of Lima was drilled in 1975 and 1976. It passed through 15,723 feet (4,796 m.) of folded and faulted strata, bottoming in Cretaceous rocks. The second test was drilled in the Centennial Valley in 1977 and 1978. It was abandoned in pre-Beltian crystalline rocks at a depth of 12,226 feet (3,729 m.).

The folding and uplift of the Blacktail-Snowcrest arch during the Late Cretaceous may have destroyed much or all of the hydrocarbons trapped in the Mesozoic and Paleozoic strata in the region. The tilted rocks on the flanks of the arch could have provided avenues for the migration of the hydrocarbons and their ultimate loss to erosion or the atmosphere. In addition, this interval of exposure would have allowed for the circulation of oxygenated fresh water into the strata. These circulating waters would have altered much of the original depositional texture (Cannon, 1971). There is, however, the possibility that stratigraphic pinch-outs and early faulting may have trapped oil and gas on the upturned flanks of the Blacktail-Snowcrest arch (Scholten, 1967).

If hydrocarbons were trapped on the flanks of the Blacktail-Snowcrest arch, the location of these accumulations will be difficult. The interference effect of two differently trending sequences of Laramide folding complicate the structural patterns on the flank of the Blacktail-Snowcrest arch. Many of the resulting northwest-trending folds would, therefore, be barren even if sufficient closure was developed (Scholten, 1967).

The similarities between the southwestern Montana region and the Overthrust Belt of western Wyoming does suggest the possibility of the development of structural traps at the updip edges of thrust slices of Paleozoic carbonates. In addition, the areally extensive Medicine Lodge thrust may conceal structural and stratigraphic configurations conducive to hydrocarbon accumulation. The nonmarine Tertiary Red Rock Valley basin does not appear to be a viable prospecting target, for much of the sedimentary strata is texturally and compositionally immature and is not likely to be a good reservoir or source.

The complexities created by the marked stratigraphic changes noted across the region, plus those of subsequent folding and faulting, suggest that the possibilities for petroleum accumulation do exist. However, these accumulations probably lie to the west of the upturned strata flanking the Late Cretaceous Blacktail-Snowcrest arch and may be concealed by early Tertiary thrusting. The application of the advanced geophysical techniques, developed for exploration in the Overthrust Belt of Wyoming and Utah, may prove useful in better defining the structural and stratigraphic framework in the southwestern Montana region and locating possible petroleum accumulations.

BIBLIOGRAPHY

Bathurst, R. G. C., 1975, Carbonate sediments and their diagenesis: Developments in sedimentology, New York, Elsevier, 658 p.

Beales, F. W., 1958, Ancient sediments of Bahama type: AAPG Bull., v. 45, p. 1845-1880.

Berry, G. W., 1943, Stratigraphy and structure at Three Forks, Montana: Geol. Soc. America Bull., v. 54, p. 1-29.

Blackwelder, Elliot, 1918, New geological formations in western Wyoming: Washington Academy of Science Jour., v. 8, p. 417-426.

Bonine, C. A., 1914, The Elliston phosphate field, Montana: U. S. Geol. Survey Bull. 580, p. 373-383.

Boutwell, J. M., 1907, Stratigraphy and structure of the Park City mining district, Utah: Jour. Geology, v. 15, p. 434-458.

Boyd, D. W., and Maughan, E. K., 1973, Permian-Triassic boundary in the Middle Rockies: in Logan, A., and Hills, L. V. (eds.), The Permian and Triassic Systems and their mutual boundary: Can. Soc. Petrol. Geol., Mem. 2, p. 294-317.

Briggs, G., and Cline, L. M., 1967, Paleocurrents and source areas of the late Paleozoic sediments of the Ouachita Mountains, southeastern Oklahoma: Jour. Sed. Petrol., v. 37, p. 985-1000.

Branson, C. C., 1939, Pennsylvanian formations of central Wyoming: Geol. Soc. America Bull., v. 50, p. 1199-1225.

Branson, E. B., and Branson, C. C., 1941, Geology of Wind River Moutains, Wyoming: AAPG Bull., v. 25, p. 132.

______________, and Gregor, D. K., 1918, Amsden Formation of the east slope of the Wind River Mountains of Wyoming and its fauna: Geol. Soc. America Bull., v. 29, p. 309-326.

