



Thrust plate stratigraphy and structure of the Many Glacier area, Glacier National Park, Montana
by Edwin Lynn Harp

A thesis submitted to the Graduate Faculty in partial fulfillment of the requirements for the degree of
MASTER OF SCIENCE in Earth Sciences

Montana State University

© Copyright by Edwin Lynn Harp (1971)

Abstract:

Roughly 7,000 feet of Precambrian Belt strata are exposed above the Lewis overthrust in the Many Glacier area of Glacier National Park east of the continental divide. Below the overthrust rocks of Cretaceous age occur. The Pre-Ravalli sequence and Ravalli Group of the Belt Supergroup are subdivided into and mapped as units of member status using criteria and nomenclature similar to that used by the Fentons (1937). The only Pre-Ravalli sequence present is the Hell Roaring Member of the Altyn Formation, which consists of 520 feet of blue-gray, variably sandy dolomite. Overlying the Altyn is the Appekunny Formation of the Ravalli Group, which is divided into a lower-Singleshot, middle-Appistoki, and upper-Scenic Point members. The Appekunny consists of roughly 2,200 feet of green argillite, siltite, and white to gray quartzite. Minor reddish strata occur in the Singleshot Member. The overlying Grinnell Formation also contains three members: lower-Rising Wolf, middle-Red Gap, and upper-Rising Bull. The Grinnell consists of 1,800 feet of red argillite and white quartzite with the Red Gap containing very little quartzite. Minor green beds are present throughout. The overlying Siyeh Formation is the only unit within the Piegan Group and consists of roughly 3,000 feet of dolomite, limestone and calcareous shale. This formation contains many laterally continuous stromatolite zones and is divisible into three stratigraphic units; however, it is mapped as an undivided unit. Underlying the Siyeh is the 150-foot thick Snowslip Formation which here consists of red and green argillite and quartzite. The youngest exposed Belt in the area is the Shepard Formation. It is comprised of dolomite, limestone and calcareous shale, only the basal 150 feet are exposed.

The structural features in the map area, besides the Lewis overthrust, are minor thrusts and associated zones of small scale folds. The folds appear to be of two main types: folds with southwest dipping axial planes ("normal") and folds with north or northeast dipping axial planes ("reverse"). Stereonet analysis of the fold orientation indicate that the "normal" folds are related to the Lewis overthrust, and that the "reverse" folds are probably related to structural adjustments after thrusting. Stereonet analysis also reveals that the most probable direction of movement of the Lewis overthrust plate was N 20-30 E. The mineralogy, lithology, thickness of the Belt, and lack of intense deformation of the overthrust plate seem to agree with Hubbert and Rubey's (1959) hypothesis that development of abnormal fluid pore pressure was responsible for thrusting and/or gravity sliding of allocthonous masses of great thickness and displacement. However, apparent porosities of quartzite units studied indicate that they would have been unsuitable media for development of abnormal fluid pore pressures. Studies of the clay from the Lewis fault zone by R. C. Wilson (personal communication, 1970) also disagree with the "abnormal pore pressure" theory.

STATEMENT OF PERMISSION TO COPY

In presenting this thesis in partial fulfillment of the requirements for an advanced degree at Montana State University, I agree that the Library shall make it freely available for inspection. I further agree that permission for extensive copying of this thesis for scholarly purposes may be granted by my major professor, or, in his absence, by the Director of Libraries. It is understood that any copying or publication of this thesis for financial gain shall not be allowed without my written permission.

Signature Edwin L. Harp.

Date 3/26/71

THRUST PLATE STRATIGRAPHY AND STRUCTURE OF THE MANY GLACIER AREA,
GLACIER NATIONAL PARK, MONTANA

by

EDWIN LYNN HARP

A thesis submitted to the Graduate Faculty in partial
fulfillment of the requirements for the degree

of

MASTER OF SCIENCE

in

Earth Sciences

APPROVED:

Milton J. Edie
Head, Major Department

William J. McMannis
Chairman, Examining Committee

K. Goering
Graduate Dean

MONTANA STATE UNIVERSITY
Bozeman, Montana

June, 1971

ACKNOWLEDGMENTS

The writer wishes to express thanks to certain individuals at Montana State University who lent assistance to this study. Special thanks go to the writer's graduate committee, particularly, Dr. McMannis, Dr. Chadwick, and Dr. Dunn for helpful review and criticism of the manuscript and for help with drafting problems. Thanks also go to the writer's advisor, Dr. McMannis, for valuable assistance and direction in the field, and discussion regarding methods and theoretical aspects of the project.

Thanks are also extended to the Earth Sciences Department of Montana State University for use of their facilities.

The writer wishes also to thank Fred Reese for assistance in the field work and typing of preliminary drafts of the manuscript; and Mrs. Garry Dent for typing of the final copy.

TABLE OF CONTENTS

	PAGE
Introduction	1
Location	1
Scope	1
Drainage, Relief and Vegetation	4
Previous Investigation	5
Stratigraphy	8
General Statement	8
Pre-Ravalli Sequence	10
Hell Roaring Member	13
Appekunny Formation	20
Singleshot Member	22
Appistoki Member	25
Scenic Point Member	25
Grinnell Formation	30
Rising Wolf Member	31
Red Gap Member	33
Rising Bull Member	33
Piegan Group	38
Siyeh Formation	38
Snowslip Formation	40
Shepard Formation	41
Precambrian Intrusive Rocks	41

	PAGE
Cretaceous	42
Pleistocene-Quaternary	43
Possible additional methods of Belt correlation	44
Structure	45
General Statement	45
Many Glacier Area	48
Lewis Thrust Fault	49
Structural features above the Lewis Overthrust.	51
Secondary Folds	52
Areal distribution	52
Directional relationships	56
Secondary Faults	61
Grinnell Point	61
Ptarmigan Fault	65
Ptarmigan Tunnel	67
Mt. Altyn	68
Mt. Allen	68
Large Scale Folds	72
Mt. Wilbur anticline	72
Mt. Henkel anticline	72
Mt. Altyn anticline	73
Interpretation	73
Mode of thrust emplacement	78

	PAGE
Summary	81
Appendix	84
References cited	89

LIST OF FIGURES

FIGURE	PAGE
1a. Generalized stratigraphic section	9a
1b. Generalized stratigraphic section	9b
2. Correlation chart	12
3. Diagram of microscopic grain relationships.	34
4. Flat-pebble conglomerate.	37
5. Secondary drag folds.	53
6. Thrust fault.	55
7. Small kink folds.	57
8. Stereographic projection.	59
9. Stereographic projection.	60
10. Stereographic projection.	60
11. Cross section	64
12. Ptarmigan fault	66
13. Thrust fault.	69
14. Thrusts on Mt. Allen	71

PLATE	PAGE
1. Regional tectonic map	2
2. Tectonic map of Many Glacier area	46
3. Geologic map of Many Glacier area	In pocket

ABSTRACT

Roughly 7,000 feet of Precambrian Belt strata are exposed above the Lewis overthrust in the Many Glacier area of Glacier National Park east of the continental divide. Below the overthrust rocks of Cretaceous age occur. The Pre-Ravalli sequence and Ravalli Group of the Belt Supergroup are subdivided into and mapped as units of member status using criteria and nomenclature similar to that used by the Fentons (1937). The only Pre-Ravalli sequence present is the Hell Roaring Member of the Altyn Formation, which consists of 520 feet of blue-gray, variably sandy dolomite. Overlying the Altyn is the Appekunny Formation of the Ravalli Group, which is divided into a lower-Singleshot, middle-Appistoki, and upper-Scenic Point members. The Appekunny consists of roughly 2,200 feet of green argillite, siltite, and white to gray quartzite. Minor reddish strata occur in the Singleshot Member. The overlying Grinnell Formation also contains three members: lower-Rising Wolf, middle-Red Gap, and upper-Rising Bull. The Grinnell consists of 1,800 feet of red argillite and white quartzite with the Red Gap containing very little quartzite. Minor green beds are present throughout. The overlying Siyeh Formation is the only unit within the Piegan Group and consists of roughly 3,000 feet of dolomite, limestone and calcareous shale. This formation contains many laterally continuous stromatolite zones and is divisible into three stratigraphic units; however, it is mapped as an undivided unit. Underlying the Siyeh is the 150-foot thick Snowslip Formation which here consists of red and green argillite and quartzite. The youngest exposed Belt in the area is the Shepard Formation. It is comprised of dolomite, limestone and calcareous shale, only the basal 150 feet are exposed.

The structural features in the map area, besides the Lewis overthrust, are minor thrusts and associated zones of small scale folds. The folds appear to be of two main types: folds with southwest dipping axial planes ("normal") and folds with north or northeast dipping axial planes ("reverse"). Stereonet analysis of the fold orientation indicate that the "normal" folds are related to the Lewis overthrust, and that the "reverse" folds are probably related to structural adjustments after thrusting. Stereonet analysis also reveals that the most probable direction of movement of the Lewis overthrust plate was N 20-30 E. The mineralogy, lithology, thickness of the Belt, and lack of intense deformation of the overthrust plate seem to agree with Hubbert and Rubey's (1959) hypothesis that development of abnormal fluid pore pressure was responsible for thrusting and/or gravity sliding of allocthonous masses of great thickness and displacement. However, apparent porosities of quartzite units studied indicate that they would have been unsuitable media for development of abnormal fluid pore pressures. Studies of the clay from the Lewis fault zone by R. C. Wilson (personal communication, 1970) also disagree with the "abnormal pore pressure" theory.

INTRODUCTION

Location

The present report presents data gathered mainly from the Many Glacier area which is encompassed by parts of three 7 1/2 minute quadrangles (Plate 3) between longitudes $113^{\circ} 30' \text{ W.}$ and $113^{\circ} 52' 30'' \text{ W.}$, and latitudes $48^{\circ} 52' \text{ N.}$ and $48^{\circ} 52' 30'' \text{ N.}$ in Glacier National Park, northwestern Montana. This area includes nearly all of the drainages of Appekunny, Ptarmigan, Swiftcurrent, Grinnell, and Canyon Creeks. Its western border is just east of Swiftcurrent Pass, while its eastern boundary is a line from Wynn Mountain to the midpoint of Appekunny Mountain. It extends as far north as Ptarmigan Tunnel and as far south as Feather Plume Falls in the Grinnell Valley. Nearly all of the field work was done in this area, however, general reconnaissance in other parts of the park established areal relationships.

Scope

The data presented in the present report were gathered mainly during weekends in the summers of 1968, 1969 and 1970. Mapping was done at a scale of 1:24,000 using preliminary copies of 7 1/2-minute quadrangle topographic sheets published by the U.S.G.S. in the spring of 1969.

The purpose of the present investigation was fourfold: to present a more detailed structural picture of the upper plate of the Lewis

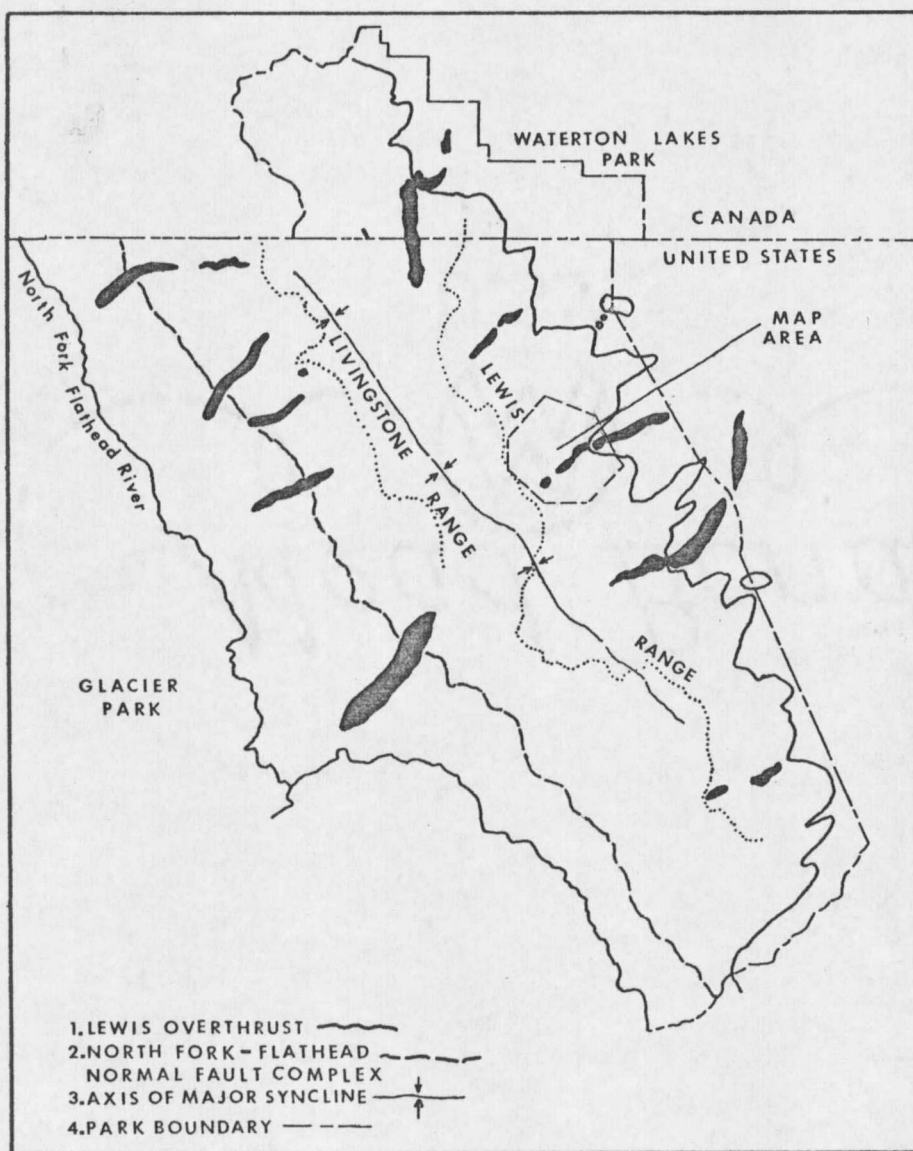


Plate 1. Tectonic map of Waterton-Glacier International Park (after Dyson, 1963).

overthrust as far as the symmetry, directions of movement, and age relations of the deformation are concerned; to establish the relationship of minor structural features to those of regional importance; to subdivide and map as formational units parts of the Pre-Ravalli sequence and Ravalli Group; and to compare the data gathered with proposed theoretical models of overthrusting.

The first objective is that of the need for increased structural detail. This is obvious in view of the scale (1:125,000) and generality of the present published map of Glacier National Park (Ross, 1959). Any study of symmetry of deformation relies upon analysis of minor structural features such as kink folds, drag folds, minor thrusts, etc., and the recently published 7 1/2-minute quadrangle maps provide a convenient base for these purposes.

The second objective results from the assumption that the minor structural features are intrinsically related to major structural features of regional importance. The Lewis overthrust is one of the greatest and earliest discovered thrust faults in western North America, however, relatively little work has been done regarding minor structural features related to it. It was obvious that an attempt should be made to determine, if possible, whether all of the minor structures are related to the great thrust, and if so, in what way. Also, if some of the minor features appear not to be related to the Lewis overthrust, then a relationship to some other feature of major

structural importance should be sought.

The third objective was undertaken in view of previous work by the Fentons (Fenton & Fenton, 1937). They subdivided the formations of the Belt Supergroup into units of member status. A mappable differentiation of the formations of the Belt is necessary for increased stratigraphic control in areas of structural complexity. For the purposes of this report, only the formations of the Pre-Ravalli sequence (Ross, 1963, p. 27) and the Ravalli Group have been subdivided. It is hoped that further study of such aspects as the lateral facies change within the individual members will be attempted so as to refine areal and regional correlation of these units.

The fourth objective stems from the fact that mechanisms and forces responsible for overthrusting still remain in conjecture. It seems appropriate that data from the present report be compared to various established theoretical models of overthrusting and forces related to them.

Drainage, Relief, and Vegetation

The previously mentioned drainages all flow into Lake Sherburne, one of the two man-made lakes extending into the park. The flow from Lake Sherburne empties into the St. Mary River which drains into Canada and eventually into Hudson Bay.