Brenner, R. L., and Davies, D. K., 1973, Storm-generated coquinoid sandstones: genesis of high-energy marine sediments from the Upper Jurassic of Wyoming and Montana: Geol. Soc. America Bull., v. 84, p. 1685-1698.

______________________________, 1974, Oxfordian sedimentation in western interior United States: AAPG Bull., v. 58, p. 407-428.

Brown, R. W., 1946, Fossil plants and Jurassic-Cretaceous boundary in Montana and Alberta: AAPG Bull., v. 30, p. 238-248.

Bushinski, G.I., 1964, On the shallow water origin of phosphorite sediments: in van Straaten, L.M.J. (ed.), Deltaic and shallow marine deposits, New York, Elsevier, p. 62-70.

Bushinsky, G. I., 1935, Structure and origin of Phosphorites of the U.S.S.R.: Jour. Sed. Petrol., v. 5, p. 81-92.

Calkins, F. C., and Emmons, W. H., 1915, Description of the Philipsburg Quadrangle, Montana: U. S. Geol. Survey, Geological Atlas Folio 196, p. 1-25.

Cannon, J. L., Jr., 1971, Petroleum potential of western Montana and northern Idaho: in Cram, I. H., (ed.), Future petroleum provinces of the United States-Their geology and potential, AAPG Mem. 15, p. 547-568.

Chamberlain, C. K., 1978, Recognition of trace fossils in cores: in Basan, P., (ed.), Trace fossil concepts, S.E.P.M., Short Coarse No. 5, p. 133-183.

Clark, D. L., Peterson, D.O., Stokes, W. L., Wardlaw, Bruce, and Wilcox, J. D., 1977, Permian-Triassic sequence in northwest Utah: Geology, v. 5, p. 655-658.

Cobban, W. A., 1945, Marine Jurassic formations of the Sweetwater Arch, Montana: AAPG Bull., v. 29, p. 1262-1303.

______________, 1955, Cretaceous rocks of northwestern Montana: Billins Geol. Soc., Ann. Field Conf. Guidebook, v. 6, p. 107-119.

______________, Imlay, R. W., and Reeside, J. B., 1945, Type section of Ellis Formation (Jurassic) of Montana: AAPG Bull., v.29, p. 451-453.

Collinson, J. W., and Hasenmueller, W. A., 1978, Early Triassic paleogeography and biostratigraphy of the Cordilleran miogeosyncline: in Howell, D. G., et al., (ed.), Mesozoic paleogeography of the western United States, Pac. Coast Paleogeogr. Symp. No. 2, p. 175-187.

Condit, D. D., 1916, Realtions of the Embar and Chugwater Formations in central Wyoming: in Shorter contributions to general geology, U. S. Geol. Survey Prof. Paper 98, p. 263-270.

____________, 1918, Relation of late Paleozoic and early Mesozoic formations of southwestern Montana and adjacent parts of Wyoming: U. S. Geol. Survey Prof. Paper 120, p. 111-121.

Cook, P.J., and McElhinny, M.W., 1979, A reevaluation of the spatial and temporal distribution of sedimentary phosphate deposits in the light of plate tectonics: Econ. Geol., v. 74, p. 315-330.

Cressman. E. R., 1955, Physical stratigraphy of the Phosphoria Formation in parts of southwestern Montana: U. S. Geol. Survey Bull. 1027-A, 31p.

____________, and Swanson, R. W., 1964, Stratigraphy and Petrology of the Permian rocks of southwestern Montana: U. S. Geol. Survey Prof. Paper 313-C, p. 275-569.

Cross, C. W., 1894, Description of the Pikes Peak Sheet, Colorado: U. S. Geol. Survey, Geologic Atlas Folio 7, p. 1-5.

Darton, N.H., 1904, Comparison of the stratigraphy of the Black Hills, Bighorn Mountains, and Rocky Mountain Front Range: Geol. Soc. America Bull., v. 15, p. 379-448.

Davies, P.J., Bubela, B., and Ferguson, J., 1978, The formation of ooids: Sedimentology, V. 25, p. 705-730.

Dawson, G.M., 1886, Preliminary report of the physical and geological features between latitudes 49° and 51°31': Canada Geol. Survey, Ann. Report, v. 1, 169p.

Dawson, J.W., 1885, Notes and news: Science, v. 5, p. 531-532.

Daetwyler, C.C., and Kidwell, A.L., 1959, The Gulf of Batabano (Cuba), a modern carbonate basin: World Petroleum Cong., 5th, New York, Proc., sec. 1, p. 1-21.

Deffeyes, K.S., Lucia, F.J., and Weyl, P.K., 1965, Dolomitization of Recent and Plio-Pleistorene sediments by marine evaporite waters on Bonaire, Netherlands Antilles: S.E.P.M., Spec. Publ. 13, P. 71-88.

Douglass, Earl, 1905, Some notes on the geology of southwestern Montana: Carnegie Musium, An. 3, p. 407-428.

_____________, 1909, A geological reconnaissance in North Dakota, Montana, and Idaho; with notes on Mesozoic and Cenozoic geology: Carnegie Museum, Ann. 5, nos. 2-3, p. 211-288.

Dunham, J. B., and Olson, E. R., 1978, Diagenetic dolomite formation related to Paleozoic paleogeography of the Cordilleran miogeocline in Nevada: Geology, v. 6, p. 556-559.

Dutro, J.T., Jr., and Sando, W.J., 1963, Age of certain post-Madison rocks in southwestern Montana and western Wyoming: U.S. Geol. Survey Prof. Paper 475-B, p. 93-94.

Eardley, A.J., 1951, Structural geology of NOrth America, New York, Harper, 624p.

Fisher, C.A., 1908, Southern extension of the Kootenai and Montana coal-bearing formations in northern Montana: Economic Geology, v. 3, p. 77-99.

Folk, R.L., 1962, Spectral subdivision of limestone types: in Ham, W.E., (ed.), Classification of carbonate rocks, AAPG Mem. 1, p. 62-84.

__________, 1975, Petrology of sedimentary rocks: Austin, Texas, Hemphill's, 170p.

__________, and Land, L.S., 1975, Mg/Ca ratio and salinity: Two controls over crystallization of dolomite: AAPG Bull., v. 59, p. 60-68.

Frenzel, Hugh, and Mundorff, M.J., 1942, Fusulinidae from the Phosphoria Formation of Montana: Jour. Paleont., v. 16, p. 675-684.

Gardner, L.S., 1959, Revision of Big Snowy Group in central Montana: AAPG Bull., v. 43, p. 329-349.

Gardner, L.S., Hendricks, T.A., Hadley, H.D., and Rodgers, C.D., Jr., 1946, Stratugraphic sections of upper Paleozoic and Mesozoic rocks in south-central Montana: Montana Bur. Mines and Geol. Mem. 24, p. 1-100.

Gebelein, C.D., 1976, Open marine subtidal and intertidal stromatolites (Florida, the Bahamas, and Bermuda): in Walter, M.R., (ed.), Stromatolites, Developments in Sedimentology No. 20, New York, Elsevier, p. 381-388.

Ginsberg, R.N., 1975, Introduction, Ancient siliciclastic examples: in Ginsberg, R.N., (ed.), Tidal deposits, New York, Springer-Verlag, p. 92.

Hagan, G.M., and Logan, B.W., 1975, Prograding tidal-flat sequences: Hutchinson Embayment, Shark Bay, Western Australia: in Ginsberg, R.N., (ed.), Tidal deposits, New York, Springer-Verlag, p. 215-222.

Hallam, A., 1975, Jurassic environments, Cambridge, Cambridge Univ. Press, 269p.

Heckel, P.H., 1972, Recognition of ancient shallow marine environments: in Rigby, J.K., and Hamblin, W.K., (eds.), Recognition of ancient sedimentary environments, S.E.P.M., Spec. Publ., no. 16, p. 226-286.

Heim, H.C., 1962, Petrography of the Dinwoody Formation: unpub. masters thesis, Indiana Univ. Geology Library.

Hoffman, P., 1976, Stromatolite morphogenesis in Shark Bay, Western Australia: in Walter, M.R., (ed.), Stromatolites, Developments in Sedimentology No. 20, New York, Elsevier, p. 261-272.

Holland, F.D., 1952, Mississippian stratigraphy in northwestern Utah and southwestern Montana: AAPG Bull., v. 36, p. 1697-1734.

Holm, M.R., James, W.C., and Suttner, L.J., 1977, Comparison of the Peterson and Draney limestones, Idaho and Wyoming, and the calcareous members of the Kootenai Formation, western Montana: Wyoming Geol. Assoc., Ann. Field Conf. Guidebook, no. 29, p. 259-270.