The maximum relief within the area is 4516 feet, with Lake

Sherburne the lowest point at 4788 feet and Mr. Wilbur the highest point at 9304 feet. Southwest facing slopes tend to be relatively gentle; the northeast facing slopes are the most precipitous.

Much of the vegetation in the Swiftcurrent and Grinnell Valleys was destroyed in the 1936 fire. Some of the mature trees remain, but most of the area below treeline is occupied by brush and young conifers.

Previous Investigation

No attempt is made here to provide a complete review of all previous work done in Glacier National Park and the surrounding area, rather, only those publications relevant to the present investigation are cited. For a more complete review of previous publications concerning Glacier Park and surrounding areas, one should consult Ross's work (Ross, 1959, p. 4-7).

Willis (1902) was the first to do geological work in Glacier National Park. He discovered the Lewis overthrust and recognized the Precambrian age of the rocks now referred to as the Belt Supergroup. He thought that the Lewis thrust was an erosion thrust and that the Flathead Valley west of the park was a graben.

C. L. and M. A. Fenton (1931 and 1937) have written several papers dealing with organic remains, sedimentary structures, and stratigraphy, of which only two are here cited. They subdivided most of the formations of the Belt Supergroup exposed in the park into units of member status but did not map them. They also described various sedimentary

structures and classified many stromatolite zones.

To attempt to provide an added criterion to regional correlation to Belt stratigraphy, Rezak (1957) described and differentiated eight stromatolite zones of the two main algal genera. He discussed the lateral continuity of the zones and included comparisons of the Belt stromatolitic morphologies and apparent environments with modern algal deposits. He also proposed a slightly different subdivision of the Siyeh Formation than that of the Fentons.

Ross (1959) has provided a rather complete summary of the general geology of the park and adjoining Flathead region. He discussed the major structural, stratigraphic, and geomorphic features and mapped them at a scale of 1:125,000. He chose not to subdivide the formations of the Belt as did the Fentons. As a result his descriptions of the stratigraphic units are too generalized to be of much use, and his locations of contacts are fairly inaccurate.

Shelden (1961, M.S. thesis, Montana State College) described the general geology of the northwest Ural quadrangle west of Eureka, Montana. He discussed the Belt stratigraphy in detail and thereby providing data useful in correlation of Belt strata in the area of the present report with Belt strata of northwestern Montana.

Childers (1963) discussed the structure and stratigraphy of the Marias Pass quadrangle. His stratigraphic description resulted in the definition of three new formation names. He defined the Snowslip Forma-

tion immediately above the Siyeh Formation and two formations below the Missoula Group as the Shields Formation and Red Plume Quartzite; these two together form his Kintla Group. He established the presence of the Roosevelt fault and named several other normal faults of the North Fork normal fault zone.

Smith and Barnes (1966) proposed regional correlation of the major carbonate sequences of the Belt Supergroup from Glacier National Park to western Montana and Idaho. They correlate the Pre-Ravalli sequence with the Prichard Formation, the Piegan Group with the Wallace Formation, and the Missoula Group with the upper Purcell Series of Canada.

Mudge, Erickson, and Kleinkopf (1968) studied the geology, geophysics, and geochemistry of the southeastern part of the Lewis and Clark Range. Their stratigraphic description of the Belt provided a basis for correlation of the Belt units between that area and the area of the present report.

Mudge (1970) proposed gravity slide phenomena as being responsible for the development of the disturbed belt of northwestern Montana and southern Canada. He stated that the initial decollement took place at a sediment depth of about 25,000 feet as the result of the development of excess fluid pore pressure at this horizon and that the initial eastward dip of Belt sediments at the time of sliding was around 8 degrees.

Wilson (1970) ran compression tests on standard clays and clays

from the Lewis thrust and concluded that no excess fluid pore pressure was necessary for thrusting or gravitational sliding. He found that previous compression of the clays alone reduced frictional coefficients to values sufficiently low for either overthrusting or gravitational sliding.

STRATIGRAPHY

General Statement

The area studied contains rocks varying in age from Precambrian to Recent. For the purposes of this investigation, only those of Precambrian age were examined, and of the Precambrian, only those of the Pre-Ravalli sequence and the Ravalli Group of the Belt Supergroup were studied in detail. The formations of the Ravalli and Pre-Ravalli were subdivided into member status using basically the same criteria as those used by the Fentons (1937). Certain differences regarding the selection of contacts were inevitable, however, due to the gradational character of the rocks near the contacts.

Most of the rocks cropping out in the map area are carbonates, argillities, siltites, and quartzites of the Belt Supergroup. The oldest exposed Belt is the Hell Roaring Member of the Altyn Formation. The youngest exposed Belt in the area is the Shepard Formation, (Plate 3), which lies stratigraphically immediately below the Kintla Group of

Generalized stratigraphic section of the Belt Supergroup			
Stratigraphic unit		General Lithology	Thick- ness Feet
Shepard Fm.		Dolomite, limestone, and calcareous argillite and shale. Dark gray to buff. Top eroded.	150
Snowslip Fm.		Red and green argillite and white to gray quartzite.	150
Piegan Group	Siyeh Fm. Gray to buff dolomite, limestone, and calcareous argillite. Contains scattered stomatolites. Dark gray massive zone of <u>Conophyton inclinatum</u> . Gray to buff dolomite, limestone, and calcareous shale and argillite. Contains scattered <u>Collenia symmetrica</u> .		3,000
Ravaali Group	Grinnell Fm.	Rising Bull Mbr.	907
		Red Gap Mbr.	726
		Rising Wolf Mbr.	163

Figure 1A.

Stratigraphic unit			General Lithology	Thick- ness Feet
Ravalli Group	Appokunny Fm.	Scenic Point Mbr.	Green argillite, siltite, and impure buff to rusty quartzite. Con- tains abundant mud chip conglomerate.	992
		Appistoki Mbr.	Green argillite and siltite. Minor quartz- ite.	688
		Singleshot Mbr.	Green and maroon argil- lite and siltite. Bed- ding thin to thick. Massive quartzite unit at base.	541
Pre- Ravalli	Altyn Fm.	Hell Roaring Mbr.	Blue-gray to white, thin to thick bedded dolo- mite. Top 300 feet variably sandy.	526
fault				
Figure 1B.				

Childers (1963, p. 147).

Underlying the Lewis overthrust are rocks of Cretaceous age. At Marias Pass, Childers (1963, p. 154) reports subthrust rocks from the lower Blackleaf Formation of Early Cretaceous age. Exposures of Cretaceous rocks below the thrust in other areas (Dry Fork Valley, Roes Creek, etc.) have been labeled as Colorado Group (Wilson, R. C., personal communication, 1970). Exposures of the Cretaceous within the map area are extremely poor due to the cover of vegetation and glacial drift. The Cretaceous is undifferentiated (Ku, Plate 3).

Pleistocene glacial deposits, landslides, and talus occupy the lower portions of the Grinnell, Swiftcurrent, Ptarmigan, and Canyon Creek drainages. These deposits are mapped only as Quaternary undifferentiated (Qu, Plate 3).

Pre-Ravalli sequence

Of the Pre-Ravalli sequence, only the Hell Roaring Member of the Altyn Formation is present within the map area. The Waterton Formation (Daly, 1912) is the oldest Precambrian Belt exposed above the Lewis overthrust. It was previously grouped with the Hell Roaring and Carthew Members within the Altyn Formation (Fenton & Fenton, 1937), but has since received formation rank. The Waterton and Altyn Formations were originally placed in the Ravalli Group; however, since the sequence is composed mainly of carbonate, it represents a depositional condition sufficiently distinct from those existing during deposition

of the argillaceous and quartzitic Appekunny and Grinnell Formations (Smith and Barnes, 1966, p. 1411-1412) to warrant removing them from the Ravalli Group. The Pre-Ravalli sequence is the first exposed carbonate phase of the cyclical Belt sedimentational units. The Pre-Ravalli carbonate sequence of Glacier National Park has been correlated with the upper part of the Prichard Formation of western Montana and northern Idaho (Smith and Barnes, 1966, p. 1410-1413). This carbonate phase of Pre-Ravalli deposition (Waterton-Altyn sequence) has also been correlated with the type Newland Formation of the Big Belt Mountains. Sheldon (1961, M.S. thesis, Montana State College, p. 14) believed that he found a western "wedge-edge" of the Altyn Formation which intertongues with the upper part of the Prichard Formation in the N.W. Ural quadrangle west of Eureka, Montana.

The Waterton Formation is reported to be composed of dark gray to reddish dolomites weathering gray, reddish, and buff. The formation contains no sand grains and is finely laminated. The maximum thickness is approximately 280 feet. This formation is exposed only in Waterton Lakes Park and the southern Clarke Range in Canada (Fenton and Fenton, 1937, p. 1882-1883).

The upper (Carthew) member of the Altyn Formation is estimated to range from 700-900 feet in thickness (Fenton and Fenton, 1937, p. 1884). It consists of limestones, dolomites, and quartzites ranging in color from blue-gray to reddish-brown with bedding thin to thick. This member

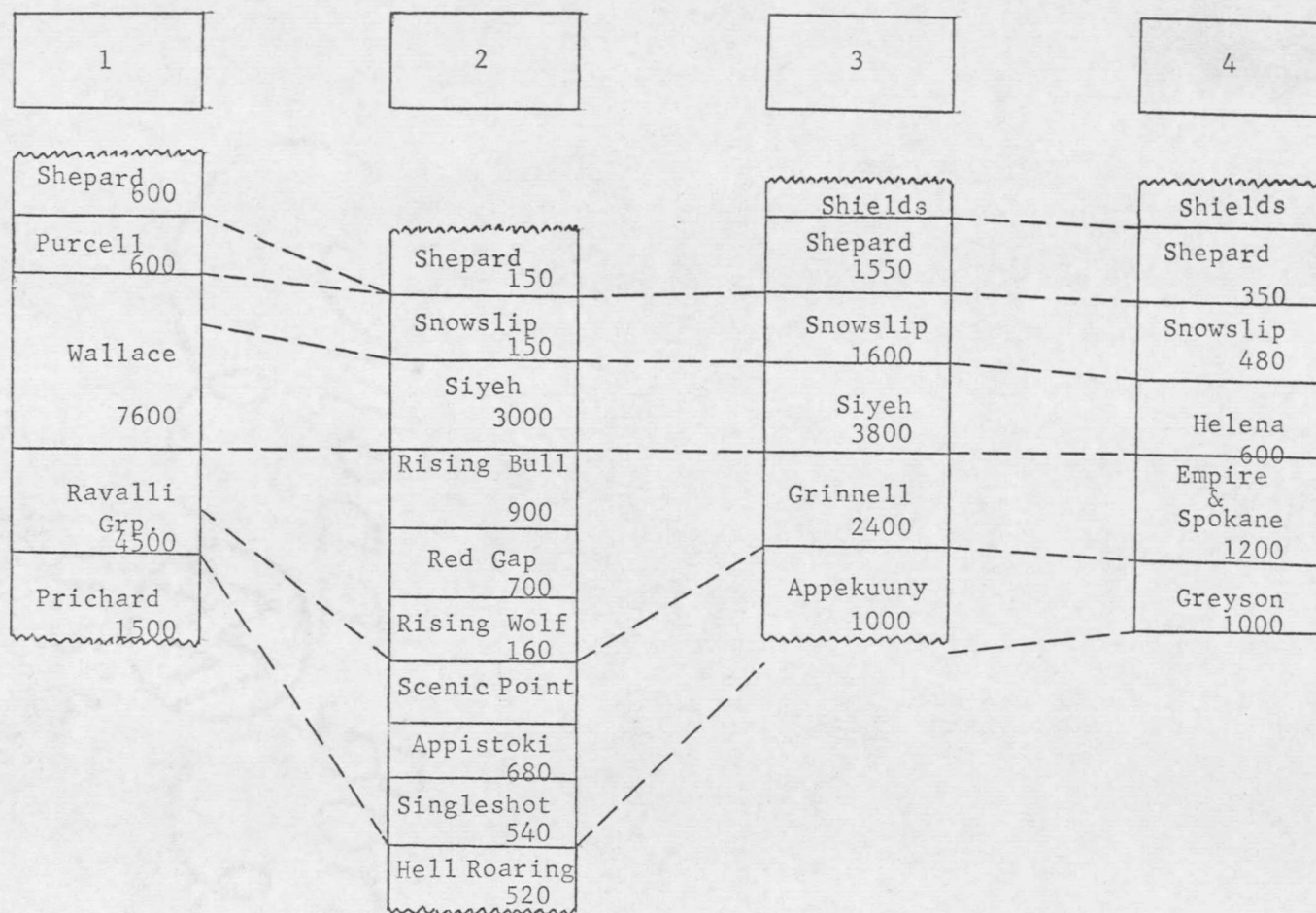


Figure 2. Correlation chart of formations of the lower Belt Supergroup. Sources: 1, Northwest Ural quadrangle (Shelden, 1961, M.S. thesis, Mont. State Coll.). 2, Many Glacier area (this report). 3, Southwest Marias Pass area (Childers, 1963). 4, Southeast Lewis and Clark Range (Mudge, Erickson, and Kleinkopf, 1968).

is supposedly eliminated by a "major subsidiary fault" (Willis' second Lewis thrust) from Yellow Mountain southward. The writer has examined the strata on Mt. Altyn and Mt. Appekunny, and in most of the Many Glacier area, and no "second Lewis thrust" (Willis, 1902) was observed. Secondary thrusting displacing the Altyn and Appekunny Formations is obvious on Vimy Peak in Waterton Lakes Park and at Scenic Point in the Two Medicine area, but no major thrusting within this stratigraphic horizon is evident in either the St. Mary or Many Glacier areas. Also, since the Carthew Member exposed upon the north face of Yellow Mountain appears to be only 200-300 feet thick, it is probable that this member of the Altyn Formation stratigraphically pinches out southward and is not removed by faulting.

Hell Roaring Member - The Hell Roaring Member (Fenton and Fenton, 1937) is the only part of the Altyn Formation exposed in the Many Glacier area. A section approximately 520 feet thick was measured (Appendix) on the southeast slope of Mt. Altyn. A sizeable portion of the Hell Roaring Member is presumably eliminated by the Lewis overthrust at this locality, since at the type section (Hell Roaring Creek in Waterton Lakes Park) it is roughly 1200 feet thick, and on Divide Mountain south of St. Mary Lake, the member is also approximately 1200 feet thick. On the slopes of Mt. Altyn it consists of blue-gray to greenish dolomite weathering gray, buff, white, and orange. It is variably sandy in all

but the lower 100 feet. The bedding varies from thin to thick. The upper 420 feet, in many places, contain quartz and chert grains up to 10 mm in diameter. Many layers contain irregular stringers of silica which appear to be fracture filling. Near the top of the member are several strata containing cross-bedded, sandy lenses. The top of the Hell Roaring Member is transitional with the overlying Singleshoot Member of the Appekunny Formation as intercalated beds of greenish argillite and siltite begin to appear near the top of the Altyn. On the north face of Mt. Wynn, sections of sandy zones of the Hell Roaring Member are repeated in a series of fault wedges, however, the lowest exposures of the Hell Roaring Member on Mt. Altyn, about five miles to the north, are relatively undisturbed and contain no apparent faults to any magnitude. The Fentons (1931) describe a 20 foot thick bed of dolomite within the upper part of the Hell Roaring Member which consists mainly of colonies of Collenia columnaris (Collenia frequens, Rezak). In the section measured for the present report, this 20 foot section is absent. Instead, four widely separated stromatolite zones (presumably Collenia frequens) were found, the largest being two feet thick (see Appendix).

The Fentons (1937) also describe carbonaceous films in the "dolomitic limestone" just above the Lewis thrust, between Appekunny and Swiftcurrent Falls as belonging to a zone of Beltina and Morania antiqua n. sp. This "carbonaceous" zone is reported to be some 200

feet thick. This zone was studied at Roes Creek near Rising Sun Camp ground where the Altyn Formation is injected with dark slickensided kaolinite along fractures. The zone near Swiftcurrent Falls is not nearly as intensely fractured and does not contain as much dark material, but the material seems to be the same as that occupying fractures in the Altyn at Roes Creek. The material has been determined by X-ray study to be kaolinite, with the black color imparted by sulfides (Wilson, R.C., 1970, personal communication).