Honkala, F.S., 1967, The Phosphoria Formation in Montana: Montana Geol. Soc., Ann. Field Conf. Guidebook, v. 18, p. 21-30.

Hoskins, C.W., 1964, Molluscan biofacies in calcareous sediments, Gulf of Batababno, Cuba: AAPG Bull., v. 48, p. 1690-1704.

Howard, J.D., 1972, Trace fossils as criteria for recognizing shorelines in stratigraphic record: in Rigby, J.K., and Hamblin, W.K., (eds.), Recognition of ancient sedimentary environments, S.E.P.M., Spec. Publ. 16, p. 215-225.

Hoyt, J.H., 1967, Barrier island formation, Geol. Soc. America Bull., v. 78, p. 1125-1135.

Huh, O.K., 1967, The Mississippian system across the Wasatch Line, east central Idaho, extreme southwestern Montana: Montana Geol. Soc., Ann. Field Conf. Guidebook, v. 18, p. 31-62.

________, 1968, Upper Paleozoic stratigraphy in southwestern Montana and central Idaho: unpub. Ph. D. thesis, Penn. State Univ., 175p.

Illing, L.V., 1954, Bahaman calcareous sands: AAPG Bull., v. 38, p. 1-95.

Imlay, R.W., 1952, Correlation of the Jurassic formations of North America, exclusive of Canada: Geol. Soc. America Bull., v. 63, p. 953-992.

________, 1957, Paleoecology of Jurassic seas in the western interior of the United States: Geol. Soc. America Mem. 67, p. 469-504.

Irwin, M.L., 1965, General theory of epeiric clear water sedimentation: AAPG Bull., v. 49, p. 445-459.

Katich, P.J., Jr., 1951, Recent evidence for Lower Cretaceous deposits in Colorado Plateau: AAPG Bull., v. 35, p. 2093-2094.

Kazakov, A.V., 1937, The phosphorite facies and the genesis of phosphorites: in Geological investigations of agricultural ores, Trans. Sci. Inst. Fertilizers and Insecto-Fungicides, no. 142, 17th Sess., Int. Geol. Congr., Leningrad, p. 95-133.

Kendall, C.G.St.C., and Skipwith, P.A.d'E., 1968, Recent algal mats of a Persian Gulf lagoon: Jour. Sed. Petrol., v. 38, p. 1040-1058.

King, R.E., 1930, Geology of the Glass Moutains, Texas, Part II, Faunal summary: Texas Univ. Bull. 3042.

Kinsman, D.J.J., and Park, R.K., 1976, Algal belt and coastal sabkha evolution, Trucial Coast, Persian Gulf: in Walter, M.R., (ed.), Stromatolites, Developments in Sedimentology No. 20, New York, Elsevier, p. 421-434.

Klein, G.deVr., 1971, A sedimentary model for determining paleotidal ranges: Geol. Soc. America Bull., v. 82, p. 2585-2592.

______________, and Ryer, T.A., 1978, Tidal circulation patterns in Precambrian, Paleozoic and Cretaceous epeiric and mioclinal shelf seas: Geol. Soc. America Bull., v. 89, p. 1050-1085.

Klepper, A.C., Honkala, F.S., Payne, O.A., and Ruppel, E.T., 1953, Stratigraphic section of the Phosphoria Formation in Montana: U.S. Geol. Survey Circ. 260, 39p.

Krusekopf, H.H., Jr., 1948, Geology of the Tendoy Range near Dell, Beaverhead County, Montana: unpub. masters thesis, Univ. of Michigan.

Kukal, Zdenek, and Saadallah, Adnan, 1973, Aeolian admixture in the sediments of the northern Persian Gulf: in Purser, B.H., (ed.), Teh Persian Gulf, New York, Springer-Verlag, p. 115-121.

Kummel, Bernhard, Jr., 1943, The Thaynes Formation, Bear Lake Valley, Idaho: Am. Jour. Sci., v. 241, p. 316-322.

____________________, 1954, Triassic stratigraphy of southeastern Idaho and adjacent areas: U.S. Geol. Survey Prof. Paper 254-H, p. 165-191.

____________________, 1957, Paleoecology of Lower Triassic formations of southeastern Idaho and adjacent areas: Geol. Soc. America Mem. 67, p. 437-468.

____________________, 1960, The Triassic of southwestern Montana: Billings Geol. Soc., Ann. Field Conf. Guidebook, v. 11, p. 233-238.

Lee, W.T., 1920, Type section of the Morrison Formation: Am. Jour. Sci., v. 49, p. 183-188.

________, 1927, Correlation of geologic formations between east central Colorado, central Wyoming, and southern Montana: U.S. Geol. Survey Prof. Paper 149, 80p.

Logan, B.W., 1961, Cryptozoan and associated stromatolites from the Recent of Shark Bay, Western Australia: Jour. Geology, v. 69, p. 517-533.

__________, Rezak, R, and Ginsburg, R.N., 1964, Classification and environmental significance of algal stromatolites: Jour. Geology, v. 72, p. 68-83.

Longman, M.W., 1980, Carbonate diagenetic textures from nearsurface diagenetic environments: AAPG Bull., v. 64, p. 461-487.

Lowell, W.R., and Klepper, M.R., 1953, Beaverhead Formation, a Laramide deposit in Beaverhead County, Montana: Geol. Soc. America Bull., v. 64, p. 235-243.

Maughan, E.K., and Roberts, A.E., 1967, Big Snowy and Amsden Groups and the Mississippian-Pennsylvanian boundary in Montana: U.S. Geol. Survey Prof. Paper 554-B, p. 1-27.

McKee, E.D., and Wier, G.W., 1953, Terminology for stratification and cross-stratification in sedimentary rocks: Geol. Soc. America Bull., v. 64, p. 381-389.

McKelvey, V.E., Williams, J.S., Sheldon, R.P., Cressman, E.R., Cheney, T.M., and Swanson, R.W., 1956, The Phosphoria, Park City, and Shedhorn formations in the western phosphate field; advanced summary: AAPG Bull., v. 40, p. 2826-2863.

McKerrow, W.S., (ed.), 1978, The ecology of fossils: an illustrated guide: Cambridge, MIT Press, 384p.

McMannis, W.J., 1965, Resume of depositional and structural history of western Montana: AAPG Bull., v. 49, p. 1801-1823.

Miller, A.K., and Cline, L.M., 1934, The cephalopods of the Phosphoria Formation of northwestern United States: Jour. Paleont., v. 8, p. 281-302.

Milliman, J.D., 1974, Marine carbonates, New York, Springer-Verlag, 375p.

Monty, C.L.V., 1976, The origin and development of crypt-algal fabrics: in Walter, M.R., (ed.), Stromatolites, Developments in Sedimentology No. 20, New York, Elsevier, p. 193-250.

Moritz, C.A., 1951, Triassic-Jurassic stratigraphy of southwestern Montana: AAPG Bull., v. 35, p. 1781-1814.

Mundt, P.A., 1956, Heath-Amsden strata in central Montana: AAPG Bull., v. 40, p. 1915-1934.

Newell, N.D., and Kummel, Bernhard, Jr., 1941, The Permian-Triassic boundary in Idaho, Montana, and Wyoming: Am. Jour. Sci., v. 239, p. 204-208.

______________________, 1942, Lower Eo-Triassic stratigraphy, western Wyoming and southeast Idaho: Geol. Soc. America Bull., v. 53, p. 937-995.

__________, Imbrie, J., Purdy, E.G., and Thurber, D.L., 1959, Oraganism communities and bottom facies, Great Bahama Bank: Bull. Am. Mus. Nat. Hist., v. 117, p. 177-228.

__________, Purdy, E.G., and Imbrie, J., 1960, Bahamian oolitic sand: Jour. Geology, v. 68, p. 481-497.

Otros, E.G., 1970, Development and migration of barrier island, northern Gulf of Mexico: Geol. Soc. America Bull., v. 81, p. 241-246.

Paine, M.H., and Suttner, L.J., 1971, Unusual lacustrine limestones in the Kootenai clastic wedge of south-western Montana: (abstr.) Geol. Soc. America Abstr., v. 3, p. 401-402.

Pardee, J.T., 1917, The Garrison and Philipsburg phosphate fields, Montana: U.S. Geol. Survey Bull. 640, p. 195-208.

__________, 1950, Late Cenozoic block-faulting in western Montana: Geol. Soc. America Bull., v. 61, p. 359-406.

Peale, A.C., 1893, The Paleozoic section in the vicinity of Three Forks, Montana: U.S. Geol. Survey Bull. 110, p. 1-56.