In microscopic study of the Hell Roaring Member, quartz, feldspar, and chert can be seen to comprise as much as 50 per cent of the rock. Quartz dominates in all sections studied, and the maximum feldspar percentage observed is 10-12. The quartz in general ranges from 0.4 to 0.8 mm in diameter and is subangular to subrounded. Much of it appears to be "stretched" (Folk, 1968, p. 70) with extremely crenulate internal boundaries within the composite grains. Many of the composite grains have subparallel extinction, and practically all grains exhibit strongly undulose extinction. This suggests a metamorphic source such as the Precambrian craton to the east. Most of the quartz grains contain vacuoles, but with no apparent pattern. Many grains have small needle-like inclusions thought to be rutile. A few inclusions of short prismatic crystals with high interference colors are possibly zircon, but they are too small to yield interference figures.

In the nine thin sections of the Hell Roaring Member that were

studied, the feldspars appear to be partly sericitized. All the feldspar grains are anhedral and most have ragged boundaries in contact with the carbonate micro-spar matrix. Some of the feldspar in the Altyn-Appekunny transition zone contains inclusions which appear to be micas altered to chlorite. The average grain size of the feldspars is slightly smaller than that of the quartz grains. Roundness and sphericity values are also somewhat less for the feldspar grains than for the quartz. Generally the thin sections contain subequal amounts of microcline and Na plagioclase. Some orthoclase has been tentatively identified, but grains suitable for a Becke line test are lacking.

The chert grains in thin section average near 3 to 4 per cent of the rock and are generally larger than other grains, having diameters of up to 3 mm. One section, however, contains as much as 20 per cent chert. In this section, the chert appears to have completely replaced many oolites and is in various stages of replacing others. Found also in this section were fragments of oolitic rock which had been silicified prior to their latest deposition. The roundness and sphericity generally appear to be lower than for other minerals. Many of the grains display reentrant boundaries in contact with the carbonate matrix and many chert grains have undergone some recrystallization to quartz.

Allochemical constituents within the rock consist of intraclasts and oolites from 0.2-2.0 mm in diameter. These intraclasts consist of micrite, and where present, generally comprise from 5-10 percent of the

rock. One specimen contains about 40 per cent intraclasts. The oolites most commonly have a nucleus of micro-spar; however, nuclei of silt sized quartz and chert are not uncommon. Some of the intraclasts and oolites have recrystallized from micrite to spar. The micro-spar which generally makes up the oolites undoubtedly is recrystallized from micrite.

The matrix is generally made up of micrite and micro-spar, with the micro-spar being the most abundant. The matrix comprises as much as 85 per cent of the rock. In some sections, faint traces of algal growth lines are present throughout the matrix. The cell structure of the algae seems to have been destroyed, possibly by the recrystallization of micrite to micro-spar. Sparite occupies as much as 25 per cent of many thin sections. It appears in patches and may be filling original pores in some cases, but generally appears to be crystallizing from an original micrite matrix. The sparry cement intrudes all detrital grain boundaries with which it is in contact. The spar is mostly about 0.1 mm in diameter or less, but occurs as coarse as 0.9 mm in diameter. It occurs both as anhedral interlocking crystals and euhedral rhombs. In some of the sandier zones the matrix contains an appreciable amount of clay which has undergone partial recrystallization to sericite. The sericite is present as randomly scattered microlites.

The last cement formed appears to be quartz as vein filling and overgrowths. The sparry cement does not breach the quartz cement

boundaries but is included within the quartz in veinlets where they are both present.

In most of the thin sections studied, there are opaques which are identified as pyrite with hematite as its alteration product. The pyrite constitutes as much as 2-3 per cent of the rock and is present as both euhedral cubes and pyritohedrons of about 0.1 mm and as disseminated clay sized particles. The euhedral crystals have penetrated boundaries of original detrital grains, but have not intruded any of the quartz cement or sparite. Therefore, it appears that the formation or recrystallization of sulfides was the first diagenetic event to take place after deposition. Harrison and Grimes (1970, p. 45) report that most of the sulfides present in Belt rocks are primary detrital grains, but have undergone varying degrees of diagenetic recrystallization. A sequence of diagenetic mineral formation would be as follows (Figure 3): (1) formation of pyrite and other sulfides, (2) recrystallization of micrite to micro-spar and sparry cement, and (3) precipitation of quartz cement in veinlets and as overgrowths. The quartz cement may be the result of tectonic deformation since it appears to be the last cement formed. This paragenetic sequence, incidentally, seems to hold true for all the Appekunny and Grinnell rocks studied in thin section as well.

The environment during Hell Roaring time was apparently shallow marine with periods of fluctuating wave energy. The presence of intra-

clasts, oolites, and quartz grains in a micrite matrix indicates the mixing of constituents from two different environments. The micrite had to have been precipitated in relatively quiet waters below effective wave base, while the quartz and allochems came from shallower, more agitated waters. The admixing of the products of the two environments could have taken place by turbidity currents or an increase in wave energy (via storms) to form the intraclasts and transport the quartz and allochems to deeper waters where they could settle to the bottom into the micrite.

The intraclasts were formed in the basin of micrite deposition by the same high energy conditions which transported the oolites and quartz grains.

A possible alternative is that the micrite was deposited in a quiet lagoonal environment or broad, shallow, barrier-protected platform while the oolites and quartz grains were blown or washed across the barrier from the foreshore probably by storms.

The rather simple mineralogy, the limitation of detrital grains to nothing larger than occasional granule, and the fair degree of rounding of the detrital grains suggest a very low terrane and stream gradient from depositional site to source areas (Harrison and Grimes, 1970, p. 1-2). The sediment was exposed to chemical weathering and reworking for fairly long periods of time as is indicated by the relative scarcity of feldspar and complete absence of ferromagnesian minerals.

The precambrian metamorphic rocks of the Canadian Shield are the likely source rocks and are for the most part granitic (Harrison and Grimes, 1970, p. 42-43); however, even granitic sources generally have associated ferromagnesian minerals, and were the terrane not of low relief and transport slow, they should be present in the sediment. Also since most granitic rocks contain subequal amounts of quartz, K-feldspar, and plagioclase, the elimination of feldspar by chemical weathering has been fairly thorough. Circulation may have been fairly restricted and reducing conditions prevalent at the time of Hell Roaring deposition as suggested by the presence of pyrite; however, the sulfide formation may only reflect the chemical environment after deposition and during diagenesis.

The Pre-Ravalli sequence is correlated with the Prichard Formation of western Montana and northern Idaho (Smith and Barnes, 1966, p. 1410-1411). The dolomite part of the Pre-Ravalli depositional sequence in Glacier Park represents the eastern edge of the Belt-Purcell basin of deposition (Smith and Barnes, 1966, p. 1412). Both the Pre-Ravalli of the park and the Prichard Formation contain disseminated sulfides. Sheldon (1961, M.S. thesis, Montana State College) records finding carbonate beds which he says represent the western wedge-edge of the Altyn Formation.

Appekunny Formation

Smith and Barnes (1966, p. 1412) have stated that the vertical

transition from the Pre-Ravalli sequence to the Ravalli Group represents a transition from "an earlier, deeper water, predominantly reducing environment to a later, shallow water, predominantly oxidizing environment where water circulation was less restricted." The Ravalli strata are generalized as "greenish, reddish, or purplish (hematitic), less evenly laminated, lacking pyrite and abundant carbonate." The Appekunny Formation as a whole in the Many Glacier area contains disseminated pyrite and possibly other sulfides throughout its vertical range. The Appekunny in general consists of greenish and reddish argillites, siltites, and quartzites. It consists of a lower (Single-shot), middle (Appistoki), and upper (Scenic Point) member. In the map area the lower two members do not contain appreciable amounts of ripple marks, mud cracks, or mud breccias. The uppermost member contains all of these sedimentary features in abundance. Therefore, the transition from an earlier, deeper water, reducing environment to a shallow water, oxidizing environment may be represented by the two lower members of the Appekunny. The upper member definitely represents a shallow water environment. The sulfides present may represent early diagenesis instead of primary precipitation, since the sulfides seem to penetrate all grain boundaries with which they are in contact (microscopic description). Presence of reddish beds in the Appekunny within the map area are confined to the Singleshot Member.

Singleshot Member - The Singleshot Member (Fenton and Fenton, 1937) of the Appekunny Formation is exposed on the lower slopes and cliffs of Mt. Appekunny, Mt. Altyn, and Mt. Allen. A section was measured on Mt. Altyn (Appendix) where it is 541 feet thick. At this locality the Singleshot Member consists of green and maroon to reddish-brown argillites and siltites with interbedded layers of quartzite.

The basal unit consists of a massive white to greenish white, medium grained quartzite 50 feet thick. The Fentons (1937, p. 1885) describe a "slight angular unconformity" between the Altyn and this unit, but within the map area, no obvious unconformity is present. The contact with the underlying Altyn appears gradational with thin quartzite beds and argillaceous dolomite immediately below the quartzite. The beds of the Singleshot Member are conspicuously lacking in mud breccias, ripple marks, and mud cracks. The argillites and siltites commonly contain dark laminae which might be carbonaceous films. The Fentons (1937) report that the Singleshot contains these features in other areas of the park, but none in any appreciable amount were found in the map area.

In the thin sections examined the quartz constitutes as much as 62 per cent and as little as 25 per cent of the rock. In the siltites, the quartz grains, being silt sized have lower values of sphericity than that in quartzites, i.e., about 0.5 (using terminology from Krumbein and Sloss, 1963, p. 111). Roundness of grains is also poor, being around

0.3. Extinction is straight to undulose, and the grains tend to display a rude alignment parallel to bedding. In the quartzite beds, the quartz grains have sphericity values of up to 0.6 and roundness values of 0.5 to 0.7. The diameter ranges from .05 to 1.0 mm with the average being about 0.6 to 0.7 mm. Extinction is generally undulose and there are many composite and semi-composite grains. Quite a few of the composite grains have the crenulate internal boundaries of stretched metamorphic type quartz. Vacuoles are abundant. Inclusions of sericite and possibly zircon are present. Pressure indentations between grains and overgrowths are common.

The feldspars were observed to constitute from 3 to 10 per cent of the quartzitic beds. The roundness and sphericity of the grains are roughly the same as for quartz. The majority of the grains are relatively unaltered; however, many are almost replaced by sericite. Many of the grains are composite with quartz and most have pressure indentations with other grains. In the sections examined plagioclase tends to dominate in content over microcline.

Chert grains comprise about 3 per cent of the sections examined. They have about the same sphericity and roundness values as quartz and feldspar. In some cases they contain some sericite as inclusions.

Intraclasts from 0.1 to 2.0 mm in length were found in one section in the amount of 1-2 per cent. The clasts are spar and micro-spar recrystallized from micrite. Some of the clasts are rimmed with chert.

A mixture of clay, sericite, and fine crystalline spar generally makes up the matrix. One section contains 75 per cent matrix. Another specimen contains carbonate only as a matrix. The lowest percentage of matrix is 25 per cent. The sericite present appears to be recrystallized from the clay. The spar likewise is a recrystallization product of micrite. There is almost no micrite left, and the spar is present as anhedral and euhedral crystals. Both the sericite and the spar penetrate grain boundaries. As in the sections of the Hell Roaring Member, these recrystallization products penetrate all constituents except the quartz overgrowths.

Pyrite and possibly other sulfides exist in the rocks in amounts as great as 1-2 per cent. They occur both as finely disseminated deposits and as cubic crystals of silt size and smaller.

The quartzites of the Singleshot Member indicate a high energy environment while the siltites and argillities suggest deposition in deeper water under quiet conditions. The scarcity of intraclasts, mud cracks, and ripple marks indicates that most of the deposition took place in a fairly low energy regime. The set of conditions during Singleshot deposition would seem to be fairly unique, that is, waves were of sufficient energy to produce rather pure, fairly well sorted sands yet they were low enough not to produce appreciable intraclasts. Of course, many of the quartzites and siltites are poorly sorted and contain large clay fractions.

Appistoki Member - The Appistoki Member (Fenton and Fenton, 1937) is a fine grained phase of the Appekunny. It consists of greenish to greenish-gray argillites and siltites in thin to thick beds. The beds weather rusty, buff, tan, and orange. The sequence contains only one 25-foot thick quartzite unit near the top. At the top of the member there is an 8-foot bed of impure sandstones containing flat pebbles. In discussing the Appistoki, the Fentons (1937, p. 1886) describe the member as containing abundant mud cracks, ripple marks, and mud chip breccias, all of which are rare in the map area. Aside from this difference their lithologic description matches that of the Appistoki of the map area, especially in the relative lack of quartzite. The most striking unit within the Appistoki is a 116-foot thick siltite which is massive, dark greenish, and weathers to buff and orange, the orange color being the striking feature. The Appistoki was not studied in thin section; however, its siltites and argillites appear very similar in hand specimen to the corresponding rock types of the Singleshott Member. The depositional environment of the Appistoki would seem to be one of sedimentation in quiet waters below effective wave base. The first signs of shallow water, high energy deposition become evident near the top of the Appistoki where it grades into the Scenic Point Member.

Scenic Point Member - On Mt. Altyn nearly 1000 feet of beds assigned to the Scenic Point Member (Fenton and Fenton, 1937) have been measured.

The Fentons (1937, p. 1887) list the thickness as ranging from 200 to 700 feet. It is probable that the contact used by the writer between the Scenic Point and Appistoki members is lower than used by the Fentons. The Scenic Point Member on Mt. Altyn contains abundant shallow water features such as mud breccias, mud cracks, ripple marks, and cross bedding practically throughout the member. The lithologic types are argillaceous, calcareous, sandstones, sandy argillites, siltites, and argillites. The colors include dark to light green, tan, buff, white and minor reddish-brown. Bedding thickness is generally thin to medium. Units which contain impure, variably friable sandstones are interbedded with argillites. As a result argillitic beds commonly display sand grain imprints on bedding surfaces. A thin zone of pyrite molds from which the pyrite has since weathered out is present 150 feet below the top of the member (Appendix). All that remains are the impressions of the pyrite cubes and pyritohedrons along with abundant hematites staining. The uppermost 10 feet of the member contains intercalated red argillite, as the contact is gradational, and for the purposes of convenient mapping, the contact is drawn at the horizon where the green color diminishes and red becomes dominant. This contact is undoubtedly a variable horizon throughout the park. On Grinnell Point the transition zone from Appekunny to Grinnell is thicker. At this locality the thickness from the horizon showing the first red argillite to the horizon where red becomes dominant is about 50 feet. The variability

of the thickness of this zone is to be expected in view of the fact that some of the beds of the Grinnell have been mottled with green because of chemical alterations presumably produced during deformation. Petrographic analysis was done on specimens from argillite zones containing mud breccias, and from zones containing calcareous sandstones. Quartz grains in the argillites comprise approximately 3-4 per cent of the rock. Two size grades are present; a size ranging from 2.0 to 0.5 mm and a size ranging from coarse silt to finer grain sizes. A few grains contain abundant vacuoles; most do not. Inclusions and microlites (probably rutile) are common. Extinction is straight to strongly undulose. The sphericity and roundness of the larger size is greater than that of the smaller size grade.

Intraclasts comprise about 20 per cent of the sections. They are made up mostly of clay with possibly some disseminated sulfides imparting a dark color. They average about 5 mm in length. They are not rounded appreciably and seem not to have traveled great distances. When deposited they apparently were in a semi-consolidated state as indicated by their "draping" over the quartz grains in response to compaction.

The matrix of the argillite breccias constitutes about 75 per cent of the rock, being composed of clay, sericite, and some carbonate. The carbonate is present as fine crystalline rhombs of spar that apparently have recrystallized from micrite. Traces of pyrite are present as indistinct blebs, largely altered to hematite.

The quartz grains comprise as much as 42 per cent of the sandy zones of the Scenic Point Member. They range from 0.1 to 1.2 mm in diameter. There are some silt sized grains present in the matrix. The larger grains have sphericities of 0.7 or better while the smaller grains have values of about 0.5. The roundness of the larger grains is 0.6 and that of the smaller fraction is 0.4. Vacuoles are abundant, and extinction is straight to undulose. Where grains are in contact with each other, there are pressure indentations. Much of the quartz is composite with both crenulate and straight intragrain boundaries of the "stretched" and "recrystallized" metamorphic types (Folk, 1968, p. 74). Overgrowths are relatively abundant.