Peck, R.E., 1957, North American Mesozoic Charophyta: U.S. Geol. Survey Prof. Paper 294-A, 44p.

Perry, D.G., 1979, A Jurassic brachiopod-oyster association, Twin Limestone, southeastern Idaho: Jour. Paleont., v.53, p. 997-1004.

__________, and Chatterton, B.D.E., 1979, Late Early Triassic brachiopod and conodont fauna, Thaynes Formation, southeastern Idaho: Jour. Paleont., v. 53, p. 307-319.

Perry, E.S., 1962, Montana in the geologic past: Mont. Bur. Mines Geol., Bull. 26, p. 1-78.

__________, and Sloss, L.L., 1943, Big Snowy Group, lithology and correlation in the northern Great Plains: AAPG Bull., v. 27, p. 1287-1304.

Peterson, G.L., 1978, Mesozoic paleoclimatic variations in the western United States and the northward drift of North America (abstr.): Geol. Soc. America, Abstr. with Programs, v. 10, p. 471.

Peterson, J.A., 1954, Marine Upper Jurassic, eastern Wyoming: AAPG Bull., v. 38, p. 463-507.

__________, 1957, Marine Jurassic of northern Rocky Mountains and Williston Basin: AAPG Bull., v. 41, p. 399-440.

Pettijohn, F.J., 1975, Sedimentary rocks, third edition, New York, Harper and Row, 628p.

Pisciotto, K.A., 1978, Basinal sedimentary facies and diagenetic aspects of the Monterey Shale, California, unpub. Ph.D. thesis, Univ. Calif., Santa Cruz, 450p.

Playford, P.E., and Cockbain, A.E., 1976, Modern algal stromatolites at Hamelin Pool, a hypersaline barred basin in Shark Bay, Western Australia: in Walter, M.R. (ed.), Stromatolites, Developments in Sedimentology no. 20, New York, Elsevier, p. 389-412.

Purdy, E.G., 1963, Recent calcium carbonate facies of the Great Bahama Bank, Pt. 2, Sedimentary facies: Jour. Geology, v. 71, p. 472-497.

__________, 1974, Reef configurations: cause and effect: in Laporte, L.F. (ed), Reefs in time and space, S.E.P.M., Spec. Publ. No. 18, New York, Elsevier, p. 9-77.

Read, J.F., 1976, Calcretes and their distinction from stromatolites: in Walter, M.R., (ed.), Stromatolites, Developments in Sedimentology No. 20, New York, Elsevier, p. 55-72.

Reineck, H.E., and Singh, I.B., 1975, Depositional sedimentary environments, New York, Springer-Verlag, 439p.

Richards, R.W., and Mansfield, G.R., 1912, The Bannock overthrust, a major fault in southeastern Idaho and northeastern Utah: Jour. Geology, v. 20, p. 681-709.

Richardson, G.B., 1913, The Paleozoic section in northern Utah: Am. Jour. Sci., v. 36, p. 406-416.

Rock-Color Chart, Reprinted 1963. Distributed by Geol. Soc. America, Boulder, Colorado, no pagination.

Rose, P.R., 1977, Mississippian carbonate shelf margins, western United States: Wyoming Geol. Assoc., Ann. Field Conf. Guidebook, v. 29, p. 155-172.

Ross, C.P., 1962, Upper Paleozoic rocks in central Idaho: AAPG Bull., v. 46, p. 384-387.

Rudwick, M.J.J., 1970, Living and fossil brachiopods, London, Hutchinson, 199p.

Ruppel, E.T., 1978, Medicine Lodge thrust system, east-central Idaho and southwestern Montana: U.S. Geol. Survey Prof. Paper 1031, 22p.

Ryder, R.T., 1967, Late Cretaceous syntectonic deposits in southwestern Montana and east-central Idaho: Diss. Abstr. Int., v. 30, p. 1205B.

__________, 1968, Beaverhead Formation, Late Cretaceous-Paleocene syntectonic deposits in southwestern Montana and east-central Idaho: Geol. Soc. America Spec. Paper 121, p. 259-260.

__________, and Ames, H.T., 1970, Lima region: tectonic events of the Cretaceous-Paleocene Beaverhead Formation: AAPG Bull., v. 54, p. 1155-1171.

__________, and Scholten, Robert, 1973, Synthectonic conglomerates in southwestern Montana: Their nature origin, and tectonic significance: Geol. Soc. America Bull., v. 84, p. 773-796.

Sadler, R.K., 1981, The structure and stratigraphy of the Little Sheep Creek area near Lima, Beaverhead County, Montana: unpub. masters thesis, Oregon State Univ., (in prep.).

Sando, W.J., Gordon, Mackenzie, Jr., and Dutro, J.T., Jr., 1975, Stratigraphy and geologic history of the Amsden Formation (Mississippian and Pennsylvanian) of Wyoming: U.S. Geol. Survey Prof. Paper 848-A, 83p.

Scholten, Robert, 1961, Sedimentation and tectonics in the thrust belt of southwestern Montana and east-central Idaho: Wyoming Geol. Assoc. Guidebook, p. 73-83.

__________, 1967, Structural framework and oil potential of extreme southwestern Montana: Montana Geol. Soc., Ann. Field Conf. Guidebook, no. 18, p. 7-19.

__________, 1968, Model for evolution of Rocky Mountains east of Idaho Batholith: Tectonophysics, v. 6, p. 109-126.

__________, Keenmon, K.A., and Kupsch, W.O., 1955, Geology of the Lima region, wouthwestern Montana and adjacent Idaho: Geol. Soc. America Bull., v. 66, p. 345-404.

Scott, H.W., 1945, Age of Amsden Formation: Geol. Soc. America Bull., v. 56, p. 1195.

Scott, N.B., 1935, Some Carboniferous Stratigraphy in Montana and northwestern Wyoming: Jour. Geology, v. 43, p. 1011-1032.

Scott, W.F., and Wilson, P.C., 1953, Revision and extension of the Sacajawea Formation in Wyoming and Montana (abstr.): Geol. Soc. America Bull., v. 64, p. 1554.

Selley, R.C., 1976, Subsurface environmental analysis of North Sea sediments: AAPG Bull., v. 60, p. 184-195.

__________, 1978, Ancient sedimentary environments, second edition, Ithaca, Cornell Univ. Press, 287p.

Shaw, A.B., 1964, Time in stratigraphy, New York, McGraw-Hill, 385p.

Shinn, E.A., 1968, Practical significance of birdseye structures in carbonate rocks: Jour. Sed. Petrol., v. 38, p. 215-223.

__________, Ginsberg, R.N., and Lloyd, R.M., 1965, Recent supratidal dolomite from Andros Island, Bahama: S.E.P.M., Spec. Publ. 13, p. 112-123.

__________, Lloyd, R.M. and Ginsberg, R.N. 1969, Anatomy of a modern carbonate tidal flat, Andros Island, Bahamas: Jour. Sed. Petrol., v. 39, p. 1202-1228.

Siever, Raymond, 1959, Petrology and geochemistry of silica cementation in some Pennsylvanian sandstones: in Ireland, H.A., (ed.) Silica in sediments, S.E.P.M., Spec. Publ., no. 7, p. 55-79.

Silberling, J.J., 1973, Geologic events during Permian Turassic time along the Pacific margin of the United States: in Logen, A., and Hills, L.V. (eds.), The Permian and Triassic Systems and their mutual boundary: Can. Soc. Petrol. Geol., Mem. 2, p. 345-362.

Skipp, Betty, and Hait, M.H., Jr., 1977, Allochthons along the northeast margin of the Snake River Plain, Idaho: Wyoming Geol. Assoc., Ann. Field Conf. Guidebook, no. 29, p. 499-515.

Sloss, L.L., 1950, Paleozoic sedimentation in Montana: AAPG Bull., v. 34, p. 423-451.

__________, and Moritz, C.A., 1951, Paleozoic stratigraphy of southwestern Montana: AAPG Bull., V. 35, p. 2135-2169.

Solien, M.A., 1979, Conodont biostratigraphy of the Lower Triassic Thaynes Formation, Utah: Jour. Paleont. v. 53, p. 276-306.

Sorby, H.C., 1879, Anniversary address of the President: Structure and origin of limestones: Geol. and origin of limestones: Geol. Soc. London, Proc. 35, p. 56-95.

Storey, T.P., 1958, Jurassic of the Williston Basin and adjacent areas: Alberta Soc. Petroleum Geologists Jour., v. 6, p. 90-104.

Strickland, J.W., 1960, Summary of Mississippian stratigraphy, western Wyoming and vicinity: Billings Geol. Soc., Ann. Field Conf. Guidebook, no. 11, p. 2170223.