The feldspars have about the same morphological relations as the quartz grains and range from 0.2 to 0.5 mm in diameter. They constitute up to 10 per cent of the rock.

The chert in the slides studied occupies around 6 per cent of the sections. The sphericity averages about 0.7 and the roundness about 0.5. Grain diameters range from 0.1 to 1.3 mm. A few chert grains are composite with sparite intraclasts indicating formation in a previous environment and subsequent reworking. Also some of the chert contains disseminated micrite as though the chert and micrite were precipitated simultaneously.

Intraclasts are present in amounts of up to 12 per cent. They range in length from 0.2 to 5.0 mm in long dimension. They consist of both

micrite and sparite and are well rounded. Some of the grains are composite with chert and quartz.

Matrix constitutes up to 30 per cent of the rock and apparently consisted of clay and/or micrite originally. The micrite is in various stages of recrystallization to sparry cement. The cement comprises as much as 20 per cent of the rock and has penetrated all grain boundaries with which it is in contact. The only constituent not intruded by the sparite is the quartz cement that occurs as overgrowths. In fact some of the overgrowths penetrate sparite boundaries. The paragenetic sequence of diagenetic minerals is the same as that for the Altyn Formation, that is, the sparite formed by recrystallization before the quartz cement was precipitated.

The association of micrite with clasts from a high energy environment is the same association as in the Hell Roaring Member of the Altyn Formation, that is, clasts from a high energy environment have been transported to areas of low energy where micrite was being deposited. Most of the sands and argillaceous deposits of the Scenic Point Member reflect very shallow water features, such as mud cracks, ripple marks, mud breccias, etc. The association of clay with grains and intraclasts suggests a mechanism which either mixes or poorly sorts the constituents. The fact that sand grains and intraclasts fill many of the cracks in mud cracked clays and leave imprints on many argillite bedding surfaces suggests that the mechanism of deposition had to be of sufficient energy to

transport the grains and intraclasts to the mud cracked surfaces, yet be sufficiently gentle so as not to rip up all of the mud. Once the mud surface was buried, the environment could increase in energy without disturbing the mud cracks. To develop a mud cracked surface, a subaerially exposed mud flat is necessary. Sand grains from a high energy environment could be either washed into the mud flat by waves during storms, or be blown in by the wind, and mixed with intraclasts, possibly developed from the curled up tops of the mud cracked surfaces.

Grinnell Formation

The main lithologic types of the Grinnell Formation in the Glacier Park area are red argillite and white quartzite with minor green argillite. The beds contain such sedimentary structures as flat pebble conglomerate, mud cracks, cross beds, ripple marks, and worm burrows. The Grinnell has been divided into three members: the Rising Wolf, Red Gap, and Rising Bull in ascending order. The main feature differentiating the members is the relative abundance of quartzite and argillite. The middle (Red Gap) member is an argillitic phase with relatively little quartzite. The lower and upper members both have sizeable proportions of quartzite and green argillite.

In a recent letter to the Geologic Names Committee from Jack E. Harrison of the U.S. Geological Survey, the name Spokane Formation was proposed to supercede the Grinnell Formation. This would be the result of adopting the nomenclature from the Belt Mountains for the eastern

part of the Belt province, which includes the Glacier Park region. No change was advocated for the rocks in the Glacier Park region stratigraphically below the Grinnell.

Rising Wolf Member - The Rising Wolf Member consists of argillite and quartzite in subequal amounts. The argillites are mainly red and green with green being the dominant color. Various shades of green and yellow-green are also present. The argillites also display abundant mud cracks. The quartzites are pink, green, white, and gray. They display ripple marks, cross bedding, and flat pebble conglomerates with red and green argillite clasts. Bedding is thin to medium. On the Going to the Sun highway just east of Sunrift Gorge, Rezak (1957, p. 136-137) describes a 10-inch bed of Collenia undosa near the base of the Grinnell in a reddish-brown sandstone interbedded with green argillite. So far this has been the only algal deposit found in the Ravalli Group. No such zone was found in the map area. In the Many Glacier area, the Rising Wolf appears to form a transitional zone between the greenish, sandy Appekunny below and the argillaceous Red Gap Member above. The greenish color within the transition zone appears to be that of primary reducing conditions either during or shortly following deposition. Some of the irregular greenish mottling within dominantly red zones may be the result of some sort of reduction process during tectonism. A section 163 feet thick was measured on the north slope of Mt. Henkel. The Fentons (1937) assign from 200 to 700 feet of beds to the Rising Wolf.

Their contact between the Appekunny and the Grinnell is probably somewhat lower than that used in this study.

Several of the quartzitic beds were studied in thin section, and the quartz fraction was found to comprise from 78 to 83 per cent of the rock. The quartz grains form a tightly interlocking mosaic with pressure strain shadows common. The size range is from medium silt up to 0.7 mm in diameter. The average sphericity is about 0.7 and the roundness about 0.2. Quartz overgrowths are common, and vacuoles are fairly abundant.

The feldspar is mostly microcline and constitutes around 2 per cent of the rock in sizes ranging from 0.1 to 0.5 mm; sphericity and roundness values are 0.5 to 0.4 respectively. Almost all grains are relatively unaltered.

Intraclasts comprise approximately 8 to 9 per cent of the sections studied. They consist of partially recrystallized, sericitic clay and range from 0.2 to 4.0 mm in length.

The matrix consists of the same type of recrystallized clay as that of the intraclasts. It comprises up to 8 per cent of the rock.

Micro-shears are filled with quartz cement which makes up as much as 2 per cent of the rock.

The quartzites of the Rising Wolf suggest a high energy environment. The overall sorting is fair, but the energy of the depositional environment was insufficient to thoroughly wash the same of all clay.

The original grains were well rounded, but with compaction and tectonic deformation, the quartz overgrowths have developed and greatly reduced the roundness.

Red Gap Member - A section of the Red Gap Member 726 feet thick was measured on the south slope of Mt. Henkel and consists of red, fissile, thin bedded argillites and minor white, thin to medium bedded quartzites containing ripple marks, cross bedding, and flat pebble conglomerates with red argillite intraclasts. The red argillite contains abundant flat mud crack polygons. There are zones of minor green argillite which appear to represent a reduction of the hematite at some time after lithification. This could have happened either during diagenesis or as a result of tectonic deformation. The green generally is present as irregular mottling and appears to spread outward from sandy layers and certain bedding planes. The quartzite in this member is much less abundant as compared with the Rising Wolf and the Rising Bull members. It would seem that a mud flat environment with subaerial exposure prevailed throughout most of the time during which the Red Gap was deposited with changes of only minor duration facilitating deposition of sandstones. Both the lower and upper contacts are transitional. The upper contact is placed where quartzite becomes a prominent rock type.

Rising Bull Member - The Rising Bull Member (Fenton and Fenton, 1937) was also measured on the south slope of Mt. Henkel. It includes 907

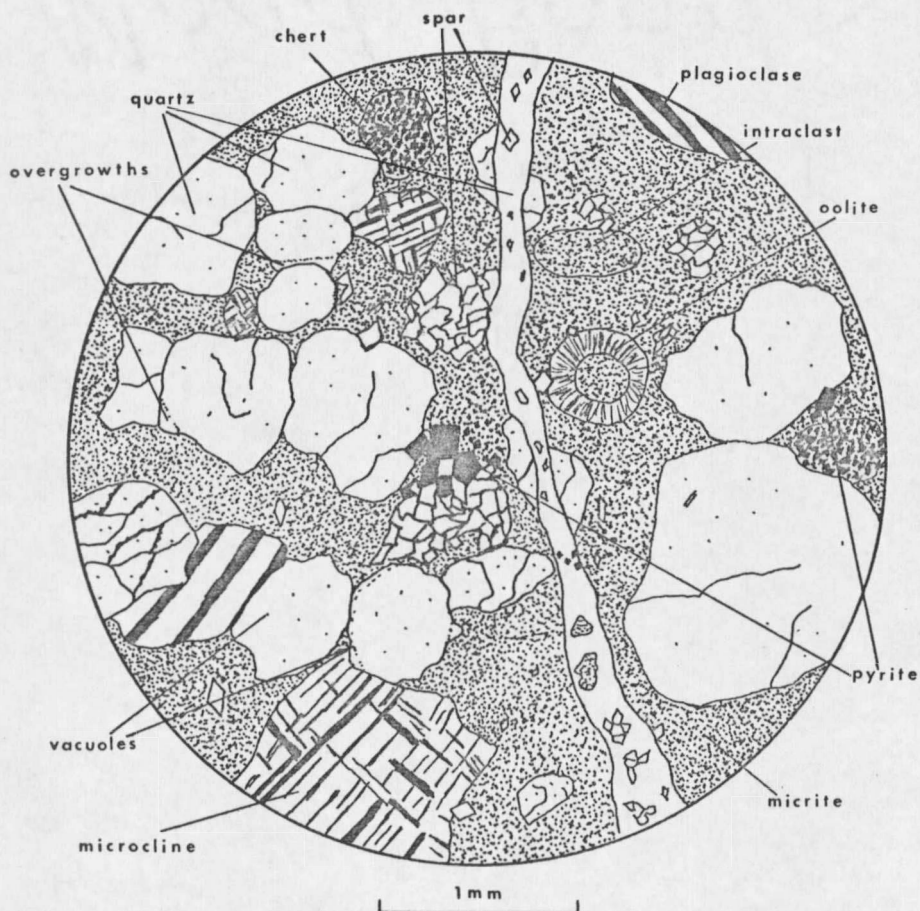


Figure 3. Generalized diagram of microscopic grain relationships typical of the Pre-Ravalli sequence and Ravalli Group. Also shows paragenetic relationship of diagenetic or secondary minerals.

feet of red and green argillite and quartzite. The red color is dominant. Argillites are thin to medium bedded and contain mud cracks, ripple marks, worm burrows, and cross beds. The quartzite is white, pink, gray, and green. The beds are thin to thick and contain flat pebble conglomerates, cross bedding, ripple marks, and mud cracks. The upper contact is gradational with the Siyeh Formation. The 50 feet of beds immediately above the Rising Bull consist of greenish, calcareous argillites and may be the eastern wedge edge of the Werner Peak Formation of the Whitefish Range.

Several of the quartzite beds were examined microscopically. The quartz fraction was observed to comprise from 62 to 96 per cent of the sections and ranges from less than 0.1 mm to as much as 2.0 mm in diameter. The grain sphericity ranges from 0.3 to 0.8 with an average of about 0.6. The roundness values range from 0.1 to 0.7 with an average of about 0.5. The smaller silt sized grains, as would be expected, have the lower values. Extinction is generally undulose. Many sections have quartz with well developed strain shadows and a tightly interlocking mosaic in those with upward of 80 per cent quartz grains. The Rising Bull Member contains the only Belt strata examined that tend toward pure quartzites. The quartz grains in general contain abundant vacuoles, some in strings within the grains. Composite grains and overgrowths are common. Inclusions are fairly common. A few inclusions were tentatively identified as tourmaline and apatite. Most of

the grains, as in the units previously discussed, reflect the metamorphic source rock of the Precambrian craton.

The feldspars appear to make up only about 6 per cent of the total rock. Microcline tends to dominate. The size ranges from less than 0.1 to 1.2 mm in diameter. Sphericity values average about 0.6 and roundness about 0.5. Most grains appear only slightly altered, but a few are extensively sericitized.

Very few specimens contain chert in amounts of up to 1 per cent of the rock. Only one section examined contains intraclasts. It has about 11 per cent intraclasts of 0.2 to 5.0 mm in diameter which consist of argillaceous micrite with some hematite.

Matrix material is generally negligible, but does in a few cases comprise as much as 15 percent of the rock. It consists of clay and micrite which is in the process of recrystallizing to spar.

The cement is mostly spar in quantities ranging from 3 to 30 per cent. Its crystal size ranges from 0.1 to 0.4 mm in diameter. It is not associated with a micrite matrix, so it could have been introduced as a cement filling pore spaces instead of being generated from a micrite. The sparry cement penetrates the boundaries of all constituents except the quartz cement which occurs as overgrowths and some veinlets. The quartz cement penetrates spar boundaries.

A few disseminated sulfides were noticed in one specimen in quantities less than 1 per cent of the whole.



Figure 4. Typical flat-pebble conglomerate of the Rising Bull Member of the Grinnell Formation showing intraclasts of red argillite in a quartzite matrix.

The degree of size sorting, lack of matrix in the quartzite beds, and the relative lack of feldspar suggests that a high energy environment (possibly beach) prevailed throughout much of Rising Bull time and that a fairly prolonged period of reworking of sands took place. The grains as originally deposited were much rounder than after overgrowth development, and in many cases, the original deposit could have been called a mature orthoquartzite. Most of the quartzitic portions of the member were surely the products of a very high energy regime as attested to by the size of some of the argillite intraclasts, many being pebble size. The transition from Grinnell deposition to that of the Siyeh carbonate indicates a gradual average deepening of the water even though extremely shallow water structures are common in the lower Siyeh.

Piegan Group

The Piegan Group of Glacier Park contains only the Siyeh Formation, and strictly speaking, cannot be called a group. Ross (1963, p. 29), however, believed that further mapping would enable workers to divide it into new formations, and therefore he retained the group name.

Siyeh Formation

The Siyeh Formation of the Glacier Park region has been correlated with the Wallace Formation of northern Idaho and western Montana and the Helena Formation of the Big Belt Mountains (Smith and Barnes, 1966,

p. 1410).

The Siyeh has been divided into members or zones by different authors using different lithologic and paleontologic criteria. The Fentons (1937) differentiated four members. Childers (1963) and Rezak (1957) both described three zones. Their zones differ only in that the middle unit of Childers contains part of Rezak's lower unit. For the purposes of this paper it seems that Rezak's subdivision is the more useful. His lower unit is some 2500 to 3000 feet thick. In the map area, the rocks corresponding to Rezak's lower member consist of argillaceous dolomite and limestone, oolitic limestone and dolomite, intraformational limestone conglomerate, sandy dolomite and limestone, and dolomitic and calcareous quartzite. The bedding ranges from thin to thick and the colors vary from blue-gray to buff. Scattered sporadically through the sequence are colonies of stromatolites, most of which are probably Collenia symmetrica. Many beds of limestone and dolomite containing "molar tooth" structure are common. One specimen of intraclastic, oolitic limestone was examined microscopically. It contains about 60 per cent clasts and 40 per cent matrix. The clasts are micrite intraclasts and oolites. Many of the intraclasts have rims of chert. The oolites have nuclei of micrite, chert, and composite chert and micrite. Most of the micrite nuclei appear to have been intraclasts originally and later to have become the oolite nuclei. Most of the matrix is now recrystallized to spar from micrite. A few quartz

grains of fine sand size are present.

The middle zone of the Siyeh is a massive blue-gray micrite unit roughly 100 feet thick which is known as the Conophyton zone. The zone is very distinctive as it is always a cliff forming unit. The zone is comprised mainly of colonies of Conophyton inclinatum with minor colonies of Collenia frequens, Collenia multiflabella, and Cryptozoan occidentale (Rezak, 1957, p. 138).

The upper zone is very similar to the lower, consisting of the same sort of impure dolomite and limestone, oolitic and intraclastic limestone, and dolomitic and calcareous argillite. Several zones of Collenia multiflabella are present within this unit (Rezak, 1957, p. 138). The upper contact is gradational with the overlying Snowslip Formation. The Piegan Group boundary is placed at the base of the Snowslip because the change in rock type indicates a significant change of depositional environment. To the west where the argillaceous Snowslip grades into carbonate deposits, the upper boundary of the Piegan is placed at the top of the Shepard Formation (Smith and Barnes, 1966, p. 1421), a significantly higher stratigraphic position.

Snowslip Formation

Rocks of the Snowslip Formation (Childers, 1963) are exposed in the map area only at Swiftcurrent Pass. The beds consist of approximately 150 feet of red and green argillite and quartzite. Mud cracks, ripple marks, and some mud breccias occur commonly. At Marias Pass,

Childers measured some 1400 feet of beds assigned to the Snowslip Formation. The formation thins rapidly northward and is no longer a mapable unit at the international boundary. The Fentons (1937) mistakenly correlated this unit with the Spokane Shale of the Big Belt Mountains. The upper contact is gradational into the overlying calcareous argillites and impure limestones of the Shepard Formation.