Suttner, L.J., 1966, Analysis of the Upper Jurassic-Lower Cretaceous Morrison and Kootenai Formations, southern Montana: unpub. Ph.D. thesis, Univ. of Wisconsin, Madison, Wisconsin.

______________, 1968, Clay minerals in the Upper Jurassic-Lower Cretaceous Morrison and Kootenai Formations, southwest Montana: Earth Sci. Bull., v. 1, p. 5-14.

______________, 1969, Stratigraphy and Petrography of the Upper Jurassic-Lower Cretaceous Morrison and Kootenai formations of southwestern Montana: AAPG Bull., v. 53, p. 1391-1410.

Thompson, M.L.; and Scott, W.H., 1941, Fusulinids from type section of Quadrant Formation: Jour. Paleont., v. 15, p. 349-353.

Vail, P.R., Mitchum, R.M., Jr.; and Thompson, S., 1977, Seismic stratigraphy and global changes of sea level, Part 4: Global cycles of relative changes of sea level: in Payton, C.E. (ed), Seismic stratigraphy-applications to hydrocarbon exploration, AAPG Mem. 26, p. 83-97.

VanHouten, F.B., 1968, Iron oxides in red beds: Geol. Soc. America Bull., v. 79, p. 399-416.

Waldschmidt, W.A.; and Leroy, L.W., 1944, Reconsideration of the Morrison Formation in the type area, Jefferson County, Colorado: Geol. Soc. America Bull., v. 55, p. 1097-1114.

Walker, T.R., 1967, Formation of red beds in modern and ancient deserts: Geol. Soc. America Bull., v. 78, p. 353-368.

Wallace, S.R., 1948, Geology of part of the Tendoy Mountains, Beaverhead County, Montana: unpub. master's thesis, Univ. of Michigan.

Wanless, H.R., 1975, Carbonate tidal flats of the Grand Canyon Cambrian: in Ginsberg, R.N. (ed.), Tidal Deposits, Springer-Verlag, New York, p. 269-278.

Wardlaw, B.R., 1980, U.S. Geol. Survey, Denver Colorado, Pers. commun., March 21, 1980.

Weed, W.H., 1892, Livingston atlas sheet (Montana): U.S. Geol. Survey, Geol. Atlas, Livingston folio, preliminary edition.

__________, 1896, Yellowstone National Park, sedimentary rocks: U.S. Geol. Survey Geol. Atlas, Folio 30.

__________, 1899, Description of the Little Belt Mountains quadrangle: U.S. Geol. Survey, Geol. Atlas Folio 56, 9p.

Williams, Howel, Turner, F.J., and Gilbert, C.M., 1954, Petrography, an introduction to the study of rocks in thin section, San Francisco, Freeman and Co., 406p.

Wilson, J.L., 1970, Depositional facies across carbonate shelf margins: Trans. Gulf Coast Ass. Geol. Soc., v. 20, p. 229-233.

__________, 1975, Carbonate facies in geologic history, New York, Springer-Verlag, 471p.

Wilson, P.C., 1962, Pennsylvanian stratigraphy of Powder River Basin, in Pennsylvanian System in the United States-A symposium: Tulsa, Oklahoma, AAPG, p. 117-158.

Woods, P.J., and Brown, R.G., 1975, Carbonate sedimentation in an arid zone tidal flat, Nilemah Embayment, Shark Bay, Western Australia: in Ginsberg, R.N. (ed.), Tidal Deposits, New York, Springer-Verlag, p. 223-232.

Wright, R.P., 1973, Marine Jurassic of Wyoming and South Dakota; its paleoenvironments and paleobiogeography: Mich. Univ., Mus. Paleontol., Pap. Paleontol., no. 2, 49p.

Wyoming stratigraphic nomenclature chart, 1977, Stratigraphic nomenclature committee (revision, 1969), Wyoming Geol. Assoc., Ann. Field Conf. Guidebook, no. 29.

Yen, T.C., 1951, Fresh-water mollusks of Cretaceous age from Montana and Wyoming: U.S. Geol. Survey, Prof. Paper 233-A, p. 1-20.

Zamarreno, Isabel, 1975, Peritidal origin of Cambrian carbonates in northwestern Spain: in Ginsberg, R.N. (ed.), Tidal Deposits, New York, Springer-Verlag, p. 289-298.