Shepard Formation

Only a small part of the base of the Shepard Formation is exposed in the map area, at Swiftcurrent Pass and on the summit of Mt. Wilbur. Exposures consist of greenish and grayish calcareous shale and argillaceous limestone and dolomite. Some clayey, calcareous sandstone is also present. The base is transitional upward into the upper Shepard which is lithologically similar to the Siyeh.

Precambrian Intrusive Rocks

Within the Many Glacier area, two forms of intrusive rocks occur: dikes and sills, both of gabbroic composition (Ross, 1959, p. 56). On the face of Mt. Gould, just south of the map area, a dark sill about 50 feet thick is made conspicuous by white alteration bands in the Siyeh Formation above and below. The sill in most places appears to have been metamorphosed slightly or to have undergone an unorthodox style of late stage crystallization. In many places, the rock is slightly pink with K-feldspar. The rock is generally medium to fine crystalline, but

varies in crystal size and texture with the locality. Also conspicuous in much of the rock is a greenish alteration product which is tentatively identified as epidote. The sill is exposed on Mt. Grinnell just below the Conophyton zone of the Siyeh. At this locality, the sill splits into several sills which are laterally discontinuous. At Swift-current Pass, the sill ascends from below the Conophyton zone to a horizon just below the base of the Snowslip and continues northward at this horizon.

The dikes observed within the map area are of diabasic texture and are much altered especially at their contacts with host rock. A dike cuts through the north shoulder of Mt. Wilbur and the Ptarmigan Wall just west of Iceberg Lake. Grinnell Point is also cut in two places by the diabase as is discussed later in the report, and Mt. Allen is crossed by a dike near its northernmost extension (Plate 3). The dikes vary in width from 10 to 30 feet or so. Copper mineralization is associated with the dikes, and abandoned mines are present on the north shoulder of Mt. Wilbur and on the southeast side of Grinnell Point at the exposure of the southernmost dike. Large crystals of calcite are associated with the contacts. The only copper mineral personally observed was bornite in the dike rock.

Cretaceous

Exposures of Cretaceous rocks within the map area are poor due to glacial and vegetative cover. Exposures outside the map area along

Kennedy Creek and Roes Creek have been visited. At both sites, black fissile shales were found. At Roes Creek, volcanic, bioclastic sandstones were also present containing abundant gastropods and pelecypods. The beds at Roes Creek are immediately below the Lewis thrust plate and were determined by R.C. Wilson (personal communication, 1970) to be Colorado Group. Childers (1963, p. 154) reports strata of the Blackleaf Formation of Early Cretaceous age below the thrust at Marias Pass. At both localities personally visited by the writer, the beds were highly contorted and bedding largely indistinguishable in the black shale.

Pleistocene-Quaternary

Mantle of glacial debris, landslides, stream deposits, and talus forms the Recent and Pleistocene alluvium or colluvium. Glacial debris fills the lower reaches of the valleys of Ptarmigan, Swift-current, Grinnell, and Canyon Creek. Landslide material is present along Canyon Creek near the mouth of its canyon. Talus is present both in parts of the lower valleys and higher up on the mountain flanks. Recent stream and lake deposits, of course, occupy the lowest elevations of the valleys. Very little study was given to these deposits and consequently they are all mapped together as Qu, and then only where those deposits are of significant thickness and breadth.

Possible additional methods of Belt correlation

Since most of the individual beds within members and formations of the Belt are commonly discontinuous over distances of tens of miles, the contacts of these units are not time stratigraphic boundaries but depend only upon lithologic characteristics. Suitable time marker beds generally have not been found within the Belt. One such useful time stratigraphic unit that does exist is the Purcell Basalt flow which has been used to correlate other units (Smith and Barnes, 1966, p. 1403) on a regional scale. This unit is within the base of the Shepard Formation at Swiftcurrent Pass, but does not appear to be present within the Shepard a few miles east of this point near the summit of Mt. Wilbur. Another possible key to regional time stratigraphic correlation may lie in greater subdivision and lateral tracing of stromatolite zones within the Siyeh Formation.

A further tool which could possibly provide precise location of time boundaries is paleomagnetism. Norris and Black (1961, p. 933-935), using samples from the Kintla Formation of the Flathead and Clarke Ranges in Canada, have conducted a paleomagnetic study with reddish, hematitic quartzites. They recorded a few magnetic pole reversals within certain units of the Kintla. If magnetic reversals are common within the Belt, they could be used in differentiating time stratigraphic boundaries. Assuming such reversals might exist in Belt rocks, paleomagnetic studies could conceivably be conducted within the Glacier

Park area since most of the stratigraphic units there contain at least some hematitic quartzite or sandstone. Reddish argillites should also be suitable since they contain fine grained hematite.

STRUCTURE

General Statement

The Lewis and Livingstone Ranges of Glacier National Park lie within the structural geologic province known as the Front Ranges (Bally, Gordy, and Stewart, 1966, p. 343). This subdivision of the region is characterized by thrust sheets, commonly in imbricate arrangement, dipping westward, generally involving Paleozoic carbonates and/or Precambrian Belt carbonate and clastic rocks. The area to the east of the Front Ranges, a small portion of which is within the boundaries of the park, is known as the Foothills, or structurally, the disturbed belt. The Foothills are characterized by Paleozoic and highly deformed Mesozoic strata involved in imbricate thrust faults dipping toward the southwest. The area as a whole is underlain by the westward continuation of the gently southwesterly dipping top of the basement of the Canadian Shield, which is relatively undisturbed (Bally, Gordy, and Stewart, 1966, p. 343).

The area mapped for this report (Plate 1) is just southeast of the middle portion of the northeast facing Clarke Range salient



Plate 2. Tectonic map of the Many Glacier area.

(Clarke, 1954). The margin of this salient in Glacier Park extends along the northeast margin of the Lewis Range. Northward in Canada the margin parallels the northeastern limit of the Clarke Range. The map area approximately borders the trace of the Lewis thrust and extends some 10 miles to the west. The two structural features of greatest significance in the Glacier Park area are the Lewis overthrust and the North Fork normal fault (Willis, 1902). Childers (1963) suggested that Willis' name (North Fork fault) be raised to group or zone status since the North Fork fault was in actuality "a complex of bifurcating and en echelon faults, commonly with associated antithetic faults" (Childers, 1963, p. 159). Presumably the Roosevelt fault (Clapp and Deiss, 1931) is one of the en echelon faults within the North Fork fault zone. Some confusion exists between geologists as to which structure the term "Flathead fault" refers. The Canadians refer to Willis' North Fork fault as the Flathead fault, but Clapp (1932) named the normal fault on the west side of the Flathead Range the Flathead fault. The trend of the fault zone as determined by Ross (1959, p. 87) ranges between N 20 W and N 45 W. The strike of the Blacktail fault (one of the component faults of the zone in the Marias Pass quadrangle) is about N 50 W (Childers, 1963, p. 159). The faults comprising this complex are regarded to have individual stratigraphic separations varying from 2,000 to 14,000 feet. An interpretation by Bally, Gordy, and Stewart (1966, p. 354-355) based on geophysical data postulates that the Flathead fault is a listric (fault merging with

bedding plane at depth) normal fault that merges with the Lewis thrust at depth. This implies that the thrust plate has been "stretched" after emplacement. This normal fault zone crops out some 40 miles west of the map area, but appears to bear a genetic relationship to many of the structures that are present within the area.

To the east of the thrust there are presumably imbricate thrusts and highly contorted structures within the Cretaceous rocks as in areas examined by previous authors; however, due to extremely limited outcrops, virtually all of the structures examined in this study were in the Belt rocks of the thrust plate (Plate 2).

Many Glacier area

Immediately west of the thrust trace are subsidiary thrusts with associated kink and drag folds. These faults undoubtedly have displacements of at least hundreds of feet since they bound intricately sheared and deformed wedges of the Hell Roaring Member of the Altyn Formation. Actual amount of movement along these thrust planes would be extremely difficult to establish due to complexity of deformation.

Further west throughout the entire map area occur zones of secondary faulting and associated folding. Many of the fold zones are associated with small scale thrusts. Both low and high angle thrusts occur.

In the extreme western portion of the map area there exists a very continuous low angle thrust with associated schuppen structure. This fault can be traced through the map area and at least 20 miles further

to the north. Several large scale anticlines are also present within the thrust plate. Their trend varies roughly from N 30 W to N 70 W and the steep limbs are to the northeast. Amplitudes vary from hundreds to thousands of feet.

Throughout the map area there exists a conjugate system of fractures whose trend and attitude varies with rock type. No important tear faults were observed in the overthrust plate in the area examined.

Lewis Thrust Fault

The Lewis overthrust was first described by Bailey Willis (1902, p. 331). It extends from as far as 90 miles north of the Clarke Range to 40 miles southeast of Marias Pass (Clapp, 1932). The surface trace of the thrust can be followed for more than 12 miles from northeast to southwest through Marias, a direction transverse to general trend of the trace. Displacement is estimated at more than 40 miles to the northeast (Erdmann, 1947; Daly, 1912). The stratigraphic separation at Marias Pass is in the neighborhood of 20,000 feet. At this point, beds of the Precambrian Appekunny Formation overlie rocks of the Cretaceous Kootenai and lower Blackleaf Formation (Childers, 1963, pl. 1, p. 157). To the north in Glacier National Park, the separation is even greater, since beds of the Waterton Formation overlie Upper Cretaceous rocks. In the map area, the lowest strata of the overthrust plate are the Hell Roaring Member of the Altyn Formation. The age of the Lewis overthrust is probably younger than Paleocene (Russell and Landes, 1940, p. 93).

In the map area, the thrust contact can be located between exposures approximately 200 feet apart. At Rising Sun, the fault contact can be limited by exposures 50 feet apart. The average strike of the thrust is approximately N 30 W (Ross, 1959, p. 76). As roughly determined from graphical calculation, the strike of the thrust in the map area is in accordance with Ross's figure. The dip of the thrust varies along its trace in the park; from nearly horizontal to as much as 20 degrees, but within the map area, the dip is estimated at less than 10 degrees to the southwest. The thrust contact is at the base of Mt. Appekunny near the northern border of the map area (Plate 1). It extends southwestward along the base of Mt. Altyn to a point just east of Many Glacier Hotel, and then southeastward along the base of Mt. Allen and Mt. Wynn. Just above the thrust on Mt. Wynn, the Altyn Formation is faulted into a number of wedges or slices about 50 to 100 feet thick. The same relationship of the sliced zone to the overlying rocks is present north of the map area on the lower half of Chief Mountain and through the bulk of Yellow Mountain. Elsewhere in the map area, there are no fault wedges or slices in the rocks immediately overlying the thrust. The angle of the intersection between the thrust and overlying beds averages less than 5 degrees. North of the map area along the international boundary, relatively high angle shearing is prevalent with intersection angles of 18 to 19 degrees (Childers, 1963, p. 157). As far as the configuration of the thrust is concerned, it is a relatively planar feature in the Many Glacier area. There appear

to be no folds in the thrust surface, but at the southern end of the Lewis Range broad, open, southwest plunging folds in the Lewis thrust surface are partly responsible for the reentrant in the thrust trace at Marias Pass (Childers, 1963, p. 158).

Structural features above the Lewis overthrust - Ross (1959, p. 81-83) refers to the thrust plate as "a mass of competent strata that would yield reluctantly to deformative stresses." He points out that the first impression one gets of gently southwesterly dipping beds is not quite the case and proceeds to describe a prominent syncline stretching from Boulder Pass south to near Tinkham Mountain, a distance of almost 40 miles. This structural feature is approximately 10 miles wide where it crosses West Flattop Mountain with a trend of N 48 W. It is a rather shallow structure with beds on either flank having dips of up to 30 degrees (Ross, 1959, pl. 1). It is bordered on both sides by poorly defined anticlinoria, with the beds to the southwest being the most highly deformed with intricate crumpling and thrusting common. In reference to fold zones to the northeast of this syncline, Ross states (1959, p. 81-83): "East of the major syncline the folds in the compound anticline are open. No extensive zones of crumpled or overturned beds are recorded outside of the locally thick zones of fracture constituting the Lewis overthrust itself. Local crinkles and minor thrusts are present..."

The secondary thrusts described as "minor" are admittedly so in displacement, however, there are many that are continuous over fairly long distances, the longest seen being some 20 miles. The zones of secondary folding to the northeast of the major syncline are in many cases traceable for several thousand feet along mountainsides. The individual folds within these zones may range in amplitude from a few inches to more than a hundred feet. These zones are indeed not so pronounced as those comprising the anticlinorium to the west of the major syncline, but they can hardly be passed off as unimportant "local crinkles". They are, however, fairly subtle in topographic expression in many cases, and have not been previously recorded.

The preceding general description of minor thrusts and zones of folding is characteristic not only of the map area, but of all areas visited north and south of the area in a direction parallel to the major syncline.

Secondary folds

Areal distribution - The folds in the block overlying the Lewis thrust plane in the map area are, as previously stated, generally not singular folds of great amplitude, but occur as scattered zones of smaller folds. These zones are commonly associated with southwest dipping thrusts of small displacement. In fact many of the secondary faults are sheared, overturned anticlines. The folds within these zones vary in amplitude

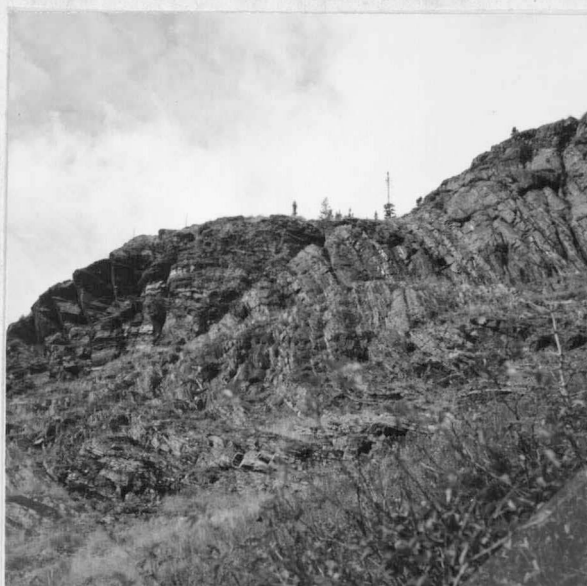


Figure 5. Secondary drag folds from the fold zone associated with the upper thrust fault on the southeast slope of Grimmell Point.

from a very few inches to as much as 100 feet. They are generally asymmettrically overturned toward the northeast and have southwest dipping axial planes.

The most prominent and continuous zones found in the map area are on Grinnell Point. There are two zones of secondary folding which trace diagonal lines up the point towards the northeast. The more southwesterly of these zones is well exposed just above the trail to Grinnell Glacier about a tenth of a mile before the first switchback. This fold zone is associated with a secondary thrust of small displacement (Figure 6). The zone follows a diagonal trace to the top of the ridge and shows a dip of approximately 40 degrees to the southwest. The thrust associated with this zone is mapped since the displacement is greater on the north side of the ridge and because the fault is so continuous.

The second prominent zone on Grinnell Point is roughly parallel to that previously described, but lies approximately 1/4 mile northeast. Other smaller scattered zones occur on Grinnell Point such as those associated with several minor thrusts of very small displacement near the base of the easternmost diabase dike that crosses Grinnell Point. These scattered zones are not mapped because of their limited linear extent.

Another zone of kink and drag folds is well displayed low on the southeast shoulder of Mt. Wilbur just north of the easternmost of the Bullhead Lakes. This zone is similar to others, but possesses no appar-

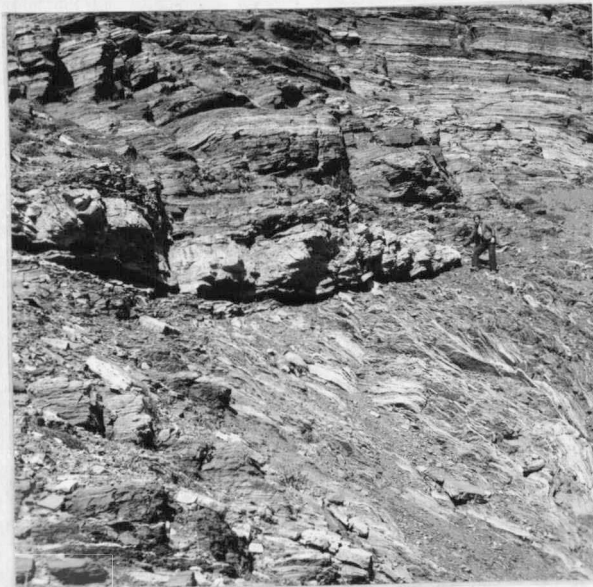


Figure 6. Upper thrust fault on southeast slope of Grinnell Point. Person standing on thrust plane.

ent relation to a thrust fault. It is possible that this zone may be a continuation of the zones on Grinnell Point, but evidence is obscured by foliage and glacial outwash of the Swiftcurrent Valley. This zone is part of the southwest limb of the Mt. Wilbur anticline which will be discussed later.

Another fold zone in the lower Siyeh Formation is exposed in the headwall of the cirque south of the southernmost switchback on the Swiftcurrent Pass trail. This zone is readily visible from the trail and consists of both large and small kink folds, the largest being about 30 feet in amplitude.

Another fold zone in the lower Siyeh Formation is exposed above the trail to Iceberg Lake about 1/2 mile above the junction of the Ptarmigan Tunnel trail. All the remaining mappable fold zones are associated with secondary thrust faults. The only fault in the map area which apparently lacks associated folds is the low angle thrust exposed on the Ptarmigan Wall and about halfway up the cliffs of Mt. Wilbur in the lower Siyeh. This fault will be discussed in detail later.

Directional relationships - As previously stated, most of the individual folds within the zones described have southwest dipping axial planes. However, an appreciable number have north or northeast dipping axial planes. Some 250 individual folds were studied for orientation of



Figure 7. Small kink folds in fold zone.. From zone associated with upper thrust fault on Grinnell Point.

axial planes and axes. The orientation of axial planes were then plotted on stereonet overlays (Figures 8, 9 and 10) to determine the existence of possible patterns and genetic relations of the two types of individual folds. Within any specific zone of the folding, many of the folds with southwest dipping axial planes tend toward the same orientation, so commonly as many as five or six orientation readings were averaged. As a result, only about 50 readings were plotted in Figure 8. Inspection of Figure 8 reveals a dominant pattern in the spacing of poles to axial planes wherein southwest dipping axial planes show a definite grouping. The poles from the folds with north to northwest dipping axial planes show no such grouping, but are more or less randomly scattered. The average strike for those folds with southwest dipping axial planes is N 26 W and for those with north or northeast dip is N 3 W. The apparent random scattering of poles to axial planes of north or northeast dip may or may not be real. It is certain that they do not group as well as those with southwesterly dip, but they may have some as yet undefined pattern. The lack of an established pattern may be due in part to the fact that relatively few folds with axial planes of northeast dip exist compared with the number of folds with axial planes of southwest dip, i.e. a larger number of measurements of these might yield some subgroups.

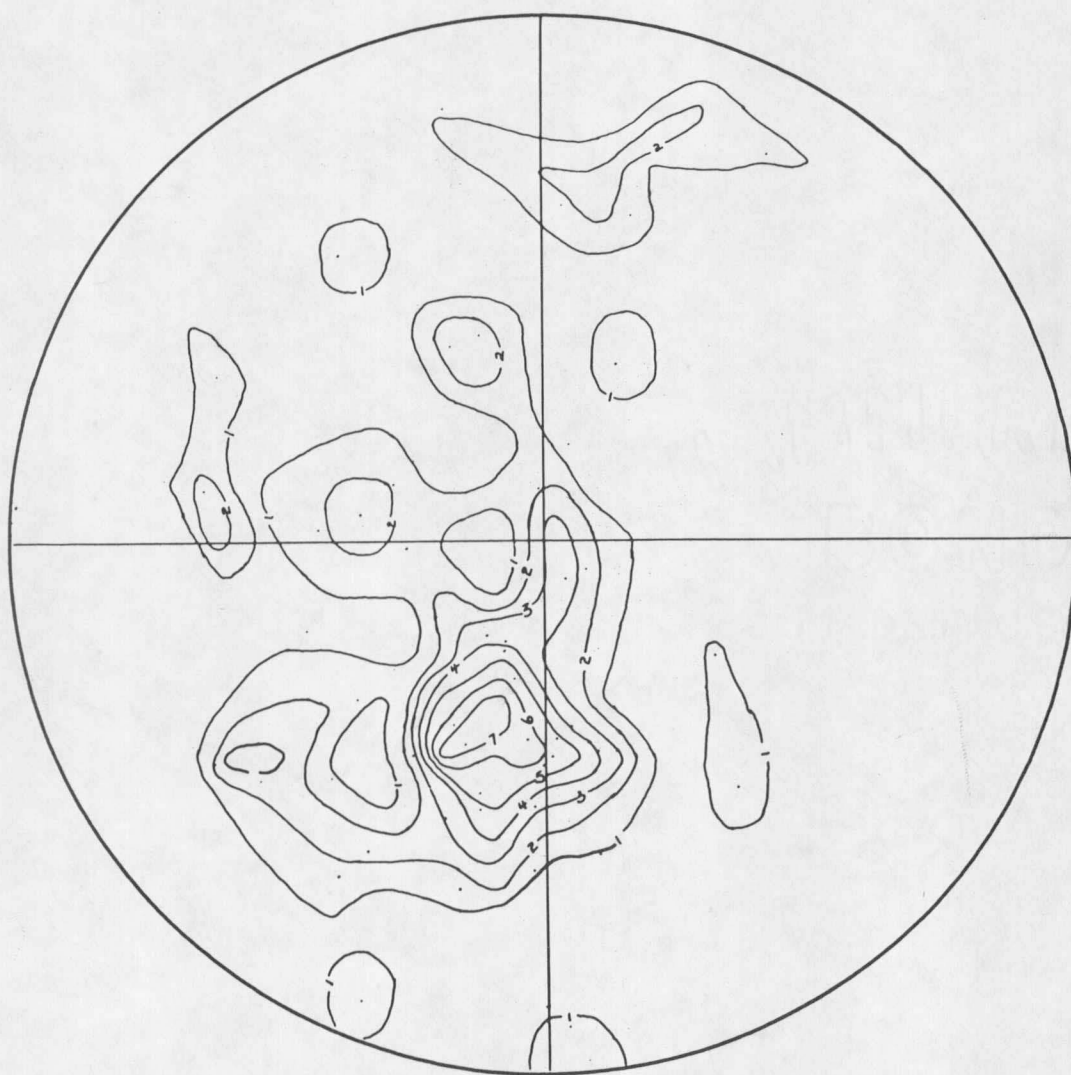


Figure 8. Stereonet plot of poles to axial planes of folds showing noticeable grouping of "normal" folds in the southwest quadrant.

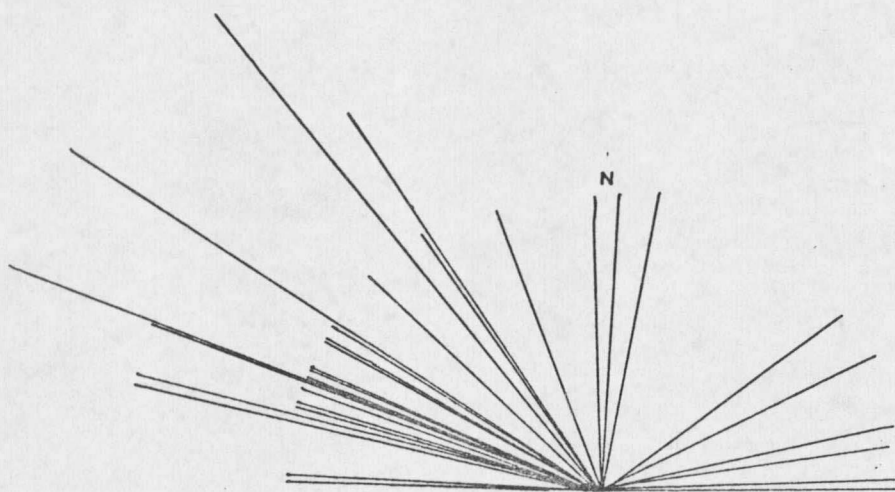


Figure 9. Stereographic projection of strikes of axial planes ("normal folds").

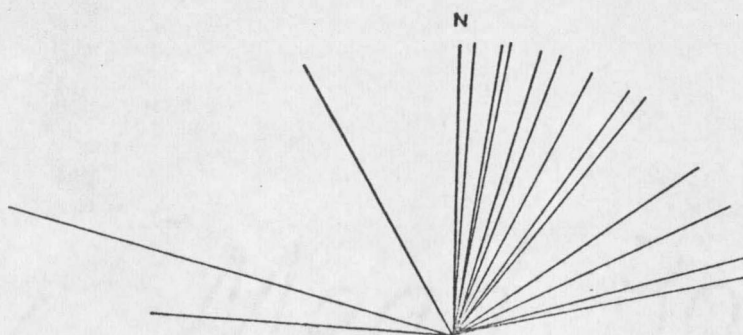


Figure 10. Stereographic projection of strikes of axial planes ("reverse folds").

Secondary Faults

In considering the complex fault wedges in the lower Altyn Formation above the Lewis overthrust, difficulty arises in deciding whether these faults belong to the Lewis fault zone or are to be considered as individual secondary thrusts. The question is admittedly a matter of choice, and the fact remains that these faults in the lower strata immediately overlying the thrust are probably directly related to the main thrust movement. Within the map area, the faulting just above the thrust is unsystematic and includes both steep and shallow dipping fault planes. Many of the fault surfaces are irregular and seem to have been folded. Within the map area of Mt. Wynn, there are no fault wedges within the Altyn Formation. This is because the thrust trace on Mt. Altyn, Mt. Allen, and most of Mt. Appekunny lies west of the zone in which the hanging wall or thrust plate strata are intensely deformed. The zone of intense deformation of strata immediately above the Lewis fault trace generally dies out within a few miles laterally (to the west) and within a few hundred feet vertically. This is, of course, not the case in Waterton Lakes Park where another major thrust overlies the Lewis thrust.

Grinnell Point - On Grinnell Point, there are two prominent faults. Both are reverse faults; the upper fault being associated with the upper zone of secondary folding previously described (Figure 5). The apparent dip is around 20 to 30 degrees to the southwest where well exposed,

but this tends to vary. As depicted by Figure 6, the displacement is negligible at this point; however, the thrust fault associated with the upper zone of folding on the northwest side of Grinnell Point has a displacement of about 20 feet and a dip of 40 degrees to the southwest. Within this zone of folding, there exists numerous smaller and less continuous faults and it is less than probable that the most prominent faults on opposite sides of Grinnell Point from this zone are, in actuality, the same fault surface. However, they do belong to the same zone and could be treated as the same fault since the faults and folds are inter-related.

The second prominent fault trace on Grinnell Point is observable from the Josephine Lake side. It is a low angle fault surface that dips gently to the southwest. It forms the main break in the cliffs of Grinnell Point above Josephine Lake and can be traced from just north of the switchback on the Grinnell Glacier trail to the end of the ridge forming Grinnell Point. Exposures of the fault plane are not available for study, however, since talus and colluvium cover its trace, so no direction evidence was found to indicate direction of movement or displacement. The fault along most of its trace is within the Red Gap Member of the Grinnell Formation and is marked by the presence of springs at various intervals. The old Josephine copper mine is located at the intersection of the fault with a diabase dike which cuts through Grinnell Point about 1/8 mile from the extreme northeastern extension of the point.

Beneath the fault at this point no trace of the diabase dike can be found. The only diabase found on Grinnell Point below this horizon is that 1/8 mile northeast at the end of the point. However, the relationship of the fault to the overlying rocks remains unclear here because the strata above the fault surface have been removed by erosion. On the north side of the point the cliffs are relatively steep even across the horizon where the fault should be exposed, but no clear evidence of any displacement or fault surface exists, the bedding being apparently conformable at this horizon. Unless the overburden could be removed along the fault trace below the Josephine mine site, the relationship between overlying and underlying strata will remain obscure. It is necessary, however, to tentatively conclude that, since the fault apparently does not pass through the north side of Grinnell Point, that displacement along it was minor whatever the direction of movement. The apparent nonexistence of the diabase dike below the fault near the mine might be explained by a pinch out or camouflaging by colluvium and vegetative cover. Since the thickness of the diabase dike is quite variable, and the contact metamorphism is such that the contact between the dark greenish-gray of the Scenic Point Member at this point is close to the color of the diabase, the presence of the dike below this contact could have gone unnoticed. The diabase dike at the northeast end of Grinnell Point and its correlative on Mt. Allen to the southeast can be easily traced through all of the exposed Appekunny Formation, so it

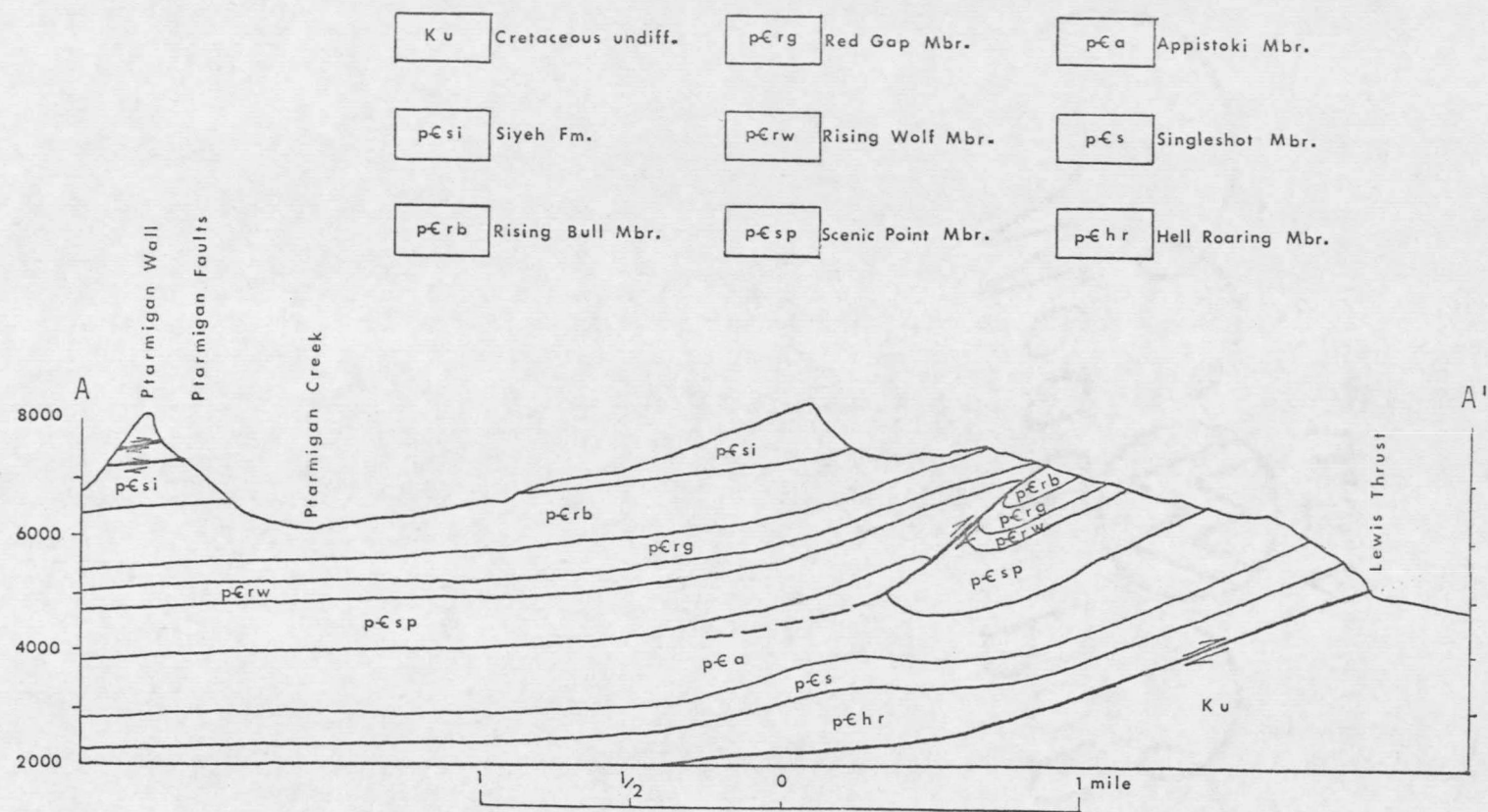


FIGURE 11. Section A - A'
(Location of section on Plate 3)

is likely that a pinch out of the dike near the copper mine is responsible for its apparent nonexistence below and to the southeast of that site. No diabase dike could be found on Mt. Allen that would correspond to the trend and position of the dike near the copper mine.

Ptarmigan Fault - A fault with significant lateral continuity is well exposed on the sides of the Ptarmigan Wall just north of Iceberg Lake. It lies about halfway up on the slopes of the wall in the lower Siyeh Formation. At a point just north of Iceberg Lake, the fault bifurcates, and the lower part maintains the same low angle dip to the southwest, while the upper bifurcation increases in dip by about 15 degrees and continues to the top of the ridge. In the cliffs west of Ptarmigan Lake, the lower fault surface strikes N 58 W and dips 9 degrees to the southwest. At this point the feature is a plane instead of a zone. In the cliffs just north of Iceberg Lake, the upper fault includes a zone of schuppen structure about 8 feet thick. This zone is an imbricate arrangement of overlapping southwest dipping slabs of limestone and calcareous shale of about 6 inches to a foot in thickness. This fault has been traced as far south as the south slope of Mt. Wilbur and as far north as Mt. Cleveland, a distance of about 20 miles. As far as the fault has been traced, it lies in the lower Siyeh Formation. The movement on the fault as determined by associated minor folds and fault zone structures is that of a thrust. The movement along the fault is probably



12. Trace of bifurcating Ptarmigan fault on the southeast slope of the Ptarmigan Wall.

not any greater than the order of tens of feet at the most. Just west of Iceberg Lake, the fault apparently cuts a diabase dike, but no displacement was observed. The dike, however, is reduced in thickness below the fault plane. Since the dike is of Precambrian age, movement along the fault parallel to the strike of the diabase may have facilitated development of the present relationship. No other exposures have been found where the thrust bifurcates as it does on the Ptarmigan Wall. Outside the map area where the fault has been traced, it seems to be present as a distinct plane rather than a zone, especially as viewed on the cliffs of Mt. Merrit and Ahern Peak some five miles to the northwest of the map area. The trace of the thrust appears not to vary greatly in either topographic or stratigraphic elevation throughout the area where it has been traced.

Ptarmigan Tunnel - Two small reverse faults occur just west of Ptarmigan Tunnel at about the same elevation. The faults are separated vertically from each other by about 30 feet of rock and occur in the transition zone between Grinnell argillite and Siyeh limestone and dolomite. The lower fault has about 60 to 70 feet of displacement, and the upper roughly 40 feet. The two faults are roughly parallel in orientation, the lower having a strike of N 68 W and a dip of 39 degrees to the southwest. The upper fault has a strike of N 76 W and a dip of 40 degrees to the southwest. Immediately east of the tunnel, a reverse fault displaces the

Rising Bull Member of the Grinnell Formation a stratigraphic distance of roughly 200 feet. This fault is an extension of the lower fault west of the tunnel. The upper fault has very little displacement at its exposure east of the tunnel. Numerous kink and drag folds are associated with these faults.

Mt. Altyn - On the ridge about midway between Mt. Altyn and Mt. Henkel, the Scenic Point Member of the Appekunny Formation is displaced upward some 400 feet against the Red Gap Member of the Grinnell Formation. The fault is high angle reverse with a dip of about 45 degrees to the southwest. The fault is covered in the cirque north of this exposure, but is present again on the ridge between Mt. Henkel and Mt. Appekunny. At this exposure, the fault has a displacement of only about 20 feet, indicating that it is dying out rapidly in this direction. The fault is not found anywhere to the south of Mt. Altyn. This fault is readily seen from Glacier Route No. 12 near Swiftcurrent Campground.

Mt. Allen - Three reverse faults of mappable dimensions occur on Mt. Allen, none having displacements greater than 10 or 20 feet, but having traces which are clearly visible from nearly any point in the Grinnell Valley. The uppermost of these faults is exposed near the summit of Mt. Allen. It dips at roughly 45 degrees to the southwest and its trace passes from the Rising Bull Member of the Grinnell upward into the lower Siyeh. It is traceable downward into the Red Gap Member of

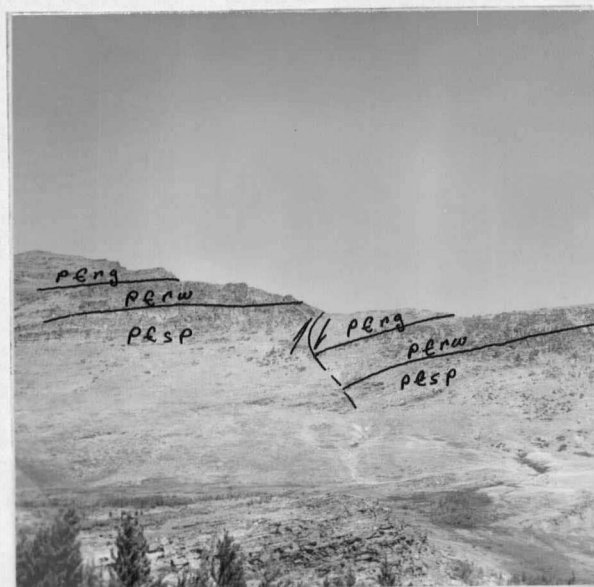


Figure 13. Thrust fault associated with ^tMy. Henkel anticline.

the Grinnell where it is then lost in the talus. Kink and drag folds are associated.

The middle reverse fault is exposed about 1/2 mile north of that previously described. It is roughly parallel to the uppermost fault, and the displacement is small.

The lowest reverse fault is found terminated by erosion at a point near the northeast end of Mt. Allen. Its displacement is small, but its plane (N 70 E, dipping 24 degrees to SE) can be traced for approximately 1/4 mile on either side of the ridge where it too is lost in the talus.

These reverse faults could possibly correlate with the faults described on Grinnell Point, since their orientation and position are roughly similar; however, they cannot be traced across the Grinnell Valley because of the alluvium and dense forest cover.

A high angle reverse fault on the cliff face of Mt. Appekunny suggests movement to the southwest. It is the result of a kink fold that sheared at its crest. The displacement is not more than 30 or 40 feet, but it is unique within the map area, since it suggests differential movement of the thrust mass in a direction roughly opposite to the faults previously described. It strikes N 75 W and dips steeply in the northeast. Back from the cliff face in the Appekunny cirque, its trace is obscured by talus.



Figure 14. Thrust faults on northeast slope of Mt. Allen.

Large Scale Folds

There are within the map area, three large scale folds. All three are anticlines with the northeast limb being the steepest. Smaller secondary folds are associated with each of these structures

Mt. Wilbur anticline - This structural feature is exposed on the southeast shoulder of Mt. Wilbur where it has an amplitude of about 1/4 mile.. Both limbs of the structure are fairly symmetrical about the axial plane with dips of up to 20 degrees on either limb. The northeast limb, however, seems to average slightly steeper dips than the southwest limb. Its axis trends N 70 W and extends from the southeast shoulder of Mt. Wilbur and plunges gently southeastward to the western tip of Redrock Lake. Its southwestern limb forms the zone of secondary folds previously described.

Mt. Henkel anticline - This fold is genetically related to the thrust previously described on the ridge between Mt. Altyn and Mt. Henkel. The anticline lies immediately west of the reverse fault. Its development undoubtedly preceded that of the fault. Its amplitude on the ridge between Henkel and Altyn, is roughly 100 feet. On the ridge between Henkel and Appekunny, the amplitude is about 20 feet. At its southern limit, the anticline has a nearly vertical northeast limb and is sheared, thereby producing the reverse fault adjacent to it. It trends N 30 W, and its trace is approximately that of the related reverse fault.

Mt. Altyn anticline - The anticlinal fold present on Mt. Altyn is best viewed from the highway approximately 3 miles east of Many Glacier Hotel. This structure is approximately 500 feet in amplitude and extends from Nahtaki Lake to the east end of Mt. Altyn. Its trend is N 70 W and the axial plane is inclined to the southwest.

Interpretation

The existence of folds as previously described, with axial planes of both southwesterly and northeasterly dip, by itself suggests that there may have been more than one stage of folding. For the sake of brevity, the folds with southwest dipping axial planes of northeast dip shall be referred to as "reverse" folds. It is remotely possible that all the folding was done during relatively the same stage, that is, during development of the Lewis thrust with differential movement or "local irregularities" within the allocthonous mass being responsible for the "reverse" folds such as suggested by Ross (1959, p. 99). If both the "reverse" and "normal" folds were developed during thrusting, then the stereonet plot of their respective poles to axial planes should show the same type of grouping pattern, or lack of it as the case may be. The stereonet plot of poles (Figure 8) to "normal" folds shows a fairly tight grouping of poles in the southwest quadrant. The plot of poles to "reverse" folds (same figure) reveals a rather random distribution as the poles are scattered throughout the northwest and northeast

quadrants. This evidence suggests that the "reverse" folds were not developed during the same structural event as that which formed the "normal" folds, namely during thrusting. In plotting the strikes of the axial planes of both "normal" and "reverse" folds on stereonet overlays (Figures 9 and 10), the difference in grouping between the "normal" and "reverse" fold axial planes is seen again. The strikes of "normal" folds in Figure 9 show a tighter grouping than the strikes of "reverse" folds in Figure 10. However, there may be some pattern of grouping of the poles of "reverse" folds, but it is vague when compared with an equal number of "normal" folds. The fact that "reverse" folds are relatively few in number as compared to the numbers of "normal" folds does admittedly inhibit statistical analysis. In any case, if the "reverse" folds are presumed to be related to "local irregularities" in the configuration of the Lewis thrust zone, then why do not the "normal" folds reflect these irregularities in their orientations, since they would have been formed in the same event? Therefore, it is not likely that the "reverse" folds are related to local flexures of the thrust plate. It is also unlikely that they formed during the same structural episode that produced the "normal" folds.

Another possibility is that the "reverse" folds are folds related to a set of stress planes conjugate to the axial planes of the "normal" folds. This possible explanation is also unlikely again because of the difference in grouping of orientations of the "normal" and "reverse"

folds. If the "reverse" folds were a set of folds conjugate to "normal" fold planes, then they should display the same degree of grouping per given number of folds as "normal" folds on the stereonet plots, since they would have been formed simultaneously. Since they do not, this explanation does not appear applicable.

Another possible explanation is that these "reverse" folds may well be related to the tensional phase of deformation, that is, the North Fork normal fault zone bordering the western limit of the Livingstone Range. This fault zone has reportedly been active since the Oligocene (Ross, 1959, p. 87) and has continued with some degree of activity until the present. The zone is generally poorly exposed and is comprised of many normal faults in a zone of variable width, in places as great as 20 miles (Childers, 1963, p. 159).

Movement in the thrust plate in a direction roughly opposite to the initial thrusting during the tensional phase of structural activity may have occurred as a result of backsliding along the pre-existing Lewis overthrust zone with a rift developing in the overthrust plate at the site of the surficial exposure of the normal fault zone. Seismic evidence from reflection and refraction data from the southern Clarke Range has provided evidence of considerable horizontal movement during normal faulting:

"It must be concluded that the Flathead normal fault flattens at shallow depth and, as shown, we believe it merges with the Lewis overthrust. This suggests the Lewis overthrust has been

"stretched" or "bottlenecked" after its emplacement, an event also related to the formation of the Tertiary basins.

The conclusion is that the Flathead fault is a listric normal fault formed after emplacement of the Lewis overthrust by "back-slippage" along a pre-existing thrust during a phase of post orogenic uplifting. The predominance of east dips in the Tertiary beds west of the fault is perhaps due to rotation along a listric normal fault (Bally, Gordy, and Stewart, 1966, p. 355)."

To produce the "reverse folds", which are compressional features, the backsliding would have to have had to be accompanied by some local resistance to movement. This hypothesis encounters problems in that seismic data are lacking for the western boundary of Glacier, and consequently, the presence of a listric normal fault can only be speculated. Another problem is that most of the "reverse" folds generally occur within the same zones as those of the "normal" folds. It is difficult to see why they developed at precisely the same spots during backsliding unless the zones of "normal" folds previously formed provided zones of weakness with most of the later deformation being concentrated at these sites.

None of the previous hypotheses provide a completely satisfactory explanation of the formation of the "reverse" folds. As mentioned previously, the relative scarcity of "reverse" folds within the map area limits development of any statistical pattern which might be present. "Reverse" folds from a larger area would have to be studied to provide a valid statistical picture of their orientation relationships.

The significant relationship established in the present report is that of the "normal" folds to the Lewis overthrust. The genesis of the "normal" type was probably during the main stage of thrusting, gravitational gliding, or a combination of both that formed the Lewis overthrust. "Normal" folds would be associated with a gravity slide block only if there were resistance to sliding. This relationship is the most probable, since the "normal" folds' steeper northeast limbs indicate movement in that direction. The grouping of the poles to axial planes of the "normal" folds also suggests that they are related to a single, geologically brief structural event, such as the Lewis overthrust is assumed to be. This supposition is supported by Norris and Black (1961, p. 933-935) who have applied paleomagnetic studies to the Lewis overthrust plate in the Flathead, Clarke, and Lewis Ranges of Canada. Using samples from the Kintla Formation (Kintla Group of Childers), they found no appreciable divergence in declination of remnant magnetic poles between the samples from the locations in the Flathead Range and those in the Clarke Range even though a deflection of the thrust trace of 35 degrees from the northwest to north occurs at Kootenay Pass. Their conclusion is that the Lewis thrust plate moved in a single event as a coherent mass, and that deflections in the thrust trace are due to the original configuration of the thrust surface.

The main trend of the Lewis overthrust is N 30 W; however, the perpendicular to this (N 60 E) is not necessarily the direction in which

the thrust moved. Within the map area, the "normal" folds have a maximal occurrence of strikes at about N 70-80 W (Figures 9 and 10).

These "normal" folds, being directly related to the Lewis overthrust, should strike or trend perpendicular to its direction of movement.

Thus within the map area, the most probable direction of movement was N 20-30 E. Analysis of fold orientation in other areas is necessary to determine the average direction of movement of the overthrust on a regional scale. Folds within the map area may represent localized deformation and this analysis may not give the direction of the Lewis overthrust plate as a whole.

Mode of thrust emplacement

Within the scope of the present study, very little quantitative evidence can be offered in support of any theory of emplacement except by analogy with previous laboratory tests. Hubbert and Rubey (1959) have at length quantitatively stated the impossibility of moving any thrust plate of the dimensions of that of the Lewis thrust a distance of at least 40 miles without some sort of buoyant force acting upward on the overthrust plate. This they mathematically prove for either the static forces of thrusting or body forces of gravitational gliding. Their thesis is that overthrusting of great sediment thicknesses and distances cannot occur until the fluid pore pressure within the sediment approaches that of lithostatic pressure and produces a state of incipient flotation. They state that the ideal situation for production of these

excess pore pressure are accumulations of geosynclinal thicknesses of alternating clays and sands. The Belt Supergroup, at least during the stages of deposition and subsidence, qualified as an ideal assemblage for this development. The Belt sediments accumulated to thicknesses of more than 40,000 feet and are excellent examples of cyclical sedimentation of alternating carbonates, argillites, shales and quartzites, many of which are lensoid in form. The clays are instrumental in preventing escape of excess fluid pore pressure which develops in sand and/or clay as a result of cumulative overburden pressure and/or tectonic compression.

Powers (1967) correlated the development of abnormal fluid pore pressures to the alteration of montmorillonite to illite in clays at depths of 9,000 to 12,000 feet. According to this theory the last four monomolecular layers of water within the clay interlayers have a greater density than 1.0 g/cm^3 , and upon alteration to illite, are expelled and subsequently expand creating excess pore pressure. The possibility of this should also be considered, since the Belt strata contain large concentrations of illite (Klages, M. G., personal communication, 1970), and the depths of burial previous to overthrusting were certainly adequate. The "sand" bodies of the Belt before and during thrusting are now quartzites, and their porosities do not appear high. Therefore, excess pore pressure may have never been present within the quartzites, but may have been developed and sustained within the argil-

lite or clay. The ability of argillite to sustain any abnormal pore pressure is not known however. The strata of the overlying thrust plate are intensely deformed in places; however, these are restricted to well defined zones. In other places deformation is slight to non-existent.

R.C. Wilson (1970, personal communication) has run several compressional tests on both standard clays and clays from the Lewis fault zone. His results show that the coefficient of friction of a clay is directly dependent upon compaction. With proper compaction, and even without any water content, he concludes that the coefficient of friction is reduced to values which would permit thrusting or gravitational sliding with no necessary excess fluid pore pressure in the fault zone. He also concludes that the extensive fracturing within the leading edge of the Lewis thrust plate would have led to the escape of any abnormal fluid pressure that might have existed during emplacement.

These results, however, do not rule out the possibility of the initial development of the Lewis fault zone under conditions of abnormal fluid pore pressure. They only show that the last phase of movement which put the overlying mass in its present position was, in all probability, due to a low coefficient of friction in the clays of the thrust zone and not because of abnormal pressures, at least in the eastern part of the thrust mass.

Mudge (1970) has accepted the hypothesis that the Lewis overthrust

is a gravity slide feature. He maintains that formation of abnormal fluid pore pressure at depth along with large scale regional uplift is responsible for emplacement on the thrust. By comparison with Hubbert and Rubey's (1959) data on thickness and distance of displacement, he calculates that the initial northeasterly dip of Belt strata immediately prior to sliding was 8.5 degrees.

With the data presently available there appears to be no quantitative evidence to support a conclusion as to whether the Lewis overthrust is a gravity slide feature or whether it was formed by compressional crustal shortening, since the compressional features studied in this report could conceivably have been produced by either mechanism. Further studies involving the symmetry of folding in other parts of the park, along with subsurface study of the Lewis fault zone should give additional insight as to the nature of the fault zone and the forces responsible for emplacement.

SUMMARY

In summary, it is proposed that the "normal" folds are directly related to the Lewis overthrust, and that the "reverse" folds are related to later adjustments during the tensional phase, such as those evident in the North Fork normal fault zone. The lack of a distinct pattern in orientation of "reverse" folds compared to "normal" folds

is believed to be related to the greater variation in trend within the North Fork fault zone and/or to the relatively erratic local structural adjustments that may have taken place after emplacement of the thrust during the tensional phase.

Within the map area, the orientation of "normal" folds centers about a strike of N 70-80 W. This would suggest that movement of the overthrust plate was N 20-30 E, since these folds should be nearly perpendicular to the direction of movement of the Lewis thrust. The assumption that the Lewis overthrust is a single, geologically brief structural event is supported by Norris and Black's (1961) paleomagnetic structural analysis of the Lewis overthrust in Canada.

In consideration of mode of thrust emplacement, it seems that the thickness of the thrust plate, distance of displacement, lithology, mineralogy, and relative lack of intense deformation tend to best support Hubbert and Rubey's (1959) hypothesis of thrusting by development of abnormal fluid pore pressure.

Wilson's compressional tests, however, show that at least in the eastern part of the Lewis fault zone, there was no accompanying fluid or fluid pore pressure involved. His results show that the clay in the thrust zone at its surface trace has a sufficiently low coefficient of friction, even in a "dry" state, (if subjected to previous compaction) to permit thrusting of a mass such as the Lewis overthrust plate without the assistance of abnormal fluid pore pressure. This, however, does

not exclude the possibility of the existence of abnormal fluid pore pressures along the fault zone to the west at depth. Further testing involving fluid content of clays from the fault zone at localities west of the trace and at depth is needed to compare with Wilson's results.

APPENDIX

Altyn Formation (Measured on the southeast slope of Mt. Altyn). Upper contact placed at base of 50 ft. quartzite unit.

Hell Roaring Member

	Feet
17. Dolomite, blue-gray, massive, medium to thick bedded, medium crystalline, very sandy, weathers orange, contains chalky, greenish argillite beds; also a few thin beds of brownish quartzite along with some medium to coarse grained pure quartzite; some dolomite lenses with quartz grains up to 10 mm in diameter. At 65' a lense of medium bedded quartzite	72
16. Limestone, blue-white, medium crystalline with sandy lenses up to an inch thick	4
15. Dolomite, dark blue, dense, finely crystalline; contains a small stromatolite zone with irregular stringers of silica. . .	5
14. Dolomite, medium crystalline, blue-gray, weathers buff, sandy, contains intercalated beds of greenish gray Fe-stained dolomite	74
13. Argillite, black, fissile, Fe-stained	1
12. Dolomite, medium crystalline, massive, thick bedded, sandy, exhibiting cross bedding, weathers orange	15
11. Dolomite, blue-gray, massive, thick bedded, grading upward into blue-white, medium crystalline, somewhat sandy dolomite; some sand grains up to 5 mm in diameter, at 105' a few small chalky beds; at 135' a 1 ft. thick dark purple, limonite stained sandstone	136
10. Same as 9 with abundant horizontal quartzose stringers; some cross bedding	42
9. Dolomite, blue-gray, thin bedded, fine crystalline.	10
8. Same as 7 except alternating with thin bedded finer crystalline dolomite; some thicker beds contain large grains of quartz and chert	41

	Feet
7. Dolomite, medium crystalline, massive, extremely thick bedded with stringers of silica, sandy appearance on weathered surface, 2' of stromatolite zone at top	25
6. Dolomite, blue-gray, fine crystalline, very thin bedded	7
5. Dolomite, blue-gray, medium crystalline, somewhat re-crystallized, siliceous, medium to thick bedded	20
4. Dolomite, blue-gray, thin to medium bedded, platy, sandy in places, small stromatolite zone near base; individuals a maximum of 1 inch in diameter	57
3. Dolomite, massive, stromatolitic, individuals as much as 6 inches in diameter	2
2. Dolomite, blue-gray, fine crystalline, thin to medium bedded. .	15
1. Dolomite, blue-gray, medium crystalline, medium bedded, weathers buff, surface pitted	15

Any outcrops of Altyn below this horizon are obscured by talus. —

Total thickness 526

Appekunny Formation (Measured on the southeast slope of Mr. Altyn).
Upper contact gradational with overlying Grinnell Formation.

Scenic Point Member

	Feet
4. Argillite, red and green interbedded, thin bedded, fissile, alternates with white, thin bedded quartzite	10
3. Argillite, greenish, thin bedded, fissile, contains mud chip conglomerate	137
2. Argillite, brownish, weathering rusty, thin bedded, contains a thin zone of pyrite casts near top	491
1. Argillites, siltites, and impure sandstones, light green to dark green, weathers buff, orange, and rusty, bedding thin to medium, some sandy argillites intercalated, some argillites and siltites finely laminated. At 330' a 10 ft. thick bed of brownish quartzite grading upward to whitish quartzite. Sequence contains abundant mud breccias	354
Total thickness	<u>992</u>

Appistoki Member

5. Siltite and argillite, light green to dark gray green, weathers rusty to buff, thin bedded with some minor medium and thick beds; at top of unit an 8 ft. bed of limonite stained, greenish brown argillaceous sandstone containing flat pebble conglomerate at base	132
4. Quartzite, greenish, brown, and white, medium to coarse grained, medium to thick bedded	25
3. Siltite, dark greenish, massive, thick bedded, weathers buff to orange, a few thin beds at top of unit with a trace of cross bedding	116
2. Argillite and siltite, greenish, thin to medium bedded, few if any ripple marks or mud cracks	263

1. Siltites and fine grained quartzites, yellow green to tan and greenish gray, thin to medium bedded; some layers weathering rusty	Feet 152
Total thickness	<u>668</u>

Singleshot Member

4. Siltites and argillites, mainly green, some buff, greenish gray, and tan with intercalated beds of reddish brown to maroon siltite and argillite. Bedding thin to medium, red disappears at 228'	258
3. Quartzite, white to greenish buff, massive, thick bedded, up to 5', medium to fine grained; basal layer of brownish green thick bedded siltite	71
2. Argillite and siltite, green, gray, medium to thin bedded, contains minor beds of finely laminated, maroon siltite and argillite which weathers dark purple, laminae may be carbonaceous material; or possibly sulfides	162
1. Quartzite, white to greenish white, massive, medium grained	50
Contact between Appekunny and Altyn placed at base of this unit. _____	
Total thickness	541

Grinnell Formation (Measured on the south slope of Mr. Henkel, Plate 3).

Rising Bull Member

	Feet
3. Argillite and quartzite, argillite red with irregular interbedding of green, thin to medium bedded, contains mud cracks, worm burrows, and ripple marks. Quartzite white, pink, gray and green, bedding thin to thick, contains abundant mud chip conglomerate, cross bedding, ripple marks, and mud cracks. Upper contact gradational with Siyeh Formation	907

Red Gap Member

2. Argillite, red, thin bedded, fissile, interbedded with layers of white, thin to medium bedded quartzite displaying cross bedding, ripple marks, and mud chip conglomerate of red argillite. Argillite irregularly mottled with green some 200 feet above base; also contains abundant mud cracks	726
---	-----

Rising Wolf Member

1. Argillite and quartzite, red and green in subequal amounts, thin to medium bedded. Argillite red, green, and greenish yellow displaying mud cracks. Quartzite pink, green, white and gray, showing ripple marks, cross bedding, flat pebble conglomerate of red and green argillite. Contact with underlying Appekunny Formation extremely gradational; placed arbitrarily where red becomes dominant	163
--	-----

Total thickness	<u>1796</u>
-----------------	-------------

REFERENCES CITED

- Bally, A.W., Gordy, P. L., and Stewart, G. A., 1966, Structure, seismic data, and orogenic evolution of southern Canadian Rocky Mountains: Bull. Canadian Petroleum Geology, v. 14, no. 3, p. 337-381. ✓
- Childers, M. O., 1963, Structure and stratigraphy of the southwest Marias Pass area, Flathead County, Montana: Geol. Soc. America Bull., v. 74, no. 2, p. 141-163.
- Clapp, C. H., 1932, Geology of a portion of the Rocky Mountains of northwestern Montana: Mont. Bur. Mines Geology Mem. 4, 30 p., 1 pl.
- Clapp, C. H., and Deiss, C. E., 1931, Correlation of Montana Algonkian formations: Geol. Soc. America Bull., v. 42, p. 673-696.
- Clark, L. M., 1954, Cross section through the Clarke Range of the Rocky Mountains of southern Alberta and southern British Columbia: Alberta Soc. Petroleum Geologists Guidebook, 4th Ann. Field Conf., p. 105-109.
- Daly, R. A., 1912, Geology of the North American Cordillera at the forty-ninth parallel: Geol. Survey Canada, Mem. 38.
- Dyson, J. L., 1949a, The geologic story of Glacier National Park: Glacier Nat. History Assoc., Spec. Bull. 2.
- Erdmann, C. E., 1944, Geology of dam sites on the upper tributaries of the Columbia River in Idaho and Montana, Part 2, Hungry Horse dam and reservoir site South Fork Flathead River, Flathead County, Montana: U.S. Geol. Survey Water Supply Paper 866-B, p. 37-116.
- Fenton, C. L., and Fenton, M. A., 1931, Algae and algal beds in the Belt series of Glacier National Park: Jour. Geology, v. 39, p. 670-686.
- Fenton, C. L., and Fenton, M. A., 1937, Belt series of the north: Stratigraphy, sedimentation, paleontology: Geol. Soc. America Bull., v. 48, p. 1873-1969.
- Folk, Robert L., 1968, Petrology of Sedimentary Rocks: Hemphill's, 170 p.
- Harrison, J. E., and Grimes, D. J., 1970, Mineralogy and Geochemistry of Some Belt Rocks, Montana and Idaho: U.S. Geol. Survey Bull. 1312-0, 49 p.

- Hubbert, M. K., and Rubey, W. W., 1959, Role of fluid pressure in mechanics of overthrust faulting. I. Mechanics of fluid filled porous solids and its application to overthrust faulting: Geol. Soc. America Bull., v. 70, p. 115-166.
- Krumbein, W. C., and Sloss, L. L., 1963, Stratigraphy and Sedimentation: second ed., W. H. Freeman and Co., 660 p.
- Mudge, M. R., 1970, Origin of the Disturbed Belt in Northwestern Montana: Geol. Soc. America Bull., v. 81, p. 377-392.
- Mudge, M. R., Erickson, R. L., and Kleinkopf, Dean, 1968, Reconnaissance geology, geophysics, and geochemistry of the southeastern part of the Lewis and Clark Range, Montana: U.S. Geol. Survey Bull. 1252-E, 35 p.
- Norris, D. K., and Black, R. F., 1961, Application of Paleomagnetism to Thrust Tectonics: Nature, v. 192, p. 933-935.
- Powers, M. C., 1967, Fluid release mechanisms in compacting marine mudrocks and their importance in oil exploration: Am. Assoc. Petrol. Geol. Bull., v. 51, p. 1240-1254.
- Rezak, Richard, 1957, Stromatolites of the Belt series in Glacier National Park and vicinity, Montana: U. S. Geol. Survey Prof. Paper 294-D.
- Ross, C. P., 1959, Geology of Glacier National Park and the Flathead region, northwestern Montana: U. S. Geol. Survey Prof. Paper 296, 125 p.
- Rubey, W. W. and Hubbert, M. K., 1959, Role of fluid pressure in mechanics of overthrust faulting: II. Overthrust belt in geosynclinal area of western Wyoming in light of fluid-pressure hypothesis: Geol. Soc. America Bull., v. 70, p. 167-206.
- Russell, L. S., and Landes, R. W., 1940, Geology of the southern Alberta plains: Canada Geol. Survey Mem. 221.
- Sheldon,*
Smith, A. G., and Barnes, W. C., 1966, Correlation of and Facies Changes in the Carbonaceous, Calcareous, and Dolomitic Formations of the Precambrian Belt-Purcell Supergroup: Geol. Soc. America Bull., v. 77, p. 1399-1426.

Willis, Bailey, 1902, Stratigraphy and structure, Lewis and Livingstone Ranges, Montana: Geol. Soc. America Bull., v. 13, p. 305-352.

THESES
7378
H225
Cap. 2

GEOLOGIC MAP OF THE MANY GLACIER AREA

GLACIER NATIONAL PARK, MONTANA

by Edwin L. Harp
presented to the faculty of the Department of Earth Sciences
in partial fulfillment of the degree of Master of Science
at
Montana State University
Bozeman, Montana
1971

EXPLANATION

Qu	Quaternary undifferentiated. Stream deposits, glacial drift, landslides, and talus.
Ku	Cretaceous undifferentiated. Black shales and bioclastic sandstones containing pelecypods and gastropods.
pCsh	Shepard Formation. Dolomite, limestone, and calcareous argillite.
pCsn	Snowslip Formation. Red and green argillite and quartzite.
pCsi	Siyeh Formation. Buff to gray dolomite, limestone, and calcareous argillite; contains several zones of <i>Conophyton</i> and <i>Collenia</i> .
Grinnell Formation.	
pCrb	Rising Bull Member. Red argillite and white quartzite; contains mud cracks, ripple marks, and flat pebble conglomerate.
pCrg	Red Gap Member. Red argillite and minor quartzite; contains mudcracks and ripple marks.
pCrw	Rising Wolf Member. Red argillite, quartzite, and minor green argillite.
Appokunny Formation.	
pCsp	Scenic Point Member. Green argillite, siltite, and quartzite; minor red argillite; contains mud breccias.
pCa	Appistoki Member. Green argillite and siltite.
pCs	Singleshot Member. Green and maroon argillite and siltite; massive bed of quartzite at base.
Allyn Formation.	
pChr	Hell Roaring Member. Blue gray to white dolomite and sandy dolomite; contains beds of <i>Collenia</i> frequens.
pCmd	Igneous Rocks. Metagabbro sill and diabase dikes.

BELT SUPERGROUP

PRECAMBRIAN



- Attitude of bedding (inclined, overturned)
- Attitude of fractures (inclined, vertical)
- Geologic contact (defined, approximate, assumed, uncertain)
- Fault contact, showing dip and relative movement (defined, approximate)
- Thrust fault, sawteeth in upper plate
- Anticline, arrow indicating direction of plunge
- Axial trend and dip of small anticline
- Axial trend and dip of folds too small to plot individually



3 1762 10014194 2

N378
H225
cop.2

Harp, Edwin Lynn
Thrust plate strati-
graphy and structure of
the Many Glacier area,
Glacier National Park,
Montana